ATMOSPHERIC ELECTRICITY

2nd Edition

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PERGAMON PRESS

OXFORD · LONDON · EDINBURGH · NEW YORK

TORONTO · SYDNEY · PARIS · BRAUNSCHWEIG

Pergamon Press Ltd., Headington Hill Hall, Oxford
4 & 5 Fitzroy Square, London W.1
Pergamon Press (Scotland) Ltd., 2 & 3 Teviot Place, Edinburgh 1
Pergamon Press Inc., 44–01 21st Street, Long Island City, New York 11101
Pergamon of Canada, Ltd., 6 Adelaide Street East, Toronto, Ontario
Pergamon Press (Aust.) Pty. Ltd., 20–22 Margaret Street, Sydney, N.S.W.
Pergamon Press S.A.R.L., 24 rue des Écoles, Paris 5e
Vieweg & Sohn GmbH, Burgplatz 1, Braunschweig

First edition 1957
Second edition 1967

Library of Congress Catalog Card No. 66–29669

2963/67

Preface to Second Edition

THE progress of the subject since the first edition was prepared in 1956 has necessitated a considerable rewriting and enlargement; the same general principles and arrangement have been adopted as in the previous edition.

The extension of atmospheric electrical study to "space" has posed a problem with regard to inclusion, and the decision has been reached to refer to "space electricity" only in so far as it involves an extension of methods used lower in the atmosphere, and to omit any general discussion.

I wish to give especial thanks to my colleague, DR. W. C. A. HUTCHINSON, and also to research students working under me, for help and suggestions given during the production of this edition.

I have not been able to include material which had not come to my notice by 1 August 1965.

Physics Department,
Durham University
August, 1965

J. ALAN CHALMERS

Preface to First Edition

IN 1949 the author published a small book on atmospheric electricity (CHALMERS, 1949), but even before it appeared in print it was out-of-date, largely from the publication of war-time work, particularly in Germany (ISRAËL, 1948a) and the United States. Since that time there have been many advances in various countries and it now seems desirable to produce a new and more comprehensive book. In those places where the treatment in the earlier book remains adequate, the wording is retained unaltered; the author wishes to express his thanks to the Clarendon Press for permission for this. Two important changes have been made; first, the M.K.S. system of units is introduced, as being particularly suitable for the subject; and, second, the term "field" is replaced by "potential gradient" because of the discrepancy between the sign conventions for field in atmospheric electricity and in ordinary electrostatics.

Within the last few years there have been a number of books and articles dealing with various aspects of the subject, but none as comprehensive as the present work aspires to be. Of the general works, we may mention the relevant chapters in the *Compendium of Meteorology* (MALONE, 1951), the second edition of the monograph by SCHONLAND (1953a) and review articles by WORMELL (1953a) and CHALMERS (1954). More specialized are *Das Gewitter* by ISRAËL (1950a) and *Thunderstorm Electricity* edited by BYERS (1953). Where a subject is treated exhaustively in one of these works, it has been considered suitable to give here a condensed account and to refer the reader to the more complete work for further details. Examples that may be mentioned are the radioactive substances in the atmosphere and the detailed form of atmospherics.

Certain subjects which might have been considered appropriate for inclusion have been omitted, as in the earlier work, as irrelevant to the general problems of the subject. For example, there is no mention of "earth currents" since these appear to have no connection with the other problems discussed. The exploration of the ionosphere has also not been included, since the techniques employed are so different from those in other branches, and the results do not concern electrical phenomena at lower levels. And the aurora, although it is an electrical phenomenon in the atmosphere, is similarly omitted.

It is a great pleasure to record my gratitude to the research students working with me who have read this work at various stages and have made many helpful suggestions.

The author has been unable to include material that had not come to his notice by 1 March 1956.

April, 1956 J. A. CHALMERS

CHAPTER 1

Historical Survey

1.1. Primitive Notions

The effects of thunderstorms and particularly of lightning flashes have been a source of terror to human beings from the earliest times, and the primitive religions ascribed to the vengeance of the Gods the destruction caused by "thunderbolts", which, though perhaps sometimes meteorites, were usually lightning flashes. SCHONLAND (1964) gives a very interesting account of the part played by the "thunder-god" in religion and folk-lore, and mentions that in Bechuanaland, even as recently as 1930, witch-doctors were believed to have the power of directing lightning flashes to kill individuals and to damage property.

Although the results of lightning flashes have been mainly destructive, it may be surmised that Man first obtained fire from a lightning stroke to a forest; there is evidence that even now, in the forests of the north-western United States, most forest fires are caused by lightning and there are still relics of prehistoric fires so produced.

1.2. Identification of Lightning with Electricity

The idea that electricity exists in the air originated when it was suggested that the well-known phenomena of lightning and thunder were just a large-scale manifestation of effects which can be obtained on a smaller scale in the laboratory with static electricity. The earliest reference to this suggestion seems to have been that of WALL (1708), who observed cracklings and a flash with amber held at a small distance from his finger and remarked that "it seems in some degree to represent thunder and lightning." GRAY (1735) observed an "electric fire" and considered that "it seems to be of the same nature with that of thunder and lightning." WINKLER (1746) made a more detailed comparison of the electric spark and the lightning discharge, and even went so far as to suggest an origin for the electricity in the air from the friction and collision of the various particles in the air, an idea quite close to some of the most modern theories.

1.3. Electricity from Thunder Clouds

The invention of the early electrical machines and of the Leyden jar made it possible to obtain and store larger quantities of static electricity than previously and so to begin the investigation of electric discharges. This led FRANKLIN (1750) to suggest attempting to collect electricity from thunder clouds by means of a point on a high tower or steeple and insulated from the ground; unfortunately there was no suitable tower or steeple at the time in Philadelphia, where FRANKLIN lived, and he was forestalled by the French scientist D'ALIBARD (1752), who used an iron rod 40 ft high insulated by a glass bottle and fixed by wooden masts, at Marly near Paris. On 10 May 1752, sparks from the rod to an earthed wire were noticed by an old soldier named COIFFIER, who had been left in charge of the apparatus. A month later, and before he had heard of D'ALIBARD's success, FRANKLIN (1752) obtained similar results with a kite whose conducting string ended on an insulating silk ribbon; a detailed account of FRANKLIN's experiments has been given by SCHONLAND (1964). Only 2 months later FRANKLIN fitted an iron rod to the chimney of his house and arranged a suspended brass ball to give chimes on bells when electrification occurred.

In the next year FRANKLIN collected the charge from his point in a Leyden jar and, by the use of a cork ball on a silk thread, he tested the sign of the charge in comparison with charges produced by rubbing. He found that the charge from a thunder cloud is nearly always negative. As SCHONLAND remarked, there was no better information than FRANKLIN's as to the sign of the charge in the lower parts of a thunder cloud for 170 years.

D'ALIBARD and FRANKLIN obtained only small sparks or small movements of balls, and, though these established the presence of electricity in thunder clouds, more was really needed to show the identity of a lightning flash with an electric spark. This was provided by DE ROMAS (1753) who, quite independently of FRANKLIN, also used a kite and obtained much more intense results, reaching to a spark 3 m long and 3 cm in diameter, with "more noise than a pistol shot". The most intense effect of this nature occurred in 1753 in the case of the unfortunate Professor RICHMANN of St. Petersburg (Leningrad) who was killed by a lightning flash from an insulated conductor exposed above his laboratory; it is to be presumed that an actual lightning stroke hit his conductor.

These early experimenters recognized electrification mainly by the production of sparks and therefore required effects of some magnitude. Except in RICHMANN's case, all the effects so far discussed were produced by what is now termed point discharge.

1.4. The Lightning Conductor

FRANKLIN, even before he had proved that thunder clouds contain electricity, had thought of the possibility of a lightning conductor to take the electricity harmlessly into the ground, instead of allowing it to damage buildings, ships, etc.

Since FRANKLIN's invention of the lightning conductor there has been much controversy as to its efficiency and mode of action, and this has been described in interesting detail by SCHONLAND (1964); some of the more important points will be discussed in Chapter 14.

1.5. Electricity of Fine Weather

One of the earliest imitators of D'ALIBARD was LEMONNIER, who set up a wooden pole with a pointed iron rod fixed to the top. An iron wire was attached to the rod and entered a building without making any contact, ending on a stretched silk fibre. In addition to obtaining sparks under suitable conditions, LEMONNIER found that particles of dust were attracted to the iron wire when electrified. This method is much more sensitive than that of the observation of sparks and can show electrification which is too feeble to give sparks.

By this method, LEMONNIER (1752) first found effects in fine weather. This result was quite unexpected and was perhaps more important than those of FRANKLIN and D'ALIBARD. A lightning flash can be recognized as similar to an electric spark, but there is no similar obvious effect for the electric phenomena of fine weather. DE ROMAS, almost at the same time as LEMONNIER and independently, found electrical effects in fine weather, using his kite for the purpose.

LEMONNIER also introduced a new method of obtaining electrification, using a stretched wire without points in place of the pointed conductor; in modern terminology, the wire acts as a weak potential equalizer, slowly acquiring the potential of its neighbourhood.

LEMONNIER suspected a variation in the electrical effects of fine weather with time of day, though some of his results, as he himself realized, were due to poor insulation at night. It was left to BECCARIA (1775), who also used a stretched wire, to confirm the existence of a daily variation, as a result of nearly 20 years' observations.

FRANKLIN (1752) (see § 1.3.) had shown that he usually obtained a negative charge from thunder clouds, but occasionally a positive charge. BECCARIA confirmed this and found that in fine weather his wire acquired a positive charge.

1.6. The Movable Conductor

Advances were produced by the introduction of new methods of measurement, in particular by DE SAUSSURE (1779). In place of the crude earlier methods of detecting electrification by sparks or by dust particles, DE SAUSSURE developed FARADAY's cork ball into a form of electrometer which consisted of two silver wires carrying balls of elder pith and suspended from the same support inside a glass vessel with metal casing; when the wires were charged they repelled one another, the deflection giving a measure of the charge. This instrument was more sensitive than earlier methods and also gave quantitative results.

DE SAUSSURE also introduced a new method of observation, that of the movable conductor. The mode of action is understood best in terms of the more modern concept of lines of force, starting on a positive charge and ending on a negative charge. If there are vertical lines of force in the atmosphere, with their positive ends somewhere high up, then an earthed conductor at or near the surface of the earth has the negative ends of the lines of force on its surface and so has the corresponding "bound" charge; if the conductor is moved upwards in the air, then the lines of force tend to concentrate on the conductor (see Fig. 1) and there is therefore a greater bound charge on the conductor. Now if, instead of being earthed, the conductor is connected by a conducting wire to an insulated electrometer which had been connected to earth before being insulated, then the change in the bound charge on raising the

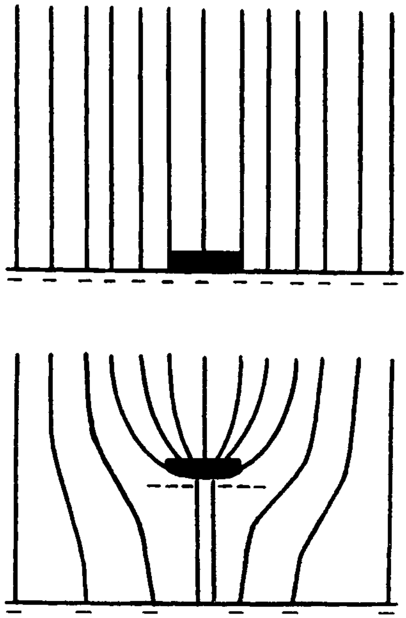


FIG. 1. The principle of the movable conductor.

conductor can come only from within the isolated system of conductor and electrometer, and so must involve the appearance of an equal and opposite charge (in this case positive) on the remote part of the system, i. e. the electrometer. The pith balls will thus obtain a positive charge and diverge by an amount which depends on the change of bound charge on the conductor, and, for a definite movement of the conductor, this depends on the density of the lines of force and so on what is now called the potential gradient. By this method DE SAUSSURE was able to compare effects at different times.

DE SAUSSURE also made the important discovery that his effect disappeared on lowering the conductor to its original position, showing that the system had received no charge from outside itself. DE SAUSSURE's method differs fundamentally from those of previous workers in which the effects depended on the actual collection of charge from the air by conduction, assisted in some cases by point discharge. If the air were a perfect insulator, then the methods of

LEMONNIER and BECCARIA would have given no effects, but that of DE SAUSSURE would still have done so. DE SAUSSURE's method may be considered as the forerunner of all methods of measuring potential gradient depending on the bound charge, and perhaps particularly of the electrostatic fluxmeter method (see § 5.26.).

DE SAUSSURE also found that he could obtain a permanent charge on his electrometer if he earthed his system when the conductor was at its highest point and then lowered it insulated; this is perhaps the forerunner of the agrimeter method of measuring potential gradient (see § 5.29.). Another of DE SAUSSURE's methods was to throw into the air a metal ball, having a wire attached to the electrometer, but breaking connection while the ball is in the air; the ball then takes its charge with it and leaves the electrometer with the charge of opposite sign. This might be considered the forerunner of the water dropper (see § 5.6.).

DE SAUSSURE realized that the effects he obtained were those of electrostatic induction, but he ascribed them, as did many after him, just to positive charges in the air, without, apparently, recognizing the necessity for the corresponding negative charges on the earth.

DE SAUSSURE appears to have been the first to discover an annual variation in the magnitude of the effects, which are greater in fine weather in winter than in summer.

1.7. Early Theories

In the time of DE SAUSSURE it was considered that the phenomena of atmospheric electricity could be explained by supposing that the air in fine weather carries a positive charge, increasing with the height above the ground. This would account for the positive charge acquired by a point or a stretched wire and also for DE SAUSSURE's induction effects. At that time it was not possible to estimate the variation of charge with height and so it was not realized that the main part of the positive charge lies well above the parts of the atmosphere close to the ground.

It was also necessary to suggest the origin of the positive charge in the region and a theory was put forward by VOLTA (1782) that the effects were caused by a separation of electric charge accompanying the change of state from water to vapour. VOLTA believed that, as well as latent heat, some amount of positive electricity would be needed for the conversion of water to vapour; this would give a negative charge to the earth (apparently the first recognition of this charge) and a positive charge in the air.

This attractive and simple theory might, even now, account for a number of phenomena in atmospheric electricity, but it has found no support from direct experiment in spite of numerous searches, starting with those of Volta himself, for electrical effects accompanying the change of state from water to vapour (see § 3.21.).

1.8. The Flame Technique

Volta (1782) was the first to make use of a new technique of measurement. He found that if he placed a candle or lighted fuse in contact with an exposed conductor, the latter acquired a charge of the same sign as that obtained by the use of a point, but of increased magnitude. He proved that this could not be due to any electrical effect of the combustion because he found no charging in a control experiment carried out in a closed room. Also, after the conductor is earthed, when there is a flame the charge reappears almost immediately, instead of taking some time, as is the case in the absence of the flame. In modern terms, the flame acts as a potential equalizer, providing charged particles and ions which quickly bring the conductor to the potential of its surroundings, while without the flame this takes a time of the order of an hour. The use of a flame first made possible the measurement of changes in potential gradient over periods as short as minutes.

1.9. Theory of the Charge on the Earth

The idea that the phenomena of atmospheric electricity could be accounted for by the earth being charged, without any charge being necessary in the lower air, was first brought forward by Erman (1804), who showed that similar effects to those of de Saussure could be obtained by moving a conductor horizontally from an exposed to a sheltered position. But Erman did not develop his ideas very far, nor does he seem to have realized that there must be, somewhere in the air, a positive charge equal and opposite to that on the earth.

It was left to Peltier (1842) to confirm the ideas of Erman. He put forward again the hypothesis of an original permanent negative electrification of the earth, without at first discussing its origin or realizing the necessity of discussing its replenishment. Peltier showed how the results of de Saussure and of Erman could be accounted for by the induction from a negatively charged earth, and he pointed out the importance of realizing that an electrometer measures only the difference between the charges (we would now say potentials) on the cage and on the moving system (gold-leaf or pith balls), thus explaining why an earthed electrometer does not show any effect due to the negative charge on the earth.

PELTIER attempted to account for the phenomena by the idea that water vapour leaving the earth might carry with it some of the earth's negative charge, thus giving an explanation of annual and diurnal variations, the electrical effects of clouds, etc. But this theory never found any experimental support and there are many facts not covered by it.

1.10. Kelvin's Contributions

Important contributions to the subject were made by W. THOMSON, later LORD KELVIN. (For convenience, he will be referred to as KELVIN, even in connection with work carried out before his elevation to the peerage.) KELVIN had made considerable advances in the theory of electrostatics and applied the same ideas in atmospheric electricity. He (KELVIN, 1860a) introduced the idea of "potential" and showed that the flame, as used by VOLTA (see § 1.8.) and the water-dropper (see § 5.6.) which he himself invented (KELVIN, 1859b) serve to equalize the potential of a conductor with that of the air in its neighbourhood. He also explained the connection between the two methods of measurement in atmospheric electricity at the time, that of the potential equalizer and that of the movable conductor (see § 1.6.). KELVIN perfected several new methods of measurement, including electrometers more sensitive than those used previously.

KELVIN (1860a) pointed out that it is not possible to determine the location of the charges responsible for the phenomena of atmospheric electricity merely from measurements carried out at the earth's surface; for example, a uniform potential gradient over the whole earth would imply a uniform layer of positive charge at some height, but could give no indication of what height. KELVIN (1860b) put forward the argument that the positive charges, corresponding to the vertical potential gradient in the atmosphere and to the negative charges on the earth's surface in fine weather, must exist actually in the atmophere and not at infinity, because it is known that at low pressures, such as exist in the upper atmosphere, air becomes a conductor and, as lines of force are not able to penetrate a conductor, the positive charges must be on the inner side of such a conductor if not below. This appears to be the first suggestion of a conducting layer in the upper atmosphere.

Since the potential gradient over the earth's surface is not uniform but varies both with time and place, KELVIN (1860a, b) concluded that there must exist space charges in the lower atmosphere even in fine weather. He proposed an "electro-geodosy" to investigate these by means of simultaneous observations at a number of different places; it is remarkable that this suggestion has never been completely carried out and the only approaches to it are quite recent (see §§ 2.35., 5.36.). As KELVIN realized, even such a programme would not determine unambiguously the location of the charges and he suggested that the only really satisfactory approach would be to use balloons.

Another suggestion of KELVIN's (1860b), that of investigating the electrification of rain, was not feasible at the time but was carried out later (see Chapter 10). KELVIN (1860c) appears to have been the first to observe changes of potential gradient caused by lightning flashes; he also first observed the electrical effects of steam from a locomotive (see § 5.57.).

KELVIN (1859a) was very insistent on continuous recording of the potential gradient and appears to have been the first (KELVIN, 1860b) to suggest the use of photographic recording.

KELVIN (1860b, c) believed that the lowest layers of the atmosphere contained negative charges, and he found evidence for this in some cases in disturbed weather when the potential gradient near the earth's surface became negative, while at some 30 m higher it remained positive; a similar result for greater heights has been found at the Eiffel Tower by CHAUVEAU (1900) and has not yet received any satisfactory explanation (see § 5.65.).

1.11. Absolute Values of Potential Gradient

The use of a potential equalizer, such as KELVIN's water-dropper involves some disturbance of the lines of force by the equalizer and its housing. EXNER (1900) realized that such a method could not give an absolute value of the potential gradient or, from this, of the charge per unit area of the earth's surface, and he devised methods of obtaining the "reduction factor" or "exposure factor" to convert observed values to corresponding absolute values over level ground (see § 5.18.).

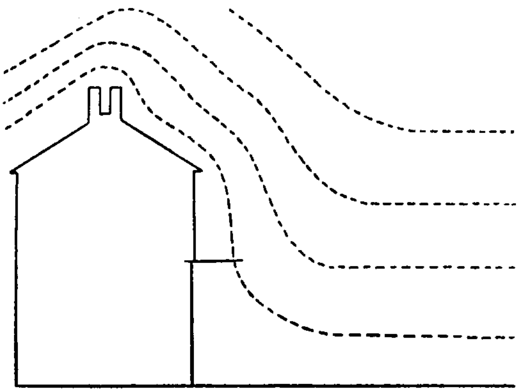


FIG. 2. The reduction factor.

More modern work has shown that a reduction factor is not very satisfactory, especially if space charges are present, and it is preferable to measure under conditions when it is not required.

1.12. Relation with Humidity

Following on PELTIER's theory that water vapour carries negative charge from the earth, there would be a relation between potential gradient and humidity, and this EXNER believed he had established. But such a theory would involve negative charges on the water vapour in the lower atmosphere and so would predict an increase of potential gradient with height, as a result of lines of force with their lower ends on these charges.

KELVIN (1881) proposed the use of two water-droppers one 10 ft above the other, attached to a balloon. The earliest measurements by LECHER and TUMA appeared to give an increase of potential gradient with height, but later results by LE CADET (1898) and TUMA (1899) gave definite evidence that above, at most, a few hundred feet, the potential gradient decreases with height and the space charge is therefore positive.

The theory of PELTIER must therefore be abandoned and the connection between potential gradient and humidity is only indirect, since both are dependent on the effects of *austausch* (see § 2.32.).

1.13. The Fine-weather Potential Gradient

A great deal of the continuous measurement that has been carried out has been concerned with the potential gradient and has shown that in fine or fair weather this is nearly always positive, giving a negative charge on the surface of the earth. The actual value of the potential gradient shows variations with time of day and time of year (see §§ 5.42., 5.43., 5.44., 5.47.). In addition to the fine-weather potential gradient, there is a fine-weather conduction current which brings positive charge to the earth.

Although the potential gradient is the feature of atmospheric electricity which is most simple to determine, and which therefore has been most often measured, it has been realized (see § 2.24.) that its significance is often not very fundamental and that the air–earth conduction current has more importance.

1.14. Conductivity of the Air

Among the earlier experimenters, COULOMB (1795) and MATTEUCCI (1850) had established that the air is a conductor of electricity, but the importance of this was not realized until the observations of LINSS (1887), who pointed out that if the rate at which charge leaks from the earth owing to the conductivity of the air were maintained, then the whole negative charge on a portion of the earth would leak away in a period of about 10 min; or, to put it slightly differently, the positive charge reaching a portion of the earth's surface by conduction amounts, in a period of about 10 min, to the negative charge permanently present there. And yet the earth retains its negative charge!

This has raised what has become one of the most important problems of atmospheric electricity, the question of how the earth's negative charge is maintained, and various theories have been put forward to account for a regular passage of negative electricity to the earth, often termed the "supply current". At one time it appeared possible that something outside previous knowledge in physics might be required, such as creation of negative charge within the earth, or a current to the earth of negative particles moving so fast as to be undetectable; SWANN (1955) discussed such ideas critically. But it is now generally agreed that it is in stormy weather that the negative charge comes to the earth to balance the positive charge arriving in fine weather.

Much work has also been done on the nature of the conductivity of the atmosphere. ELSTER and GEITEL (1899a, b) and WILSON (1900) demonstrated the existence of "ions", particles of molecular size or larger, carrying positive or negative charges. Following on this discovery, LANGEVIN (1905) found large ions and POLLOCK (1951 b) and others have found ions of intermediate size; further work has been concerned largely with equilibrium conditions relating ions and uncharged particles of different sizes (see Chapter 4).

1.15. The Origin of Atmospheric Ionization

The properties of ions in the atmosphere have been investigated by many workers, and it has been discovered that ions constantly disappear by combination with ions of opposite sign or are transformed into ions of different size by combination with uncharged particles. It thus becomes evident that there must be some mechanism for the production of ions. The most obvious agent would seem to be the radioactive bodies in the earth itself, but this would lead to the prediction that the conductivity of the air would decrease as a body rises above the earth. Measurements, first by HESS (1911) and KOLHÖRSTER (1913), have, however, shown that, on rising above the earth, there is, in fact, a very considerable increase of conductivity, which cannot be caused by effects of terrestrial origin. Further work established beyond any doubt the existence of highly penetrating radiation entering the earth's atmosphere from all directions in space. These "cosmic rays" are capable of reaching the earth's surface through the atmosphere and even of penetrating an appreciable distance into the earth, and thus are an effective source of atmospheric ions at all levels, with the ionization greater at greater altitudes. Other, less penetrating, radiation reaches the earth's atmosphere from the sun; this makes an important contribution to the ionization of the highest levels of the atmosphere where the conducting layers are situated, but is absorbed before reaching the lower layers, and thus is of no importance in atmospheric electricity. The origin and nature of the primary cosmic rays and the processes by which secondary rays are produced will not be dealt with here; from the point of view of atmospheric electricity, all that matters is the fact that there is ionization which increases on rising above the earth, and that for the part of the atmosphere of concern to us the ionization is independent of solar or sidereal time. It is, however, of interest that the discovery and early history of cosmic rays derived from the subject of atmospheric electricity. Close to the earth, ionization due to radioactivity in the earth and in the atmosphere is important.

1.16. Stormy Weather Phenomena

In stormy and disturbed weather the potential gradient has been found to be larger than in fine weather, to vary rapidly and often to show negative values. In addition to the conduction current, there are other currents in the lower atmosphere, for example, currents brought down by rain, and the conductivity may be different from the normal value, by reason of ions produced in processes connected with the weather disturbance. Results showed that there must be charges in the clouds, and various investigations have been undertaken to determine the signs, magnitudes and locations of these charges.

The earliest measurements of the charge brought down by rain showed an excess of negative charge, but the majority of the later measurements have shown an excess of positive charge (see Chapter 10).

1.17. The Atmosphere as a Condenser

Although KELVIN (1860b) (see §§ 1.10. and 2.21.) had suggested the existence of a conducting layer in the upper atmosphere and the analogy of the whole atmosphere to a condenser, this idea was not brought into use in atmospheric electricity until very much later. Even in the treatise of CHAUVEAU (1922, 1924, 1925), the only mention of a region of high conductivity in the upper atmosphere does not relate it to phenomena of atmospheric electricity near the ground.

The first mention of the "Heaviside layer" in connection with atmospheric electricity in this century appears to have been due to WILSON (1920), but its importance was not recognized by all, even in 1929, when SIMPSON in discussion on a paper by WHIPPLE (1929a) argued that the potential gradient was not due to charges in the Heaviside layer. However, in SCHONLAND's (1932) book, the conducting layer is accepted as a fact to account for the constancy of air–earth current with height, and SCHONLAND distinguished between this conducting layer, whose conductivity is derived from "penetrating radiation" (cosmic rays) and the higher Heaviside layer whose conductivity is of solar origin. This distinction is now preserved in the terms "electrosphere" and "ionosphere" (see § 2.21.).

1.18. The Thunder Cloud

A considerable part of the work on atmospheric electricity during the present century has been concerned with the electrical nature of the thunder cloud and the processes by which the separation of charge within the cloud is effected. The names chiefly associated with earlier work on this subject are those of WILSON (1916) and SIMPSON (1927), and there was for some time a controversy between them as to the polarity of a thunder cloud, i.e. as to whether it was to be considered to have a positive upper part and a negative base or vice versa. SIMPSON favoured the theory of the positive base, largely on the ground that photographs of lightning flashes show that those from cloud to earth are almost always branched downwards, and that laboratory photographs of sparks show branching from the positive pole. SIMPSON accounted for this negative polarity of a cloud on the theory of the separation of charge by the breaking of drops. WILSON, on the other hand, insisted that observations by himself and others on the potential gradients and potential-gradient changes of thunder clouds could be interpreted reasonably only if the cloud usually has a positive polarity, i.e. a negative base. Spark discharges under conditions more close to those of lightning have shown (SCHONLAND and ALLIBONE, 1931) that the direction of branching is not a criterion of polarity, but rather of the geometrical arrangement, and, after this, work using balloons (SIMPSON and SCRASE, 1937) showed that the top of the cloud is positive, with a negative charge below and in some, perhaps all, cases of thunder clouds, there is a positive charge in the base as well, though this may be absent in non-thundery showers. The quantities of free charge of either sign in a thunder cloud range usually between 10 and 100 C. Much work has also been done (see Chapter 14) on the details of the lightning flash.

1.19. Military Applications

The needs of World War II stimulated advances in atmospheric electricity, notably in Germany and the United States; among the problems that arose were the hazards to aircraft and to captive balloons in thunderstorms, and the efforts on radio communications, particularly to and from aircraft. An important advance was the possibility of carrying measuring instruments in aircraft and so of realizing KELVIN's dream of a three-dimensional survey, though balloons had earlier given a start to this.

While, regrettably but understandably, some of the work carried out by service departments in several countries may have been kept secret because of possible connections with military needs, a great deal of fundamental work has been carried out by the United States Services Departments, or sponsored by them, and published. The United States Air Force has been largely responsible for two conferences which have greatly stimulated the advance of the subject (HOLZER and SMITH, 1955; SMITH, 1958).

1.20. Use of Electronics

One of the most important technical advances of the past 20–30 years has been the application of electronic methods of measurement, replacing the traditional quadrant electrometer. One consequence of the availability of electronic amplification has been that field machines have become generally used for the measurement of potential gradient.

An important application of electronic methods has arisen in the adaptation of the meteorological radio-sondes to the measurement of electric parameters, a technique which has made it possible to obtain much information about the higher levels of the lower atmosphere.

The use of computers for analysis of results and the digitization of results are recent extensions of the same advance.

1.21. International Co-operation

It is probably correct to say that there is less scope for international co-operation in atmospheric electricity than in several other branches of terrestrial and atmospheric physics, which have profited so much by the International Geophysical Year and the International Quiet Sun Year. But many of the problems of atmospheric electricity can progress by the use of results from various parts of the world and international co-operation has proved valuable. And the various international bodies have given opportunities for conferences, particularly that at Montreux in 1963 (CORONITI, 1965), sponsored by the International Association of Meteorology and Atmospheric Physics (I.A.M.A.P.) and the International Association of Geomagnetism and Aeronomy (I.A.G.A.) of the International Union of Geodosy and Geophysics (I.U.G.G.).

1.22. Recent Tendencies

As might be expected from general considerations, recent developments in atmospheric electriicty have tended to concentrate largely on the relations between this and other branches of atmospheric physics. For example, the fine-weather phenomena have shown connections with more general meteorological conditions, and much work has been done in establishing relations between the generation of charge in clouds and the other physical processes at work. While progress has continued in the elucidation of the phenomena of the lightning flashes, the radiation emitted has been very extensively used to investigate the properties of the higher layers of the atmosphere.

With increasing understanding of the origin of electric charges in clouds, the possibility of controlling them and perhaps inhibiting their production has received serious consideration although, at the time of writing, no serious practicable proposals have yet been made.

1.23. Space Electricity

Until recently, it was possible to make a firm division of electrical phenomena in the atmosphere into the two branches of atmospheric electricity and ionospheric physics. In atmospheric electricity we were concerned with the regions of the atmosphere accessible to direct measurements, i.e. up to the limit for balloons of about 22 km, with inferences only of effects at higher levels. In ionospheric physics the information found by reflections of radio waves was available down to, perhaps, 60 km on occasions.

The distinction is also possible in two other ways; in atmospheric electricity the life-time of electrons is so small that they need be considered only in cases where there is actual breakdown; but, on the other hand, it is the electrons which are responsible for the reflecting properties of the ionosphere. The ionization in the region covered by atmospheric electricity is practically unaffected by the sun, being caused almost entirely by cosmic rays and radioactivity; on the other hand, the ionization in the ionosphere is almost entirely caused by solar radiation.

The advent of rockets and satellites has made it possible to make direct measurements at much higher altitudes and has meant that the methods of atmospheric electricity, with suitable adaptations, can be used at much greater altitudes.

The question then arises as to how far the term "atmospheric electricity", for example as used in the title of this book, should be taken to extend. The decision is taken, for this book, to include extensions of more conventional atmospheric electricity, i.e. measurements of potential gradients, currents, conductivities, etc., but not to include problems concerned with the ionosphere as such, where the methods are so different. This is in accord with the contributions to the "Space" section of the *Third International Conference on Atmospheric and Space Electricity in 1963* (CORONITI, 965), which were almost entirely from those who had progressed upwards from conventional atmospheric electricity, rather than from those who were dealing with problems of the production, structure and motion of the ionosphere.

CHAPTER 2

General Principles and Results

2.1. Scope of the Chapter

This chapter is intended to serve several purposes. In the first place, the M.K.S. system of units is introduced and described for the benefit of those unfamiliar with it, as it is used throughout the book; various principles are enunciated and worked out so that they can be applied at various points in the book. The general relations between the different phenomena, which are separately discussed later, are dealt with here so that each subsequent chapter will not require references to other chapters for the relations. Also included are various matters from other branches of physics and various items of experimental techniques, as well as brief discussions of the relations of atmospheric electricity with other phenomena.

2.2. Systems of Units

In electrical theory and practice, there have been three sets of units in use, the E.M.U. system, the E.S.U. system and the system of practical units based on the E.M.U. system with factors of powers of 10.

The subject of atmospheric electricity is one in which the question of units is particularly awkward because here, more than perhaps in any other branch of electricity, there is an interconnection between phenomena of static and current electricity; if the different systems of units are used where they seem to be most appropriate, there very often arises the need of changing from one system to the other, and therefore of remembering, or looking up, the appropriate conversion factor.

It is clearly desirable to use one set of units throughout and there is the choice between the E.S.U. system, which could be applied to all phenomena, and the system of the practical units, which can be applied by making a few simple changes. All electrical instruments and standards are calibrated in terms of the practical units, and if these are abandoned for atmospheric electrical measurements we still have to remember, or look up, the factors to convert volts, ohms, farads, etc., to E.S.U.; as a further difficulty, we would have to invent new names for the units, or else use the clumsy "E.S.U. of potential" or the misleading "statvolt".

The balance of argument is in favour of retaining the familiar practical units and making the necessary adjustments.

2.3. The Rationalized M.K.S. System

If we are to use the practical units of potential and current, the volt and the ampere, then the unit of power is the watt (or volt-ampere). If we retain the second as the unit of time, we retain the coulomb as the unit of charge and the unit of energy is the joule.

Since the joule is 10^7 ergs, where the erg is the unit of energy based on the centimetre, gram and second, it follows that we shall have to choose new units of length and mass. There are various possible choices, but by far the most convenient is that of the metre and the kilogram, which, it is simple to see, agrees with the joule and so with the volt and the ampere. It is a fortunate accident that makes it possible to choose the two units as those of which the material standards are preserved, and it might well have happened that the original definitions of the practical electrical units would not have fitted with both the metre and the kilogram.

For a system of units which is to include electrical phenomena, four fundamental units must be defined, and in the M.K.S. system these are the metre, kilogram, second and ampere. It is somewhat more convenient, in electrostatic theory, to consider the coulomb as the fourth unit, and, if it were a question of a material standard, it might be better to choose the ohm. From some points of view there are advantages in selecting the volt, as well as the ampere, as a fundamental unit and leaving the kilogram as a derived unit.

At the same time as the change is made from E.S.U. to M.K.S., it is convenient to make another change which is actually quite independent of the change of units, and either change could be made separately. This other change consists in altering the position of "4π" in formulae, and is called "rationalizing". The factor 4π is incorporated in a constant in such a way that it disappears from many formula, e.g. GAUSS's theorem, and appears only where spherical shape would lead one to expect it.

With the M.K.S. system, it is no longer necessary to remember the various conversion factors, but there is introduced one figure, the value of the "permitivity of free space" or "electric space constant", denoted by ε_0 and having a value of $8{\cdot}854 \times 10^{-12}$ (or approximately $1/36 \times 10^9$) farads per metre.

In atmospheric electricity we shall be concerned only with electrostatic formulae which are modified by the introduction of the M.K.S. system; formulae of current electricity are unaltered and formulae of electromagnetism do not enter into our subject. Further details of the M.K.S. system are given in a book by PIDDUCK and SAS (1947), an excellent article by NICHOLSON (1951) and in many other works.

Just as in older systems, multiples and submultiples of the fundamental units are often convenient, so it is not necessary to insist in the M.K.S. system on always using metres and kilograms where the centimetre or gram, or other multiple or submultiple, may be more convenient.

To enable the reader to compare the results given here in M.K.S. units with those given elsewhere in E.S.U., a list of conversion factors is given in Appendix I.

It may be noted that, in many cases, if ε_0 in the formulae given here is replaced by $1/4\pi$, the E.S.U. formula is obtained.

The following abbreviations will be used: A = ampere, V = volt, C = coulomb, F = farad, Ω = ohm and the prefixes p or $\mu\mu$ = $= 10^{-12}$, $\mu = 10^{-6}$, m $= 10^{-3}$, k $= 10^3$ and M $= 10^6$; k and M when used alone represent kΩ and MΩ respectively. Less frequently, n = nano $= 10^{-9}$ and G = giga $= 10^9$ are also used.

It is a good principle to use either a prefix or else a power of 10, but not both, e.g. a capacitance of $3{\cdot}5 \times 10^{-8}$ F can be expressed as such or as 35 nF, but preferably not as $3{\cdot}5 \times 10^{-2}$ μF or $3{\cdot}5 \times 10^4$ pF.

2.4. Electrostatic Formulae

In the M.K.S. system the unit of force is the newton (Nw), defined as the force required to give 1 kg an acceleration of 1 m/sec^2. It is clear that the form of the law of force between electrostatic charges must differ from that on the older system. The force, in Newtons, between two charges e and e' coulombs distant apart r metres *in vacuo* is:

$$ee'/4\pi\varepsilon_0 r^2,$$

where ε_0 is the permittivity of space ($8{\cdot}854 \times 10^{-12}$ F/m) and incorporates both the change from E.S.U. to M.K.S. units and also the rationalization, with the 4π, which appears here, where there is a spherical "spreading out" of the effect, and which disappears in cases of plane geometry.

In place of "total normal induction", the quantity "total electric flux" is now used; this is defined as ε_0 times the normal component of electric field strength (force per coulomb on a small test charge), multiplied by the area. Cases in which the medium is other than a vacuum or air do not occur in atmospheric electricity and so need not be considered here. Then GAUSS's theorem is stated in the form that the total electric flux over a closed surface is equal to the total charge in the volume bounded by the surface, the 4π having disappeared here; GAUSS's theorem can be thought of as a statement of the fact that lines of electric flux can start or end only on electric charges; it is thus not directly connected with the inverse-square law, and, in fact, the inverse-square law itself then is deduced from GAUSS's theorem, or the assumption behind it, and the three-dimensional nature of space (CHALMERS, 1963).

2.5. Potentials

The idea of the electric "potential" at a point, first introduced by KELVIN, is very important in atmospheric electricity. The definition of the potential difference between two points is that it is the mechanical work (in joules or newton-metres) per coulomb needed to move a small positive charge from one point to the other; it follows that, unless there is a current flowing, there can be no potential differences between any points of a conductor. In electrostatics, and in its application to atmospheric electricity, it is always the potential difference and not any absolute value of potential that is of concern, but it is often convenient to choose some zero of potential and measure potential by the work done in moving a small charge from the position of zero potential to the point in question. In theoretical electrostatics the zero of potential is conveniently chosen as the potential of a point far removed from all electric charges, so that an isolated, uncharged body has zero potential. But in the case of atmospheric electricity such a choice would not be suitable since we have, at present, no knowledge of electrical forces outside the ionosphere (the conducting layers in the upper atmosphere) and hence no method of obtaining the potential of the ionosphere relative to the conventional zero.

In atmospheric electricity, which is concerned only with matters between the two conductors, the earth and the electrosphere (see § 2.21.), it is convenient to measure potentials relative to the earth as zero, but it must be realized that this zero is not the potential of a point at an infinite distance from charges. Our present ignorance of the true potential of the electrosphere relative to an infinitely distant point has no effect upon matters between the earth and the electrosphere, and, even if rocket and satellite measurements give a value for this potential, it will still remain much more convenient, for atmospheric electricity, to measure potentials relative to the earth.

2.6. Potential Gradients and Fields

If there is a potential difference between two points, then work must be done in moving a charged body from one point to the other. If a positively charged body is moved from the point at lower potential to that at higher potential, then a positive amount of work is done, so that the body is moved against a force; the force is thus in the opposite direction to the difference of potential. The ratio of the force acting on the body to the charge on the body is known as the "electric intensity" or the "electric field strength", often abbreviated to "field".

If the force is measured in newtons and the charge in coulombs, then the field is measured in newtons/coulomb; but, since volts × coulombs = joules = newtons × metres, the field can also be measured in volts/metre.

The field has a direction and so is a vector quantity. It is easy to see that the direction of the field is opposite to that of the rate of change of potential, or potential gradient; the component of the field in any direction is equal in magnitude but opposite in sign to the rate of increase of potential in that direction.

Expressed mathematically,

$$\mathbf{E} = -\text{grad}\, V \quad \text{and} \quad E_r = -\partial V/\partial r.$$

In most of the phenomena of atmospheric electricity, the equipotentials are horizontal and the fields are vertical; at the horizontal surface of the conducting earth the field must always be vertical, whatever may be its direction higher up.

In normal fine-weather conditions the potential V increases with height.

2.7. Sign Convention for Field

The field is a vector and, as such, has a direction but not, of necessity, a sign until some convention of sign is laid down.

If we take the fairly obvious convention that height is measured positively upwards, then, in fine weather, $\partial V/\partial h$ is positive, and so E must be negative. This is to be interpreted that, in fine weather, a positive charge tends to move downwards.

Unfortunately, in most of the literature on atmospheric electricity, the term "field" has been used to denote $\partial V/\partial h$ and not $-\partial V/\partial h$, when h is measured positively upwards; this probably goes back to the time before theoretical electrostatics was formalized, and then the fine-weather field was conventionally taken as positive. It is now far too late to attempt to re-educate workers in the subject and rewrite the literature to use the word "field" with its true electrostatic meaning and sign, and, if it were seriously attempted, signs in the literature would become chaotic; an example of this is seen in a paper by GUNN (1954b) in which he used "field" with its correct electrostatic meaning and so obtained field-time curves which must be inverted to be compared with those given by other workers.

Although "field" is a much less clumsy term than any alternative, such as "potential gradient", the difference in usage as between ordinary electrostatics and atmospheric electricity seems to make it necessary to abandon the use of the word "field" except where there is no question of sign involved.

Much time and effort have been wasted in discussion of the sign convention, wasted because the whole question is merely one of stating what is meant, and not one of physical principles. DOLEZALEK (1960) has argued that it could be justified to keep to the old meaning of "field"; this would amount to considering the vertical space direction to be measured vertically *down*wards (see § 2.9.). ISRAËL (1962) suggested it would be better, initially, to describe conditions as "a field in the normal fair-weather direction" or in the opposite direction, without, at first, ascribing any sign.

It is now time, the author believes, to cease these unprofitable discussions and to reach the decision that any worker may use whichever sign convention he pleases, provided that: (a) he is consistent in its use, and (b) he makes it clear to others what his convention is, preferably by reference to normal fine-weather conditions.

In accordance with these principles, in the present work, $F = \partial V/\partial h$ is termed the "potential gradient" and is positive in normal fine-weather conditions.

2.8. Lines of Force, Space Charges and Surface Charges

The phenomena of static electricity are very easily pictured in terms of lines of force which commence on a positive charge and end on a negative charge. The density of lines of force across any area gives a measure of the potential gradient and this can be made precise if it is considered that one line of force commences on each unit positive charge (1 C). If, as we move vertically, there is a change in the density of lines of force, but these remain everywhere vertical, then there must be charges present on which the extra lines of force have commenced or on which the missing lines of force have ended. Thus a change of potential gradient with height involves space charges.

Mathematically,

$$\varepsilon_0 \, d^2V/dh^2 = \varepsilon_0 \, dF/dh = -\varrho,$$

where V is the potential in V, F the vertical potential gradient in V/m, h the vertical distance from the earth's surface, ϱ the volume density of charge in C/m^3 and ε_0 the permittivity of free space. Similarly, when lines of force end on a conductor, for example, the surface of the earth, there must be a surface density of charge σ in C/m^2, given by:

$$\varepsilon_0 F = -\sigma.$$

It should be noted that these formulae differ in sign from the usual formulae in electrostatics, because F is not the electric intensity but the potential gradient.

It is perhaps advisable to point out that the formula $\varepsilon_0 F = -\sigma$ is applicable only when the charge σ resides on a conductor, so that lines of force go in one direction only; it is not applicable, for example, to the total charge in a cloud, even a widespread cloud, as lines of force from this can go both upwards and downwards.

Figure 3 shows the lines of force and the charges at their ends in a typical case of a positive potential gradient which decreases with height above the earth.

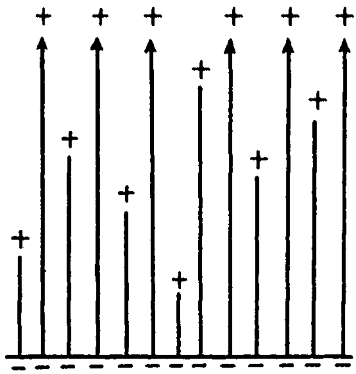


FIG. 3. Lines of force in the earth's normal potential gradient.

2.9. Sign Convention for Currents

In the discussion of sign conventions for field and potential gradient (§ 2.7.), it was considered suitable to measure the h-coordinate vertically upwards. It would therefore seem logical to measure currents in the atmosphere also upwards. But, in fact, actual measurements are almost always of the current coming *to* the earth and it has become customary to consider a positive current to represent a positive charge coming to the earth. As in the case of "field" it is too late to re-educate workers and rewrite the literature to use a more consistent convention.

With the fine-weather field taken conventionally as positive and a fine-weather conduction current giving, on the above convention, a positive value also, no apparent difficulty occurs with OHM's law and no inconsistency was apparent.

In the present work we shall retain the usual sign convention for currents, namely, that a positive current brings positive charge to the earth; if F is the potential gradient, as discussed in § 2.7., then there is no difficulty of sign if one uses OHM's law in the form $i = \lambda F$. It can be pointed out that if the positive direction of h is taken as *down*wards, then F is the field and i the current in this direction, and some of the difficulties disappear; it is, however, contrary to the usual custom to measure downwards as positive. DOLEZALEK (1960a) considered the sign convention for currents as well as for fields.

As in the case of the potential gradient or field, it it essential that authors should state clearly what they mean by a positive current.

2.10. Location of Charges

The problem of the location of charges in the atmosphere from electrostatic measurements can best be understood in terms of lines of force. Charges exist at the ends of lines of force and electrostatic measurements consist essentially in finding the direction and density of lines of force. Therefore, to find out whether lines of force end within any given volume it would be necessary to measure the number of lines entering and leaving the volume over the whole of its boundary. So to get a complete knowledge of the distribution of charge in the atmosphere it would be necessary to make measurements at all points of the atmosphere simultaneously.

Although it is impossible to determine uniquely the distribution of charge corresponding to a limited number of potential-gradient measurements, it is possible, often, to guess the distribution of charge and to show that this is not inconsistent with the measurements; other charge distributions are not precluded, but in many cases they would be unreasonable from other grounds. When it is possible to make measurements at different times as well as at different places, reasonable assumptions in regard to the movement of charges can also be verified.

2.11. Numerical Values

It is convenient to give here some numerical values relating potential gradients and charges.

From the formula $\varepsilon_0 F = -\sigma$ a potential gradient of 100 V/m corresponds to a surface density of charge of $-8{\cdot}854 \times 10^{-10}$ C/m^2 on the earth, i.e. -1 C on an area of about 1130 km^2. At the same time, there must be a corresponding positive charge somewhere above.

If a charge of 1 C is placed evenly on a sphere, of whatever radius less than 1 km, centred at 1 km above the ground, the potential gradient at the ground due to it, at a point directly below it, is

$2/(4\pi\varepsilon_0 \times 10^6)$ V/m, taking into account the electrical image (§ 2.19.). This is $1{\cdot}8 \times 10^4$ V/m. Thus the normal potential gradient of 100 V/m would be entirely neutralized by a charge of $-5{\cdot}5 \times 10^{-3}$ C centred at 1 km vertically above the point of measurement.

2.12. Ionic Conductivity

The conductivity of the atmosphere is produced by ions which carry charges of either sign and which move in the field. As is true in most cases of current electricity, OHM's law is usually obeyed, and the current between two equipotential surfaces is proportional to the potential difference. If i is the current density and X the potential difference, then $i/X = 1/\varrho$, where ϱ is the specific resistance. $1/\varrho$ can be replaced by λ, the "specific conductivity", measured in reciprocal ohms or Ω^{-1} (the term "mho" is sometimes used); "conductivity" is often used where specific conductivity is meant.

In considerations of atmospheric electricity, it is often more convenient to discuss conductivity rather than resistance; this can be readily understood when it is realized that different types of ions, or even individual ions, can be considered to act as conductors in parallel, so that conductivities, but not resistances, can be added simply; thus we can obtain the total specific conductivity by adding together the separate conductivities due to the different ions.

2.13. Mobility

In a given potential gradient an ion moves with a certain velocity which depends on the properties of the ion. The velocity acquired in a unit potential gradient (1 V/m) is termed the "mobility" of the ion; the velocity in a potential gradient of X V/m is wX m/sec, where w is the mobility in M.K.S. units.

At first sight it might appear than an ion, carrying an electric charge and situated in an electric field, would experience a definite force and would therefore be accelerated, acquiring a velocity which would not be constant, but would continually increase. However, this neglects the fact that the ion experiences incessant impacts with the molecules of the air and so loses, at each impact, some or all of the momentum it has acquired since the previous impact. There will thus be an average velocity of travel of the ion in the direction of the potential gradient, and proportional to the potential gradient; clearly, the longer the mean free path of the ion, the greater is the mobility.

Instead of considering the matter from the molecular viewpoint, the ion can be thought of as a small particle moving in a viscous medium; with a constant force on the particle it acquires a steady velocity, such that the viscous force tending to retard the motion just balances the accelerating force. In the case of a small spherical particle falling under gravity the result is the familiar STOKES's law. An ion moving in an electric field will be an exactly similar case and an ion of a particular size and shape will acquire a definite velocity in a given potential gradient. If the forces are such that the ordinary laws of viscosity apply, the steady velocity is proportional to the force acting, and so to the potential gradient.

It might be thought that it would be possible to calculate the size of an ion from its mobility, using STOKES's law or a suitable modification thereof, but this cannot be done because the charge on the ion introduces extra retarding forces arising from polarization and these are not calculable simply.

Mobilities ought to be measured in the M.K.S. system of units of m/sec for a potential gradient of 1 V/m, but the customary unit has been that of the C.G.S. system, namely cm/sec for 1 V/cm, and, to avoid the change of units, the old unit will often be used here, but described as "C.G.S. unit". It must be remembered, for insertion in M.K.S. formulae, that the mobility to be used is 10^{-4} times that in C.G.S. units. It may be noted that the C.G.S. unit is neither the E.S.U. nor the E.M.U. value for mobility.

2.14. Diffusion Coefficient

Diffusion is a process by which any kind of particle tends to reduce its concentration gradient, since more particles tend to move in one direction than in the other.

The diffusion coefficient, D, for particles in a gas at rest is defined by:

$$dN/dt = -D\, dZ/dx,$$

where dN/dt is the number of particles crossing unit cross-section in unit time, perpendicular to the concentration gradient dZ/dx.

EINSTEIN (1905) derived from his theory of the Brownian movement that:

$$D = kTw/e,$$

where k is BOLTZMANN's constant, T the absolute temperature, w the mobility and e the charge.

Diffusion in the sense of the above equation, is not of importance in the free atmosphere, since there are other processes which produce similar effects and are much more important.

The diffusion coefficient can, however, be used for measurement of the properties of large ions and condensation nuclei, and the radius of a particle can be evaluated from its diffusion coefficient; this is probably not very accurate for particles carrying charges, since electrical forces reduce the mean free path.

2.15. Eddy Diffusion

In the free atmosphere the movement of particles in a concentration gradient is not only that of simple diffusion, in which the motion is that described by the kinetic theory of gases; there is also movement of particles by reason of larger scale motion by convection. There is a "coefficient of eddy diffusion", usually much larger than the ordinary coefficient of diffusion, and defined by a similar formula.

The magnitude of the coefficient of eddy diffusion depends on the meteorological conditions, particularly on the temperature gradient, and varies not only from one time to another, but also from one place to another at the same time.

The phenomenon of eddy diffusion is that which brings about the mixing within the *austausch* region (see § 2.32.).

Eddy diffusion causes the movement of any unbalanced space charges in the atmosphere, but cannot produce any vertical electric current in the atmosphere unless there is a variation of space charge density with height. Since the space charge depends on d^2V/dx^2 or dF/dx, an electric current of eddy diffusion depends on d^3V/dx^3 or d^2F/dx^2.

2.16. Ionic Current

If there are n_1 positive ions per m^3, each carrying a charge e_1, and having a mobility w_1, then in a potential gradient X the current due to positive ions crossing 1 m^2 in the atmosphere is:

$$i_1 = n_1 e_1 w_1 X,$$

which can be written as $\lambda_1 X$, where λ_1 is the specific conductivity for positive ions.

Similarly, for negative ions, the current is:

$$i_2 = n_2 e_2 w_2 X = \lambda_2 X,$$

the suffix 2 referring to the negative ions.

The total current density is

$$i = i_1 + i_2 = (\lambda_1 + \lambda_2)\, X.$$

If the positive ions consist of groups of different mobilities

$$\lambda_1 = \Sigma\, n_1 e_1 w_1,$$

and similarly for negative ions.

It is found that the ionic current in the atmosphere is carried almost entirely by the small ions, whose mobilities are between 1 and 2 C.G.S. units; the part played by larger ions is usually negligible as far as carrying current is concerned, although they play an important part in the question of the number of small ions present.

Since positive ions moving downwards and negative ions moving upwards both contribute to a positive current in a positive potential gradient, it follows that the product ew must be positive; so, if e_2 is negative, w_2 should also be taken as negative.

2.17. Velocities of Ions

It is important to realize how small is the velocity of even a small ion in the normal atmospheric electric field. In the normal fine-weather field of about 100 V/m, the velocity of a small ion in the field is only about $1{\cdot}5 \times 10^{-2}$ m/sec. This can be compared with an average velocity, on the kinetic theory, of an air molecule of about 4×10^{2} m/sec, and, even if an ion is to be considered as composed of a number of molecules, its random velocity is still very large compared with that due to its charge.

Even a light wind of 5 m.p.h. corresponds to a velocity of 2·2 m/sec, so that this is large compared with the velocity of an ion in a field. In addition, ions will share any eddy motion of the air.

2.18. Saturation Current

Under ordinary conditions the ionic current obeys OHM's law and the conductivity is defined as in § 2.16. But there are cases in which OHM's law is not obeyed because the supply of ions is insufficient. If a graph is drawn of the ionic current in such a system against the potential difference, the result will be as shown in Fig. 4; for low values of potential difference there is strict proportionality, the slope giving the conductivity, but since the number of ions present is limited, there is a limit to the current, namely, the number of ions produced in the volume concerned in unit time or otherwise entering the volume, e.g. by an air current; this is termed the "saturation current".

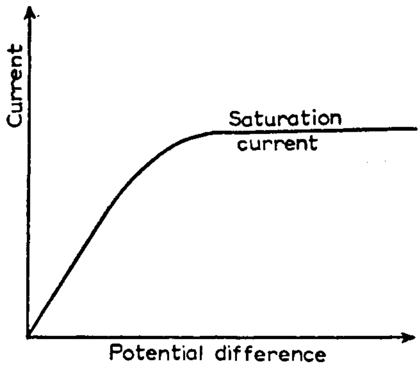


FIG. 4. Ionic and saturation current.

In an enclosed space, the saturation current corresponds to the rate of production of ions in the space and this may be reached with quite small potential differences, so that it is seldom that a true conductivity can be measured in an enclosed space.

2.19. Electrical Images

The earth behaves as a good conductor in all considerations of atmospheric electricity. Therefore lines of force must end on the earth's surface and must proceed at first vertically from the earth's surface.

When considering the potential gradient at the earth's surface due to charges in clouds or even on earth-connected bodies above the surface, it is necessary to take into account the effect of charges induced on the earth. Textbooks of electricity justify the use of the method of images, according to which the effect of the induced charges can be reproduced by adding to the charges above the surface of the earth equal and opposite charges at the same distances below the surface; it can easily be seen that the potential gradient at the surface will then be everywhere vertical. Effects of the curvature of the earth's surface are neglected.

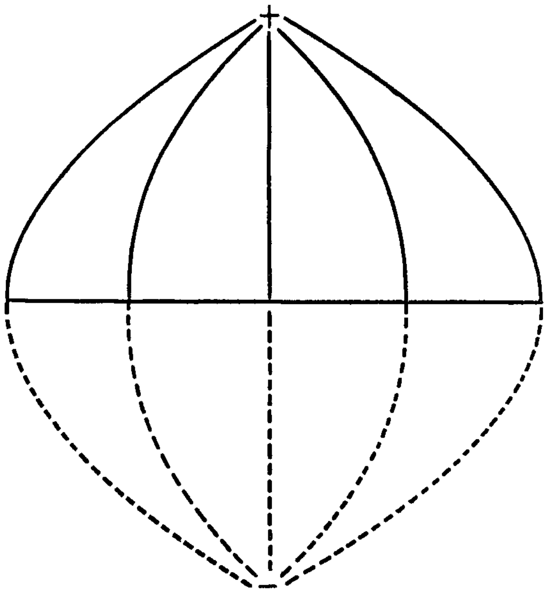


FIG. 5. Electrical images.

It would, perhaps, be more complete also to include images in the conducting electrosphere, and then multiple images in both earth and electrosphere; but there should then be a further complication of the image effects of the space charges arising from alterations of conductivity. As far as is known, these extra complications have not yet been dealt with in calculations of potential gradient.

2.20. Conductor in Potential Gradient

A conductor placed in an electric field will have charges induced on it, corresponding to lines of force ending on it. In addition, the conductor may have a resultant charge of its own.

The distribution of charges over the surface of the conductor can be calculated if the conductor has a simple shape and if its potential or total charge is known. The principles of such calculations are, first, that the conductor must be all at the same potential; and, second, that potentials add. Thus the potential at any point of the surface of the conductor must be the sum of: (a) the potential due to the external charges which give rise to the field, and so the potential which would exist at that point in the absence of the conductor; and (b) the potential due to the charges on the conductor. The condition that now has to be satisfied is that the sum of (a) and (b) is the same for all points on the surface of the conductor. For simple shapes this problem can easily be solved, and the reader is referred to textbooks of electricity.

Because potentials add and potentials are proportional to charges, it can be shown that the distribution of charges on a conductor can be represented by the sum of: (1) the charges on the conductor in the field if its total charge is zero and so the charges if its potential is the same as that of its surroundings; and (2) the charges on the conductor if it has its resultant charge in no external field.

Considering the usual case in atmospheric electricity of a vertical potential gradient and assuming a conductor with a horizontal plane of symmetry, we can choose two portions of the surface of the conductor which are symmetrical with respect to this plane. We then see that the charges due to (1) above will be equal but opposite in sign, while those due to (2) will be equal and of the same sign. Thus, if we now take the difference of the charges on the two surfaces, we get a value proportional to the external potential gradient and independent of the charge or potential of the conductor; while if we take the sum of the charges, we get a value proportional to the charge on the conductor or to the potential difference between the conductor and its surroundings. These results are made use of in the double field mill (see § 5.32.).

2.21. The Electrosphere

KELVIN (1860b) was the first to suggest the existence of a conducting layer in the upper atmosphere, in relation to atmospheric electricity, basing the idea on the fact that gases at low pressures act as conductors; although this is no longer considered as the true reason for the existence of a conducting layer, that does not detract from the interest of the early suggestion. KELVIN also put forword the idea that the earth and this layer would act as a gigantic condenser.

Definite evidence that the atmosphere at heights over about 100 km is a good electrical conductor has come from observations on the reflections of radio waves. The existence of such a layer was postulated to explain long-distance transmissions almost simultane ously by HEAVISIDE (1902) and KENNELLY (1902), and the region has often been spoken of as the "Heaviside layer". With the discovery that there is not just one such conducting layer, it has been considered preferable to adopt some other name for the whole region which reflects radio waves and the term "ionosphere" is now generally used.

From the point of view of atmospheric electricity, the interest in a conducting region in the upper atmosphere is not concerned with the reflection of radio waves, but with the conductivity being so high, in relation to the currents flowing, that the whole region is at effectively the same potential. While the conductivity of the ionosphere is certainly high enough to achieve this, it turns out that the height at which the equipotential condition is reached is around 50 km (see § 11.18.) and it is therefore desirable to distinguish this region from the ionosphere, as defined by reflections of radio waves. Some workers have used the term "equalizing layer", but a better term is "electrosphere"; in much of the literature of atmospheric electricity (including the earlier edition of this book), the term "ionosphere" is used for what we now call the "electrosphere". One advantage of distinguishing the electrosphere from the ionosphere is that it is now possible to define atmospheric electricity as electrical phenomena between the electrosphere and the earth, thus excluding the properties of the ionosphere itself.

The electrosphere is all at a definite potential, different from that of the earth (see § 11.2.). It is believed that the potential difference between the electrosphere and the earth fluctuates with the time of day and probably with the season. We shall also see (§ 11.6.) that the total charge on the inner side of the electrosphere, regarded as a conductor, is zero.

Since the electrosphere is a good conductor, it acts as a perfect electrostatic shield. Thus thunderstorms or other electrical phenomena close to the earth can have no effect outside the electrosphere, since any lines of force from them must terminate at the lower side of the electrosphere. In the same way, any charges arriving at the ionosphere or electrosphere from outside can have no effect inside unless they penetrate right through the electrosphere. This explains, for example, why sunspots, which are believed to involve the emission of large numbers of electrified particles towards the earth, have little, if any, effect on the phenomena of atmospheric electricity, although there are marked effects on the reflection of radio waves and on terrestrial magnetism, both of which depend on charges in the ionosphere, well above the electrosphere. On the other hand, cosmic rays, which can actually penetrate inside the electrosphere, can produce effects in the region which concerns atmospheric electricity.

Kasemir (1965a) has given reasons for believing that some of the current between the earth and the electrosphere does, in fact, penetrate the electrosphere and the ionosphere, particularly in high latitudes. On the other hand, Frenkel (1949) considered all the lines of force from the upper pole of a thunder cloud to return to the earth, rather than to the electrosphere.

2.22. Potential of the Electrosphere

In fair or fine weather there is a positive potential gradient in the atmosphere, so the potential with respect to earth increases on rising within a fair-weather column, giving a positive potential difference between the electrosphere and the earth wherever there is fine weather. The electrosphere and the earth are both good conductors and there are only small currents flowing in each, so that the whole of each is practically at the same potential. Since there is always some part of the earth experiencing fine weather, it follows that the electrosphere is always at a positive potential with respect to the earth, though the actual value of this potential may vary with the time of day and time of year. Measurements have shown that the actual value of the average potential of the electrosphere is about $2{\cdot}9 \times 10^5$ V (see § 11.2.).

2.23. Quasi-static State

Although a current is flowing in the atmosphere, and so the actual charges do not remain static, yet in steady conditions many of the phenomena of atmospheric electricity show a "quasi-static" state, charges which have been moved from any region being replaced by other charges which have arrived in this region, so that instantaneous pictures of the distribution of charge taken at different times would be the same. In these quasi-static conditions (sometimes referred to as conditions of "dynamic equilibrium"), it is always assumed that the laws of electrostatics can still be applied, and there is no reason for doubting this assumption.

When a quasi-static state exists, we can consider not only that the distribution of charges remains the same, but also that there is continuity of current. For the simple case where all motion of charge is vertical, this means that there must be the same vertical current flowing at all levels; where charges are moved by different mechanisms at different levels, some important conclusions can be drawn from this application of "the principle of the quasi-static state", as it may be called.

Care is necessary when attempts are made to extend the principle of the quasi-static state to cases when conditions are not steady, but are altering. After any change, conditions approach a new quasi-static state at a rate which depends upon the relaxation time (see § 2.27.), so that, if we are concerned with changes over times long in comparison with the relaxation time, the conditions are close to quasi-static all the time, but for changes which are more rapid, it will not be correct to use results for the quasi-static state.

Not only can the principle of the quasi-static state be used when conditions are stationary or altering slowly, but it can also be applied, at any rate in certain circumstances, to average results over a long period of time for phenomena which are changing rapidly.

A simple application of the ideas of the quasi-static state is in the use of OHM's law. For steady conditions the conduction current in the atmosphere is proportional to the potential gradient and for slow changes the current follows the potential gradient almost exactly; if we consider the diurnal variation of potential gradient, we find that the current maximum lags behind the potential-gradient maximum by a time equal to the relaxation time, a difference which is inappreciable under the conditions of measurement.

2.24. Fundamental Formulae for Fine Weather

Provided that, first, conditions are sufficiently steady for the principle of the quasi-static state to be used and, second, that the transfer of charge is entirely by conduction currents, then OHM's law can be applied to give important results.

If V is the potential of the electrosphere with respect to earth and R is the "columnar resistance", i.e. the resistance of a column of the atmosphere of 1 m^2 cross-section from the earth to the electrosphere, then the conduction current density i is given by:

$$i = V/R.$$

If r is the resistance of the lowest metre of this column, the potential drop across this metre must be F, the average potential gradient over the lowest metre, and thus:

$$F = ir = Vr/R.$$

Or, using λ, the specific conductivity of the lowest metre, rather than r,

$$F = V/R\lambda.$$

Thus variations of F in fine weather may be ascribed to variations in any of the three quantities, V, λ or R; in most circumstances it is λ which has the greatest amount of variation and so measurements of the potential gradient usually merely give information mainly about changes in the local conductivity and not about anything more fundamental. Measurements of i, however, do not involve changes in λ, except in so far as these produce changes in R.

Care must be taken to use these results only in conditions when they are correctly applicable, as discussed above. Otherwise (see, for example, § 11.2.) errors can arise.

2.25. Variation of Potential Gradient and Conductivity with Height

Measurements of the potential gradient in fine weather at different heights above the earth's surface have shown that there is little change in the first few metres, in spite of what would be expected from the electrode effect (§ 2.31.); we shall leave until later (§ 8.15.) the discussion of the non-appearance of the electrode effect. At heights above a hundred or so metres the potential gradient shows a progressive decrease with height.

Measurements of the conductivity in fine weather show that the sum of the positive and negative conductivities is nearly constant for the first few metres, but shows a marked increase with height above about a hundred metres. There are two causes for this increase; in the first place, the ionization due to cosmic rays increases with altitude; also, on rising, there is less pollution and so fewer nuclei which replace ions of high mobility by larger and slower ions.

The phenomena of the decrease of potential gradient and increase of conductivity with height are not independent, but are related by the fact that the product of the potential gradient and the conductivity gives the current density, and, as we have already seen, in quasi-static conditions the current density must be the same at all levels. The fundamental fact is the change in conductivity, and the change in potential gradient is a consequence of this.

The above discussion in terms of current density gives all the information that might be required in regard to the values of the potential gradients at different heights, without needing to take any account of space charges. However, to obtain a complete picture of the location of charges we can consider lines of force, the density of which gives the potential gradient. If a positive potential gradient decreases on rising, there must be fewer lines of force higher up than lower down, and the missing lines of force must have positive charges at their upper ends, giving a positive space charge and hence an excess of positive over negative ions.

It is interesting and instructive to consider how a change of conductivity with height could actually bring about the change of potential gradient with height and the positive space charge, starting with an original condition of uniform potential gradient all the way from the earth to the electrosphere. Then lines of force would be continuous and there would be no space charge. Now, because the conductivity increases with height, there will be a larger vertical current higher up than lower down, and so, if ions of both signs carry the current, there will be more positive ions moving down at a high level than at a lower level and so the number of positive ions in any volume will increase; in the same way, since negative ions move upwards, the number of negative ions in a volume will decrease. Any volume in the air thus acquires a positive space charge which produces a difference in potential gradient as between the top and bottom of the volume. This process will continue until the numbers of positive and negative ions entering and leaving the volume are equal, the differences in conductivity at the top and bottom being balanced by differences in potential gradient.

2.26. Relation between Space Charge and Change of Conductivity

From § 2.8., the space charge ϱ is given by:

$$\varrho = -\varepsilon_0 \, dF/dh,$$

where F is the potential gradient, provided that the equipotentials are horizontal.

If i is the current density and r the specific resistance,

$$F = ri.$$

If there are quasi-static conditions, i is the same at all levels, so that:

$$dF/dh = i\,dr/dh,$$

and

$$\varrho = -\varepsilon_0 i\,dr/dh.$$

If we consider a boundary between two regions of specific resistances r_1 and r_2, then, by integration across the boundary, the total charge per unit area on the boundary is:

$$Q = \varepsilon_0 i(r_1 - r_2).$$

If we consider conductivities rather than specific resistances:

$$\varrho = \varepsilon_0 \frac{i}{\lambda^2} \frac{d\lambda}{dh},$$

and

$$Q = \varepsilon_0 i(1/\lambda_1 - 1/\lambda_2).$$

These results were obtained by REITER (1955b) in a slightly different way.

2.27. Relaxation Time

If a portion of a conductor, of area S, carries a charge Q and is exposed to the atmosphere, the charge Q is gradually dissipated by ionic conduction and the rate at which this occurs gives a "time constant" or "relaxation time" for electrical phenomena in the atmosphere.

The surface density of charge is Q/S if the charge is uniformly distributed and so the potential gradient close to the surface is $Q/S\varepsilon_0$; the current, of negative ions if Q is positive, is thus:

$$\lambda_2 Q/S\varepsilon_0 \qquad \text{per unit area}$$

and

$$\lambda_2 Q/\varepsilon_0 \qquad \text{for the whole area.}$$

Thus

$$\frac{dQ}{dt} = -\frac{\lambda_2 Q}{\varepsilon_0}$$

which integrates to:

$$Q = Q_0 \exp(-\lambda_2 t/\varepsilon_0),$$

giving a time τ to reach $1/e$ of the original value Q_0 where:

$$\tau = \varepsilon_0/\lambda_2.$$

Similarly, for a negatively charged body

$$\tau = \varepsilon_0/\lambda_1 .$$

These results are valid only if λ_2 (or λ_1) remains constant, i. e. if OHM's law is obeyed, and will not be true if the current dQ/dt is sufficient to deplete appreciably the number of ions present.

If, instead of considering a charged body, which can attract ions of one sign only, we consider a horizontal plate in a vertical field, then ions of both signs can arrive at the plate and so there will be a relaxation time $\tau = \varepsilon_0/\lambda$, where $\lambda = \lambda_1 + \lambda_2$.

Actual values of the relaxation time are: for air near the earth's surface, from 5 to 40 min, depending on the amount of pollution; for air at 18 km, 4 sec; for the electrosphere at 70 km, 10^{-8} sec; and for the earth itself, 10^{-6} sec or less.

2.28. Capacitative Effects

For a potential difference maintained constant the current depends on the resistance of the medium, but when the potential difference is varying, capacitative effects are also involved and OHM's law cannot be applied in the simple form.

A unit cube of air has a resistance of $1/\lambda$, where λ is the specific conductivity, and it can be considered to have also a capacitance as though there were conducting plates on two opposite faces; from the simple formula $C = \varepsilon A/d$ the capacitance of the cube is ε_0.

If the resistance and capacitance are considered in parallel, the system has a "time constant" of RC, which is ε_0/λ, in agreement with the result of § 2.27. for the relaxation time.

The relative importance of resistive and capacitative effects can be considered for a sinusoidal change in potential difference, $E = E_0 \sin pt$. The capacitance of a unit cube behaves as a reactance of $1/p\,C$, which is $1/p\varepsilon_0$ and this is to be compared with the resistance $1/\lambda$ so, if $1/p\varepsilon_0$ is large compared with $1/\lambda$, capacitative effective effects are small; this is so if $1/p$ is large compared with ε_0/λ, which is τ, the relaxation time. Now, the period of the sinusoidal variation is $T = 2\pi/p$, so the condition for small capacitative effects, when all the current flows through the resistance and OHM's law is obeyed in its simple form, is that $T/2\pi$ is large compared with τ. On the other hand, if $T/2\pi$ is small compared with τ, capacitative effects preponderate and the current is 90° out of phase with the applied potential difference. Thus slow fluctuations give small capacitative effects but rapid fluctuations give appreciable effects.

2.29. Sudden Changes

Suppose that the lower atmosphere is in a quasi-static state, with a current flowing and space charges as required by differences of conductivity. Now suppose there is a sudden change produced, for example, by a lightning flash, and that, after this, the atmosphere again takes up in time a new quasi-static state. The question arises as to how the atmospheric electric elements vary during the transition, and this was discussed in detail by KASEMIR (1950).

A very simple case is sufficient to illustrate KASEMIR's arguments. Suppose that suddenly a fresh charge appears in a cloud, to give an effect additional to that which exists already. This will immediately produce a potential gradient at the ground, calculable purely electrostatically and visualized as the termination of the lines of force from the charge. But, also immediately, the current flow will be altered from that occurring previously, charges will move and ultimately a new quasi-static state will be set up; some of the lines of force which, immediately after the change, ended on the ground will now end on space charges in the atmosphere. As KASEMIR pointed out, all movements of charge are governed by the relaxation time and hence all changes must depend on this. If F_1 is the potential gradient at the earth's surface immediately after the change and F_2 that when conditions are again quasi-static, then KASEMIR showed that the potential gradient F at any intermediate time is:

$$F = F_1 \exp(-t/\tau) + F_2[1 - \exp(-t/\tau)],$$

where τ is the relaxation time of the atmosphere close to the ground.

If i is the conduction current flowing in the atmosphere, $i = \lambda F$, and so changes in i follow changes in F. But this does not, of course, apply to the current in the ground flowing from a portion of the earth's surface, since this includes the "displacement current" $\varepsilon_0\, dF/dt$.

In the same way, the space charge in any region follows a similar law, but the value of τ is then that at the position of the charge, not at the ground.

If there is a sudden change, not of the distribution of charges but of conductivities at different places, then very much the same considerations apply, except that there is no immediate electrostatic change and F_1 will now refer to the potential gradient before any change and τ is the relaxation time after the change. A sudden change of conductivity would, however, cause a sudden change in the conduction current. The matters here discussed have been carried further by KASEMIR (1963, 1965b).

2.30. Validity of Ohm's Law

DOLEZALEK (1960b) considered in some detail the question whether OHM's law is valid in atmospheric-electric phenomena. Apart from its extension to include capacitative effects, as discussed in § 2.28., DOLEZALEK considered other possible deviations from OHM's law, namely, any dependence of ion density or of mobility on potential gradient or current density; of these, the only one which might be significant is the influence of current density on ion density and this is unlikely to matter in the free atmosphere.

DOLEZALEK (1960b) also pointed out that care must be taken in applying OHM's law to the vertical current in the atmosphere, both in regard to capacitative effects and also in regard to the possible existence of horizontal currents if there are vertical columns of different resistances.

2.31. The Electrode Effect

By the term "electrode effect" is meant the non-uniformity of electrical conditions close to an electrode, and in atmospheric elec-

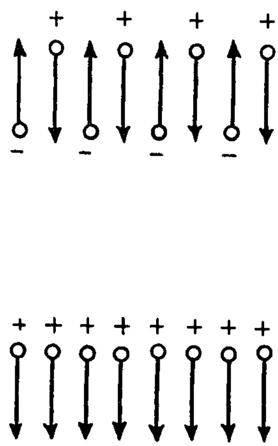


FIG. 6. The electrode effect.

tricity it is usually the earth's surface which is the electrode in question.

If there is a volume of still air between two electrodes and if ions are being produced uniformly throughout the volume, then it might be expected that the electrical conditions would remain uniform throughout the volume when a potential difference is applied between the electrodes. This, however, is not the case.

Consider a volume bounded at one end by the negative electrode and extending into the central region between the electrodes. Positive ions can enter the volume from the central region and leave it to the negative electrode and, in the initial state when there is a uniform potential gradient between the electrodes and a uniform distribution of ions, these two positive-ion currents will be equal. Negative ions can leave the volume into the central region but no negative ions can enter the volume except in the case when they are actually produced at the negative electrode. Thus the volume under consideration becomes depleted in negative ions and so acquires a positive space charge; this must alter the potential gradients, increasing that near the electrode and decreasing that near the centre, and finally a quasi-static state is attained in which the same current flows through all cross-sections, but neither the potential gradient nor the ionic distribution remains uniform. The exact extent to which the potential gradient near the electrode would differ from that in the central region depends on the effects of large ions and nuclei, as will be discussed in § 8.15.

The simple electrode effect assumes quasi-static conditions in still air; if there is turbulent diffusion, the effect becomes more complicated. The simple discussion also assumes a uniform rate of ionization, and if this is not so, the matter is, again, more complex.

The electrode effect can be recognized in various ways, by the change of potential gradient, by the space charge or by the deficit of ions of sign opposite to the potential gradient.

Bent and Hutchinson (1966) proposed a general definition of the electrode effect, as follows: in atmospheric electricity the electrode effect is the modification of elements such as space charge distribution, conductivity and potential gradient near an earthed electrode, which may be a raised object or the surface of the earth itself, because in the prevailing field ions of one sign are attracted towards the electrode, whilst those of the opposite sign are repelled from it.

2.32. "Austausch"

An important phenomenon in the meteorology of the lower atmosphere is that described by the word *austausch* (borrowed from the German), meaning exchange; modern work in atmospheric electricity has shown increasingly the importance of this phenomenon in electrical manifestations. *Austausch* consists in the continual and fairly complete mixing of the air and its contents in the whole of a region extending from the ground up to a definite level which is defined as the top of the *austausch* region. The height of the top of the *austausch* region depends upon meteorological conditions, in particular upon the variation of temperature with height and may have any value up to about 3 km. Within the *austausch* region, there are small-scale convection currents which carry the surface air, together with the ions, nuclei, etc., which it contains, up to the top of the region and ensure a thorough mixing. The upper boundary of the *austausch* region is usually easily visible as the level of fine-weather cumulus clouds or of the top of haze and is shown very markedly by electrical measurements (see § 7.14.).

Where there is a temperature inversion, i. e. where temperature increases with height instead of the usual decrease, *austausch* cannot occur, so that a temperature inversion may coincide with the top of the *austausch* region, and when the temperature inversion is at ground level, there can be little *austausch.*

2.33. Disturbed Weather Phenomena

Whereas in fine weather it is usually possible to consider the lines of force to be vertical and equipotentials horizontal, and to consider that all currents are simple conduction currents, involving Ohm's law, in disturbed weather these assumptions are by no means legitimate. Further, in fine weather the various electrical quantities remain constant or vary only slowly, but in disturbed weather there are rapid alterations.

Thus many of the ideas used in the discussion of fine-weather phenomena are no longer applicable; for example, $\varepsilon_0 \, d^2V/dh^2 = -\varrho$ is no longer true if horizontal variations of potential are taken into account; in addition to the ordinary conduction currents, there may be currents carried by precipitation and by lightning flashes and also alterations of conductivity due to fresh ions produced by point discharge, by lightning and perhaps by other processes; and it is seldom that it is appropriate to apply the principle of the quasi-static state (§ 2.23.).

It is not even correct to assume that the quantities of charge of the two signs in the clouds are equal. Although cloud charges are produced by separation of positive and negative charges by some process or processes, it is not likely that the dissipation of these charges to the electrosphere and to earth will be equal, and so an inequality of the quantities of the two signs will then arise.

We have already seen (§ 2.10.) that observations at the earth's surface cannot definitely locate the charges responsible for the potential gradients measured, but they can show that some charge distributions are possible and others not.

2.34. Relation between Atmospheric Electricity and Meteorology

Although there was a period in the history of atmospheric electricity when meteorological phenomena were considered to be a "disturbing" influence on the "normal" atmospheric electrical phenomena, it has been increasingly realized that, in fact, the phenomena of atmospheric electricity are very closely bound up with meteorological phenomena and that each can contribute to the understanding of the other. ISRAËL (1955b) has pointed out that much information can be obtained by using the meteorological synoptic methods for the discussion of atmospheric electricity; he has also emphasized that it is highly desirable to measure not only one element, e.g. potential gradient, but also a second, either conductivity or air–earth current, and preferably all three, as this makes it possible to distinguish local effects from those of wider range and significance.

2.35. Electric Climatology and Synopsis

ISRAËL (1957) has pointed out that observations in atmospheric electricity can yield two different types of result. If measurements are made of electric elements over a long period of time, and, in some way, "undisturbed" observations are selected, then one can obtain mean values and the variations with time of day and with season; such results give what ISRAËL calls an "electric climatology". On the other hand, if simultaneous measurements are made at a number of places, one can obtain an "electric synopsis" and hope to get the consequences of disturbing influences.

DOLEZALEK (1958) discussed the requirements for an adequate electric synopsis and pointed out that electrical measurements might be able to help in the elucidation of meteorological problems since there may be influences which produce similar effects on some meteorological observations, but could be distinguished by their electrical effects.

2.36. Relation between Atmospheric Electricity and Phenomena in the Upper Atmosphere

There are a number of phenomena in the higher regions of the atmosphere which might be expected to affect the electric state of the lower atmosphere if the electrosphere does not act as a complete electrostatic shield; in particular, magnetic storms and aurorae could be suggested.

GISH (1931, 1944b) discussed the evidence for any relations between these phenomena and any elements of atmospheric electricity and came to the conclusion that any correlation is very small. ISRAËL (1947) quoted results suggesting an effect of the aurora (see § 5.56.). HOGG (1955) looked for an effect of sunspots on the air–earth conduction current, but failed to find any relation (see § 8.25.).

2.37. Correlation of Meteorological and Physiological Phenomena

While many physiological phenomena can be directly connected with meteorological factors, of which temperature and humidity are the most obvious, there are some cases, for example, of patients in closed hospital rooms maintained in constant conditions, where the outside, and unperceived, weather appears to have definite effects physiologically and, perhaps in consequence, psychologically. Various suggestions have been made as to the physical mechanism by which such effects might occur, and one of these is that the factor is one concerned with atmospheric electricity, perhaps the ion content. Some evidence for this has been discussed by SCHILLING and CARSON (1953) and SCHILLING and HOLZER (1954) (see § 7.24.).

Further work on this subject and the related phenomenon of the effect of electrical parameters on plant growth has been carried out by HICKS (1956), KORNBLUCH and GRIFFIN (1955), KRUEGER (1962) and KRUEGER, KOTAKA and ANDRIESE (1962), among others.

2.38. Measurements of Atmospheric Electricity

Various quantities concerned with atmospheric electricity have been observed and measured for longer or shorter periods or spasmodically.

The greatest number of observations has been made on the vertical potential gradient at the surface of the earth, some of them extending over a considerable number of years. These have given information on the diurnal and annual variation of potential gradient at different localities, but, as we have seen in § 2.24, deductions from the observations cannot establish much more than the variation of the local conductivity, except where it is known that this remains constant. Measurements of potential gradient have also been made in aircraft, in gliders and in balloons.

Measurements of conductivity, and also in many cases the conductivities due to ions of each sign, have also been carried out fairly extensively. Combining these with measurements of the potential gradient, the air–earth current is obtained by the "indirect" method. Conductivity measurements have been made in aircraft, gliders and balloons.

The number of ions of either sign, within various ranges of mobility, can also be measured. If the mobilities are also measured, this forms an indirect method of arriving at the conductivity.

The space charge in the atmosphere can be measured by direct methods or it can be deduced from measurements of the potential gradient at different heights.

The air–earth current has been measured by direct methods, with correction or compensation for potential-gradient changes.

The number of condensation (or Aitken) nuclei has been measured, also the proportion of these which are charged.

Attempts have been made to measure the rate of production of ions, but direct methods require correction for the effect of the vessel used.

All the above measurements are concerned with fine- or fair-weather phenomena, but some, e.g. potential gradient, have also been carried out in stormy weather, though others clearly cannot.

Other measurements during stormy weather include precipitation currents, both in bulk and for single drops, point-discharge currents, potential-gradient changes due to lightning, actual lightning currents and the photographic study of lightning with moving cameras.

Atmospheric Electricity

A great many of the observations have been made in or near large towns, where the conditions are disturbed by atmospheric pollution which varies with the time of day and with the season of the year, but some important measurements have been carried out at sea, notably on the cruises of the *Carnegie*, and at other places where the atmosphere is unpolluted. An extensive series of observations has been carried out at Watheroo in Western Australia and another at Huancayo in Peru, as well as less continuous observations in polar regions and on mountains.

In the post-war period there has been less emphasis on long-continued measurements and more on those of shorter duration intended to solve particular problems. But the range and type of measurements have been extended, and, in particular, more measurements have been made above ground level.

A mountain-top observatory for atmospheric electricity at Mount Withington in New Mexico is a very significant addition to facilities available.

2.39. Variability of Results

In contrast to many other branches of physics, where the same conditions may be reproduced almost exactly and the same experiment repeated, in atmospheric electricity the conditions can never be twice quite the same, owing to variations in meteorological factors which are not under control. Consequently, the measurements of any atmospheric-electrical quantity are subject to a variability which depends not only on observational and instrumental inaccuracies, but also on factors which are uncontrollable and may not even be recognized as of importance.

To obtain results of general application, the alternatives are to use a large number of observations, spread over very different conditions, and take a mean value, or else to use only measurements under very carefully specified conditions, and realize that the results are not applicable to any other conditions. In the first case it may be necessary to exclude some kinds of observations, such as those under disturbed weather conditions, and some criterion is then needed for exclusion; even then, with the wide variability of conditions, a very wide spread of results will be likely, and any individual observation may depart from the mean quite widely. In the other case, when conditions are closely specified, the observations can be repeated only infrequently and there may be other unspecified conditions which are altering and giving a variability to the results.

Where general numerical results are given, as in Appendix II, these are mean values over a period of time, based on a large number of observations. In some cases, where the results vary according to the place on the earth, it is quite possible that observations have not yet been taken at enough places for a fair sample to be available, so that the results may not be representative of the earth as a whole and such generalizations as have been made may not be valid. This is particularly the case for stormy weather phenomena for which observations from tropical regions are sadly meagre.

2.40. Application of Statistical Methods

Because of the wide variation of results, due to variability of conditions, some care is necessary in the application of statistical methods to the results; methods which depend upon the assumption of a "normal" distribution may give incorrect results in cases where the conditions provide a distribution which is far from normal.

It should be pointed out that, even if statistical methods show a high correlation coefficient between two quantities A and B, it cannot be stated that the variation of A is the cause of the variation of B or vice versa; it may well be that, physically, they are quite unconnected and that the variation of each is connected, independently, with the variation of some other quantity C.

Another difficulty may arise in the use of methods for obtaining the "best straight line" connecting two quantities. In the usual method it is assumed that y is measured for known values of x; but in atmospheric electricity, it is frequently the case that there is as much variability in x as in y, and attempts to use the usual procedure give very different results if the variables are interchanged, treating y as known. A procedure for dealing with such a case has been devised by MORGAN (1960).

2.41. Autocorrelation

In ordinary statistical calculations it is assumed that the observations are all independent of one another; but in all atmospheric conditions, except those of thunderstorms, conditions change only comparatively slowly and the result of a measurement at any one time is more likely to be close to that of a previous measurement than very far from it.

An "autocorrelation function" can be defined which measures the correlation of measurements at any one time with those at other times, and from this one can obtain an idea of a "correlation interval", such that observations within this interval are not to be considered independent.

The main effect of this on the statistical discussion of observations is that, in calculating standard deviations, the number included in the formula as the number of observations is to be taken, not as the total number, but as the number of independent observations, i. e. the total divided by the number in each correlation interval.

The matter is considered in detail by Awe (1964).

2.42. Insulation

One of the most serious problems in any investigation in atmospheric electricity is that of insulation. In any measurement which involves the collection of a charge there is a leakage which takes place with a time constant of CR, where C is the capacitance of the collector and R the leakage resistance to earth. In the actual measurement, it is often the potential which is measured and as, for a given quantity of charge Q, the potential V is Q/C, it is necessary to reduce C as much as possible for the greatest sensitivity. But a reduction of C reduces the leakage time constant and so it is essential to make R as large as possible.

For indoor insulation amber has been very widely used in the past and also sulphur if the surface can be kept warm enough to prevent condensation.

For outdoor insulation, apart from the obvious necessity of protection from rain, etc., and the need to keep surfaces warm to prevent condensation, there is the difficulty of spiders' webs, which are insulating when dry but which conduct when moisture condenses; spiders, not being insects, cannot be discouraged by insecticides such as DDT and the only sure protection is the removal of webs when formed; most spiders build their webs in the early morning hours and these may not become conducting until moisture condenses in the evening, so that the removal of webs during the day may be adequate. Von Kilinski (1958) suggested the breaking of the webs as soon as they are formed, using a slow rotation of one part of the apparatus relative to another.

In most cases of outdoor apparatus it is necessary to provide some form of heating to avoid condensation, but it must be realized that the warmth attracts spiders and insects. If the heating is by an alternating electric current, efficient earthed shielding must be provided between the heater and the rest of the apparatus; this is particularly important if there is some mechanism for making and breaking contact, since if there is any induced potential from the heating current this is varying with the frequency of the current and its effect at the breaking of contact will depend on the exact phase when contact is broken and so will vary from one occasion to another.

For connections from outdoor apparatus to indoor measuring instruments it is usually essential to use coaxial cable with polythene insulation for the signal output. This involves certain problems; in the first place, if polythene is subjected to strain or movement, electric charges can arise, perhaps by piezoelectric effects, and so the cable must be very firmly fixed; then it is necessary to keep warm the exposed end of the cable, otherwise there will be condensation on the polythene and so leakage; also, coaxial cable has a quite considerable capacitance and in many cases this makes the input to the measuring apparatus of too low an impedance; there are two ways of obviating this difficulty, either with a cathode follower stage (see § 2.46.) at the outdoor apparatus, or by applying the potential of feedback (see § 2.48.) to the sheath of the cable.

For an insulator which needs mechanical working, polystyrene is very good, but the surface needs to be cleaned with alcohol, polished and kept warm. If an insulator is required which bears little mechanical strain, paraffin wax is good; it can be cut to shape or melted and poured and is less liable than other insulators to collect moisture on its surface. Porcelain or similar ceramics are also sometimes suitable. For surface insulation in damp conditions, the water-repellent silicones may be useful but, as the silicone needs heating after application, this is not practicable in some cases, e.g. for coating polythene or polystyrene. Fluorine-containing insulators, such as polytetrafluorethylene, have also been used and are less liable to collect moisture than some other materials.

With highly insulating materials there is the danger that they may collect charges on their surfaces during assembly of the apparatus, and these charges may persist for some time with no opportunity to leak away. This can often be remedied by discharging by means of a β-ray source to ionize the air temporarily.

With good modern insulating materials, leakage through the air may become important, particularly if the apparatus is such that it is well exposed to the air, with good ventilation, for there will then be the natural conductivity of the air, with a relaxation time (see § 2.27.) of 10 to 40 min; it turns out, rather surprisingly, that polluted air gives less leakage than clean air. If the air is enclosed, saturation conditions (see § 2.18.) are likely to be reached and the total current can then reach only the rate of production of ions in the volume; this leads to another surprising conclusion, that there is less leakage the smaller the volume of air concerned.

2.43. Earth Connections

Earthing is an important matter in all apparatus for measurements in atmospheric electricity. Care must be taken to use one earth connection only for each apparatus, since different earth connections may be at different potentials, either because of electrolytic potential effects or because of earth currents. Also it is necessary to avoid any possibility of potential differences being set up by currents flowing through a part of an earthing system; if there are alternating currents, or currents with rapid variations, induced currents must be avoided.

Where possible, it is desirable to connect different parts of the apparatus separately to an earthed chassis, earthed at one point only, and to take care that where there is a complete current circuit, the part through the chassis is short and not close to other connections. When the outer coating of coaxial cable is to be earthed it is necessary to avoid current through this, since there would be induced currents in the core.

The usual precautions with electronic apparatus to avoid pick-up must be taken, and it must be realized that an unscreened lead to outdoor apparatus may act as an aerial and pick up radiation, the electronic apparatus then perhaps rectifying a spurious signal.

2.44. Contact Potentials

In some measurements in atmospheric electricity the presence of contact potentials causes difficulties, and it is important that the phenomenon should be clearly understood.

When two metals are in contact the free electrons in the two metals have very nearly the same energy on either side of the boundary; actually, what are equal are the "thermodynamic potentials" on either side of the boundary, or, perhaps more precisely still, the "electrochemical potential" of an electron is the same in each metal. The difference in energy of electrons on either side of the boundary corresponds only to millivolts, as seen in the Peltier effect, and can be neglected here. To remove an electron from a metal to a point outside requires an amount of energy described as the "photoelectric work function" with a value of some electron-volts (the electron-volt is a convenient unit of energy measurement, being the amount of work required to move an electron through a potential difference of a volt). The work function differs from one metal to another, and thus there is a difference of energy between electrons just outside two metals in contact, a difference of the order of electron-volts. If we consider a parallel-plate air condenser with plates of the two metals, then, when the plates are connected directly together, there will be a difference of energy between points on either side of the air gap, so that there appears an electric intensity and a potential difference across the gap. This difference of potential appears with no source of potential in the connecting wire and is described as the "contact potential difference"; it is equal to the difference in the photoelectric work functions for the two metals, divided by the electronic charge.

In measurements in atmospheric electricity, where the potential gradients concerned amount, in fine weather, to some hundreds of volts per metre, contact potential differences of the order of volts will give effects comparable with the natural potential gradients if the metals are separated by distances of the order of centimetres.

A particular difficulty in eliminating the effects of contact potentials arises from the fact that the work function of a metal is much affected by the condition of its surface, so that contact potentials are not eliminated by the use of the same metal throughout the apparatus, and do not remain constant.

Attempts have been made to eliminate contact potentials by the use of chromium-plated nickel, gold or stainless steel surfaces, but even these have not been completely successful.

2.45. Measuring Instruments

In a great many of the measurements in atmospheric electricity it is necessary to measure small currents. In all the earlier work, the most suitable instrument for this purpose was some form of electrometer, one of the most popular being the quadrant electrometer of DOLEZALEK (1901); this measures potential difference but if connected to a system of low capacitance, the rate of deflexion of the electrometer gives a method of measuring small currents. A complete description of many types of electrometer has been given by IMIANITOV (1958).

In recent years, electronic methods have almost completely ousted the electrometer and various forms of direct-current amplifier have been used, in many cases with an "electrometer valve" as the first stage. It is now no longer necessary for the worker in atmospheric electricity to master the temperamental quadrant electrometer, but more and more he needs to become an electronic expert. Since it is possible to amplify to any desired extent, the size of the collector of current can be reduced considerably, but there is, of course, the limit of the "noise level", especially in the first stage. IMIANITOV (1958) has given a detailed description of electronic methods of measurement.

The final instrument used to display or record the output is often a moving-coil galvanometer, and this often must be used in a "dead-beat" form. It is usually desirable for the galvanometer to be "critically damped" as in this condition it responds most closely to the current. A useful device for alteration of sensitivity is the "Ayrton shunt" as illustrated in Fig. 7; $P + Q + R$ is the necessary resistance for critical damping. When the connection is

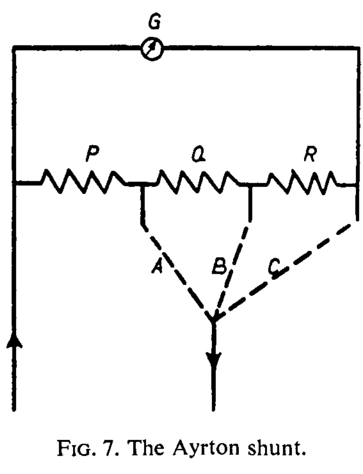


FIG. 7. The Ayrton shunt.

made in position A, the current through the galvanometer is $iP/(P + Q + R + G)$, in positions B and C it is $i(P + Q)/(P + Q + R + G)$ and $i(P + Q + R)/(P + Q + R + G)$, respectively, giving three conditions of sensitivities proportional to P, $(P + Q)$ and $(P + Q + R)$.

2.46. Use of Cathode Follower

When electronic methods are to be used for amplification of currents or voltages produced at some outdoor measuring apparatus, it is often inconvenient for the whole of the amplifier to be situated with the measuring apparatus. On the other hand, if the signal is to be brought indoors, this requires a cable of some considerable capacitance, which will decrease the input impedance of the first stage of the amplifier to such an extent that the signal voltage is too small.

The difficulty can be overcome by the use of a cathode-follower stage situated with the apparatus, and having an input capacitance as small as possible. The output from the cathode follower gives the same signal voltage as the input, but across a much lower impedance, which can consist of the cable from the cathode follower to the remainder of the amplifier. Electronic details can be found in any suitable textbook and will not be given here.

2.47. Direct-current Amplification

Some of the measurements in atmospheric electricity are concerned with small currents, which, by their nature, are unaffected by the resistance in the measuring apparatus; a simple example is that of the measurement of the current brought down by precipitation. The question of electronic amplification in such cases differs considerably from the more usual case of the amplification of a definite voltage.

If the current to be measured is passed through a high resistance, then a potential difference is generated across this resistance, but the amplification of this potential difference by electronic methods is not as simple as might appear.

One method, which in recent years has become the most reliable method, is to use a vibrating-reed electrometer, in which the potential difference to be measured is converted into an alternating current at the frequency of the reed vibrations. This alternating current can then be amplified with circuits tuned to this frequency and later rectified in the correct phase by the use of the same reed (THOMAS and FINCH, 1950). This method requires commercially made apparatus and is therefore expensive.

If it is desired to use direct amplification it is necessary to use a specially designed valve, as ordinary valves have grid currents which are too large, and the current to be measured would flow through the valve instead of aross the resistance. It is necessary to use some form of "electrometer valve" with low grid current, obtained by using reduced potentials, to avoid production of positive ions, and other precautions.

For measurements over a period of time the question of the stability of the circuit becomes important and this can be improved by the use of negative feedback (see § 2.48.). Another device is to use a bridge circuit with two valves which are equally affected by any change in supply voltages.

2.48. Negative Feedback

An important electronic method of considerable use in atmospheric electricity is that of "negative feedback".

The principle of negative feedback can be seen from Fig. 8. An input current i flowing through a resistance R generates a potential difference $V(= Ri)$; if, as in (a), the end Q of the resistance is connected to earth, then P is at a potential $V_0(= V)$ above earth. The amplifier, with a gain of G, gives a potential $V_a(= GV_0 = GRi)$ at the point S. Thus the measured output potential depends linearly on the gain of the amplifier and fluctuates in proportion to fluctuations of G.

If, however, as in (b), the points S and Q are connected, we still have $V_a = GV_0$; the potential drop between P and Q must still be $V = Ri$. So $V_0 - GV_0 = V$, giving $V_0 = V/(1 - G)$ and so $V_a = GV/(1 - G)$. Now, if it is arranged that the amplifier gives a negative output for a positive input, G becomes negative and if G is large, we get $V_a \sim -V$ and virtually independent of G; thus the output potential remains constant, even if the gain of the amplifier fluctuates, giving much increased stability.

Another advantage of the use of feedback is that the potential at P is small, being only $V/(1 - G)$ instead of V; thus if P is connected to a collector this is maintained at a potential quite close to that of the earth, so that there is little distortion of field. The value of G may be some hundreds, so this effect may be quite important.

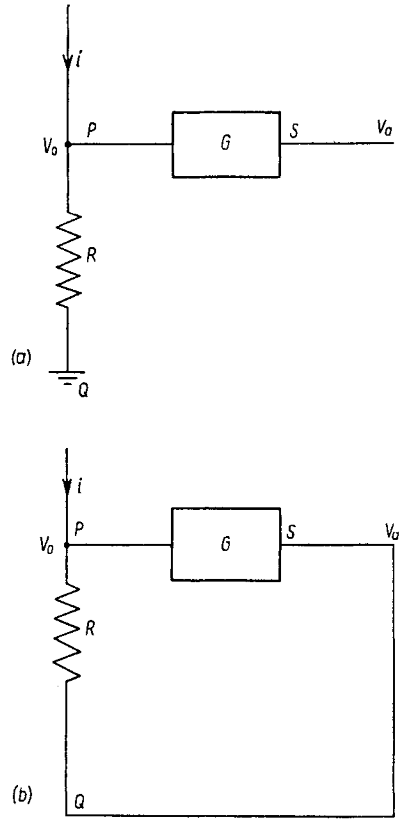


FIG. 8. Negative feedback.

Since a change in the applied potential produces only a fractional change in the potential of P, this implies a reduction in the leakage current in the cable from P to earth, etc., and it also implies a reduction in the capacitance between P and earth, so that, in some cases it is possible to dispense with the cathode follower stage situated with the collector.

Yet another advantage of this feedback circuit is that the small variation of the potential of P means that there is only a small change in the grid potential of the first stage of the amplifier, and so the range of input currents for which the amplification is linear is much greater than if the full potential V were applied to the first grid.

The feedback can also be applied to the sheath of the cable carrying the current i; since the potential difference between the current-carrying wire and the sheath is only a fraction of that between the wire and earth, the effect of leakage is reduced, and since this potential difference depends on the current flowing, the effective capacitance is also reduced.

2.49. Differentiating and Integrating Circuits

In some measurements the feature of interest is not the actual quantity which is first obtained, but its differential or integral. The methods most suitable for the purpose depend on the nature of the initial output, i. e. whether it is in the form of a potential difference or a current; even though these may be convertible, one to the other, that may prove inconvenient.

For differentiating a potential difference this can be applied across a condenser and the current immediately gives the differential; if a potential output is required, then the current can flow through a resistor and the potential difference across the ends of this is used.

For differentiating a current the simplest method is usually the use of a transformer and the output from the secondary can be used either as a current or as a potential difference.

To integrate a potential difference this is used to provide a current through a resistor charging a condenser and the potential difference across the condenser gives the integral.

To integrate a current it can directly charge a condenser. A current output from an integrating circuit is not feasible simply, as this would discharge the condenser. Because of leakage, integration is practicable only over short times.

2.50. Radio-sonding

The ordinary meteorological radio-sonde transmits to the ground signals which give the temperature, pressure and humidity. The measuring element produces a mechanical effect which alters the frequency of a modulating oscillator, and it is this frequency which is received, recorded and interpreted.

Various workers have replaced one or more of the meteorological elements by elements which give signals of electrical phenomena, e. g. potential gradient, point-discharge current, conductivity, etc., and have used various methods to convert the electric current to be measured into a change of frequency. The fact that the standard radio-sondes are easily convertible has meant that it has not been worth while to attempt to transmit the electric currents more directly, and the transmission in terms of a frequency change is convenient in that there is no effect of distance.

Radio-sonding with balloons has several advantages over other methods; a radio-sonde balloon can be sent into clouds in which an aircraft would not venture, it travels more slowly than an aircraft and so gives longer times for measurement and it probably interferes less with electrical conditions. Since the information is transmitted back immediately, there is no loss of results if the instruments are not recovered. On the other hand, it is not possible to direct exactly where a balloon shall go in disturbed conditions, nor is it always certain where it has been.

A critical account of some of the radio-sonde apparatus used for atmospheric-electrical observations has been given by MÜHLEISEN and FISCHER (1958).

2.51. Continuous Recording

In order to obtain a true picture of the phenomena of atmospheric electricity it is necessary to record them continuously and, until recently, the alternatives were either continuous photographic recording or mechanical registration at intervals.

In the photographic method, suggested by KELVIN (1860b), light reflected from a mirror attached to a galvanometer or electrometer is focused on to photographic paper on a rotating drum; if a slit is illuminated the image of the slit can be converted into a point image by a cylindrical lens. A refinement is to have a semi-transparent scale in front of the recorder and a weak fogging lamp, so that a scale appears on the record; time markings can also be obtained by switching off or intensifying the fogging light at regular intervals. The photographic record is continuous but has the disadvantages that the record cannot be examined at the time, but only after development, and also that the recording apparatus must be housed in the dark.

In BENNDORF's (1902) mechanical method the moving needle of a meter is made to give a mark of its position at regular intervals, for example, every 2 min; while this method allows the record to be inspected at the actual time of recording, there is the grave disadvantage that occurrences between the times of marking are lost.

With the use of electronic amplifiers for most of the measurements, it is now possible to obtain sufficient current to actuate a continuous pen-recorder which combines the advantages of both the previous methods.

For rare phenomena, methods of continuous recording involve the waste of a great deal of recording paper; to avoid this, CLARENCE and MALAN (1951) have used a magnetic tape recorder, from which the portion of interest is "played back" and recorded permanently, and the tape is then erased and used again.

MILNER and CHALMERS (1961) were interested only in occasions when the potential gradient was outside certain limits; they therefore arranged a less sensitive galvanometer in series with that recording the potential gradient and focused light from the galvanometer mirror on to a photo-transistor; when the potential gradient was outside the specified limits, the light did not fall on the sensitive area and a system of relays was operated, to switch on the recording lights and the camera motor.

2.52. Digitization

The analysis of results by the use of a computer requires that the results shall be put into the form of whole numbers. The measurement of records to give this information may be a very laborious process, and it is therefore desirable in certain cases that the results should be directly converted to digits.

The general principle involved is that the signal should first be converted into a frequency, as is done for radio-sonde transmissions; then the frequency is counted over a period of time and thus converted into a number, which can be punched on to tape ready for processing by a computer.

If relations between various quantities are to be investigated, these must be sampled at frequent intervals and the difficulties of non-continuous recording reappear.

2.53. The Electrolytic Tank

There are a number of instances in atmospheric electricity where it is desirable to know the distribution of electric fields and charges in particular conditions; a simple example is the reduction factor (see §§ 1.11., 5.18.).

The mathematical form for electrostatic problems in a medium of constant permittivity is exactly the same as that for current-flow problems in a medium of constant conductivity, and so the electrostatic problem can be solved by making a scale model, putting it into a tank of conducting liquid and measuring the current flowing; in the example of the reduction factor, the current gives directly the number of lines of force in the electrostatic case. It is necessary to use alternating current to avoid polarization effects in the liquid.

In some cases, a two-dimensional model is sufficient for the purpose, but a tank can be constructed to give results with a three-dimensional model.

2.54. Cylindrical Condenser Measurements

Apparatus with the same general principles has been used for the measurement of mobilities by ZELENY (1900), for the counting of ions by EBERT (1901), and for the measurement of conductivity by GERDIEN (1905b). It is convenient to give the general theory here and refer to the specific applications as they arise.

The apparatus consists of a hollow cylinder of radius a, with a smaller coaxial cylinder or wire of radius b; by means of a fan or pump, air is drawn through between the two cylinders with a velocity u. The central cylinder, or part of it, is connected to some form of measuring apparatus and a potential difference is applied

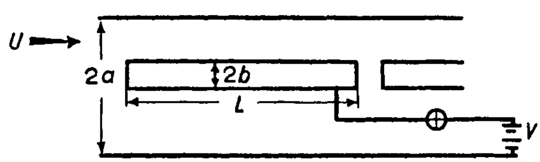


FIG. 9. The cylindrical condenser.

between the two cylinders. In one early form of the apparatus the central cylinder and an electroscope were charged to a definite potential, with the outer cylinder earthed, and measurements were made of the rate at which the potential fell while the apparatus was working. Modern practice uses electronic methods to measure the current.

If V is the potential difference between the two cylinders, the electric intensity at a point distant r from the axis is:

$$E = \frac{V}{r \log_e a/b}.$$

An ion of mobility w thus moves a radial distance $dr = wE\,dt$ in time dt.

If we now consider ions which start at the outer cylinder and move inwards towards the inner cylinder, the time taken to traverse a radial distance dr is:

$$dt = \frac{dr}{wE} = \frac{r \log_e a/b\,dr}{wV},$$

and the total time to get from the outer to the inner cylinder is:

$$t = \frac{(a^2 - b^2)}{2}\,\frac{\log_e a/b}{wV}.$$

During this time the air will have moved a distance ut along the cylinder. If this distance is less than the length, L, of the part of the inner cylinder connected to the measuring instrument, then the whole of the inward-moving ions will be collected; the necessary condition is:

$$u < L/t \quad \text{or} \quad u < \frac{2wVL}{(a^2 - b^2)\log_e a/b}.$$

This is the use in the ion-counter.

If u is much larger than this, then only a fraction of the ions will reach the inner cylinder, namely, those ions which started nearer the axis than a certain critical distance, R, given by:

$$u(R^2 - b^2)\log_e a/b = 2wVL.$$

This is the use in the measurement of conductivity.

If the portion of the inner cylinder which is connected to the measuring instrument lies beyond a distance L from the point of entry of the ions, then ions are received only if $u > L/t$ or

$$u > \frac{2wVL}{(a^2 - b^2)\log_e a/b}.$$

This is the use for measurement of mobility.

It is not necessary, in some of the uses, to assume that the flow of air is uniform across the cross-section of the opening, and we take the quantity Q of air flowing to replace $u\pi(a^2 - b^2)$. For example, provided that

$$Q < \frac{2\pi\, wVL}{\log_e a/b},$$

the apparatus counts the whole of the ions of mobilities greater than w, and also some of lower mobility (see § 4.11.).

CHAPTER 3

Clouds, Water and Ice

3.1. Properties of Clouds, Water and Ice

There are various properties of clouds, and of the water and ice which they contain, which are of greater or less importance in atmospheric electricity, particularly of disturbed weather. It is convenient to collect together these results here, so as to be able to refer to them as occasion arises.

3.2. Cloud Types

The various types of cloud have very different electrical effects and it will be convenient to summarize some of the features of the clouds.

From the point of view of atmospheric electricity, the most important cloud type is the cumulo-nimbus, which is the storm or shower cloud, characterized by great vertical height, limited horizontal extent, large vertical air currents and turbulence, and heavy rain, with, in the storm cloud, the production of lightning and thunder. The development of thunder clouds is dealt with in § 3.8. The cumulo-nimbus cloud indicates great instability of the atmosphere.

The nimbo-stratus cloud, like the cumulo-nimbus, is a cloud of great vertical height and gives rain; though, in general, the rain is less heavy than from the cumulo-nimbus, it usually lasts longer. The nimbo-stratus cloud is usually of considerable horizontal extent and has a smaller vertical air current and less turbulence than the cumulo-nimbus. The nimbo-stratus cloud is often produced by the gradual raising of moist warm air over a denser mass of colder air at a "warm front".

Less important from the point of view of atmospheric electricity is the cumulus cloud, which ranges from the small fine-weather cumulus clouds up to clouds which are developing into cumulonimbus. The cumulus cloud usually gives no precipitation, but indicates instability and convection.

The other low clouds, stratus and strato-cumulus are also of small importance in atmospheric electricity. They are clouds of small vertical thickness and wide horizontal extent and seldom produce rain, though there may be drizzle.

High clouds, cirrus, cirro-stratus and cirro-cumulus, and medium clouds, alto-cumulus and alto-stratus, are also of small importance in atmospheric electricity, though early workers suggested electrical effects connected with cirrus-type clouds (see § 16.22.).

SCORER (1963) has proposed a new and more useful classification of cloud types.

3.3. Cloud Physics

The phenomena of charge separation within a cloud must depend upon the various physical processes which are at work within the cloud, in particular those which concern the cloud particles and the precipitation particles, and perhaps most of all those concerned with the growth of the precipitation particles. It is not possible to give here any sort of summary of the results in this field, but various aspects are discussed where they appear to be of especial interest. Apart from these, the reader is referred to review and compendium articles: MASON and LUDLAM (1950), the relevant chapters in MALONE (1951), WEICKMANN (1953), BROWNE, PALMER and WORMELL (1954) and, in particular, the book by MASON (1957).

An important factor in the processes involved is that of temperature; one of the cases where this features particularly is that of the freezing of supercooled water drops, where latent heat is released, so that the temperature rises and under certain conditions not all the drop freezes; in every case there is a rise of temperature.

It is convenient here to define terms to be used in discussing changes of state of water. For the water–vapour change of state, the terms "evaporation" and "condensation" will be used; for the ice–water change, "melting" and "freezing" and for the ice–vapour change we shall use the terms "sublimation" and "deposition", using the former only for the production of vapour from ice and not for the reverse process. By the term "riming" is meant the production of ice by complete and almost instantaneous freezing of a supercooled water drop on impact with a cold body; in clouds this must be an ice particle but in laboratory experiments a metal rod or plate is often used. Another process, in German called *Vergraupelung*, must be distinguished from riming; in this a supercooled water drop coming into contact with a cold body forms a water layer on the body and this later freezes forming glazed ice; this process we shall term "glazing". The distinctions between riming and glazing are two; first, in riming the ice is crystalline, while in glazing it is amorphous and glassy; also in riming the whole of the water drop is frozen while in glazing some water may splash off, particularly if there is relative motion between the drop and the cold body. Whether the process that occurs is riming or glazing depends on the temperature, on the size, heat capacity and conductivity of the cold body and on the ventilation which governs the transfer of heat to the air; REYNOLDS (1954) suggested that, under natural conditions in a cloud, the process would be riming at temperatures below about $-15°C$ and glazing at higher temperatures. The successive layers of a large hailstone can be identified as due to riming at higher levels and glazing at lower levels as the hailstone rises and falls in the air currents present.

The temperature of $-41°C$ appears to be fundamental in cloud physics as this is the lowest temperature at which water droplets remain supercooled and below which they freeze spontaneously; thus it is only at temperatures above $-41°C$ that both solid and liquid particles can co-exist.

3.4. Electrification of Clouds

By methods which will be described later, it has been shown that storm clouds usually, and probably always, have a concentration of positive charge in their upper regions, with a lower negative charge. In many cases of thunder clouds, perhaps all, there is another concentration of positive charge in a limited region of the base. This suggests that perhaps there are two distinct processes producing the separation of charge and operating at different levels, the upper one giving an upper positive and lower negative charge, while the lower process gives separation in the opposite direction.

In the case of non-stormy rain clouds, there is less definite evidence, but it is probable that nimbo-stratus clouds, giving continuous rain, carry mainly a negative charge; certainly their bases usually appear to carry a negative charge.

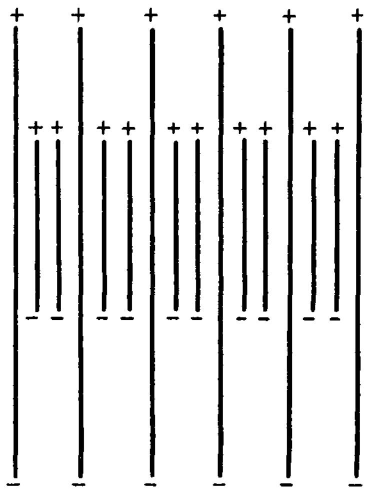


FIG. 10. Lines of force with non-raining cloud.

Even a cloud without any precipitation and without any process of charge separation will show charges resident at the top and bottom, because its conductivity is less than that of the air outside. A space charge then arises in exactly the same way as discussed in § 2.25. in connection with the variation of conductivity with height; in the case of the normal positive potential gradient, the charge must be positive at the top and negative at the bottom, exactly the opposite of the induced charges to be expected if there were a conductor in place of the cloud. These charges are somewhat analogous to the magnetic poles on the surface of a diamagnetic material. The potential gradient is positive both inside and outside the cloud, and so a change in conductivity alone can never produce a reversal of the potential gradient, either within the cloud, nor at the earth's surface; if there is no effect other than the change of conductivity, there can nowhere be a current in the reversed direction and therefore nowhere a reversed potential gradient.

The charges produced at the edge of the cloud can be regarded as analogous to a "traffic jam" at the edge of a town in which traffic moves slowly, where there will therefore arise an accumulation of the traffic moving inwards. So one can consider the change in conductivity arising from ions becoming attached to cloud droplets and so moving more slowly and giving an accumulation of charge.

3.5. The Initiation of Precipitation from Ice

BERGERON (1933) put forward the theory that precipitation originates as ice particles. According to this theory, as the water droplets in a cloud rise in the up-draught, being cooled by adiabatic expansion in the process, some of them freeze, either spontaneously or by reason of special nuclei. These, comparatively few, ice particles then grow at the expense of the surrounding water drops because of the difference in vapour pressure between solid ice and liquid water. As the ice particles grow they will reach such a size that they will fall against the up-draught, and in so falling they will capture and freeze the water droplets they meet, thus growing still further. If, while falling ,they get to a temperature above 0°C, they will melt and reach the ground as rain.

For many years it was thought that all precipitation is initiated by the above process, but evidence has accumulated that there are many cases in which this cannot be so.

3.6. The Initiation of Precipitation by Coalescence

There is ample evidence that heavy rainfall can occur from clouds which are completely at temperatures above the freezing point (BROWNE, PALMER and WORMALL, 1954), so that the ice mechanism cannot operate. In such cases it was suggested by BOWEN (1950) and LUDLAM (1951) that coalescence of droplets can cause growth to the size of raindrops if the cloud is sufficiently large, the process being started by a few larger droplets, perhaps arising from giant sea-salt nuclei. A considerable amount of theoretical work has been carried out to determine the conditions of coalescence, and the conclusion of HOCKING (1959) was that droplets must reach the radius of at least 19 μ before coalescence can occur; several attempts have been made to extend this work, but many are subject to the criticism that, while they extend HOCKING's work mathematically, this may well go beyond the limitations of the physical assumptions. The present position may be summarized that it is certain, physically, that the coalescence process can be initiated by droplets less than 19 μ—we do get rain when there are very few larger droplets at first—but nobody has yet produced convincing arguments as to how this comes about; possibly the electric charges on the droplets assist (DAVIS, 1965).

BATTAN (1953) produced evidence that some of the thunder clouds investigated in the Thunderstorm Project (BYERS and BRAHAM, 1949) may have had their precipitation initiated by the coalescence process, and this idea has been developed by REYNOLDS (1954). With such clouds, of course, the up-draught takes the precipitation to temperatures below the freezing point before it falls out, so that freezing is still a factor in the storm.

3.7. Contamination in Clouds and Precipitation

Some of the theories concerning charge generation in clouds depend upon the amounts of contamination existing in the particles involved, and it is therefore desirable to discuss the matter. By the term "contamination" is meant the small amounts in the atmosphere of various substances, mainly inorganic salts and acids, produced by industrial and natural processes, which are soluble in water; the evidence appears to be that insoluble particles, such as sand, dust or soot, play little, if any, part either in the processes leading to clouds and precipitation or in ionic electrical phenomena. The most important contaminants are, firstly, sodium chloride derived from the evaporation of sea spray, secondly, calcium carbonate and then nitrogen and sulphur compounds, particularly acids, from industrial fumes.

The condensation nuclei on which drops form in clouds are probably composed of soluble contaminants and it may be surmised that the initial process of growth of a cloud droplet is assisted by the reduction of vapour pressure by the dissolved substance (see § 4.6.). Thus cloud droplets are contaminated by the nuclei; when cloud droplets grow by coalescence, the percentage amount of contamination will remain constant; also, when a precipitation particle grows by accretion of cloud droplets, the percentage amount of contamination approaches that of the cloud droplets. But, when an ice particle grows by the process envisaged by BERGERON, it receives only pure ice and therefore becomes less contaminated than the cloud droplets.

3.8. Development of Thunder Clouds

The investigation of thunderstorms (BYERS and BRAHAM, 1949) has shown that the essential unit in a thunderstorm is not a single storm but a "cell", and that a storm consists of one or more cells in various stages of development.

The cell shows three distinct stages in its development. In the first, or "cumulus" stage, the vertical air currents are everywhere upwards, and the cloud is still obviously growing vertically. The vertical air current may reach values of 15 m/sec. No precipitation is falling out of the cloud during this stage, though it is probable that the process of the initiation of precipitation has already started.

The second or "mature" stage commences when the precipitation particles have grown sufficiently large to fall out of the cloud. The up-draught of air has increased beyond the values in the cumulus stage and reaches as much as 30 m/sec in the centre of the cell. The precipitation particles cannot, of course, fall against such velocities as these, but fall where the up-draught is less. As the precipitation falls, it carries air with it and there then exists a down-draught as well as an up-draught. It is in this stage that lightning occurs and the other phenomena of the thunderstorm reach their greatest intensity.

In the final or "dissipating" stage the up-draught has ceased and the motion of the air is downwards over the whole cell. The complete life-time of a cell is usually somewhat under half an hour, by which time all the effects, electrical and otherwise, have come to an end, although another cell may meanwhile have become active. The diameter of a cell is of the order of a few kilometres.

WORKMAN and REYNOLDS (1949b) have described their thunderstorm observations in terms of smaller cells, or "turrets" as they have since been called, with diameters of the order of 100 m; a number of these turrets make up the larger cells described above.

Thunderstorms may be divided broadly into "heat" and "frontal" types. The heat storms are caused by local heating, usually of land masses during the local early afternoon. Frontal storms are associated with cold fronts and may occur at any time of day and over land or sea equally.

BYERS (1965) has added other important facts about thunderstorms in relation to electrification; he emphasized the importance of the down-draught connected with the heaviest rain, and pointed out that the up-draught and down-draught areas cover only about one-seventh of the whole area of a cell. The turrets represent successive steps of increasing height where the up-draught goes progressively higher into clear air.

3.9. Developments of Nimbo-stratus Clouds

Nimbo-stratus clouds are most frequently formed at a warm front, when warm air rises as it reaches a region of cold air; the boundary between the warm and cold air forms a surface with a slope of around 1 in 100, so that there is a horizontal motion of about 100 times the speed of the vertical motion, applying both to the air and to the cloud droplets when formed. The vertical motion causes the air to cool and hence to condense, forming the cloud, when the relative humidity reaches 100 per cent.

A warm front is usually associated with a depression and hence the general wind direction is anti-clockwise round the centre of the depression. The front itself may be moving, not necessarily in the same direction as the wind at cloud level.

In the electrical phenomena of the nimbo-stratus cloud, it is probably the vertical motion that is more important than the horizontal motion, though the latter ought not to be left out of consideration.

3.10. Radar Study of Clouds

Radar waves are reflected by water droplets according to the sixth power of the diameter, and therefore rain drops give a much stronger effect than do clouds. Ice particles give a less effect than water drops of the same size, but ice precipitation is more effective than cloud droplets. The use of radar in following the development of thunder clouds is discussed in § 12.23.

The "bright band" phenomenon occurs at the level where precipitation melts; the reflections are more intense at this level than above, because the melted water reflects better than the ice; and the reflections are more intense than below because the water drops have not yet reached their terminal velocity, which is greater than that of the ice particles.

3.11. The Capture of Ions by Water Drops

Wilson (1929) first considered that water drops polarized in an electric field could, under suitable conditions, capture selectively ions of one sign rather than the other and that this process might lead to a building up of the charges in clouds. The capture of ions by falling drops must also be a feature of the charging of precipitation.

If a water drop, initially uncharged, is situated in a vertical electric field, its two halves will carry charges of opposite sign. If the potential gradient is positive, the upper part of the drop carries a negative charge and the lower part a positive charge. If the drop is at rest with respect to the air, then negative ions moving upwards are attracted to the lower half of the drop and positive to the upper, so that the drop acquires a resultant charge only if there is a difference in the numbers of ions of either sign reaching the respective halves of the drop. If the drop is falling through the air less rapidly than the positive ions move downwards (case of "fast" ions), then it turns out that the consequences are the same as for the drop at rest. However, if the rate of fall of the drop is greater than the downward velocity of the positive ions (case of "slow" ions), then these cannot catch up to the top of the drop and are repelled from the positively charged lower half of the drop; thus the positive ions do not reach the drop, but the upward-moving negative ions are still attracted to the lower part of the drop and so the drop as a whole acquires a negative charge. As the negative charge on the drop increases, so a larger fraction of the drop becomes negatively charged and the slow positive ions, which the drop overtakes, can reach the drop, while the negative ions are repelled by the negative charge. A condition is reached in which the drop is receiving equal numbers of ions of both signs, the exact value of the charge depending on the relative numbers of ions of both signs present.

The distinction between fast and slow ions is important, since selective capture occurs only if the positive ions are slow. The potential gradient at which the change from fast to slow occurs depends on the size of the drop. For a drop of raindrop size with a speed of a few m/sec, the small ions with a mobility about 1·5 C.G.S. units become fast when the potential gradient reaches values of the order of 10^4 V/m, but breakdown would occur before the potential gradient becomes high enough for the large ions to become fast. For the small droplets that exist in clouds, the small ions become fast at much lower values of the potential gradient.

The details of the process have been worked out by WHIPPLE and CHALMERS (1944) for the case of a drop whose motion through the air follows STOKES's law. If ions of negative sign alone are present, the final charge obtained by the drop is $-12\pi\varepsilon_0 Xa^2$, where X is the potential gradient and a is the radius of the drop. When the conductivities due to ions of either sign are equal, the final charge is $2{\cdot}06\pi\varepsilon_0 Xa^2$. WHIPPLE and CHALMERS also worked out the rate of charging and KRASNOGORSKAYA (1960) extended this to the case of unequal conductivities.

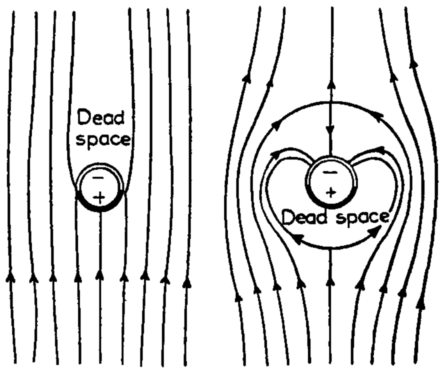


FIG. 11. The charging of a water drop. Paths of negative and slow positive ions $Q = 2{\cdot}06\,\pi\varepsilon_0 X a^2$, $V = 2wX$. (From WHIPPLE and CHALMERS, 1944, Fig. 4b and c, p. 107).

CHALMERS (1947) has shown that the results of the detailed analysis of WHIPPLE and CHALMERS can be obtained by assuming that the current into a drop is given by the ordinary conductivity formula, from the polar conductivity, for the area of the surface of the drop concerned, except that when there are slow positive ions the positive charge on the lower part of the drops screens off an equal amount of negative charge on a higher portion of the drop. Thus if a drop carries a positive charge $+Q_1$ on the lower part and a negative charge $-Q_2$ on the upper part, then the resultant current of positive ions is to be considered due to a charge of $-Q_2 + Q_1$; if this is worked out it is found to be $\lambda_1(Q_2 - Q_1)/\varepsilon_0$, while there is a current of negative ions equal to $-\lambda_2 Q_1/\varepsilon_0$. Then ,if $\lambda_1 = \lambda_2$, these two currents will be equal if $Q_2 = 2Q_1$. On working out the areas of the drop surface having charges of either sign when the drop is polarized and also carrying a resultant charge $-Q_2 + Q_1$, the result quoted above is obtained.

WORMELL (1953a) worked out the consequences of these results when λ_1 and λ_2 are not equal. GOTT (1933, 1935) showed experimentally that, under the conditions suggested by WILSON, water drops certainly do acquire charges of the right sign and of the right order of magnitude; MÜHLEISEN and HOLL (1953) obtained further evidence that the capture process acts in accordance with the formulae of WHIPPLE and CHALMERS (1944).

MÜLLER-HILLEBRAND (1954) extended the theory to take account of the additional attraction of ions to the drop by reason of the force between an ion and its image charge in the conducting drop. He showed that the factor 2·06 would become 2·44. PAUTHENIER and COCHET (1953) made calculations in which the ions are replaced by charged cloud droplets or ice particles, which can be polarized themselves, and found that, in some cases, the efficiency of capture is much increased.

CHALMERS (1947) showed that there can be somewhat similar selective ion capture by ice particles in which case dielectric polarization replaces the polarization of a conductor; charges arriving at the insulating surface remain where they arrive, instead of being distributed over the whole surface as in the case of a conductor.

3.12. Thermoelectric Properties of Ice

Ice has a small electrical conductivity which decreases as the temperature falls; this has been shown to be caused by the dissociation of some ice molecules into positive and negative ions, possibly H^+ and OH^-, or possibly more complex; as the temperature falls, fewer molecules are dissociated and hence the conductivity is less. The positive ions have a mobility around ten times that of the negative ions and so are mainly responsible for the conductivity (EIGEN and DE MAEYER, 1958).

If a slab of ice has its two sides maintained at different temperatures, more positive ions will tend to leave the warmer side than the cooler, just because there are more positive ions there. Thus the cooler side receives a positive charge and this increases until the electrical potential difference set up is sufficient to balance the tendency for positive ions to move to the cooler side.

The matter can be looked on in terms of the thermoelectric effects, and the potential difference set up is what is known as the "homogeneous thermoelectric power" of ice. If P is the homogeneous thermoelectric power, π the contribution of the ice to the Peltier effect at a junction and σ the Thomson effect coefficient, then $P = \pi/T$ and $\sigma = -T\,dP/dT$, where T is the absolute temperature.

Measurements of the thermoelectric power can be made, normally, only with a complete circuit of two materials, but the values for ice are expressed in mV/°C, while those for metals are μV/°C, so that the influence of the particular metal chosen is negligible. The measurements of LATHAM and MASON (1961a) with pure water gave a potential of $-2{\cdot}0$ mV/°C, representing charges of $\pm 1{\cdot}65 \times 10^{-12}\,(dT/dx)$ C/m^2, where dT/dx is the temperature gradient in °C/m; BRYANT and FLETCHER (1965) measured the thermoelectric power with different concentrations of HF and NH_3 present and found negative potentials (negative charge on the warmer side) for pH values between 1·5 and 7·7 and positive potentials for pH values from 7·7 to 12·5. There was good general agreement with the theory of JACCARD (1959, 1963).

Confirmation of the view that the metal concerned has negligible effect was obtained by LATHAM (1964b), who suspended an ice needle with its ends at different temperatures and measured its electric dipole moment, obtaining results agreeing with the thermoelectric results.

While there seems no doubt that there is separation of charge in a slab or needle of ice in air, it does not seem so easy to be certain of similar effects in, for example, an ice shell on a drop of water freezing, where the warmer surface is not in contact with the air.

3.13. Electrical Effects of Ice Impact

SIMPSON (1919) first suggested that the impact of ice crystals on one another might generate a separation of charge which could account for the high potential gradients found in blizzards, and he returned to this idea (SIMPSON and SCRASE, 1937) in connection with charge separation in a thunder cloud.

Observational evidence on the subject has been very confusing; STÄGER (1925a, b) found quite contradictory results in two sets of observations with natural snow. PEARCE and CURRIE (1929) found that when an air blast eroded a block of snow, the larger eroded fragments acquired a negative charge while the air carried away a positive charge; but CHAPMAN (1953) found that, when natural snow flakes fall on to snow in a dish, the air acquired a negative charge and the dish a positive. CHALMERS (1952d) rubbed fallen snow together and found that the larger fragments obtained a negative charge. NORINDER and SIKSNA (1954) found that, when snow was poured or blown on to a target, there were important effects depending on the nature of the target surface concerned, so that the results of such experiments are not directly applicable to phenomena inside a cloud where no such surfaces are present. Further experiments by NORINDER and SIKSNA (1955) have been concerned with the charges carried in the air when the snow is blown, and have shown that there are produced small, invisible, snow particles carrying charges, usually positive, of several hundred electronic units, as well as ions of various sizes.

MAGONO and SHIOTSUKI (1964) found an effect of the air bubbles, ice particles with more air becoming negative on rubbing. HOBBS (1964) suggested and LATHAM (1965c) confirmed that this was merely a temperature effect, the air bubbles altering the effective area and conductivity.

The complexity of these observations may be ascribed to the fact that there are several different processes at work including the following: charge transfer between two specimens of ice during a short contact, effects of a rubbing contact and effects of breaking a piece of ice.

3.14. Ice Contact

If two pieces of ice at different temperatures are brought into contact for a limited period of time, charge will be transferred, the colder piece becoming positively charged; if the contact lasts too long, thermal conductivity tends to equalize the temperatures and the charge transfer decreases. LATHAM and MASON (1961a) deduced that the maximum transfer should occur for a time of contact of $8{\cdot}5 \times 10^{-3}$ sec and should amount to $3{\cdot}05 \times 10^{-3}$ (ΔT) ESU/cm^2 or $1{\cdot}02 \times 10^{-12}$ (ΔT) C/m^2. Experiments showed reasonable agreeement with these calculations, except when the temperatures approach 0°C. LATHAM (1963c) found that, in such cases, there was a surface effect.

When two pieces of ice are rubbed together, it has been found that there is charge separation if the rubbing is asymmetric, so that a larger area of one piece is concerned than of the other. REYNOLDS, BROOK and GOURLEY (1957), LATHAM and MASON (1961a) and LATHAM (1963b) found that the "rubber", using less area, always acquired negative charge and explained this by the fact that the rubber is heated to a higher temperature by the friction. LATHAM (1963b) found the measured charge to be in reasonable agreement with theoretical predictions based on the theory of LATHAM and MASON (1961b).

With direct momentary contact, LATHAM (1964b) found the charge transfer to be of the same order as that predicted but rather smaller. LATHAM and STOW (1965a) found that the charge transfer on impact depends on the velocity of impact and on the surface irregularities of the specimens.

LATHAM (1964a) has discussed also the phenomena observed in snow storms and blizzards.

3.15. Electrical Effects of Freezing

WORKMAN and REYNOLDS (1950) found that when water freezes a "freezing potential" of some hundreds of volts is set up, the water being positive and the ice negative when the water contains the dissolved substances usually found in rain. Since any contact-potential measurement involves at least three junctions (see § 2.44), it is not certain what part of this potential is to be ascribed to the actual ice–water contact. The magnitude, and even the sign, of the freezing potential depends upon the dissolved substances in the water and for very pure water GILL and ALFREY (1952) found the ice to become positive; they ascribed a large part of the potential to charges literally frozen into the ice. GROSS (1965) found that the freezing potential depends mainly on the negative ion present except when ammonium salts exist.

MASON and MAYBANK (1960) observed the sudden freezing of small supercooled drops of water, often accompanied by shattering and often by the production of splinters. They found that when the residue of a drop was more than half the original drop it usually acquired a negative charge, but when less than half usually a positive charge. EVANS and HUTCHINSON (1963) obtained somewhat similar results. LATHAM and MASON (1963b) explained the results of MASON and MAYBANK in terms of the electrical properties of ice as follows: as a supercooled drop is freezing the outside is colder than the inside and so acquires a positive charge; when the pressure of freezing inside causes the drop to shatter the smaller parts of the residue come from the outside and so will be positively charged. A numerical comparison of the predictions of the theory and the experimental results shows that the charges observed are quite frequently appreciably larger than predicted. This may be explained if, in addition to charge separation by the temperature gradient in ice, there is also charge separation at a water–ice interface.

KACHURIN and BEKRYAEV (1960) and STOTT and HUTCHINSON (1965) found that the charge separation depends on the nature of the breaking and tended to give negative charges on the residues if liquid water is ejected from a breaking spicule and positive if a solid spicule breaks, agreeing with the suggestion of charge separation at a water–ice interface.

TAKAHASHI (1962) investigated the breaking of a piece of ice in which a temperature gradient exists, and found that a noticeable effect could be obtained with polycrystalline ice but not with single crystals; if the polycrystalline ice contained many bubbles of air, he found the colder portion to be charged positively, to a greater extent than predicted by the theory of LATHAM and MASON (1961 a).

LATHAM (1963 a) investigated the charging produced by blowing air over frost deposits, breaking off splinters, and found that the sign and magnitude of the charge depended on the temperatures of the air and the deposit, giving good agreement with the theory of LATHAM and MASON (1961 a).

3.16. Effect of Electric Field on Freezing

PRUPPACHER (1963 a, b) has summarized the work on the effect of an electric field in causing freezing of supercooled water drops and has carried out investigations himself. His conclusions were that an effect exists only when there is a solid surface present and so is not a factor in actual clouds in the atmosphere.

3.17. Electrical Effects of Riming

During riming, FINDEISEN and FINDEISEN (1943) found the deposit to be charged positively while the corresponding negative charge was carried off either on ice splinters or as ions in the air. On the other hand, REYNOLDS (1954) found that there was very little charging on riming, provided that there were no ice crystals present as well as water drops, and this was confirmed by MAGONO and TAKAHASHI (1963). LATHAM and MASON (1961 b) found the rime deposit to acquire a negative charge, as would be expected on their theory and on the results of MASON and MAYBANK (1960) if the water droplets freeze and leave as rime their larger portions, smaller portions going into the air.

If ice crystals as well as water drops are present during the process of riming, the electrical effects are very different from those in the absence of ice. REYNOLDS (1954) and REYNOLDS, BROOK and

GOURLEY (1957) found that, when both ice crystals and water drops are present, there is a very considerable rate of charging, the deposit obtaining a negative charge. Their explanation is that riming releases latent heat and so warms the rimed deposit to a temperature higher than that of the ice crystals; contact between the crystals and the deposit would then give a negative charge to the latter. However, the magnitude of the charging effect due to temperature differences that would be needed to account for these results is several powers of ten greater than that actually found by LATHAM and MASON (1961a); HUTCHINSON (1960) found no detectable effect with apparatus which would easily have measured the effect assumed by REYNOLDS.

MAGONO and TAKAHASHI (1963) also found that the presence of ice crystals affected the charging during riming. They found that the sign and magnitude of the charge depended on the temperature and on the rate of riming, the charge being negative except for low rates of riming and high temperatures. They considered that the process of charge separation may be the tearing off of rime from the deposit, by the impact of the ice crystals, as well as a possible effect of temperature differences. It is only when the conditions of riming are such as to produce a fine structure that the negative charge is left on the rime.

LATHAM and MASON (1961b) ascribed their charging effect on riming to the removal of splinters and it may be that the conditions of their experiment were such as to cause splinters to be broken off by the air currents, without any ice crystals, while in other experiments, e.g. REYNOLDS (1954) which showed no charging with riming in the absence of ice crystals, conditions were not favourable to the breaking off of splinters.

The effects of velocity of impact and of roughness of surface (LATHAM and STOW, 1965a) may help to remove the discrepancy between the results of REYNOLDS (1954) and of LATHAM and MASON (1961b). LATHAM (1965a) has attempted to assess the relative importance of different processes under different conditions.

3.18. Electrical Effects of Glazing

Electrical effects accompanying glazing are complicated by effects due to impact between water and ice. For water which is not supercooled, it has long been known that impact with ice gives a positive charge to the ice and a negative charge is carried off, either on the water or as ions in the air (FARADAY, 1843; SOHNCKE, 1886; FINDEISEN, 1940). LUEDER (1951a) whirled a metal rod in a fog consisting of natural supercooled water droplets and obtained evidence, rather indirect, that the growth of ice by glazing gives it a negative charge, the positive charge being presumably carried away by the water splashing off. MEINHOLD (1951) found a strong negative charge to be produced on the surface of an aircraft flying through a cloud of supercooled water droplets, while glazing was taking place.

As it is possible that the effects of LUEDER and MEINHOLD might be influenced by the metals on which the deposit is formed, WORKMAN and REYNOLDS (1953) used a suspended hailstone and investigated the charges produced when supercooled water droplets fell on it; they found the effects to depend considerably on small amounts of contamination in the water; for most of the contaminants likely to be found in rain water, the hailstone acquired a negative charge and the water splashing off a positive. On the other hand, LATHAM and MASON (1961b) found a change of sign when riming was replaced by glazing, i. e. when the supercooled water droplets did not wholly freeze, so that the ice now received a positive charge. MAGONO and TAKAHASHI (1963) also found a positive charge for the conditions of low water content or high temperature, when riming would be replaced by glazing; they interpreted the effect as a water skin, rather than ice particles being torn off the surface. For normal conditions the distinction between riming and glazing would occur at about $-10°C$ to $-15°C$.

3.19. Electrical Effects of Melting

DINGER and GUNN (1946) found that ice, for example in the form of snow flakes, which contains entrapped air, gives a separation of charge on melting, the resulting water showing a positive charge, while the escaping air shows a negative charge; the effect is, however, sensitive to impurities, among them carbon dioxide. A similar effect on melting has been found by MAGONO and KIKUCHI (1963), KIKUCHI (1965) and by MACCREADY and PROUDFIT (1965b), but MATTHEWS and MASON (1963) failed to obtain any charging; DINGER (1964) attributed the result of MATTHEWS and MASON to the presence of carbon dioxide, but MACCREADY and PROUDFIT also did nothing to eliminate carbon dioxide; they suggested that the speed of air flow might be important.

BENT and HUTCHINSON (1965) found space charges which they attributed to charge released in the process of the melting of snow on the ground.

MAC CREADY and PROUDFIT (1965a) observed the charges on individual particles inside clouds and found changes appearing at the melting level. But their results gave a negative charging on melting, in contrast to the previously discussed results, all of which showed positive charges on melting.

Results for the charges on continuous snow and rain and the corresponding potential gradients (see § 13.4.) would all agree with the precipitation obtaining a positive charge on melting.

Many of the above results could be explained by a separation of charge which is only indirectly connected with melting. If, for example, molten snow forms water drops big enough to be shattered by the air stream present, the actual process would correspond with that of splashing (§ 3.24.). Or there might be effects dependent on the contact between ice and water (see § 3.15.).

In view of the possible importance of melting in the production of the lower positive charge in thunder clouds (see § 16.36.) and of the charge on continuous rain (see § 16.39.), it is highly desirable that more experiments on melting effects should be carried out, particularly in conditions approaching those in the atmosphere.

3.20. Electrical Effects of Sublimation and Deposition

FINDEISEN and FINDEISEN (1943) found that small ice splinters are formed both when ice sublimes directly and when vapour deposits to form ice. They found the splinters to be charged, normally negatively during deposition and positively during sublimation, but in the presence of a strong potential gradient the charges on the splinters were always of the sign to be expected from induction. KRAMER (1949) and KUMM (1951) repeated some of these experiments and found in some cases very little electrification and in others charges of opposite sign to those found by FINDEISEN and FINDEISEN.

It has been pointed out that FINDEISEN and FINDEISEN used an ice layer on a metal plate, and the latter may have some influence. Also, as REYNOLDS (1954) pointed out, the plate was maintained at $-60°C$, so that conditions were not similar to those in clouds. It may be that effects of temperature differences (see § 3.14.) come into play.

LATHAM and STOW (1965b) found positive charge in the air and negative charge left on the residue when ice sublimes.

3.21. Electrical Effects of Evaporation and Condensation

VOLTA (1782) suggested that the phenomena of atmospheric electricity could be explained if charges were liberated on evaporation and various workers, starting with VOLTA himself, have tried to establish the phenomenon experimentally. Of those who thought to have found evidence for it, EXNER (1886–90) and MÜHLEISEN (1958) found effects corresponding to one unit electronic charge liberated for about 5×10^{13} molecules evaporated. However, BLAKE (1883) and LENARD (1914) found zero effect and GILBERT and SHAW (1925) summarized the evidence and pronounced against the existence of the effect. ISRAËL and KNOPP (1962) made further investigations and found that any charge liberated on evaporation could be at most one-thousandth of that postulated by EXNER and MÜHLEISEN and probably less.

BLANCHARD (1964) found that appreciable separation of charge occurs with very violent evaporation and he suggested that the significant process is the disruption of water surfaces; he discussed the production of charge when sea water falls on to molten volcanic lava.

3.22. Evaporation of Charged Droplets

As a charged droplet evaporates, its radius decreases and so the electric field strength at its surface increases. Ultimately, the condition is reached when the disruptive forces of mutual repulsion of the charge overcome the cohesive force of surface tension.

To investigate this experimentally, DOYLE, MOFFETT and VONNEGUT (1964) used charged droplets suspended in a field, as in the Millikan experiment. They found that, as expected, the field required to keep the droplet steady became less as the droplet evaporated, but at intervals there was a sudden increase with a simultaneous ejection of from 1 to 10 very small droplets and a loss of charge on the drop.

Calculations showed that, before the droplet became unstable, the field outside its surface was greater than the normally accepted dielectric strength of air of about 3×10^6 V/m.

The evaporation of cloud droplets in the air could be expected to give rise to similar phenomena, giving large numbers of very small, highly charged droplets.

3.23. Electrical Effects with Water Droplets

Barklie, Whitlock and Haberfield (1958) investigated the charges on droplets produced by spraying water into a closed chamber, by drawing the cloud past a suspended cylinder. The cylinder picks up the charges on the larger particles, those greater than about 10 μ in diameter, while the smaller particles are carried round in the air stream. The results immediately after spraying were variable, probably because of the electrical effects of the actual spraying process; but if the cloud was allowed to remain for a period of 10–15 min before being drawn past the cylinder, the charge acquired was found to be positive. They found that the approach to the final state could be much accelerated by the proximity of a β-ray source, showing that small ions must be the agents by which the particles acquire their charges. When the water droplets were replaced by oil, the charge acquired by the cylinder was negative. These results may be compared with those obtained by the same workers from an aeroplane in a natural cloud (see § 13.7.).

Mühleisen (1958) measured the space charge in a closed room and found that, when the air was very moist, the space charge became negative; this might be the same effect as above, if the larger particles, with positive charge, fall out. Mühleisen also found that, when the air was dried, a positive space charge develops; this is not so easy to explain in terms of the results quoted above. Later, Mühleisen (1961 b) attempted to explain the results in terms of an increased life-time of free electrons in conditions of high humidity.

3.24. Electrical Effects of Splashing

Lenard (1892) found that, when water splashes, the larger particles of water obtain a positive charge, while the air and smaller particles carry a negative charge. Simpson (1909), Nolan and Enright (1922) and others found that the same effects occur if water drops are broken up, not by splashing, but by the effect of air currents. The magnitude and sign of the charge depends on substances dissolved in the water. The results suggest that the charge produced depends on the area of new water surface produced, but it is necessary that there is violent breaking, as no electrical effects occur on a gradual production of a new surface.

ADKINS (1959b) investigated the splashing of drops in an electric field and concluded that ions of both signs are formed, but those of one sign only, opposite to the potential gradient, are released into the air. He also found a dependence on the energy of the falling drop. SMIDDY and CHALMERS (1959) found that their measurements of space charge during rain could be explained in terms of the production of negative charges in the air from splashing, whatever the potential gradient, thus agreeing with the results of LENARD rather than with those of ADKINS.

MAGONO and KOENUMA (1958) measured the charges on water droplets produced by fragmentation of a larger drop in an air blast; they found that, if fragmentation took place in a field-free region, the droplets carried a positive charge, but if there is an electric field of 2000 V/m then droplets of both signs are formed.

MATTHEWS and MASON (1964) also measured the charges produced on drops during fragmentation. They used large drops breaking when falling freely in air, and found that the charges separated depend upon the electric field; in a field of 3×10^4 V/m, corresponding to that to be expected in a thunderstorm, separation of charge of the order of 1 E.S.U. ($3{\cdot}3 \times 10^{-10}$ C) per cm^3 of water was found; this is about 100 times that found in the absence of a field, and could be important in the lower positive charge of a thunder cloud.

CHAPTER 4

Ions and Nuclei

4.1. Small Ions

COULOMB (1795) appears to have been the first to notice the conductivity of the air, in particular the leakage through the air of the charge from an insulated electrified body. He considered that air or dust particles could receive charges by impact with the charged body and would then be repelled away from it. This view was held until, almost simultaneously, ELSTER and GEITEL (1899a, b) and WILSON (1900) discovered the existence of "ions", i. e. particles of approximately molecular size but carrying positive or negative charges, usually of the magnitude of the electronic charge; these ions are produced by some "ionizing" process by which a molecule is split into a positive and a negative portion; often, probably always, the ionizing process consists in the removal from the molecule of one electron, which then attaches itself to a neutral molecule forming the negative ion. The leakage of charge from an insulated body comprises the attraction to the body of ions of opposite sign of charge to that on the body and the repulsion of ions of like sign. In the absence of wind, ions travel along lines of force.

Many laboratory experiments have been carried out to investigate the properties of ionic conduction, using, most frequently, not natural ions from the atmosphere, but ions produced by other agencies, e.g. X-rays; it is found that such ions do not differ much from the natural small ions in the atmosphere which are responsible for the major part of the conductivity. For the detailed description of such experiments and their results the reader is referred to treatises on the subject, such as that of THOMSON and THOMSON (1928, 1933).

The results show that there is a class of ions, the "small" ions, having mobilities between 1×10^{-4} and 2×10^{-4} m/sec for a potential gradient of 1 V/m at N.T.P. Mobilities are most usually quoted in terms of C.G.S. units and the mobility of a small ion is between 1 and 2 cm/sec for 1 V/cm. The actual value of the mobility depends on various factors, among them the pressure, the humidity, the presence of impurities and the age of the ion. Under similar conditions it is found that negative ions have mobilities rather greater than those of positive ions.

Free electrons would have mobilities much greater than those of small ions, but electrons cannot remain unattached for an appreciable time in the conditions existing in the parts of the atmosphere of concern in atmospheric electricity.

4.2. Nature of Small Ions

The mean free path of an ion is not the same as that of a molecule of the same size, since the charge on the ion can produce polarization in nearby molecules and consequent attraction, giving collisions which otherwise would not have occurred. Thus it is not possible to calculate the size of the ion simply from measurements of mobility or of diffusion coefficient.

If a small ion were merely a molecule with a single electron in excess or deficit, then we should expect no effects of humidity or impurities and we should expect the same mobility for positive and negative ions. The fact that the mobility does depend on various conditions suggests that an ion is more complex than a single molecule.

The general result of investigations on the small ions is that an ion consists of a single ionized molecule with other molecules clustered round it, and kept together by the charge. WRIGHT (1936) considered the mass of the small ion to be equal to that of 10 or 12 water molecules, while TORRESON (1939) considered an ion to be of the size of about 4 molecules of oxygen. Since water molecules are easily polarized, it is probable that, where they exist in sufficient numbers, they will be more readily attached to the ionized molecule than will other molecules in the neighbourhood. If one assumes a V-shape for the water molecule, with the base of the V the negatively charged oxygen atom, then it can be seen that water molecules can cluster more readily round positive ions, and in greater numbers, thus giving a lower mobility to the positive ion than to the negative ion in a humid atmosphere; the effects of humidity on mobility are also accounted for by this picture.

4.3. Large Ions

The "large" ions, discovered by LANGEVIN (1905), have mobilities between 3×10^{-4} and 8×10^{-3} C.G.S. units. There is some evidence that the mobilities of the large ions in the atmosphere fall into distinct groups, corresponding to different sizes. For rough calculations it is often sufficiently correct to take the average mobility of the large ions to be about 1/500 of that of the average small ions.

Whereas the small ions are not much larger than molecular size, the large ions are considerably larger, and are charged particles of the same kind as the "nuclei" on which AITKEN (1880) found condensation of moisture. It is probable that they are particles of such substances as evaporated sea salt or soluble substances from industrial smoke. The large ions and nuclei are much more abundant in urban than in rural areas.

MCGREEVY and O'CONNOR (1962) found that nuclei can be produced by the action of sunlight on trace gases such as hydrogen sulphide or sulphur dioxide and also on various natural and artificial solids such as pine needles and polyethylene.

Work on the artificial production of large ions for laboratory measurements will be discussed only in so far as it is directly concerned with atmospheric measurements.

4.4. Intermediate Ions

Ions of mobilities between those of the small and large ions are classified as "intermediate" ions. Some workers have found that the ions in the atmosphere can be divided into separate groups, each group comprising a definite range of mobilities, with no ions with intervening values of mobility, while other workers have found a continuous range of mobilities.

POLLOCK (1915b) first discovered a group of intermediate ions, with mobilities between 10^{-1} and 10^{-2} C.G.S. units. He found that such ions could exist only under conditions of low humidity and would disappear at higher humidity, presumably becoming large ions. WAIT (1935b) confirmed the existence of intermediate ions, but did not find the same effect of humidity.

HOGG (1939a) investigated intermediate ions in detail and found quite definite groups of ions with mobilities from 2×10^{-1} to 10^{-3} C. G. S. units. Interpreting them as particles of sulphuric acid, he was able to calculate their radii and found that the volumes of the ions were integral multiples (up to 15) of a unit size (radius $3{\cdot}6 \times 10^{-9}$ m), comprising some 2000 molecules. The mobility of an ion of definite radius is not constant, but increases both with increase of temperature and with decrease of humidity. HOGG found no evidence of the disappearance of the intermediate ions at high humidity. HOGG gave no theory to account for the unit size. If his interpretation of the ions as particles of sulphuric acid is correct, then they could be found only where industrial processes produce sulphuric acid.

On the other hand, WEISS and STEINMAURER (1937) and YUNKER (1940) found no evidence for distinct intermediate ions, and YUNKER found ions to occur for all values of mobility between those of the small and large ions. The discrepancy may be due to the need for special substances, such as sulphuric acid, to produce intermediate ions.

4.5. Distinction between Large and Small Ions

Quite apart from the difference in mobility there is an important physical difference between large and small ions, which can be expressed by stating that the essence of a small ion is its charge, while the charge is only incidental to the large ion. Although a small ion comprises a group of molecules, when the charge is removed the molecules no longer remain attached, since the charge itself is the mechanism by which they congregate together; when a small ion loses its charge, no trace of its existence remains. On the other hand, when a large ion loses its charge it becomes an uncharged nucleus (see § 4.6.) and remains easily detectable; it can readily acquire another charge of either sign. By this criterion the intermediate ions would be classified as "large" rather than as "small".

4.6. Condensation Nuclei

In addition to the ions of both signs and of different sizes, there are also uncharged particles in the atmosphere, of the same general size as the large ions. These are usually called "condensation nuclei" or "AITKEN nuclei", but, with the use of the word "nucleus" in a very different connection in atomic physics, another word for the atmospheric particles would be very desirable; the author (CHALMERS, 1954) suggested the word "kernel" (the German word is *Kern*), but this has not gained acceptance. It is to be hoped that those who study atmospheric nuclei will not be expected to produce nuclear power nor be blamed for nuclear weapons! The nuclei are the particles on which COULIER (1875) and AITKEN (1880) found water to condense when the air becomes saturated with moisture. It should be pointed out that condensation of water does not take place on coarser smoke or dust particles, but only on nuclei, including the charged nuclei, or large ions. The nuclei consist of substances soluble in water; in some cases, at least, the nuclei are hygroscopic and condensation may take place when the relative humidity is less than 100 per cent. A detailed account of the properties of nuclei, as known to that date, was given by LANDSBERG (1938).

The mechanism of the condensation can be pictured as follows: the equilibrium vapour pressure over a curved surface of radius of curvature r is greater than that over a plane surface of the same liquid by an amount $2T/r$, where T is the surface tension. Thus, in an atmosphere saturated over a plane surface of water, a small droplet would evaporate, and so there would be no condensation. But the presence of a dissolved substance reduces the equilibrium vapour pressure, so that a small droplet of dissolved substance may have an equilibrium vapour pressure less than that of a plane water surface, and would then grow if the relative humidity is 100 per cent.

Taking into account both the effects of curvature and of dissolved substances, it is possible to calculate the equilibrium vapour pressure for any radius for any particular amount of dissolved substance. For a given mass of substance the curve goes from very low vapour pressure for very concentrated solutions, to a maximum at some radius with vapour pressure over 100 per cent relative humidity, and to just 100 per cent for very large drops which are very dilute.

If there is a particular relative humidity less than the maximum, there will be an equilibrium condition with drops of a definite size for each mass of dissolved substance. But if the relative humidity is greater than the maximum of the curve, drops will continue to grow and their size will be limited only by the amount of water available for condensation.

4.7. Original Method for Counting Nuclei

The total number of nuclei, charged and uncharged, in a given volume of air can be counted by a "nucleus counter". The original nucleus counter was designed and perfected by AITKEN (1888–89). The principle of the counter is simply that of adiabatic expansion of moist air, causing condensation of water on to the nuclei; the droplets produced then fall on to a ruled glass window and observation with a microscope is used to determine the number falling on to 1 mm^2. Most samples of air to be investigated contain too many nuclei for the counting to be carried out unless the air is mixed with a volume of clean air, and the AITKEN counter has an arrangement for obtaining suitable proportions; WAIT (1932) pointed out errors in the calibration of counters made by AITKEN and according to his specifications. Since some nuclei do not act as

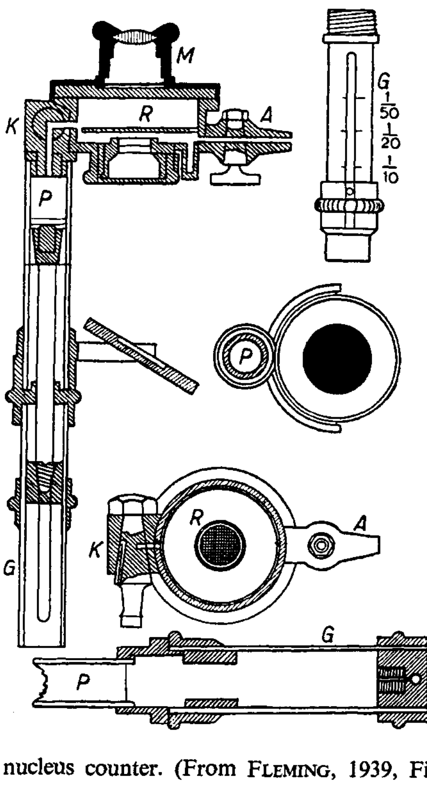


FIG. 12. The nucleus counter. (From FLEMING, 1939, Fig. 17, p. 255.)

centres of condensation on the initial expansion, further expansions are usually necessary. The AITKEN counter was developed and modified by SCHOLZ (1931 a), and POLLAK (1952) used photographic recording to give increased accuracy.

4.8. Photoelectric Counting of Nuclei

Instead of counting the droplets formed on the nuclei, after they have fallen on to a glass plate, it is possible to measure, photoelectrically, the obscuration of a beam of light passing through the cloud formed by the droplets. This method was first used by BRADBURY and MEURAN (1938) and, independently, by POLLAK and MORGAN (1940), and has been further developed in various ways; POLLAK and MURPHY (1952) used a version as a mobile instrument and VERZÁR (1953) and MURPHY (1958) arranged them to give records every quarter of an hour. Calibrations with AITKEN-type counters were made by NOLAN and POLLAK (1947) and by POLLAK and METNIEKS (1959), and in more recent versions by POLLAK and O'CONNOR (1955b) and POLLAK and METNIEKS (1957) some inaccuracies in earlier types have been removed. RICH (1955) carried out the expansion in two stages, and thus hoped to control the size of particle which can act as a nucleus, thereby making it possible to determine the distribution of such particles with size.

In the versions of the counter with which POLLAK has worked, the air in the chamber is kept saturated by means of moist blotting-paper, so that the droplets, when formed, do not evaporate; in the VERZÁR version the air is made saturated before entering the apparatus and there is no additional moisture, so the droplets evaporate after formation; various advantages are claimed for either form.

NOLAN and O'CONNOR (1955a) used the counter, without the blotting-paper, as a hygrometer, finding the relative humidity from the amount of expansion required to give condensation.

A review account of the photoelectric nucleus counter with various modifications, and the uses of its results, has been given by POLLAK (1959).

4.9. Continuous Nucleus Counter

The apparatus so far described can be used to make measurements only at intervals, because of the need for expansions.

For an apparatus which can be used to give a continuous record, HOLL and MÜHLEISEN (1954) utilized the fact that the vapour pressure over a solution is less than that over pure water. The air under examination is passed over water so as to become saturated, and is drawn into the observation chamber. At the same time, HCl gas, also saturated with water vapour, is passed into the chamber; if a drop of pure water starts to form, it will not grow in the saturated water vapour, because of the effect of the curvature; if, however, the drop contains HCl, the equilibrium vapour pressure is reduced and the drop, if formed on a nucleus, can grow and continues to do so until it falls out. Thus the number of drops formed depends on the number of nuclei present and can be measured photoelectrically by the same method as used by NOLAN and POLLAK (1947). The apparatus needs calibration with an AITKEN-type counter. By altering the concentration of HCl gas, nuclei of different sizes can be included or excluded.

4.10. Charged and Uncharged Nuclei

The numbers of the nuclei which are or are not charged can be determined in two ways; in the first, measurements are made of the total number of nuclei and of those which are uncharged. The actual counting of the nuclei is carried out by one of the methods described earlier, and counts are made with the air in its natural state and with the air after the charged nuclei have been removed. The charged nuclei are removed by passing them through a cylindrical condenser in which the rate of flow of air and the potential difference are adjusted so that all ions of mobility greater than a very low value shall be removed. It is found that the cylindrical condenser removes some of the uncharged nuclei, either by accidental impact with the walls, or by the polarization of the nuclei in the field and their subsequent motion in the inhomogeneous field and attraction by the image charges. To correct for this effect it is possible to use two cylindrical condensers in series and compare the numbers of nuclei passing through; assuming each condenser removes the same fraction of uncharged nuclei, a correction is found.

The second method uses somewhat similar apparatus, but instead of measuring the number of uncharged nuclei getting through, measurements are made of the charge reaching the cylindrical condenser. If, and only if, each charged nucleus carries a unit charge, the number of charged nuclei can be deduced. The apparatus used is, in fact, the ion counter (§ 4.11.). By comparison of results obtained by the two methods, it is possible to determine whether nuclei are multiply charged and if so what proportion (see § 4.16.).

4.11. The Ion Counter

The ion counter, designed by EBERT (1901), uses the cylindrical condenser, as described in § 2.54. and makes use of the first arrangement, in which $u < L/t$, so that all ions of mobility greater than $[u(a^2 - b^2)\log_e e\, a/b]/2VL$ are collected. If n is the number of such ions of the appropriate sign per m^3 of the air, the number passing through the apparatus per second is $n\pi(a^2 - b^2)\, u$, so that, if e is the charge per ion, the current flowing to the inner cylinder is $n\pi(a^2 - b^2)\, ue$; this, it may be noted, is independent of the applied potential difference, provided this is sufficient to maintain the condition $u < L/t$. OHM's law is not obeyed and we have the condition of saturation current (see § 2.18.).

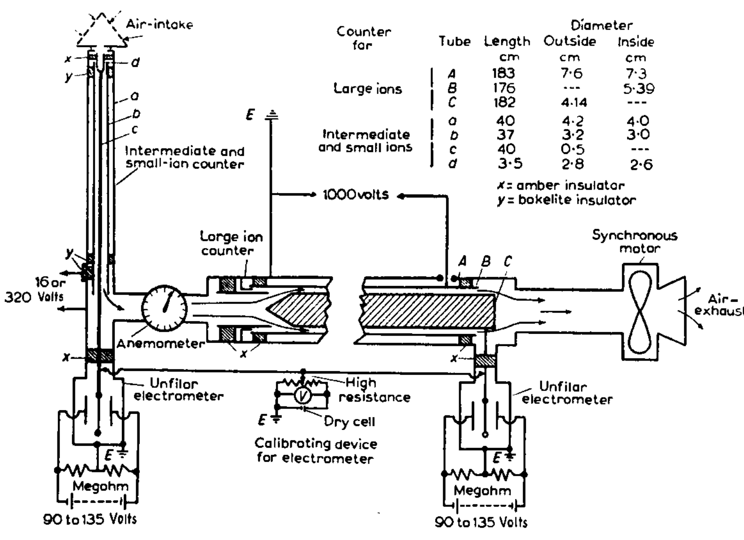


FIG. 13. The ion counter. (From FLEMING, 1939, Fig. 19, p. 259.)

In a typical example of the actual use of the apparatus, u is 2 m/sec, L is 30 cm, a is 1 or 2 cm, b is a few mm and V is 200 V. t is then of the order of $10^{-2}/w$ sec, where w is the mobility. For small ions, L/t is about 30 m/sec and for large ions about 0·03 m/sec.

Under these conditions the counter collects all the small ions and very few large ions. By increasing V or decreasing u it is possible to arrange that the large ions as well as the small ions are collected. By a method similar in principle to this, NOLAN, BOYLAN and DE SACHY (1925) measured first both small and large ions, and then small ions alone, obtaining the number of large ions by subtraction.

It is also possible to count the large ions alone if there is a "filter" to remove the small ions. The filter consists of a cylindrical condenser similar to an ion counter, with the conditions such that $u < L/t$ for small ions.

A counter records all ions with mobility greater than a specified value, but it also records ions of lower mobility if they have entered the counter close to the axis. ISRAËL (1931) has shown how to use graphical methods to give the correct number of ions if measurements are made with different applied voltages or with different air velocities.

The ion counter does not, in fact, count the ions, but, rather, measures the total charge on the ions, so that it is necessary to assume that each ion carries a unit charge.

NORINDER and SIKSNA (1949) discussed in detail the errors to which the ion counter is subject and described improvements to the apparatus.

MISAKI (1950) devised a method of recording the ion spectrum by dividing the inner cylinder into two parts and measuring the current to each.

4.12. Results of Ion Counting

The results for the number of small ions per unit volume vary very widely even at sea-level, quite apart from variations with altitude, as shown by measurements on mountains, in balloons or in aircraft. The lowest recorded value appears to be that of 40 per cc found by POLLOCK (1915a) at Sydney; MCCLELLAND and KENNEDY (1912) found values of less than 100 per cc. The highest value at sea-level is around 1500 per cc.

The number of large ions varies from about 200 per cc over or near the sea to about 80,000 per cc in large towns. As would be expected, large numbers of large ions and nuclei are associated with small numbers of small ions and vice versa.

There is always an excess of positive over negative small ions in a volume of the free atmosphere near the ground, the ratio of the numbers being, on the average, about 1·22 (NOLAN and DE SACHY, 1927). This excess may be ascribed to the electrode effect (§ 2.31.) or to the greater diffusion coefficient of the negative ions (see also § 4.23.). On the other hand, the ratio of the number of large ions of either sign is quite close to unity under most conditions.

At higher levels, e. g. at the tops of mountains, the number of small ions is greater, values of over 2000 per cc having been reported. The excess of positive over negative small ions is more marked at higher levels. The increase in the number of small ions implies an increase in conductivity and it follows from the arguments of § 2.25. that there will be a positive space charge which is supplied by the excess of positive small ions. In general, the number of large ions decreases with altitude.

The ionic content of the atmosphere shows regular annual and daily variations which are of small magnitude and can be accounted for by changes in the number of nuclei. There are also important variations with meteorological conditions; the number of small ions is found to be exceptionally large when the visibility is good and when the wind is from the sea or from high mountains, all these cases being conditions of few large ions. The number of small ions is reduced by fog or mist or even by high humidity, or by a land wind, particularly from an industrial area, bringing large ions and nuclei; NOLAN and NOLAN (1931) found the concentration of large ions at Glencree was about 1500 ion pairs per cc when the wind was from the direction of Dublin, but only 314 when winds were in other directions.

Variations in ion content near the ground during thunderstorms have been measured by NORINDER and SIKSNA (1951), who found a rise in the concentration of small ions of both signs, particularly the negative. Effects on large ions were not so marked. ADKINS (1959b) found that splashing produced ions in heavy rain, the sign of the ions released being opposite to that of the potential gradient; light rain gave no effect (see § 6.11.).

4.13. Ion Counting in Aircraft

SAGALYN and FAUCHER (1954) measured the large-ion content of the air up to 15,000 ft by carrying a large-ion counter in an aircraft, the counter being adjusted to measure ions of mobilities between 0·7 and 2·0 × 10^{-4} C.G.S. units. They found the number of large ions to vary irregularly with height within the *austausch* region, though on rather more than half the occasions the number decreased with height. Above the upper boundary of the *austausch* region, the number of large ions became very small. Though there were no simultaneous measurements of the number of small ions, the conductivity was measured, and this is mainly due to small ions; as would be expected, there was an inverse relation between large-ion content and conductivity.

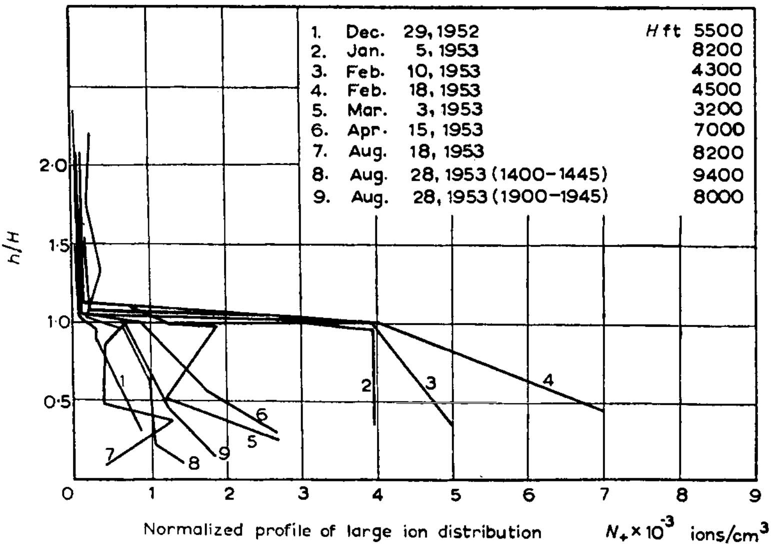


FIG. 14. Results of ion counting in aircraft. (From SAGALYN and FAUCHER, 1954, Fig. 8, p. 260.)

Within the *austausch* region, horizontal flights showed some abrupt deviations of large-ion content above an average value, the deviations being more marked near the ground and again near the top of the region. These increases of large-ion content coincided, as expected, with decreases in conductivity and also with increases in relative humidity. The results suggested turbulent convective motion of parcels of air containing large ions and water vapour in excess of the normal.

4.14. Balloon Measurements at Heights over 20 km

KROENING (1960) measured small-ion densities from balloons and found values much less than expected from simple theory.

PALTRIDGE (1965) made further similar measurements and found that the aerodynamic conditions were such that the ion-counter did not function according to theory; however, even with a correction, the small-ion density was found lower than would be expected. This was attributed to dust present in the stratosphere, originating from volcanoes.

PALTRIDGE (1965) also made some measurements with a constant-level balloon at 30.5 km and found the ion density to be only one-third of that expected.

4.15. Measurements in Large Enclosed Sphere

GUNN and ALLEE (1954) described a 3000 m^3 spherical chamber, and measurements of the numbers of ions of both signs were made by PHILLIPS *et al.* (1955). Such measurements differ, on the one hand, from those in the free atmosphere because there are no effects of potential gradients (e. g. electrode effect) and no effects of wind. On the other hand, they differ from those in small enclosed chambers because wall effects are reduced and the ionization is close to that in the free atmosphere.

The results showed that when nuclei are present there is always an excess of positive small ions, but when nuclei have been removed there are equal numbers of ions of both signs. The latter result is what is to be expected, since equal numbers of ions of both signs are produced, and if small ions can be removed only by recombination there can never arise a difference; but, when nuclei are present, the fact that negative ions attach themselves more readily to nuclei than do positive ions must give an excess of positive small ions. Since the result in the presence of nuclei agrees with that in the free atmosphere, it follows that the positive excess of small ions normally found is not due entirely to the electrode effect or other consequence of the potential gradient (see § 4.12.). Results of conductivity measurements are given in § 7.18.

4.16. Multiply Charged Nuclei

A comparison of the two methods of obtaining the number of charged nuclei (§ 4.10.) by NOLAN, BOYLAN and DE SACHY (1925) led to the conclusion that nuclei are always singly charged. However, measurements by HOGG (1934a) showed that this is not true when the total number of nuclei is small; it appears from these results, and others analysed by ISRAËL (1941), that when the total number of nuclei is greater than about 10^4 per cc, they are all singly charged, but for smaller numbers, some carry multiple charges.

Artificially-produced nuclei often carry multiple charges (NOLAN and O'KEEFFE, 1933; NOLAN and O'CONNOR, 1955b) and appear to show anomalous values of the mobility, which might be accounted by the effect of charge on mobility.

4.17. The Production of Ionization

There are various processes which may be concerned in the production of ions in the atmosphere. Certain of these are of little importance and can be dismissed shortly. It is well known that ionization can be produced by the photoelectric effect, but if the solar radiation contains wave lengths capable of producing a photoelectric effect among the molecules of the air, such radiation would be absorbed in the upper parts of the atmosphere and is unlikely to penetrate to the portions of the atmosphere with which we are concerned in atmospheric electricity in sufficient amount to produce appreciable ionization. While the photoelectric effect is certainly of importance in the production of ionization in the ionosphere, it can be neglected as regards the lower atmosphere. There is no evidence for any appreciable ionization by the photoelectric effect on the earth's surface.

LENARD (1892) showed that there is a production of charge on the air and the water, due to the breaking of water drops, for example at waterfalls. It seems improbable that such a production of ionization can be of great importance in the atmosphere except in the immediate neighbourhood of waterfalls or perhaps the seashore. The possible importance of this effect in the electrification of clouds will be discussed later (§ 16.21.).

EBERT (1904) found that air diffusing from the ground contains ions, presumably derived from radioactivity in the earth's crust. The rate at which this occurs depends upon the conditions and varies widely (see ZEILINGER, 1935).

VOLTA (see § 1.7.) and PELTIER (see § 1.9.) thought that evaporation of water gave electrification, and this would presumably be in the form of ions. But no evidence has been found for such a process (see GILBERT and SHAW, 1925).

Ionization by collision occurs in lightning flashes and in the process of point discharge, but these occur only in limited regions of the air and for limited times in stormy weather.

Dust storms are highly electrified, presumably with production of ions, and the same process may also occur on a smaller scale. The same may apply to blizzards.

"Man-made" ions may also be important at some places and times. For example, steam from locomotives produces a positive potential gradient, probably due to positive ions (ISRAËL, 1950b) (see § 5.57.) and CHALMERS (1952a) found negative potential gradients in mist and fog, ascribed to negative ions produced at high-tension cables (see § 5.59.). Other similar effects have been described by MÜHLEISEN (1953) (see § 5.57.).

4.18. Ionization due to Radioactivity

Radioactive substances in the atmosphere and in the earth's crust are important in the production of ions in the atmosphere. If measurements are made in a closed vessel there may also be ionization due to radioactive contamination of the material of the vessel; it is clear that this last effect must be eliminated before it can be possible to use measurements in the closed vessel to obtain the rate of production of ions in the free air.

The effect, in a closed vessel, of radioactivity of the earth's crust can be successively reduced by interposing sheets of lead below the vessel. In the same way the effect of the radioactivity of the atmosphere outside the vessel can be eliminated by surrounding it with lead. However, the residual ionization in a closed vessel is not the same as that in the free air, since radioactive substances produce ionization not only at the places where they are situated but also at some distance.

Ionization from radioactive substances can be produced by α-, β- and γ-rays. The α-rays have the greatest energy, but the least power of penetration; so the ionization in the free air produced by radioactive substances in the atmopshere is largely due to the α-rays, but the walls of a measuring vessel will stop α-rays from outside unless the walls are very thin; thus the ionization in a closed vessel due to α-rays from the atmosphere is less than in the free air.

The ionization due to radioactivity in the earth is due more to β- and γ-rays than to α-rays because β- and γ-rays can emerge into the atmosphere from depths from which α-rays are stopped. Thus the walls of the measuring vessel have less effect upon the measurements of radioactive effects from the ground than from the air.

4.19. Ionization due to Cosmic Rays

When the vessel used for the measurement of ionization is completely shielded by thick sheets of lead, effects of radioactivity could come only from the air in the vessel itself or from the walls of the vessel. In either case, careful removal of possible sources of radioactivity should reduce the effect; yet, as was demonstrated first by ELSTER and GEITEL (1901) and WILSON (1901), there still remains a definite ionization produced in the vessel. The nature of the gas in the vessel does not affect the amount of ionization, and a change of pressure alters the ionization in proportion to the pressure; if the effect were due to α-rays from the material of the walls, an increase of pressure would be expected to give a slight decrease of measured ionization since the ions produced by the α-rays would be closer together, giving greater opportunity for recombination. Also different vessels show differences only in proportion to their volumes, so that there can be no doubt that the residual ionization must be caused by penetrating radiation from outside the vessel. At first it was thought to be "ultra-γ" radiation from the earth, but HESS (1911) and KOLHÖRSTER (1913) found an increase on rising above the earth, showing that this radiation could not be terrestrial in origin.

No attempt can be made here to give any kind of summary of the knowledge that has been obtained about the "cosmic rays". As far as atmospheric electricity is concerned, all that is of importance is that they produce ionization which is greater at higher levels in the atmosphere than lower down. The amount of ionization shows little dependence on solar or sidereal time, but does show some change with geomagnetic latitude, being less at the magnetic equator than at higher magnetic latitudes.

4.20. Measurement of Rate of Ionization

In order to measure the rate of production of ions in a closed vessel it is necessary to apply a large field, sufficient for saturation current, so that all the ions are measured as soon as they are produced, without allowing recombination.

HOGG (1934a, b, 1935) used a 50 l. vessel, with walls of 0·4 mm thickness, for continuous observations, refilling the vessel every hour. WAIT and MCNISH (1934) used a 25 l. vessel with a cellophane wall of 0·03 mm thickness; even this wall corresponds to an absorption equivalent to 1·5 cm of air, while HOGG's walls would reduce the α-ray effect to about 20 per cent of that in the free air. KAWANO (1958b) used a 27 l. vessel with walls only 7 μ thick and allowed a continuous flow of air.

A closed vessel gives the rate of ionization in free air due to cosmic rays and to radioactivity in the ground to a fair degree of accuracy, particularly if it is at a height greater than the range of the α-rays from the ground (less than 10 cm). For the effect of radioactivity in the atmosphere itself, it is probably more accurate to collect the radioactive matter and deduce from it the expected rate of ionization than to make measurements in a closed vessel.

KAWANO (1958b) made the surprising observation that, during a 90 per cent eclipse of the sun, the rate of production of ions at a height of 1 m rose to double the normal value, although before and after the eclipse the values agreed with the normal. No explanation of this has yet been offered.

It must be realized that measurements of the total number of ions produced, by the use of a thin-walled vessel and a field sufficient for saturation, will give a higher value for the rate of production of ions than the value which is appropriate for considerations of equilibrium, since some ions are separated by the field and counted, while in natural conditions they would have recombined while still close together.

4.21. β-Ray Ionization

HESS, PARKINSON and MIRANDA (1953) discussed the ionization due to β-rays and reached the conclusion that the effects are small compared with those of γ-rays, except in the case where there is extremely little mixing in the lowest layers of the atmosphere. Some of the β-ray effects are to be ascribed to surface contamination by radioactive fall-out from nuclear explosions.

More recently, MIRANDA (1958) measured β-ray effects and found them to be greater, by a considerable factor, than can be accounted for by the measured radioactive substances in the atmosphere. He suggested that the effects are due to very soft cosmic rays, of the same range of energy as β-rays but not arising from radioactivity. This must mean that the total effect of soft radiations is greater than can be estimated by collecting the radioactive matter in the atmosphere.

4.22. The Total Ionization of the Air

In discussions of the rate at which ions are produced in the air it is customary to use the unit "I", representing one pair of ions per cc per second. In air at normal temperature and pressure and at sea-level the production of ions by cosmic rays amounts to about 1·8 I at geomagnetic latitudes over 50° and to about 1·5 I nearer the equator. The ionization due to cosmic rays increases with altitude above sea-level.

Over the sea the contribution due to radioactivity is negligible and so the total ionization in free air or in a closed vessel with uncontaminated walls is 1·8 I or 1·5 I according to latitude. But over the land there is also the effect of the radioactive material in the air and the earth; the number of ions produced by radioactivity depends upon the amount of radioactive material in the earth's crust at the place of measurement and varies widely from one place to another, so that, for a general survey, only average values can be given.

The average value, over the land, of the effect of radioactivity in the ground amounts to about 4 I. For the radioactive substances in the atmosphere, measurements with a very thin-walled vessel with correction for absorption, or by collection of radioactive substances, the ionization in a volume of free air would be about 5 I, giving a total ionization over land of around 11 I. In some places, however, higher values have been obtained. WAIT (1931), at Washington, and PARKINSON (1948) made measurements which showed a variation with local time, perhaps caused by differences in the rate of efflux of gas containing radioactive matter from the ground. WAIT found an average value at Washington of about 16 I and PARKINSON at Huancayo (altitude 3300 m) about 40 I.

It is possible to deduce values of the rate of production of ionization from the consideration of ionization equilibrium (see § 4.35.).

4.23. Ionization Close to the Ground

Very close to the earth's surface the above results are not correct, because of effects of α- and β-ray emitters in the ground and on the surface. PIERCE (1958) made use of various measurements to derive a curve showing the variation of the rate of production of ions with height between 1 cm and 1 m. His results show that the rate alters from about 60 I at 1 cm to about 8 I at 1 m. PIERCE also pointed out that radioactive fall-out is likely to be increasing the effect of β-rays at the surface.

The importance of this variation of ion production with height in connection with the potential gradient and conductivity in the lowest metre of the atmosphere is discussed in §§ 7.9. and 8.20.

4.24. Ions from the Ground

The mobility and diffusion coefficient of negative ions are both greater than those of positive ions, so that air diffusing out of the soil, having been ionized by radioactivity in the earth, will contain an excess of positive ions which have not been so readily drawn into the walls of the interstices of the ground as the negative ions have been. This effect was made the basis of a theory of the maintenance of the earth's charge by EBERT (1904) but it was not found to be of sufficient magnitude. The effect, however, does exist and may be of some importance in the ionization of the lowest regions of the atmosphere in natural conditions.

4.25. Ions from the Sea

BLANCHARD (1961) carried out extensive measurements on charged particles produced at the sea surface by the action of wind and waves. He found a quite appreciable current of positive charge upwards over the sea surface from "jet drops", i. e. drops produced by the breaking of air bubbles and projected upwards. The charges are carried on particles usually larger than of large-ion size.

4.26. Equilibrium of Ionization

In any volume of air, if conditions remain steady, the total number of ions in the volume must be constant. This is extended to say that, for both signs of ion separately and for each size of ion that is considered distinct, there is a balance between, on the one hand, the number entering the volume and the number produced in the volume, and, on the other hand the number leaving the volume and the number destroyed in the volume.

Unless we are dealing with a volume very close to the earth's surface, then in normal fine weather the number of ions of any one kind entering a volume will be equal to the number leaving the volume and these terms can be left out in considering the equilibrium. Close to the earth, where the rate of production of ions is varying rapidly with height, this condition will no longer be true.

The normal processes of the production of ions give small ions and not directly any large ions. Large ions can be produced only by the attachment of small ions to nuclei.

4.27. Destruction of Ions

Small ions can be destroyed in three ways: by recombination with a small ion of opposite sign, giving neutral molecules; by combination with a large ion of opposite sign, giving a neutral nucleus and neutral molecules (more than one if the small ion has attached molecules); and by combination with a nucleus to give a large ion. If multiply charged large ions exist, there will also be the process of combination with a large ion of the same sign to give an ion of greater charge.

Large ions cease to exist as such by combination with small ions of opposite sign or by combination with large ions of opposite sign. In each case neutral nuclei are formed.

We can expect the number of combinations that occur in unit time in a given volume to be proportional to the numbers present of the two types of particle concerned; the factor of proportionality defines a "combination coefficient" for the process concerned.

4.28. Discussion of Equilibrium

In the discussion of the equilibrium of ionization, it is assumed that the processes at work are: (1) the production of small ions (but not large ions directly) by radioactivity, cosmic rays and possibly other agents; (2) the combination of small ions and uncharged nuclei to form large ions; (3) the recombination of small ions (often negligible); (4) the combination of large ions with small ions of opposite sign; (5) the combination of large ions with small ions of the same sign to give multiply charged ions (neglected in the simpler discussion); and (6) the combination of large ions of opposite sign (usually negligible). The intermediate ions and their effects have been left out of consideration, and it is assumed that all the large ions are of the same size, etc.

The symbols n_1, n_2, N_1, N_2 represent the numbers of small ions and large ions of the two signs in unit volume, N_0 the number of uncharged nuclei and $Z = N_0 + N_1 + N_2$ the total number of nuclei. q is the rate of production of small ions per unit volume per second.

The recombination coefficient for small ions of both signs is α; for combination between small ions of one sign and large ions of opposite sign the coefficients are η_{12} and η_{21}, the first suffix referring to the small ions and the second to the large; for combination between small ions and uncharged nuclei the coefficients are η_{10} and η_{20}, and, finally, for combination between large ions of both signs the coefficient is γ.

Considering small positive ions,

$$\frac{dn_1}{dt} = q - \alpha n_1 n_2 - \eta_{12} n_1 N_2 - \eta_{10} n_1 N_0$$

where q is the rate of production of small ions.

For small negative ions,

$$\frac{dn_1}{dt} = q - \alpha n_1 n_2 - \eta_{21} n_2 N_1 - \eta_{20} n_2 N_0.$$

For the large ions,

$$\frac{dN_1}{dt} = \eta_{10} n_1 N_0 - \eta_{21} n_2 N_1 - \gamma N_1 N_2$$

and

$$\frac{dN_2}{dt} = \eta_{20} n_2 N_0 - \eta_{12} n_1 N_2 - \gamma N_1 N_2.$$

The value of γ is so small that $\gamma N_1 N_2$ can nearly always be neglected and usually n_1 and n_2 are small compared with N_1 and N_2, so that the term $\alpha n_1 n_2$ is often also negligible.

When equilibrium is attained, the variations with time are zero and we get

$$q = \eta_{12} n_1 N_2 + \eta_{10} n_1 N_0 = \eta_{21} n_2 N_1 + \eta_{20} n_2 N_0,$$

$$\eta_{21} n_2 N_1 = \eta_{10} n_1 N_0,$$

and

$$\eta_{12} n_1 N_2 = \eta_{20} n_2 N_0.$$

These equations were first deduced by NOLAN and DE SACHY (1927), improving on the earlier work of NOLAN, BOYLAN and DE SACHY (1925) who had not differentiated between the properties of the ions of the two signs.

Usually, $N_1 = N_2 = N$ and so

$$\frac{N_0}{N} = \left[\frac{\eta_{12}\,\eta_{21}}{\eta_{10}\,\eta_{20}}\right]^{\frac{1}{2}} \quad \text{and} \quad \frac{n_1}{n_2} = \left[\frac{\eta_{21}\,\eta_{20}}{\eta_{12}\,\eta_{10}}\right]^{\frac{1}{2}}.$$

If the values of the η's are constant, then the ratios N_0/N and n_1/n_2 should remain constant, and any variations in the ratios would be ascribable to alterations in the η's and hence in the sizes, etc., of the large ions.

These equations can be simplified if we neglect the differences between the two signs; then if

$$n_1 = n_2 = n, \quad \eta_{12} = \eta_{21} = \eta \quad \text{and} \quad \eta_{10} = \eta_{20} = \eta_0$$

we get

$$q = \eta n N + \eta_0 n N_0$$

and

$$\eta N = \eta_0 N_0.$$

4.29. Extension to Multiply-charged Nuclei

If we can no longer neglect the possibility of a positive small ion becoming attached to a positive large ion, forming a doubly-charged large ion, then we must introduce extra terms in the equations and we get, for example:

$$\frac{dN_1}{dt} = \eta_{10} n_1 N_0 - \eta_{21} n_2 N_1 + \eta_{211} n_2 N_{11} - \eta_{11} n_1 N_1,$$

where the meanings of the new symbols η_{211}, η_{11}, N_{11} are clear from the context.

If, now, we neglect the differences between positive and negative particles and put $\eta_{211} = \eta_{122} = \eta^{21}$ and $\eta_{11} = \eta_{22} = \eta^{12}$ then

$$\frac{dN}{dt} = \eta_0 n N_0 - \eta n N + \eta^{21} n N_{11} - \eta^{12} n N.$$

We can also get, as we could before,

$$\frac{dN_0}{dt} = 2\eta n N - 2\eta_0 n N_0.$$

In equilibrium, besides $\eta N = \eta_0 N_0$ still, we get

$$\eta_{21} N_{11} = \eta_{12} N.$$

Similarly,

$$\frac{\eta_{32}}{\eta_{23}} = \frac{N_{11}}{N_{111}}.$$

These results were obtained by BRICARD (1949).

4.30. Application of Boltzmann's Law

KEEFE, NOLAN and RICH (1959), following up a suggestion by NOLAN (1955), treated the problem of charge equilibrium in terms of BOLTZMANN's law.

The simplest statement of BOLTZMANN's law for the present purpose is that if there are P particles having an energy E and P' an energy E', then

$$P'/P = \exp[-(E' - E)/kT],$$

where k is the BOLTZMANN constant and T the absolute temperature.

If there is a charged particle of radius r carrying x elementary charges (each e), then its electrical energy is

$$\frac{x^2 e^2}{8\pi\varepsilon_0 r} = \text{say, } x^2 ykT,$$

and this energy must be added to any other energy which the particle possesses.

Thus

$$N_x/N_0 = \exp(-x^2 y).$$

The total number of particles must then be

$$Z = N_0 + 2N_0 \exp(-y) + 2N_0 \exp(-4y) + \ldots$$

which can be shown to give approximately

$$Z = N_0 \left(\frac{\pi}{y}\right)^{\frac{1}{2}} = N_0 \left(\frac{8\varepsilon_0 rkT}{e^2}\right)^{\frac{1}{2}}.$$

This result for Z/N_0 was obtained in a quite different way by GUNN (1955b).

4.31. Theoretical Treatment of Combination Coefficients

Many attempts have been made to deduce the values of the combination coefficients theoretically, one of the earliest being that of LANGEVIN (1903) for the value of α; this will be discussed in § 4.33., where it is used in conditions where it is more likely to be applicable than in LANGEVIN's use. A criticism that LANGEVIN's theory neglects diffusion was answered by HARPER (1940) who showed that diffusion has a surprisingly small effect.

Various attempts have been made to derive values for the η's, particularly η_0, and it has become apparent that there are two limiting conditions, when the large ions or nuclei are large or small compared with the mean free path of the small ions. If, in the immediate neighbourhood of the large ion (i. e. within a small multiple of the radius), small ions suffer few or no collisions (this is for small sizes of the large ion or for low pressures), then their acceleration is proportional to the force acting and the chance of capture will be the geometrical one, enhanced by any electrostatic forces acting. On the other hand, if, in the neighbourhood of the large ions, the small ions suffer very many collisions, it is their velocity which is proportional to the force acting and the chance of capture will depend on the same factors as the diffusion coefficient and the mobility.

For large nuclei many workers have used the simple "diffusion-mobility" formula:

$$\eta_0 = 4\pi r D,$$

where r is the radius of the nucleus and D the diffusion coefficient of the small ions.

Using the relation $D = kTw/e$ (see § 2.14.), where w is the mobility of the small ions and e the ionic charge,

$$\eta_0 = \frac{4\pi r k T w}{e},$$

or, using the notations of § 4.30.

$$\eta_0 = \frac{ew}{2\varepsilon_0 y}.$$

For small nuclei, Wright (1936) obtained

$$\eta_0 = \pi r^2 u,$$

where u is the average velocity of the small ions.

Bricard (1949) and Keefe, Nolan and Rich (1959) have modified the formulae by taking into account the force of attraction between a charged small ion and a neutral nucleus due to the electrostatic image. This gives a factor $(1 + (\pi y)^{\frac{1}{2}})$.

Keefe and Nolan (1962) and Keefe and Nolan (1961) have produced a general formula for all sizes of nuclei, which reduces to the above cases for large and small nuclei.

Using the diffusion-mobility formulae for large nuclei and considering other coefficients than η_0, various workers, e.g. BRICARD (1949), GUNN (1954c), have obtained such values as

$$\eta = \frac{ew}{\varepsilon_0(1 - 1/l^2)}$$

$$\eta^{12} = \frac{ew}{\varepsilon_0(l^2 - 1)} \qquad \eta^{21} = \frac{2ew}{\varepsilon_0(1 - 1/l^4)}$$

where $l = \exp(e^2/8\eta\varepsilon_0 rkT) = \exp y$.

On the other hand, as KEEFE, NOLAN and RICH (1959) have pointed out, the formula from BOLTZMANN's law give

$$\eta/\eta_0 = l, \eta^{21}/\eta^{12} = l^3, \quad \text{etc.}$$

These ratios of the η's are different from BRICARD's, but tend towards equality as l approaches unity, i.e. as r becomes large.

But the diffusion-mobility results give

$$\eta/\eta^{12} = l^2, \quad \eta^{21}/\eta^{23}, = l^4, \quad \text{etc.,}$$

and combining these with the BOLTZMANN results we can produce a sequence

$$\eta^{34}, \eta^{23}, \eta^{12}, \eta_0, \eta, \eta^{21}, \eta^{32}$$

which forms a geometrical progression of ratio l, for large radii.

4.32. Nolan and de Sachy's Assumption

NOLAN and DE SACHY (1927) made the assumption that the ratios η_{12}/η_{10} and η_{21}/η_{20} are equal. This means that small ions of the two signs behave similarly as regards their preference for combining with oppositely charged rather than neutral nuclei. If these ratios are g, then § 4.28. shows that $N_0/N = g$. It is easy to see that this is reasonable, since the larger is g, the more readily will small ions combine with charged rather than uncharged nuclei and so the larger will be the number of uncharged nuclei remaining, compared with the charged.

If $\eta_{12}/\eta_{10} = \eta_{21}/\eta_{20}$, it follows that $\eta_{21}/\eta_{12} = \eta_{20}/\eta_{10}$. If this is p, then § 4.28. shows that $n_1/n_2 = p$. This is reasonable, since the larger is p the more readily do the negative, rather than positive, small ions combine with oppositely charged or neutral nuclei, and so the fewer negative small ions will remain, compared with the positive.

The actual process of the combination of small ions with uncharged nuclei must involve mainly the accidental collision of the particles, and hence the difference between η_{10} and η_{20} must be mainly a difference in the speeds with which the small ions move. NOLAN and DE SACHY suggested that the relative speeds for the small ions in random motion would correspond with their relative speeds under the influence of an electric field, and therefore that p should depend on the ratios of the mobilities; from the actual values, they suggested that p is the ratio of the mobilities raised to the $\frac{3}{4}$ power, but there has been no theoretical justification for this figure.

4.33. Whipple's Theory

WHIPPLE (1933) proposed a theory to obtain a relation between η and η_0, based on the theory of LANGEVIN (1903) for obtaining the value of α.

WHIPPLE assumed that η_0 represents the number of collisions between small ions and nuclei, whether charged or uncharged, due merely to their kinetic motion, independent of any effect of the charges such as image forces. In addition to these collisions which occur equally for charged and uncharged nuclei, there will be extra collisions for the charged nuclei because of electrostatic attraction.

In comparison with the small ions, the large ions are taken as being at rest. A small ion at a distance R from a large ion of opposite sign will experience a force of attraction $e^2/4\pi\varepsilon_0 R^2$ and so will move towards the large ion with a velocity $we/4\pi\varepsilon_0 R^2$, where w is the mobility of the small ion. Thus, in unit time, the number of small ions which enter a sphere of radius R round a large ion of opposite sign, by reason of the electrostatic attraction, is

$$\frac{we}{4\pi\varepsilon_0 R^2} \times 4\pi R^2 n = \frac{new}{\varepsilon_0},$$

which is independent of the value of R. If the conditions are now chosen so that an ion entering the sphere always combines with the large ion and if there are N large ions per unit volume, then the extra combinations due to the electrostatic forces are

$$\frac{nNew}{\varepsilon_0}.$$

Thus

$$\eta - \eta_0 = \frac{ew}{\varepsilon_0}.$$

It will be seen that this theory can give also:

$$\eta_{12} - \eta_{10} = \frac{ew_1}{\varepsilon_0} \quad \text{and} \quad \eta_{21} - \eta_{20} = \frac{ew_2}{\varepsilon_0}.$$

These results can agree with the assumption of NOLAN and DE SACHY only if

$$\frac{\eta_{20}}{\eta_{10}} = \frac{\eta_{21}}{\eta_{12}} = p = \frac{w_2}{w_1} \qquad \text{also}.$$

If we use the results of § 4.31, that for large values of r

$$\eta_0 = 4\pi rkTw/e$$

and

$$\eta/\eta_0 = l$$

so that

$$\eta \sim \eta_0 (1 + y)$$

then, approximately,

$$\eta - \eta_0 = \eta_0 y = \frac{ew}{2\varepsilon_0},$$

which differs from WHIPPLE's formula by a factor 2.

4.34. Observational Values of Combination Coefficients

The value of α, the combination coefficient for small ions of both signs, is usually taken as $1{\cdot}6 \times 10^{-6}$ cc/sec; however, LUHR and BRADBURY (1931) gave $1{\cdot}23 \times 10^{-6}$, SAYERS (1938) $2{\cdot}4 \times 10^{-6}$ and NOLAN (1943) $1{\cdot}4 \times 10^{-6}$. The units of cc/sec are here used throughout.

The above equations would give values of the η's from knowledge of N_0/N, n_1/n_2, w_1 and w_2, provided that the ionization is in equilibrium. The values of N_0/N (see § 4.37.) show much divergence, probably through lack of equilibrium, but the majority of the results lie between 2 and 3. The values of n_1/n_2 are also not consistent and might be influenced by the electrode effect (§ 2.31.); NOLAN and DE SACHY (1927) found a value of 1·24 for atmospheric air, but preferred the value of 1·11 found for indoor air, and SCRASE (1935a) found 1·51 for atmospheric air. BARKLIE (1956) has rewritten the equation as

$$N_2/N_1 = (n_2/n_1)^2 \frac{\eta_{20}\ \eta_{21}}{\eta_{10}\ \eta_{12}}$$

and measured N_2/N_1 for air which has been very highly ionized, so that n_1 and n_2 are nearly equal and obtained a result of 1·6, which corresponds to $n_1/n_2 = 1{\cdot}26$ for $N_2 = N_1$.

NOLAN and DE SACHY (1927) used their assumption (§ 2.31.) and $n_1/n_2 = 1{\cdot}11$, and obtained values of the η's. SCRASE (1935a) used WHIPPLE'S (1933) theory and $n_1/n_2 = 1{\cdot}51$. GISH and SHERMAN (1940) and PARKINSON (1949) have given detailed discussions of the derivation of values for the η's.

Using BARKLIE'S (1956) result, it would seem that the η's must lie within the limits

$\eta_{10} = 1{\cdot}20 \times 10^{-6}$	and	$\eta_{10} = 2{\cdot}20 \times 10^{-6}$
$\eta_{12} = 3{\cdot}97 \times 10^{-6}$		$\eta_{12} = 4{\cdot}97 \times 10^{-6}$
$\eta_{20} = 1{\cdot}67 \times 10^{-6}$		$\eta_{20} = 3{\cdot}00 \times 10^{-6}$
$\eta_{21} = 4{\cdot}56 \times 10^{-6}$		$\eta_{21} = 5{\cdot}89 \times 10^{-6}$
$N_0/N = 3{\cdot}0$		$N_0/N = 2{\cdot}1$

Because of the large number of small ions present, equilibrium would be attained rapidly and BARKLIE'S figures are probably the most reliable. But if it is confirmed that WHIPPLE'S result needs amendment by a factor of 2, revision of these figures would be necessary.

For γ, the combination coefficient for large ions of both signs, values of the order of 10^{-9} were found by KENNEDY (1913, 1916) and NOLAN (1941). HOGG (1934b) found a value about 10 times greater, but WAIT and PARKINSON (1951) have criticized this result. WAIT (1946) found that the value of γ is proportional to the $\frac{3}{4}$ power of the mobility of the ions concerned, giving

$$\gamma = 1{\cdot}25 \times 10^{-6} w^{\frac{3}{4}},$$

where w is measured in C.G.S. units. For a mobility of $3{\cdot}2 \times 10^{-4}$, γ becomes $3{\cdot}1 \times 10^{-9}$. In discussions of equilibrium, terms in γ are always negligible.

4.35. Rate of Ionization

From measurements on the numbers of ions, and assuming equilibrium of ionization, it is possible to calculate q, the rate of production of small ions.

From the assumptions of NOLAN and DE SACHY (1927), including the assumption that $N_1 = N_2$, GISH (1939) derived the result:

$$q = \alpha' n^2 + \beta' nN,$$

where

$$\alpha' = \alpha\eta_{10}/\eta_{20}$$

and

$$\beta' = 2\eta_{12}/(2 + \eta_{12}/\eta_{10}).$$

Using, as is essential in this treatment, the η-values of NOLAN and DE SACHY, GISH found for Washington, D. C., values of 14 I and 18 I, while the value found directly was only 6–8 I. On the other hand, using the results of BUILDER (1929) at Watheroo, the value of q amounts to only 2·46 I.

If, however, WHIPPLE's theory is used (see § 4.33.), then when $N = N_1 = N_2$, from § 4.28.:

$$q/N = \eta_{12}n_1 + \eta_{21}n_2$$

and

$$q/N_0 = \eta_{10}n_1 + \eta_{20}n_2.$$

Hence

$$q = \frac{N_0}{\varepsilon_0}\frac{(n_1 e_1 w_1 + n_2 e_2 w_2)}{(N_0/N - 1)} = \frac{N_0}{\varepsilon_0}\frac{(\lambda_1 + \lambda_2)}{(N_0/N - 1)},$$

where λ_1, λ_2 are the conductivities for positive and negative small ions respectively.

For Kew, SCRASE (1935a) used this formula and obtained $q =$ 4·5 I, but CHALMERS (1949) pointed out that, in obtaining $\lambda_1 + \lambda_2$, SCRASE had assumed that air–earth current measurements give $F\lambda_1$ (see §§ 2.31., 8.13., 8.19.); correcting for this, q becomes 2·5 I, much too small to be a true value. WAIT (1935a) used the measurements of TORRESON and WAIT (1934) for Washington, D.C., and obtained $q = 4{\cdot}5$ I.

It may be surmised that the lack of agreement between the directly observed and the indirectly calculated values of q is due to the fact that the latter are based on measurements which are assumed to be under equilibrium conditions but which, in fact, are not (see § 4.36.).

4.36. Rate of Approach to Equilibrium

Under conditions when the numbers of ions of each sign can be taken as equal, the equations of § 4.28. can be discussed approximately by the simplification of neglecting all differences between ions of different signs and putting $n_1 = n_2 = n$, $N_1 = N_2 = N$, $\eta_{12} = \eta_{21} = \eta$ and $\eta_{10} = \eta_{20} = \eta_0$.

The equations then become

$$\frac{dn}{dt} = q - \alpha n^2 - \eta n N - \eta_0 n N_0$$

and

$$\frac{dN}{dt} = \eta_0 n N_0 - \eta n N$$

with the neglect of the term in γ and the assumption that there is no production of large ions.

Israël (1948b) integrated the first equation, assuming constant values of N and N_0 and found that n approaches its final value with a half-value time of about 1 sec, using ordinary values of the quantities involved. Stergis (1954) found that this time would be considerably increased if the rate of production of ions (q) was increased or if there was little pollution (small N and N_0), and confirmed these predictions both in laboratory experiments and in the free air.

Israël (1948b) also considered the equation for dN/dt for times at which n could be considered to have reached its final value, and with $Z = N_0 + 2N$ constant, and he found that N would approach its final value with a half-value time of about $5\frac{1}{2}$ min. Nolan (1950) independently reached a similar result for the variation of N, but used constants giving a half-value time of 10–15 min.

The matter has been taken further by Rich, Pollak and Metnieks (1962), making use of the theoretical relation between η_0, diffusion coefficient and radius, and the application of Boltzmann's law to the consideration of the numbers of charged and uncharged nuclei (Keefe, Nolan and Rich, 1959). They deduced differential equations which were solved by a computer for various conditions; the conclusions were similar to those of the earlier workers, namely that n reaches its final value quite quickly—the difference is inappreciable after a few minutes, while N takes much longer and is still significantly different from its final value after an hour; in 3 cases quoted, N reaches half its final value in 4 min, 10 min and 12 min.

The fact that equilibrium conditions for N take so long to be reached means that it is probable that conditions in the free air are seldom sufficiently steady to give large-ion equilibrium and so the measured values of N_0/N are unlikely to give reliable information in regard to the η's. It is, however, found that air from the sea is in equilibrium (O'Connor and Sharkey, 1960).

4.37. Ratio of Charged and Uncharged Nuclei

If the theory of § 4.28. holds, then the ratio N_0/N gives information as to the values of the η's; further information can be obtained from the assumption of Nolan and de Sachy or from Whipple's theory. A considerable amount of attention has therefore been given to measurements of this ratio or of quantities related to it. If there is equilibrium and if the η's remain the same, the ratio N_0/N should be always the same.

In addition to the simple ratio N_0/N, other quantities sometimes measured are $Z/(N_1 + N_2)$ or $(N_0 + N_1 + N_2)/(N_1 + N_2)$, often denoted by P and also Z/N_0 or $(N_0 + N_1 + N_2)/N_0$ or $P/(P - 1)$.

Neglecting differences between ions of different signs, then, as Nolan (1950) pointed out, equilibrium conditions give $N_0/N = \eta/\eta_0$. From the arguments of Whipple's theory (§ 4.33.), η_0 is necessarily less than η, so that $N_0/N > 1$. This gives $P > \frac{3}{2}$, or $Z/N_0 < 3$. If results do not fall within these limits, it must mean that the basic assumptions are not justified; in particular, those which may be incorrect are that the nuclei are not multiply charged or that equilibrium conditions exist.

Israël (1941) has given a list of 19 sets of observations, some in free air and some in the laboratory; the average values of P quoted vary from 1·64 to 5·4. The general results appear to show an increase of P with the number of nuclei, and also a decrease of P with increasing relative humidity.

Israël pointed out that the "scatter" of the single values of P in the separate sets of observations is very different according to the location of the observing station, in particular the distance from the nearest large town, which is likely to be the source of the nuclei; this suggested that the results were due to the lack of equilibrium, though it was not realized at that time how long it takes for nuclei and large ions to reach equilibrium.

In view of the doubt thrown upon the true meaning of measured values of P in the free air, we shall not discuss these results further.

Measurements have been made of the relations between charged and uncharged nuclei stored in the laboratory to give equilibrium, and the results so obtained are in reasonably good agreement with expectation (see NOLAN, 1950). But such results are of little significance in relation to atmospheric phenomena, and we shall not discuss them.

POLLAK and METNIEKS (1962) compared values of N_0/N found experimentally with those predicted by BOLTZMANN's law (see § 4.30.) and found divergences for radii of nuclei less than $1{\cdot}4 \times 10^{-6}$ cm, the actual ratio being less than the theoretical; an apparent divergence in the opposite direction for larger radii can be explained.

4.38. Ratio of Small to Large Ions

In equilibrium, it is clear that the more large ions and nuclei are present the smaller will be the number of small ions.

Various workers have used the relation:

$$nN^x = \text{constant},$$

and found x empirically. NOLAN (1929) found $x = \frac{1}{2}$, NOLAN and GUERRINI (1935) $x = \frac{1}{4}$ and WRIGHT (1936) $x = \frac{1}{5}$. More recently, HOLL and MÜHLEISEN (1955) found $x = 1/2{\cdot}3$ and were able to give an explanation of a value about $\frac{1}{2}$ in terms of the distribution of sizes of nuclei. NOLAN (1956) also discussed the problem and gave a somewhat different explanation.

The above results were obtained from measurements in the free atmosphere; JONES, MADDEVER and SANDERS (1959b) in sealed-room experiments found $x = 1/2{\cdot}2$.

If n_0 is the number of small ions that would be present if there were no pollution ($Z = 0$), then RUHNKE (1961b) showed that n/n_0 depends on the quantity Zr^2, where r is the average radius of the nuclei concerned.

4.39. Removal of Nuclei

From industrial and other sources, nuclei are continually entering the atmosphere, yet the total nucleus content of the atmosphere is not continually increasing; therefore there must be some mechanism by which the nuclei are removed from the atmosphere. Extremely few nuclei are found at high altitudes, so the removal of nuclei cannot occur at the top of the atmosphere and must occur at the ground. The changes in charge, characterized by the combination coefficients, are not relevant to this problem.

The removal of nuclei can be investigated for nuclei stored in an enclosed vessel and involve processes of coagulation, diffusion and sedimentation; in coagulation, the total number of nuclei is reduced and, in addition, the nuclei alter in properties and become more liable to sedimentation, being now larger; diffusion is an important factor if nuclei which hit the walls of the vessel have a tendency to stick there; because the nuclei are heavier than air, they tend to move downwards, the larger nuclei doing so more readily.

The nucleus concentration in an enclosed vessel can be represented by:

$$-dZ/dt = \gamma Z^2 + \lambda Z,$$

where γ is a "coagulation coefficient" and λ is a term involving both diffusion to the walls and sedimentation (NOLAN and KUFFEL, 1955).

The experiments of POLLAK and O'CONNOR (1955a) and others have shown that γ and λ are not truly constant, but a theory due to FÜRTH (1955) has made it possible to use the results to obtain values for the diffusion coefficient and also for the sedimentation effect, and so, from theoretical results, to deduce the size, mass and density of the nuclei. JONES, MADDEVER and SANDERS (1959b) found results agreeing with the above formula, for closed-room observations.

Since nuclei stored in an enclosed vessel are not under the same conditions as in the free atmosphere, it is a matter of doubt how far, if at all, the results of one are applicable to the other.

SAGALYN and FAUCHER (1956) measured the nucleus concentration at levels above 700 ft in the atmosphere at different times during the night, when no fresh nuclei were brought into the region by convection. They showed that their results could be accounted for in terms of coagulation.

4.40. Sizes of Nuclei

Theoretical results for the value of η_0 (see § 4.31.) have been used to obtain the sizes of the nuclei concerned; a modern example is the work of KEEFE and NOLAN (1962).

RUHNKE (1961 b) measured n, n_0 and Z, and used these to find the radius, using the result quoted in § 4.38. He also found that the visibility gives a way of measuring Zr^2.

The diffusion coefficients of the nuclei depend on their radii; measurements by a "dynamic" method (NOLAN and GUERRINI, 1935) and by a "static" method (POLLAK and O'CONNOR, 1955 a) give results in reasonably good agreement.

Another method, used by O'CONNOR and FLANAGAN (1961), is to measure the mobilities of the large ions; this allows different groups to be separated and showed that stored nuclei form two distinct size groups.

The results for the sizes of nuclei range from 1 to 2 $\times$ 10^{-6} cm up to 9 to 10 $\times$ 10^{-6} cm.

4.41. The Dissipation Coefficient

In the formula

$$dn_1/dt = q - \alpha n_1 n_2 - \eta_{12} n_1 N_2 - \eta_{10} n_1 N_0$$

the term in α can often be neglected and the equation can be rewritten as

$$dn_1/dt = q - \beta n_1$$

where

$$\beta = \eta_{12} N_2 + \eta_{10} N_0,$$

and is known as the "dissipation coefficient".

SCHWEIDLER (1918, 1924) used two methods for the determination of β. In his first method he allowed an ionization chamber to remain for some time to reach equilibrium and then swept all the ions to the electrodes by means of a strong field applied for a short time. If q is the rate of production of ions in the chamber used (not the same as in the free atmosphere), then, in equilibrium, the number of small ions is q/β. SCHWEIDLER then measured q by applying a sufficient field all the time to remove the ions as they were formed.

In his second method, SCHWEIDLER used a field not sufficient to remove the ions immediately they were formed, but allowing time for combination. If the apparatus has a capacitance C and the applied potential difference is V, then there is a charge $Q = CV$ on the plate and this is lost at a rate (see § 2.27.):

$$i = dQ/dt = -n_1 e w_1 Q/\varepsilon_0.$$

For a steady state in unit volume,

$$q = \beta n_1 + i/e.$$

So

$$qe = i - \beta\varepsilon_0 i/w_1 CV,$$

giving a linear relation between i and i/V, and the slope of the graph yields β.

SCHWEIDLER originally considered the current to be carried by ions of both signs and used $(w_1 + w_2)$ in place of w_1. WHIPPLE (1933) showed that this cannot be true close to the electrodes and unless there is an "electrode effect" (see § 2.31.), giving a space charge and consequent change of n_1, SCHWEIDLER's formula is wrong. WHIPPLE considered that convection currents would keep the number of small ions uniform, avoiding the electrode effect, and then the above reasoning, using w_1 alone, holds.

O'BROLCHAIN (1931) used both of SCHWEIDLER's methods and found the value of β from the second method to be about twice that from the first; WHIPPLE's correction of SCHWEIDLER's theory removes this discrepancy.

It must be realized that these measurements refer to the air within the ionization chamber, and in particular to the nuclei there, which have been stored for some time, so that the results are not the same as would be obtained if β could be measured in the free atmosphere.

4.42. The Life of an Ion

Using the results of the last section, we see that βn_1 is the number of positive small ions destroyed in unit volume in unit time, so that β is the chance that any one small ion is destroyed. The average life of an ion is thus $1/\beta$. Since, for equilibrium, $q = \beta n_1$, β can be obtained from values of q and of n_1 for the conditions in the free atmosphere. β clearly depends upon the number and size of the charged and uncharged nuclei.

Over land areas, $q = 11$ I is an average value; if we take $n_1 = 550$ as an average value where pollution is small, $1/\beta$ is about 50 sec. For Kew, where pollution is considerable, $n_1 = 200$ and $1/\beta$ is less than 20 sec. Over the sea, n_1 is about 600 and q is less than 2 I, so that $1/\beta$ is over 300 sec. The difference in the values of β is accounted for by the difference in the number of nuclei. It is, however, merely an accident that the values of n_1 over land and sea agree fairly closely; the difference in the values of β is nearly balanced by the difference in the values of q.

4.43. Ionization up to 100 km

The electrical properties of the atmosphere up to about 20 km can be explored by means of aircraft and balloons. Down to, in some cases, 60 km, the properties can be deduced from the reflection of radio waves. The intermediate region has been explored only very sparsely by rockets.

WHIPPLE (1965) discussed the probable properties of the region between 20 km and 60 km, choosing the upper limit as that below which there is little or no ionization by radiation from the sun. He emphasized the need for developing reliable techniques for measurements in this region.

COLE and PIERCE (1965) were the first to attempt to deal with the whole of the atmosphere up to 100 km on a unified basis. They pointed out that there must be an important contribution to the balance in the atmosphere at heights above about 40 km from free electrons, and so the equilibrium equations will be very different from those of § 4.28. At these levels, condensation nuclei and large ions need not be considered.

Adapting to the notation of § 4.28, COLE and PIERCE's equations are:

$$dE/dt = Q - AE - BEn_1 + Fn_2,$$
$$dn_1/dt = Q - BEn_1 - \alpha n_1 n_2,$$
$$dn_2/dt = AE - Fn_2 - \alpha n_1 n_2,$$

where E is the number of electrons per unit volume, Q is the rate of production of positive ions and electrons, A is the attachment coefficient of electrons to form ions, B is the recombination coefficient between electrons and positive ions, α is the usual recombination coefficient between positive and negative ions and F is the detachment coefficient of electrons from negative ions. All the quantities Q, A, B, α, E and F are functions of height and Q and F have different values according to the presence or absence of solar radiation.

COLE and PIERCE (1965) discussed values to be assigned to each of these quantities at different heights and from these they were able to obtain curves giving electron and ion densities at different heights by day and night, assuming equilibrium conditions.

It must remain for suitable measurements to confirm, or otherwise, the theoretical predictions of COLE and PIERCE. WHIPPLE (1965) quoted results showing a decrease in ion density above 40 km and he suggested "dust" particles as the cause.

COLE and PIERCE also discussed the application of their results to radio-wave absorption and to the propagation of low-frequency waves. The effect upon the columnar resistance of the atmosphere is small, since most of this lies in the lower region where properties are well-known.

4.44. Radioactivity of the Atmosphere

The measurement of radioactive gases (radon, thoron and actinon) in the atmosphere has been discussed in detail by ISRAËL (1951), and the reader is referred to this work for a complete account. Measurements of radon are best made by direct "emanometry" of the gas absorbed or condensed. Thoron and actinon have shorter lives and so can best be measured by the "induction" method, in which the active deposit is obtained on a negatively charged wire.

The results are given in terms of the Curie(c), the unit of radon, being the amount of radon in equilibrium with 1 gm of radium. The total rate of ionization produced by 1 c, if the charges could all be separated, would give a current of $9{\cdot}22 \times 10^{-4}$ A. The results show that the average amount of radon in the atmosphere is about 10^{-16} c/cm^3 over continents and about 10^{-18} c/cm^3 over oceans; in certain places, e.g. Innsbruck, the values reach 4×10^{-16} c/cm^3. Results for thoron are less accurate and suggest rather less thoron than radon, and the amount of actinon is only a few per cent of that of radon. Radon is exhaled from the ground at a rate of about 4×10^{-17} c/cm^2 sec and this fits well with the measured amounts of radon at different heights in the atmosphere.

Variations in the amounts of radon with meteorological conditions have, in general, been satisfactorily explained, and it has been suggested that radioactive content might be used as a method of classifying and identifying air masses. Precipitation, particularly in thunderstorms, shows radioactivity from the air.

CHAPTER 5

The Vertical Potential Gradient

5.1. The Potential Gradient at the Earth's Surface

Since the surface of the earth is a conductor, it follows that the lines of force must reach the surface in a normal direction. Thus, where the earth's surface is horizontal, the lines of force are vertical at the surface and so equipotential surfaces close to the ground must be horizontal and the potential gradient vertical. Where the earth's surface is not horizontal there will be a corresponding modification of these conclusions.

By the term "the earth's vertical potential gradient" is meant either the actual value of dV/dx at the surface, or else the potential at a height of 1 m, depending on which type of measurement is used. Fortunately, it is very rarely that there is any material difference between results concerned with the two meanings.

As discussed earlier (§ 2.6.), the potential gradient is referred to by many workers as the "field", but this brings in an ambiguity of sign, according to which convention is used, that of traditional atmospheric electricity or that of theoretical electrostatics. By using the term "potential gradient", there is no ambiguity, since this can mean nothing but the rate of change of potential with height.

The potential gradient at the earth's surface has proved to be the simplest of the elements of atmospheric electricity for measurement and, by reason of the large number of records, some over long periods, it is probable that the importance and significance of the potential gradient have been emphasized too much.

5.2. Types of Measurement

Various methods of measurement of the potential gradient have been used and nearly all can be clearly classified into two types.

The first type consists of methods of measuring the potential difference between two points at different heights, one of these usually being the earth's surface. The potential is found, in most cases, by some form of potential equalizer; most potential equalizers disturb the natural vertical potential gradient and a correction has to be made (see § 5.18.).

In the second type of method the potential gradient is obtained by measuring the "bound" charge on a portion of the earth's surface or on an earth-connected body; this can be pictured as at the lower ends of the lines of force. For a flat portion of the earth's surface, the surface density of charge, σ, is given by $\sigma = -\varepsilon_0 F$, where F is the potential gradient. For any other earth-connected body this corresponds to a certain area of the flat earth, so that there is a "reduction factor" (see § 5.18.).

5.3. Potential Equalizers

In order to find the potential of a point in the air it is necessary to have some device by which a conductor can be brought to the same potential as the air in its neighbourhood. It is then possible to measure the potential difference between the conductor and earth with some form of electrostatic or electronic electrometer or voltmeter, provided that the measuring device has a small capacitance and does not consume an appreciable current. By reason of the conductivity of the air, an isolated conductor gradually acquires the potential of its surroundings, and this was the method used in early measurements of this type by LEMONNIER (1752), BECCARIA (1775) and others; but the process is very slow, taking place at the rate of the "relaxation time" (see § 2.27.), which is 5–20 min at the earth's surface; so, for any recording of changes of potential gradient over periods of less than hours, it is necessary to accelerate the process by some form of "collector" or "equalizer".

If a conductor is at a different potential from its surroundings it can be brought to the same potential only if there is some means by which charge of one sign is carried to the conductor, or charge of the other sign carried away. This can be achieved by actual particles leaving the conductor, as in the water-dropper, or by ionizing the air close to the conductor, so that ions carry the charges, as in the radioactive equalizer. The flame or fuse appears to make use of both of these effects. When the double field mill (§ 5.32.) acts as an equalizer the necessary charge is supplied from an external source.

It is convienent to call the agent a "potential equalizer" or simply "equalizer"; the term "collector" in also used, but since this is, more appropriately, also used for a receiver of charge by conduction or precipitation, it is better avoided here.

5.4. Effective Resistance of Potential Equalizer

If the conductor has a potential V while that of its surroundings is V_0, then the rate of change of V depends upon $V_0 - V$. If the conductor has a capacitance, C, then the current, i, to it is given by

$$i = dQ/dt = CdV/dt.$$

When the ratio of dV/dt to $V_0 - V$ is constant, say k, then

$$i = Ck(V_0 - V),$$

so that OHM's law is obeyed and we can put $1/Ck = R$ and call R the effective resistance of a potential equalizer.

If the ratio of dV/dt to $V_0 - V$ depends on the value of V, then OHM's law is not obeyed and it is not possible to speak correctly of an effective resistance.

When the conductor is connected to carth directly or through a resistance small compared with R, then $V = 0$, so that $i = CkV_0 = V_0/R$. Thus the current to a potential equalizer depends not only on V_0 but also on R.

The speed of response of a potential equalizer to a change of ambient potential depends on the effective resistance and the capacitance.

If
$$dV/dt = k(V_0 - V)$$

this integrates to

$$V - V_0 = (V_1 - V_0)\exp(-kt),$$

where V_1 is the initial potential of the conductor.

Thus the potential difference $V_1 - V_0$ is reduced exponentially to zero with a time constant $1/k$, which is equal to CR, where C is the capacitance of the system and R the effective resistance of the equalizer.

It is therefore seen that the lower the effective resistance can be made, the more rapidly does the system respond to changes in potential gradient. A decrease in the capacitance of the system has a similar effect.

DOLEZALEK (1960c) pointed out that a radioactive equalizer cannot be expressed simply in terms of an effective resistance; the true equivalent circuit must be a "4-pole" and the whole description of any kind of potential equalizer is very complicated.

5.5. Effects of Leaks

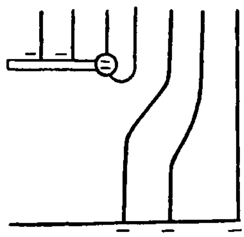
If there is a leak to earth from a potential equalizer, it is clear that the potential measured will not be the potential of the atmosphere near the equalizer; instead, the potential acquired will be such that the leakage current is equal to the current to the equalizer. If R is the effective resistance of the equalizer and S is the resistance of the leak, then the potential acquired by the equalizer is $S/(R + S)$ of the true value; if S is large compared with R, the error is small.

5.6. The Water-dropper

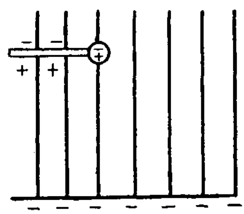
The water-dropper as a potential equalizer is due to KELVIN (1895b) and was used by him in the first attempt to obtain accurate continuous records of the potential gradient over periods of time.

The operation of the water-dropper can be described in two different ways, and as each gives some insight into the principles at work, both will be given here.

Let us consider a dropper which is initially at the same potential as the earth, with the drops falling off at a place exposed to the normal atmospheric fine-weather field. There are then lines of force with their negative ends on the exposed parts of the dropper and on the drop (see Fig. 15a). The corresponding positive charge has moved away to earth. If the dropper is now insulated from the earth and drops fall off, the first drop will carry off a negative charge; when the next drop forms, it too must have a negative charge and so the remainder of the dropper becomes positively charged. As more drops carry away negative charges the dropper becomes more charged positively; the effect of this positive charge is to hinder the charging of the drops negatively and in the end the drops will fall uncharged. When this is so, the dropper is at the same potential as the surrounding air and the drop does not affect the lines of force, which pass through undisturbed, with a positive charge on the lower half of the drop as well as a negative charge on the upper half (see Fig. 15b).



(a)



(b)

FIG. 15. The water dropper.

Expressed mathematically, the drop must be at a potential composed of V_0 which is that of the surrounding air and Q/D, which is due to its charge Q and capacitance D. If there is no charge flowing between the dropper and the drop, their potentials must be equal and so the potential of the dropper $V_d = V_0 + Q/D$. When Q becomes zero, $V_d = V_0$.

The water-dropper can be described in another way. If the potential of the dropper differs from that of the surrounding air, then this difference of potential implies a potential gradient, with lines of force having one end on the drop. If the potential of the dropper and drop is less than that of the surrounding air, there must be lines of force with their negative ends on the drop and their positive ends on the electrosphere or on space charges in the air. When the drop falls off, it carries away the negative charge and so when the next drop is formed the potential of the dropper and drop is increased, bringing it closer to that of the surrounding air. The new drop will be at a higher potential than the previous one but still below that of the air, so it will again remove negative charge, the process continuing until the drop has reached the potential of the air surrounding it; then there will no longer be lines of force leaving the drop and so no charge on it.

5.7. Effective Resistance of Water-dropper

The effective resistance of the water-dropper depends upon the size of the drops and the rate at which they leave the dropper.

If V is the potential of the drop, this is composed of two parts, namely, V_0, the potential at that place in the absence of the drop, and Q/D, where Q is the charge on the drop and D its capacitance. If the capacitance of the whole dropper system is C, then the removal, by the drop, of a charge Q, alters the potential by Q/C:

$$V = V_0 + Q/D \quad \text{or} \quad Q = D(V - V_0).$$

If n drops fall per second,

$$dV/dt = nQ/C = nD(V - V_0)/C.$$

Then, by reference to § 5.4., we see that the effective resistance is $1/nD$. The capacitance D of the drop is approximately $4\pi\varepsilon_0 r$, where r is the radius of the drop. For the same quantity of water, n is proportional to $1/r^3$, so that, in order to reduce $1/nD$ and obtain as rapid a response as possible, it is desirable to make the drops as small as possible.

It has been found possible to construct a water-dropper which reaches the potential of the air in its vicinity in a time of about 30 sec. Thus the water-dropper is not suitable for the measurement of changes of potential gradient over times of less than minutes.

5.8. The Use of the Water-dropper

The water-dropper as used in practice consists of a reservoir of water kept insulated inside a building, with a horizontal pipe projecting outside the building; the flow of water is regulated so that the water breaks into a fine stream of drops just at the end of the tube. The receiver is connected to some form of electrometer and measurements are made of the potential; a suitable method to use is to connect the reservoir to the needle of a quadrant electrometer, keeping a fixed difference of potential between the quadrants. It would also be possible to use some form of electronic method of potential measurement.

As will be discussed later (§ 5.18.), the presence of the building must affect the potential gradient and so the arrangement cannot be used to find absolute values.

With a sufficient quantity of water, it is possible to use the water-dropper for continuous measurements, using a recording device as described in § 2.51.

5.9. Electronic Method with Water-dropper

LECOLAZET (1945) described the theory of, and also (1946) the use of a water-dropper with a three-stage direct-current amplifier, using the output to provide a potential to screens, etc., in the same way as GISH and SHERMAN (1929) (see § 5.12.) provided a potential manually from a potentiometer, to reduce leaks and reduce the input capacity, as well as giving a final output suitable for a pen-recorder. This was a fore-runner of the electronic method with the radioactive equalizer (see § 5.13.) of which the circuit is given in Fig. 17.

5.10. The Radioactive Equalizer

The radioactive equalizer, first suggested by WITKOWSKI (1902), consists of a disc or wire on which is spread a small quantity of some radioactive substance emitting α-rays; there is usually a very thin protective covering over the source. The α-rays produce ionization in a small volume of air round the equalizer and if there is a difference of potential between the equalizer and its surroundings the ions move in the field of force until the equalizer, which is insulated, has received sufficient charge to bring its potential to that of its surroundings.

It is clear that the current to the equalizer is, approximately at least, proportional to the potential difference between the equalizer and the surroundings unless this is very large; thus the equalizer can be said to have an "effective resistance" as defined in § 5.4. This is not calculable simply, as in the case of the water-dropper, but can be obtained experimentally from the rate of change of potential of the equalizer, if the capacitance of the equalizer system is known. SCRASE (1934) found a value of about $2 \times 10^{10}\,\Omega$ for a new polonium equalizer.

The main disadvantage of a radioactive equalizer is that it produces ions, some of which move away from the equalizer and produce a disturbance of the electrical conditions in the air, so that the potential measured by the equalizer does not give a correct indication of the potential gradient, as it would be in the absence of the equalizer; the extent of this disturbance is probably affected by the wind speed and so differs from one time to another. The disturbance can be reduced by reducing the strength of the radioactive source, but this slows down the response of the equalizer.

A radioactive equalizer usually has to be connected to some form of electrometer inside a building; a long length of cable for this would increase the capacitance and thus slow down the response of the equalizer to changes in potential gradient. Therefore, usually a radioactive equalizer has to be placed close to a building and there is a "reduction factor" (see § 5.18.).

5.11. Radioactive Material for Equalizer

The most suitable substance for use as a radioactive equalizer is one which emits only α-rays; β-and γ-rays produce ionization at greater distances from the source, giving more disturbance without a corresponding increase in the speed of action. Not only should the choice of a material for an equalizer be limited to α-ray emitters, but it is also preferable to exclude those α-ray emitters whose decay products are β- and γ-ray emitters. On these grounds, one of the most suitable choices is polonium, which also has the advantage that it can be deposited electrolytically on the disc or wire. There is, however, the grave disadvantage that the half-value period of polonium is only 136 days, so that the equalizer has to be replaced frequently and has a strength (and so a speed of response) which alters during its period of use. Radio-thorium of -halflife 1·9 years and ionium (8×10^4 years) have also been used; while not having the advantage of polonium that the decay product is stable, these both have α-ray emitters as their immediate descendants, but both are more expensive and less easy to handle than polonium. Radium is sometimes used, because of its comparative abundance, but it is a singularly poor choice; not only does it produce β-and γ-ray emitters quite quickly, but also great care must be taken to avoid the escape of the emanation which is the first disintegration product; if this were to escape, it would give rise to ionization at a considerable distance and also produce an ionizing "active deposit" spread round the neighbourhood.

5.12. Leak-free Method with Radioactive Equalizer

To avoid the difficulty of a leak which is likely to be variable, Gish and Sherman (1929) used a leak-free method in which they surrounded the insulator of their equalizer, not with an earthed metal sheath, but with a sheath connected to a potentiometer, so that it could be maintained manually at the same potential as the equalizer. By means of an electrometer, the necessary potential could be found (see Fig. 16). Since the sheath round the insulator

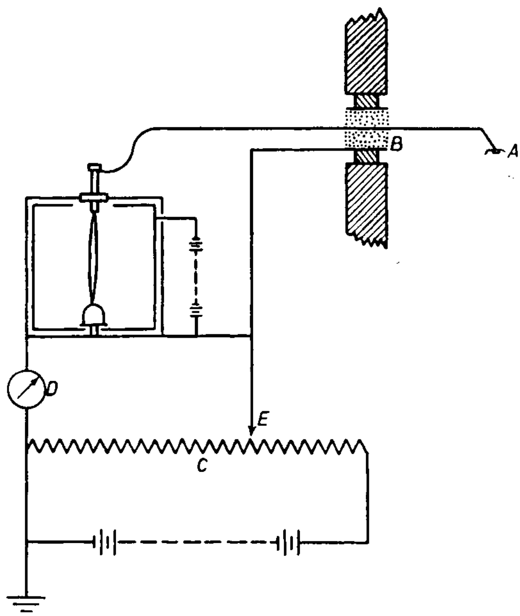


FIG. 16. Leak-free method for measurement of potential gradient. (From GISH and SHERMAN, 1929, Fig. 1, p. 232.)

connections is maintained by a battery, it can be kept to the required potential in spite of a possible leak to earth, as the battery has a very low "effective resistance". Since the equalizer system is at the same potential as its surroundings, there can be no leak from it.

The method has the further advantage of being a "null" method, so that a sensitive electrometer can be used; the disadvantage is that the apparatus requires continuous adjustment and so cannot be self-recording over long periods of time. In principle some form of servo-mechanism could be used to replace the manual control, but this does not yet appear to have been attempted, though BREWER's (1953) method (see § 5.13.) used negative feedback in place of a mechanical servo-system or manual adjustment.

5.13. Electronic Method with Radioactive Equalizer

Instead of using an electrometer for measuring and recording the potential acquired by an equalizer, MÜHLEISEN (1951) and BREWER (1953) have used electronic methods and were able to display the output on robust meters. They used electrometer valves with cathode-follower circuits. In BREWER's (1953) arrangement (see Fig. 17), there is negative feedback, which gives a voltage amplification of close to unity and a linear voltage range of ± 500V; the feedback is also connected to guard rings, giving a "leak-free"

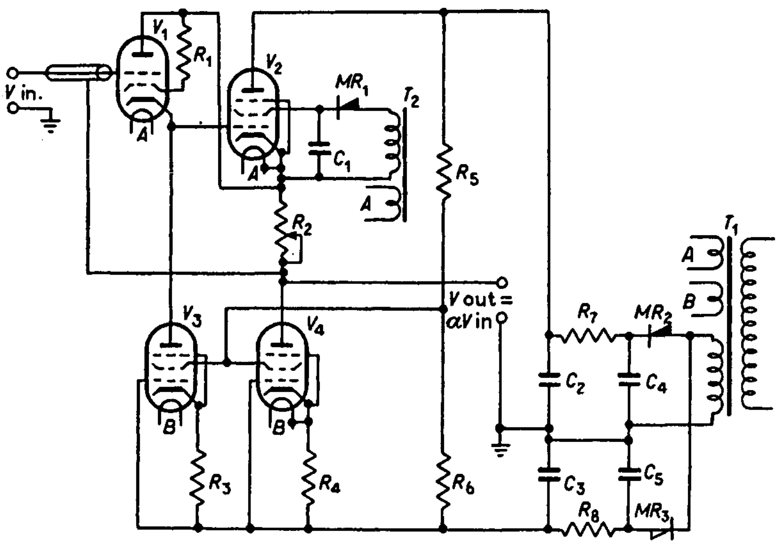


FIG. 17. Circuit for electronic method with radioactive equalizer. (From BREWER, 1953, Fig. 2, p. 90.)

$R_1 = 10$ kΩ
$R_2 = 5$ kΩ variable–to adjust zero
$R_3 = 33$ kΩ adjusted on test to give maximum ranges
$R_4 = 3{\cdot}3$ kΩ
$R_5 = 1$ MΩ, 2 W
$R_6 = 470$ kΩ
R_7, $R_8 = 10$ kΩ
$C_1 = 2$ μf, 250 V

C_2 to $C_5 = 2$ μf, 700 V
V_1 = Ferranti B.M4.a
V_2, V_3, V_4 = Cossar MS Pen T
T_1 = mains transformer outputs 4 V 2 A, 4 V 2 A and 500 V, 12 mA
T_2 = "bell" transformer (or use 200 V winding on T_1)
MR_1 = Standard Telephones K3/8
MR_2, MR_3 = Standard Telephones K 2/25

arrangement similar to that of GISH and SHERMAN (1929) (see § 5.12.); this has the further advantages that the effective input capacitance is considerably reduced, increasing the rapidity of response, and also that very little current is taken from the equalizer. If a radioactive equalizer is to be used for the measurement of potential gradient, it would be difficult to improve on this type of arrangement.

5.14. Current from Radioactive Equalizer

Instead of measuring the potential attained by the equalizer when insulated, some workers have connected the equalizer to earth through a galvanometer and measured the current. Using the result of § 5.4., this current is V_0/R, where R is the effective resistance of the equalizer and V_0 is the potential of the neighbourhood. So long as R remains constant, the current will be proportional to V_0 and so gives a measure of relative values of potential gradient, though a calibration is needed to give absolute values. But, since R depends on the motion and position of the ions produced by the radioactive substance, it is clear that there are many factors which can alter R; for example, wind may blow away ions which would otherwise reach the equalizer, so altering the current; a film of moisture over the equalizer may reduce the range of the α-rays and hence reduce the number of ions produced; and so on. In the normal use of the equalizer, where its potential is measured when insulated, any alteration of R merely alters the rapidity of response, as discussed in § 5.7., but this does not alter the reading obtained unless there is an appreciable leak.

SCHAEFER (1955) claimed the utmost simplicity in the use of a radioactive equalizer by measuring the current through an exposed point covered with radioactive material and connected to earth. BENT (1955) used this method, with an equalizer of radium sulphate, at Mount Washington; for the reasons described above, particularly in the very exposed situation, it is probable that the current measured depends on other factors besides the potential gradient; as discussed in § 5.10., the use of radium has disadvantages. In spite of its simplicity, the method of measuring the current from a radioactive point cannot be relied on to give anything more than a rough indication of the value of the potential gradient and is not to be recommended. This method is sometimes called the "radioactive corona point current" method, but at low values of potential gradient there is in fact no corona current (point-discharge current); for higher potential gradients, the same point does give corona current and under these circumstances the radioactive substance has no effect. Perhaps a more suitable description would be "radioactive *and* corona point current".

KASEMIR (1951a) used a method which is intermediate between those of this section and the previous one. He connected a radioactive equalizer to the grid of an electrometer valve with a grid leak to earth; if the resistance of the grid leak is small compared with R, the effective resistance of the equalizer, then this method amounts to measuring the current and suffers under the same disadvantages as that of SCHAEFER; on the other hand, if the grid leak is large compared with R, it is the potential V_0 which is measured and the method is similar to that of BREWER. Actually, KASEMIR used different values of grid leak for different ranges of sensitivity and did not publish the value of R. He was interested, with this apparatus, in long-period averages of potential gradient and therefore slowed down the variations by a large condenser between the grid and earth; the apparatus was arranged to be portable.

If the current through a radioactive equalizer could be calibrated in terms of the potential gradient and wind speed, and if it could be verified that other factors are not important, then such an arrangement would be available to measure the potential gradient in conditions where high insulation could not be obtained or where facilities for elaborate apparatus are not available. An attempt to carry this out by GROOM and CHALMERS (1964) did not give consistent results, probably because of films of inconstant thickness on the surface of the equalizer.

5.15. Limitations of Radioactive Equalizer

MÜHLEISEN (1951) measured the potentials acquired by two radioactive equalizers separated by a distance of some metres, and found that under certain circumstances the effect of the wind may be considerable; in low winds the potentials, which should be equal, may differ by as much as 30 per cent when strong sources are used. The ions produced by the sources are able to cause large effects on the potential gradients. MÜHLEISEN suggested precautions to minimize the effects, but his results clearly show that the radioactive equalizer, even under good conditions, is not very suitable for the measurement of potential gradient. Also, as has been previously mentioned, there is always a time lag between the occurrence of a change in potential gradient and its complete effect on the equalizer. If other measuring apparatus is set up near the equalizer, the ions from the latter may have a disturbing effect which will vary with the wind direction. WICHMANN (1952a) reached similar conclusions and also pointed out that the effective resistance of a radioactive equalizer is not a constant, but tends to increase with the potential gradient. DOLEZALEK (1956) discussed, in considerable detail, the best conditions for obtaining satisfactory results with a potential equalizer.

Another source of difficulty with a radioactive equalizer is the effect of rain, which can bring its own charge to the equalizer and produces other effects by splashing and dripping. Raindrops falling off the equalizer itself would act in the same way as a water-dropper, but dripping from elsewhere would give spurious effects.

5.16. The Flame and Fuse

The use of a flame as a potential equalizer originated with VOLTA (1782); it probably acts in the same way as the water-dropper in providing particles which leave the equalizer and carry charge with them, and also similarly to the radioactive equalizer in producing ionization. It has been found that a glowing fuse is a much more satisfactory equalizer than an actual flame, for it lasts for a longer time, is less easily extinguished and also produces less mechanical and electrical disturbance of the air. A suitable fuse consists of paper or string impregnated with a strong solution of lead nitrate.

For certain purposes a fuse has the advantage over a radioactive equalizer that it can be readily obtained and attached where it is required. But it does not last long enough for continuous recording and the point at which it is effective alters as it burns.

5.17. The Passive Antenna

In the very early days of atmospheric electricity, an exposed probe reached the potential of its surroundings by the normal conduction processes in the atmosphere, but this proved much too slow for following rapid changes, having a time constant corresponding to the conductivity of the atmosphere, namely between 5 and 40 min according to the locality. Thus there arose the need for potential equalizers to speed up the process.

But, with modern techniques, this is no longer necessary and CROZIER (1963b) has devised a "passive antenna" system which uses no radioactive or other equalizer but which brings a probe very quickly to the potential of its surroundings. The principle is the use of feedback, as in the method of BREWER (1953) (see § 5.13. and Fig. 17), to reduce the effective input capacitance to a very low value; with a head amplifier of extremely low grid current, and very good insulation, the small current available from natural ionization is sufficient to bring the probe to the potential of its surroundings very quickly. The method is available for use only in dry weather, as rain would charge the antenna and this would take a long while to be removed; DOLEZALEK (1963) pointed out that convection currents could have the same effect.

This method avoids any disturbance of the conditions, as is inevitable with equalizers.

5.18. The Reduction Factor

As has already been mentioned, the close presence of a building seriously disturbs the lines of force and makes it impossible to obtain absolute values of the potential gradient from measurements made near it; thus continuous records with the water-dropper or the radioactive equalizer do not give absolute values. BENNDORF (1900, 1906) made some calculations to give the effect of neighbouring conductors in distorting the lines of force, but these are possible only when the conductors are very simple in shape.

In order to use records made with equalizers, it is necessary to find a "reduction factor" or "exposure factor" by which the reading with the recording equalizer can be multiplied to give the absolute value of the potential gradient, i.e. the value that would occur over perfectly level ground.

BENNDORF (1906) showed theoretically that if a wire 6 m long is stretched between two posts 1 m high, then the posts, even if conductors, alter the potential gradient at the mid-point by less than one per cent. If a fuse or other equalizer is attached to the mid-point, accurate values of the potential gradient can be obtained, and if there are no objects nearby to distort the lines of force, it is possible to obtain a reduction factor for a recording equalizer by simultaneous measurements. An electrostatic voltmeter or an electrometer can be used to measure the potential of the stretched wire.

To obtain the reduction factor for a particular position of an equalizer it is necessary to take simultaneous readings for the equalizer in this position and for the stretched wire system; it is not satisfactory to take readings at different times, as the potential gradient is likely to have altered. The greater the distance between the two equalizers, the less likely is it that the variations of potential gradient at the two places will correspond exactly and so the greater will be the fluctuations of the individual values of the reduction factor. If it is desired to compare two positions of the stretched wire system or to find the effect of nearby objects on it, this can be done by comparison of each in turn with simultaneous readings of a fixed equalizer, preferably one which gives continuous recordings.

WATSON (1929) made a series of measurements involving reduction factors at Kew, and found that the reduction factor was practically constant and depends little on the value of the potential gradient. SCRASE (1934) has further discussed the reduction factors applicable at Kew for different observations.

ARNOLD, PIERCE and WHITSON (1965) considered, theoretically and experimentally, the reduction factor produced by a nearby tree.

The reduction factor represents the ratio of the potential gradient over perfectly level ground to that at the place of measurement; this can be considered as the ratio of the number of lines of force ending on unit areas at the two places. If the other ends of all the lines of force are at great distances, then it is clear that the reduction factor should be the same for different values of the potential gradient, since it is defined by purely geometrical considerations.

But if there are space charges, on which lines of force end, existing at low levels, then it can easily be seen that the reduction factor is no longer a constant. To take an extreme example, suppose that there is a condition with no space charges in the lower atmosphere and a potential gradient measured as F over level ground; at another, shielded, place of measurement, suppose the measured potential gradient is $0{\cdot}4\,F$, corresponding to a reduction factor of 2·5. Now suppose a uniform low-level space charge is introduced and the potential gradient over level ground becomes $0{\cdot}7\,F$; this means that lines of force corresponding to $0{\cdot}3\,F$ are intercepted by the space charge, and so, if there is the same effect of the space charge at the shielded site, the measured potential gradient would be $0{\cdot}1\,F$ and the reduction factor would appear to be 7 instead of 2·5. This rather extreme example shows the effect that space charges might have on the reduction factor and emphasizes the desirability of making the measurements over level ground and avoiding a reduction factor.

5.19. Measurements by Point Discharge

The earliest workers in atmospheric electricity used points to collect charges, relying on point-discharge currents to provide some equalization of potential. More recently, point-discharge currents through an earth-connected point have been used to give a measure of the potential gradient. The whole subject of point discharge will be dealt with in Chapter 9; from the point of view of its action as a potential equalizer, a point at which there is point discharge produces ions and so acts somewhat similarly to a radioactive collector. But, if the point is insulated, then, as its potential approaches that of the air in its neighbourhood, the production of ions decreases and for a certain potential difference of around 7×10^3 V ceases altogether; thus the point never reaches the potential of its surroundings and so cannot be as satisfactory as other types of equalizer; in addition, the ions liberated in the point-discharge process will themselves produce effects (see § 5.68.).

For a point connected to the earth, the current through the point can be measured; if this can be related to the potential gradient at the ground, the latter can be deduced. Earlier workers used the empirical formula of WHIPPLE and SCRASE (1936): $I = a(F^2 - M^2)$, where I is the point-discharge current, F the potential gradient close to the ground and a and M are constants. More recent work has shown that this result is not correct and that the wind speed has also to be considered (see § 9.16.).

The alti-electrograph (§ 9.20.) has made use of point-discharge currents to give a measure of the potential gradients above, in and below clouds. There is a doubt as to the calibration procedure (see § 9.21.).

5.20. Measurement of the Bound Charge

The surface density of the "bound" charge on the earth's surface is proportional to the potential gradient at the surface, as $F = -\sigma/\varepsilon_0$. The direct measurement of the bound charge is not well suited for giving continuous records of the potential gradient, since it is possible to measure the bound charge only by releasing it, and this cannot easily be arranged continuously, but only at intervals, most simply by placing an earthed cover over the portion of the surface under investigation.

This type of method measures the number of lines of force which end on the surface above which the earthed plate is placed, and gives the same value for the potential gradient as is obtained by measurements of the potential difference between points at different heights, provided that there is no appreciable space charge within the region considered.

In such measurements it is necessary to isolate a portion of the earth's surface, or to replace it with a test plate, and connect it, suitably insulated, with some form of measuring instrument. In order that the apparatus shall not disturb the field the surface must be maintained at earth potential.

Machines using the bound chargeare discussed later on (§ 5.24).

5.21. The Universal Portable Electrometer

WILSON (1906) used an instrument called by him the Universal Portable Electrometer for the measurement of the bound charge and also of the air–earth current. This is shown diagrammatically in Fig. 18.

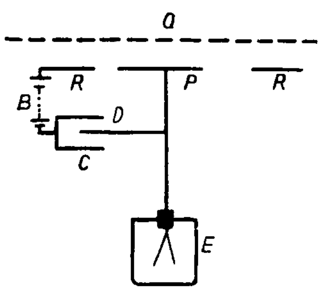


FIG. 18. The universal portable electrometer.

P is a flat circular plate surrounded by a guard ring, *R*, the whole being, if possible, in the plane of the earth's surface, so that the rest of the instrument should be housed underground. *E* is some form of electrometer which is observed from outside the case (not shown) which contains the whole instrument. *C* is a cylindrical condenser known as the compensator, and can be moved parallel to its length, so that it is possible to vary the length of the rod *D* which is inside *C*.

In making a measurement, the earthed cover, Q, is placed over the plate P, the electrometer is earthed and the zero position of the electrometer is observed. Then the cover is removed and, in a normal positive potential gradient, a negative charge is induced on the plate P; the corresponding positive charge passes to the electrometer, which therefore shows a deflection. If the condenser C is now moved inwards, a positive charge is induced on D (since C is charged negetivaly by the battery B) and it is possible to adjust the position of C so that there remains no charge on the electrometer, the positive charge on D just balancing the negative charge on P. If the condenser C has been calibrated, the amount of motion of C measures the bound charge on P and hence the potential gradient.

As changes in the potential gradient occur, the compensator can be moved so as to keep the electrometer always in the zero position. Then, after an interval, the cover Q can be replaced and the compensator again adjusted to give zero deflection of the electrometer; the difference between the initial and final positions of the compensator gives a measure of the charge which has arrived at the plate P during the time it was exposed, and so of the air–earth current.

Very often it is not possible to house the apparatus below ground level and it is mounted on a tripod, with consequent distortion of the field. The amount of this distortion has been discussed by WATSON (1929) who found the reduction factor. WATSON also made careful comparisons between the values of the potential gradient obtained by the Universal Portable Electrometer in a pit and by a fuse on a stretched wire, and came to the conclusion that they agreed within the limits of observational error. WATSON used a slightly different arrangement from WILSON, making use of the deflections of the electrometer, rather than the motion of the compensator, and thus avoiding the assumption that the voltage of the battery B remained constant, and avoiding the need to calibrate the compensator. On the other hand, the plate is not kept at earth potential, and so it will distort the field somewhat. NOLAN and NOLAN (1937) provided the compensation, not by moving C, but by altering the potential of B.

It can be seen that it would not be easily possible to adapt the original instrument for continuous readings or recording of the potential gradient, since the compensator has to be moved, and the potential gradient can be measured only by removing or replacing the cover; some form of servo-mechanism might be used, but this does not yet appear to have been attempted.

5.22. The Capillary Electrometer

WILSON (1916) adapted the Universal Portable Electrometer for continuous observations by replacing the compensator and electrometer with a capillary electrometer. This consists of a small bubble of sulphuric acid between threads of mercury in a horizontal capillary tube (see Fig. 19). If one column of mercury is connected to

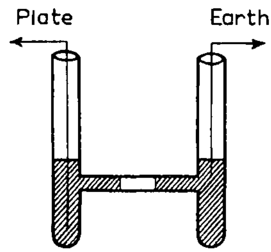


FIG. 19. The capillary electrometer.

the plate and the other to earth, then any charge moving from the plate to earth passes through the electrometer and causes the bubble to move. The mechanism of the action of the capillary electrometer is probably as follows: there is a thin layer of acid surrounding the mercury and this carries a definite charge per unit area on the double layer of the mercury–acid interface; any additional charge arriving at the interface can be taken up by the interface only if this increases in size, and this can be achieved by a motion of the bubble. The increased charge on the double layer at one side of the acid bubble is accompanied by a decrease of the charge in the double layer at the other side, and hence results in the passage of a charge to earth through the electrometer.

The most important feature of the capillary electrometer is that it gives an automatic adjustment of the potential of the plate to that of the earth, while at the same time giving a record of the charge passing.

5.23. Electronic Method for Bound Charge

In place of an electrometer for measuring the charge passing when the plate is covered, CHALMERS (1952a) used a simple electronic circuit. The plate is connected directly to the grid of a triode, and the charge of potential gives an impulse which passes through a condenser connected between the anode and the cathode. The galvanometer used to register the impulse can be a quite insensitive, robust instrument. The area of plate needed for such measurements is quite small and the whole has been arranged as transportable equipment.

The circuit is shown in Fig. 20. Simple theory shows that if a charge Q arrives on the grid, then a charge $\dfrac{C\mu R}{K(R + R_a)}\,Q$ passes through

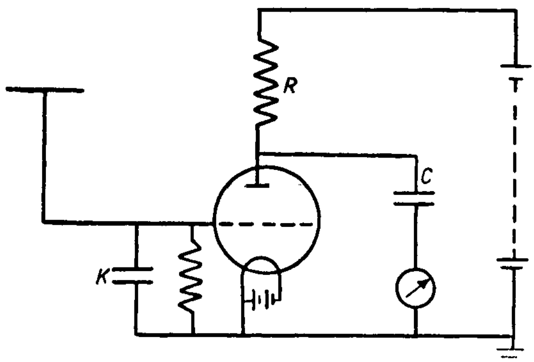


FIG. 20. Circuit for electronic method for bound charge.

the galvanometer where μ is the amplification factor and R_a the anode conductance resistance of the triode, provided that the grid resistance is sufficiently high for no leakage of charge to occur while the charge Q is arriving; as Q leaks later, an equal and opposite charge passes through in the opposite direction, but if Q leaks slowly enough, this will not affect the initial throw of the galvanometer. The grid resistance itself does not enter into the theory, but it, with K, determines the rate of leak of Q and so must remain high. If the plate has an area A,

$$Q = -\varepsilon_0 F A,$$

where F is the potential gradient, and so the throw of the galvanometer when the plate is covered or uncovered measures the potential gradient.

5.24. Field Machines

The term "field machine" can be used as a general description of the various types of apparatus used to obtain a continuous measurement of the potential gradient at the ground from the bound charge. In order to obtain a continuous signal using the bound charge it is necessary to have some kind of periodic shielding and unshielding of a conductor, and field machines can be divided into 4 groups according to the way in which this is done. Nearly every worker in the subject has devised his own type of field machine and has given it a different name. In what follows, in order to distinguish the 4 groups, we shall use some of the names that have been used before, but we shall not always be describing the machine by the same name as used by the inventor.

Two of the groups of field machines give outputs of alternating current, and these two groups are distinguished according to whether it is the cover or conductor which moves. In the other two groups the output is intermittent direct current and again the distinction is according to whether the cover or conductor moves.

The first group of field machines we shall call the "field mills" and in these an earthed cover alternately exposes and shields a conductor, which is connected to earth through the measuring apparatus.

In the second group, which we shall call the "electrostatic fluxmeters", the cover is fixed and the conductor moves relative to it, being alternately exposed and shielded; in some cases, a number of conductor are used, following one another past the position of exposure.

In both "field mills" and "electrostatic fluxmeters" the conductor is connected to earth, but in the other two groups of machine, the conductor is connected to earth when exposed and to the measuring apparatus when shielded.

In the instruments we shall describe as "induction voltmeters", the cover moves and the conductor remains stationary, while in the "agrimeters" a conductor, or a set of conductors, moves and the cover is fixed.

In addition to their use in atmospheric electricity, field machines have been used for the measurement of high-tension voltages, having the advantage that they take no current from the high-tension apparatus and that they do not provide extra opportunities for spark breakdown.

5.25. The Field Mill

In the field mill, as we here use the term, there is a fixed test plate which is alternately exposed to and shielded from the lines of force of the earth's field by a moving earthed cover. The plate is connected to earth through a resistance and capacitance in parallel and the alternating current to earth is amplified, usually with a circuit tuned to the frequency of the shielding, rectified and measured. In many cases, it is necessary to provide some method for the determination of the sign of the potential gradient (see § 5.27.).

The earliest machine of this type was probably that of HARNWELL and VAN VOORHIS (1933), who considered it as an adaptation of

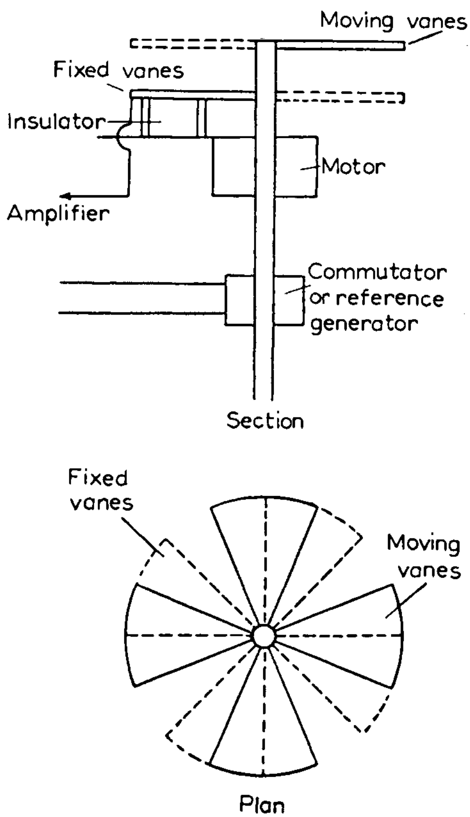


FIG. 21. The field mill.

the existing electrostatic fluxmeters (see § 5.26.), but needing no commutator. They used a fixed circular plate as the test plate with, above it, a fixed plate of two 90° sectors of a circle. Between the two plates is a "rotor", connected to earth, also consisting of two 90° circular sectors; the rotor moved, in a horizontal plane, at a constant speed, so that, when the rotor sectors were below those of the upper plate, the test plate was partially exposed to the field, while in the intervals it was completely shielded. Thus an alternating current appears in the connection from the test plate to earth. In the simple form of the instrument the upper sectors can be connected to earth, but HARNWELL and VAN VOORHIS also used it with a potential applied to the upper sectors, adjusting this potential until there was zero output from the machine, in which case there must be a uniform field at the surface of the test plate, partly from the external field and partly from the upper sectors. By the use of what is essentially a negative feedback circuit, HARNWELL and VAN VOORHIS stated they were able to control the potential on the upper sectors electronically; but, as they had no method of discrimination of sign in their output, it is difficult to follow how the sector potential could be controlled. Since they knew, independently, the sign of the potential gradient, there was no need for any method of determining this. A similar machine, but with sign determination, as discussed in § 5.27., has been used by ADAMSON (1958). One advantage of the negative feedback circuit is that the range over which the output is linear is increased.

A similar arrangement to that of HARNWELL and VAN VOORHIS, but with the upper plate at earth or other fixed potential, was used by LUEDER (1943), but other workers have preferred the simpler design of a number of equal sectors of a circle, often four 45° sectors, with a rotor of exactly similar shape above it; this design has been used by TRUMP, SAFFORD and VAN DE GRAAFF (1940), WADDEL (1948), CLARK (1949), VON KILINSKI (1950), CROSS (1953), MAPLESON and WHITLOCK (1955) and others. The output is approximately triangular in form; the theory of a field mill of this form has been discussed in detail by DAHL (1951) and by MAPLESON and WHITLOCK (1955).

In order to give a sine wave output, the plate and rotor have been suitably shaped by VAN ATTA *et al.* (1936), RANGS (1942) and ACCARDI (1952). MACKY (1937) used a clockwork mechanism to cause an earthed cover to move over and off the test plate at regular intervals. MALAN and SCHONLAND (1950) and MAPLESON and WHITLOCK (1955) used a set of studs, arranged in a circle, as the test plate and rotated over them a plate with corresponding holes.

The usual type of field mill cannot be used for the measurement of very rapid changes of potential gradient because the test plate is shielded for half of each cycle. To avoid this, SMITH (1954) divided his test plate into two sets of sectors which would be shielded alternately. By the use of pulsed "clamping" circuits he was able to add the two outputs, and so obtained a resultant which was sensitive continuously and could record very rapid changes of potential gradient.

The theory of the field mill, both with triangular and sinusoidal output, has been discussed by MAPLESON and WHITLOCK (1955), with particular reference to the choice of circuit components and to the elimination of effects of rain currents, etc.

TAKEUTI, ISHIKAWA and TAKAGI (1958) have reported an unexplained discrepancy between the readings obtained with a field mill in artificial and natural potential gradients.

5.26. The Electrostatic Fluxmeter

The term electrostatic fluxmeter is used here for the type of field machine in which the test plate is moving, not fixed as in the field mill proper. The test plate is connected to earth or to a second test plate and the current amplified and measured.

The first instrument of this type appears to have been that of MATTHIAS (1926), in which two test plates were used, forming two halves of a cylinder split along a diametral plane. The cylinder rotates about the axis, which is placed horizontally, i. e. perpendicular to the potential gradient. The charge induced on either semicylinder varies with time according to a sine wave, the two being 180° out of phase. If the two cylinders are connected together, a sine-wave current between them is obtained and this is amplified and measured. The same method has been used by KIRKPATRICK and MIYAKE (1932), KIRKPATRICK (1932), HENDERSON, GOSS and ROSE (1935), SCHUCHARD (1935), FELDENKRAIS (1937), NEUBERT (1938) and KASEMIR (1944).

An early method of a similar type was that of GUNN (1932); although the published description is that of an electrometer for the measurement of charge, the same method was adapted for the measurement of potential gradient.

Gohlke and Neubert (1940) varied the exposure of their test plate by vibrating it up and down through a corresponding hole in a flat earthed plate and later (1942) used a test plate consisting of a number of parallel strips and vibrated them horizontally below an earthed plate with similar slits.

Gunn (1954a) used a test plate consisting of a ring on the end of a rod; this moved in a vertical plane, the lower half of its motion being within an earthed case.

The theory of this type of instrument has been discussed by Schuchard (1935).

Kasemir (1964) claims that the effects of contact potential are much less in this type of instrument than in the field mill.

5.27. Determination of Sign of Potential Gradient

The field mill and electrostatic fluxmeter in their simplest forms give an alternating current output, proportional to the magnitude of the potential gradient but giving no indication of its sign.

A number of the investigators mentioned either were not concerned with the sign of the potential gradient (e. g. Macky, 1937; Lueder, 1943) or knew it otherwise (e.g. Matthias, 1926; van Atta *et al.*, 1936).

One method of determining the sign is to rectify the alternating output by a mechanical commutator synchronized with the current, usually by running it on the same shaft as the moving parts of the field machine; this method was used by Kirkpatrick and Miyake (1932), Kirkpatrick (1932), Henderson, Goss and Rose (1935), Schuchard (1935), Feldenkrais (1937), Waddel (1948), Clark (1949), Cross (1953) and Adamson (1958). The disadvantage of this method is that the signal from the machine has to be taken into the amplifier, returned to the commutator and then back again to the recording apparatus, and this may be a serious problem when the machine is situated some distance from the amplifier.

By means of an auxiliary generator synchronized with the output, the zero can be displaced so that the output is always of the same sign; this method was used by Rangs (1942), von Kilinski (1950) and Accardi (1952). Rangs used an "electromagnetic" output from a dynamo on the same shaft as the field mill and obtained a zero indication by periodically shorting out the "electrostatic" output.

MALAN and SCHONLAND (1950), who used 18 main studs, had, in addition, two subsidiary studs mounted closer to the axis than the main studs and along the same diameter as two of the studs; corresponding holes were made in the moving cover, so that every ninth impulse was made more pronounced. They displayed their output on a cathode-ray oscilloscope and so could tell whether the augmented effect occurred at a "peak" or "trough", thus identifying the sign of the potential gradient. COLLIN (1962) used a field mill with the rotor normally earthed; at intervals of $\frac{1}{2}$ min, a positive potential was applied to this, giving an increase, in the positive direction, to the output from the field mill; the direction of the impulse on the rectified output indicated the sign.

Instead of using a mechanical commutator or a simple electronic rectifier to give the final direct-current output, it is possible to use some form of phase-sensitive rectifier, with a reference voltage synchronous with the field-machine output. The earliest use appears to be that of SMITH (1951a) whose field mill was driven from the 50 cycle mains by a synchronous motor; he obtained his reference voltage by squaring and differentiating the mains voltage. MAPLESON and WHITLOCK (1955) used phase-sensitive rectifying circuits with reference voltages derived from generators working on the same shafts as the field mills. CROSS (1953), GUNN (1954a) and IMIANITOV *et al.* (1956) also used phase-sensitive rectifiers and SMITH's (1954) "clamping circuits" are similar in principle.

ISHIKAWA (1961) used a different method; on the same shaft as the rotor of his field mill he had a screen with the same number of slits as the field-sensitive electrodes; a lamp on the axis then gave pulses of light synchronized with the field-mill output and these operated a photoelectric cell whose output controlled the "gate" of a detector, thus removing one half of each cycle.

5.28. The Induction Voltmeter

In the field mill and the electrostatic fluxmeter the test plate is permanently connected to the measuring apparatus; but in the induction voltmeter and the agrimeter the plate is connected to earth when exposed and to the measuring apparatus when shielded. The measuring apparatus can be either a sensitive galvanometer which registers a direct current, or an electrometer which is charged by the successive contacts. In the induction voltmeter the plate is fixed and the cover moves, while in the agrimeter the plate moves.

The induction voltmeter was first described by REYNOLDS and NEILL (1955) and was also used by VONNEGUT and MOORE (1958b). A cover moves over and off a fixed plate and, at appropriate times, the same mechanism connects the plate to the measuring instrument and to earth.

5.29. The Agrimeter

The first instrument of this type was described by RUSSELTVEDT (1925) and it can be almost certainly claimed as the first field machine for continuous operation, having been first made and used in 1909. In this instrument a plate moves in a cylindrical fashion round a horizontal axis; in its highest position it is exposed to the field and is earthed; at a lower position, where it is shielded, it is connected to an electrometer which receives the charge which had been "bound" at the upper position. WORKMAN and HOLZER (1939) used the same principle with a number of horizontal vanes rotating round a vertical axis and coming underan opening in the cover; when a vane was under the opening, it was connected to earth and later, when shielded, to an electrometer. CHALMERS (1953e) used the horizontal axis method of RUSSELTVEDT, with

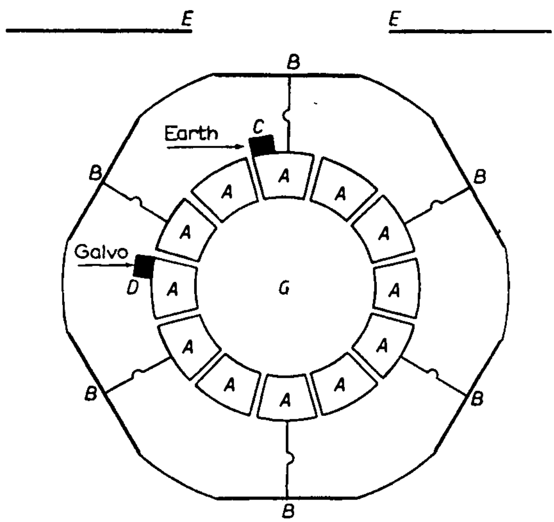


FIG. 22. The agrimeter (From CHALMERS, 1953, Fig. 2, p. 125.)

6 plates and, as an alternative to the connection to an electrometer, connected the output to earth through a galvanometer, so that each plate sends a small quantity of charge to be measured; with high-speed rotation this intermittent current appears as a direct current. GOTO (1952), GUNN (1954a) and VONNEGUT and MOORE (1958b) have described machines working on similar principles.

VONNEGUT, MOORE and HARRIS (1961) increased the sensitivity of the agrimeter, used to provide a potential rather than a current, by adding an insulated "Faraday cage", into which the plate passes before being discharged. As in the VAN DE GRAAFF generator, the whole charge on the plate can pass to the cage, irrespective of the potential the latter has already acquired. In the instrument used by VONNEGUT, MOORE and HARRIS, imperfect shielding limited the potential acquired by the cage, but it was about 25 times the potential of the plates.

5.30. Theory of the Agrimeter

The general theory of all types of field machine is very similar, and that of the agrimeter is chosen here because the conditions are simple, with successive impulses, and because there are two different output arrangements that have been used.

If a plate of area A m^2 is exposed to a potential gradient F V/m, the bound charge is:

$$Q = FA\varepsilon_0 .$$

Let C be the capacitance of the plate alone, at the time when it is in contact with the collector, and let D be the capacitance of the collector, the cable connected with it and any other connected apparatus. Also let R be the resistance between the collector and earth.

After receiving the charge Q when in contact with the earth, at the position of maximum exposure, the plate is unconnected for a time T_3 and is then in contact with the collector for time T_1. It must be assumed that there is no leakage from the plate during the time T_3.

If the collector system carried a charge P at the commencement of the time T_1, there will now be a charge $(P + Q)$ on the collector and plate, with capacitance $(C + D)$ leaking through a resistance R for a time T_1.

After the time T_1 the charge remaining is:

$$(P + Q) \exp [-T_1/(C + D) R] = (P + Q) \alpha$$

if

$$\exp [-T_1/(C + D) R] = \alpha.$$

Of this, a fraction $D/(C + D)$ remains in the collector system, the rest moving on with the plate. If T_2 is the time before the next plate makes contact, the charge remaining on the collector system leaks for a time T_2 and leaves

$$\frac{D}{(C + D)} (P + Q) \alpha \exp (-T_2/DR)$$

$$= \frac{D}{(C + D)} (P + Q) \alpha\beta.$$

if

$$\exp (-T_2/DR) = \beta.$$

This we can call $P + \partial P$.

During the times T_1 and T_2, the charges which have flowed through R are, respectively,

$$(P + Q) (1 - \alpha)$$

and

$$\frac{D}{(C + D)} (P + Q) \alpha(1 - \beta),$$

so the average current I is given by

$$I(T_1 + T_2) = (P + Q) (1 - \alpha + \frac{D}{(C + D)} \alpha(1 - \beta)).$$

If the potential gradient is steady, Q will be constant and P must reach a steady value so that δP is zero.

This gives

$$\frac{D}{(C + D)} (P + Q) \alpha\beta = P.$$

From this:

$$I(T_1 + T_2) = Q\left(1 - \frac{C\alpha}{C + D - D\,\alpha\beta}\right).$$

The average potential across the resistance R is

$$V = IR = \frac{QR}{(T_1 + T_2)}\left(1 - \frac{C\alpha}{C + D - D\,\alpha\beta}\right).$$

If the resistance R is small, so that $T_1/(C + D)\,R$ and T_2/DR become large, α and β become small and we get

$$I_0 = Q/(T_1 + T_2).$$

On the other hand, if R is large, so that $T_1/(C + D)\,R$ and T_2/DR are small, we can put:

$$\alpha = 1 - T_1/(C + D)\,R \quad \text{and} \quad \beta = 1 - T_2/DR$$

and if, further, C is small compared with D,

$$I_R = Q/(CR + T_1 + T_2)$$

and

$$V_R = QR/(CR + T_1 + T_2).$$

When R is very large, so that $CR \gg (T_1 + T_2)$,

$$V_\infty = Q/C.$$

The results I_0 and V_∞ correspond to the uses of the agrimeter with current and potential output respectively. It can be pointed out that the current output method does not depend upon very good insulation in the machine; provided that the leakage to earth has a resistance large compared with the galvanometer resistance, no difficulty occurs; on the other hand, the potential output is independent of leakage only if R remains much greater than $(T_1 + T_2)/C$, which in the Durham agrimeter amounts to about $2 \times 10^8\,\Omega$. Consistency of results for the current output depends upon a constant rotation frequency, but this does not enter into the potential output.

It is also possible to determine the rate at which the output responds to changes in potential gradient.

We had

$$P + \delta P = \frac{D}{(C + D)}(P + Q).$$

Here δP is the change of P in a time $(T_1 + T_2)$, so that, for times long compared with $(T_1 + T_2)$,

$$\delta P = \frac{dP}{dt}(T_1 + T_2) .$$

Thus

$$\frac{dP}{dt}(T_1 + T_2) = P\frac{D\alpha\beta}{(C + D)} - 1 + Q\frac{D\alpha\beta}{(C + D)} .$$

When R is fairly large we can use the previous approximations for α and β and get, with $C \ll D$,

$$\frac{dP}{dt}(T_1 + T_2) = -\frac{P}{DR}(CR + T_1 + T_2) + Q$$

which is solved by

$$P = \frac{QDR}{CR + T_1 + T_2}\left(1 - \exp\frac{-(CR + T_1 + T_2)\,t}{DR(T_1 + T_2)}\right)$$

if $P = 0$ for $t = 0$, giving a time constant of

$$\frac{DR(T_1 + T_2)}{(CR + T_1 + T_2)} .$$

With R large, this becomes $(D/C)\,(T_1 + T_2)$ with smaller values for smaller R.

In practice, if the output is passed through a high-sensitivity galvanometer or an electrometer, the rapidity of response depends on that of the recording instrument rather than of the agrimeter itself.

5.31. Calibration and Reduction Factor for Field Machines

A radioactive, or other, equalizer in its simple form gives a potential which is that of the atmosphere in its neighbourhood, so that no calibration is required beyond that of the instrument with which the potential is measured. From the potential, the potential gradient can be obtained and a reduction factor applied as necessary.

With a field machine of any type, however, the output depends upon the amplification, or, in the case of the agrimeter, upon other electrical constants, and so a calibration is needed. If the instrument can be placed so that its opening lies in the plane of the earth's surface, the rest of the machine being in a pit below ground level, then a large horizontal plate can be erected over the machine and raised to various potentials above earth; it is sometimes preferable to have a second horizontal plate, with a suitable sized hole, resting on the ground. This makes it possible to have a direct calibration of the machine; if the situation is in a large area of level ground, no reduction factor is required.

If a field machine is to be used either above the ground on a stand, or inverted or in a situation where there is any distortion of the equipotentials, then a reduction factor is needed; this can be provided by a stretched wire system as discussed in § 5.18., but it is found preferable to use a field machine in a pit as discussed above. The procedure is then as follows: the field machine in the pit is first calibrated as described. Then the machine, for which the reduction factor is required, is set up in its working situation and simultaneous readings are taken from it and from the machine in the pit, the horizontal cover now, of course, being removed. It is desirable to carry out this procedure under steady conditions and preferably when the potential gradient is fairly high.

If it is required to use the field machine under conditions of higher potential gradient than can be used for the calibration, this can be done by a "relative" calibration. An insulated plate is placed parallel and close to the opening of the field machine and various potentials applied between it and the machine cover, at earth potential; lower potentials will give outputs which correspond to calibrations by the previous method and it is then known that the applied potential gradient is proportional to the plate potential; thus the calibration can be extended.

In these calibrations with plates it is advantageous to increase both the plate distance and the applied potential, so as to reduce any effects of contact potentials, but this cannot be taken too far, since it is essential that the lines of force remain strictly vertical; the larger the plate, the farther can it be moved from the machine without distortion.

5.32. Double Field Machines

A single field machine cannot be used to measure the potential gradient at points above the earth's surface because, as can be seen from § 2.20., the charge on a portion of a conductor depends on both the external potential gradient and on the potential difference between the conductor and its surroundings.

But, if two field machines are used, at places on a conductor symmetrical with respect to a horizontal plane, then, as discussed in § 2.20., the sum of the charges, and so the sum of the outputs of the machines, will be proportional to the potential difference between the case (and covers) of the field machines and the surroundings; and the difference of the outputs will be proportional to the actual potential gradient, independent of the effect of the potential of the machines. An apparatus on this principle has been described by DEMON (1953), with the machine at earth potential.

The theory on which the above method is based is satisfactory for an isolated system and the method has been used for measurements in aircraft (see § 5.38.), but it must be realized that the apparatus must distort the lines of force. Where there has to be other apparatus close, or where there is any concern with space charges or other effects of similar nature, this distortion may be serious. To avoid this, SMIDDY and CHALMERS (1958) used the sum of the outputs to actuate a servo-mechanism to reduce this sum to zero, thus bringing and keeping the machine to the potential of its surroundings and minimizing the distortion. The double field machine thus behaves as an equalizer, acquiring the potential of its surroundings; at the same time, the output from one half gives the potential gradient.

JONES, MADDEVER and SANDERS (1959a) used a double electrostatic fluxmeter attached to a radio-sonde balloon; the difference in the two outputs gave the potential gradient, independent, within quite wide limits, of the actual potential of the mill.

FORD (1958) described a very simple instrument with similar properties. To measure the vertical potential gradient a collecting plate is set up in a vertical plane with a parallel cover rotating about a horizontal axis; near the periphery of the cover, a hole is cut and the cover balanced by suitable weighting. As the hole exposes different parts of the plate, the latter carries a charge whose variation depends on the vertical component of the potential gradient, while the charge corresponding to the horizontal component normal to the plate remains constant.

CURRIE and KREIELSHEIMER (1960) used a double field mill in which the two rotors are set at 90°; the two fixed plates are connected together and the output automatically subtracts the effects of the self-charge of the instrument, while adding the effects of the external field. The apparatus, designed to be used in a balloon with radio-sonde, does not, of course, bring the mill to the potential of its surroundings, but does, like that of JONES, MADDEVER and SANDERS (1959a) eliminate the effect of the mill's own potential.

KOBAYASHI and KYOZUKA (1962) used an agrimeter with two openings and two sets of contacts, to be carried up in a radio-sonde, and used as in the cases referred to above to eliminate the effects of self-charge.

5.33. Piezoelectric Measurement of Potential Gradient

IMIANITOV (1949b) proposed a method in which the potential gradient is measured by the bound charge but without moving parts. If the plate for the measurement of bound charge is covered by a slab of material whose permittivity is ε, then the bound charge per unit area is $-\varepsilon F$, where F is the potential gradient. If the slab is of a piezoelectric material, such as barium titanate, then its value of ε can be altered by applying an electric field, in a horizontal direction so as to avoid interfering with the natural field. If the applied horizontal field is varied sinusoidally at a definite frequency, then ε will alter at double this frequency. Thus the bound charge alters with the same frequency as ε and with an amplitude which depends upon F; with suitable amplification and rectification the value of F can be obtained. This method can be regarded as giving a partial screening of the plate by the slab, the extent of the screening being controlled by the horizontal field.

5.34. Measurements of Changes of Potential Gradient

In some investigations the interest lies not in the actual values of the potential gradient, but in its changes; this is particularly the case for the very rapid changes of potential gradient caused by lightning flashes. For such purposes equalizers are always much too slow in action and it is necessary to use some method based on the bound charge. Of the field machines, only those of MALAN and SCHONLAND (1950) and SMITH (1954) have been used for potential-gradient change measurements and other observers have measured the charges flowing to or from an exposed conductor. WILSON (1916) made use of the very rapid response of the capillary electrometer, both with the simple exposed plate and with the raised sphere (see § 5.35.); to avoid the effects of rain, WOR-

MELL (1939) used an inverted test plate, also with a capillary electrometer (see § 12.4.). ISRAËL (1943) used an electronic amplifier to measure the current to and from an exposed plate in his Feld-variograph.

WORMELL (1930) and WHIPPLE and SCRASE (1936) have used point-discharge current changes to give indications of potential-gradient changes.

5.35. The Raised Sphere

For the measurement of potential-gradient changes which are too small for accurate measurement by the change of bound charge on a plate close to the ground, WILSON (1916) used a sphere 1 ft in diameter mounted on insulators on an iron pipe 5 m long; the pipe could be lowered so that the sphere passed into a case to shield it from the lines of force.

If the sphere is at a height h where, in the absence of the sphere, the potential would be V, then when the sphere is earthed in this position it must carry a charge q of such a value as to make the potential zero. The potential consists of three parts: (1) V due to its position; (2) $q/4\pi\varepsilon_0 r$ due to a charge q on a sphere of radius r; and (3) $-q/8\pi\varepsilon_0 h$ due to the image charge (see § 2.19.). If F is the potential gradient at the ground when the sphere is removed, then $V = Fh$.

So, $Fh + q/4\pi\varepsilon_0 r - q/8\pi\varepsilon_0 h = 0$ giving q proportional to F and greater than the value $\varepsilon_0 AF$ for a small area A on the earth's surface. Thus the change in q for a given change in F is correspondingly magnified.

5.36. Measurement of Time Intervals

If two instruments of the field-machine type are used at a distance apart that is not too great, it is often found that the same pattern of potential gradient is found on both records, but with a short time interval. A reasonable assumption, though not the only one that is mathematically possible (see § 2.10.), is that there is some disturbing charge in motion, and the time interval should give the velocity of the disturbance resolved along the line joining the two instruments. A further assumption, that the disturbance is moving with the velocity of the wind, makes it possible in some cases to estimate the height of the disturbance and then to obtain some idea of the magnitude of the charge concerned.

5.37. Measurements in Balloons

In order to make measurements of the potential gradient at points more than a few metres above the earth's surface, it is necessary tu use some form of aircraft. Because it can remain nearly stationary and because there are no engines which may produce separation of charges, a balloon has been found very suitable for such measurements. However, the surface of the balloon is at least partially conducting and so the field is distorted by the presence of the balloon and corrections are needed. If the balloon could be considered as an uncharged conductor, the problem of the distortion of the field could be solved quite simply theoretically, but in most cases a balloon carries a resultant charge, derived in part from the surface charge on the earth when the balloon leaves and in part from effects due to the dropping of ballast, so that the total charge on the balloon is unknown.

LINKE (1904) attached a potential equalizer to the surface of the balloon to bring it to the same potential as the air and then made measurements of the difference of potential shown by equalizers at two levels; he employed droppers as the equalizers, using alcohol in place of water when the temperature was below the freezing point. EBERT and LUTZ (1908) used a model to determine the best positions for the equalizers, to minimize the disturbance of the

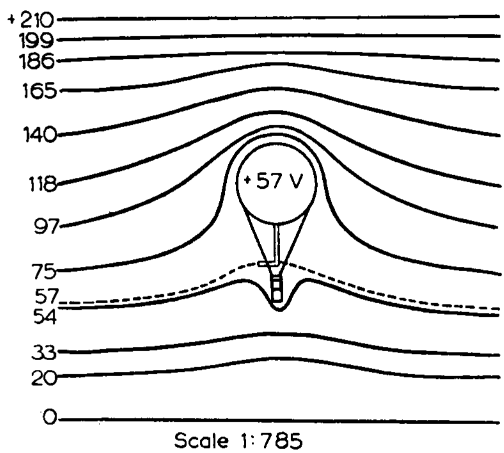


FIG. 23. Equipotentials near balloon. (From CHAUVEAU, 1925, Pt. II, Fig. 24, p. 142.)

field; they put the equalizer connected to the balloon 7 m above the basket and the two equalizers for the measurement of potential gradient at 8 m and 10 m below the bottom of the basket; they found it necessary to wait for some minutes after each change of height before taking the readings.

Excepting the earliest observations, all measurements have shown that in fine weather the potential gradient decreases with height from 100 m or so upwards, the results at lower levels being less definite. As discussed in § 2.25., this involves a positive space charge and is related to the increase of conductivity with height.

VON SCHWEIDLER (1929) summarized the results of 57 balloon flights over central Europe in the empirical formula

$$F = 90 \exp(-3{\cdot}5z) + 40 \exp(-0{\cdot}23z),$$

where F is the potential gradient in V/m and z the height in km.

The measurements described above were carried out with manned balloons, but in more recent years similar measurements have been made using radio-sonde balloons and adapting a standard radiosonde to transmit back measurements for potential gradient, obtained by using two radioactive equalizers. The method was first used by KOENIGSFELD and PIRAUX (1950) and later by VENKITESHWARAN, DHAR and HUDDAR (1953), LUGEON and BOHNENBLUST (1956) and STERGIS, REIN and KANGAS (1957a). HATAKEYAMA *et al.* (1958) used a machine similar to the agrimeter (see § 5.29.) and JONES, MADDEVER and SANDERS (1959a) a type of double field mill (see § 5.32.) instead of the radioactive equalizers.

On account of the increase of conductivity with height, due to the greater intensity of cosmic rays, it is to be expected that the potential gradient would decrease continuously to low values as the height is increased. However, KOENIGSFELD (1953) and VENKITESHWARAN and HUDDAR (1956) found that at heights of 15,000 to 30,000 ft (about 5–10 km) the potential gradient remained constant at 20 to 50 V/m. On the other hand, STERGIS, REIN and KANGAS (1957a) and HATAKEYAMA, *et al.* (1958) found the expected decrease to low values; it is not clear how the earlier results acquired what must be an error.

KOENIGSFELD (1955b) found marked changes in potential gradient at cloud level and also at certain temperature levels in clear air. JONES, MADDEVER and SANDERS (1959a) also found changes of potential gradient on entering and leaving clouds.

PALTRIDGE (1964) in measurements at 10–30 km, dispensed with the radioactive equalizers, since the conductivity is sufficient without them. By using collecting discs of large area, he was able to get rapid response. Results agreed with those expected, giving an exponential fall of potential gradient with height.

PIERCE and WHITSON (1964) measured the variation of potential gradient with height over a region with abnormally high radioactivity and found an increase by a factor of 3 to 4 in potential gradient at 1 km as compared with that at ground level; these results agree with simple approximate calculations.

5.38. Measurements in Aeroplanes

An aeroplane is very liable to become charged itself and so to disturb the lines of force, thus preventing potential-gradient measurements from being made simply by a single field machine. The rapid motion prevents the use of the same methods to remove the self-charge as are applicable with a balloon (see § 5.37.).

GUNN *et al.* (1946), GUNN (1948) and GISH and WAIT (1950) used two field machines, above and below the wing and, by combination of the outputs, were able to eliminate the effect of the self-charge and obtain the true value of the vertical potential gradient.

CLARK (1957) used three field machines, one at the bottom of the aircraft and one at each wing tip; the relative reduction factors were found by measurements when banking at an angle of 45° and the absolute reduction factor for the bottom machine by direct comparison between measurements on the ground and during low flights over the same site. He was thus able to determine the vertical potential gradient and the self-charge.

PELTON *et al.* (1953) used 4 field machines to obtain the vertical and horizontal potential gradients and the self-charge; by a suitable combination of the 4 outputs they obtained a resultant output proportional to the self-charge and they attempted to use this to actuate a discharging mechanism, functioning by means of point discharge, to neutralize the self-charge.

KASEMIR (1951 b, 1964) described a type of electrostatic fluxmeter which gave all three components of the potential gradient and also the self-charge.

FITZGERALD and BYERS (1962) also used 4 field machines and fed the signals into an analogue computer to give the three components and the aircraft charge. Since there is a "reduction factor" for the potential gradient arising from the aircraft being a conductor, a correction is required; FITZGERALD and BYERS considered the aircraft as an ellipsoid, and so determined the form of the corrections, but the numerical values were determined experimentally.

VONNEGUT, MOORE and BLUME (1957) arranged two similar installations of radioactive equalizers at the wing tips. In level flight there is no differential effect, but on banking the aircraft the vertical potential gradient can be measured.

VONNEGUT and MCCAIG (1960) mounted a pair of radioactive equalizers so that they would be in the same equipotential surface when the aircraft was charged but in no external field; in an external field, they would indicate the potential gradient in the line joining them. With three such pairs of equalizers they were able to obtain the three components of potential gradient. Since neither equalizer of the pair is at the potential of the aircraft, a simple amplifier cannot be used, but VONNEGUT and MOORE (1961) have used a coupling device to make it possible to use an amplifier.

5.39. Measurements in Gliders

LECOLAZET (1948a) made measurements of potential gradient from a glider, using a radioactive equalizer. He neutralized the charge on the glider itself by the use of a jet of water, acting as a water-dropper. His measurements were made in fine weather with fine-weather cumulus cloud present. He found a large increase in potential gradient as the glider passed down from the clear air above cloud level into the air affected by *austausch* (see § 2.32.). He also found effects due to cumulus clouds through or near which he flew. The results were explained in terms of differences of conductivity between pure air, polluted air and clouds; such differences have been found directly by SAGALYN and FAUCHER (1954) (see §§ 7.14., 7.19.).

ROSSMANN (1950) used a radioactive equalizer on a mast 1·17 m high on a towed glider, which he considered less liable to self-charging than a motored aircraft. He made measurements mainly in fine weather and found the expected decrease of potential gradient with height together with effects to be ascribed to mist and haze layers.

5.40. Measurements at Sea

Measurements of the potential gradient at sea are made difficult by the fact that a vessel does not remain steady and hence the more usual methods are not readily applicable.

ANGENHEISTER (1914) and SWANN (1914b) attempted to overcome these difficulties by using the change of charge on a conductor which is suddenly moved, the method being similar to the early method of DE SAUSSURE (§ 1.6.) and differing from that of WILSON with the raised sphere (§ 5.35.) only in that in WILSON's method the conductor is moved from a covered position to an exposed position, while in these methods it is moved from one exposed position to another.

RUTTENBERG and HOLZER (1955) used a field mill mounted on the flying bridge in such a position that disturbance by the ship was at a minimum.

5.41. Choice of Periods for Analysis

When records of potential gradient have been made over a period of time it is then necessary to analyse these to investigate regular variations. In the great majority of cases it is the fine- or fair-weather, "normal" or "undisturbed" potential gradient which is investigated and it is therefore necessary to make some choice as to whether particular readings should or should not be included in the analysis. ISRAËL and LAHMEYER (1948) have pointed out that different observers have used very different methods of choice of periods to be included and excluded in analysis; so the results of analysis by different observers may not be strictly comparable, and in cases where the same observational material has been differently selected there are conspicuous differences in the resulting variation curves.

Usually the method of selection has been to choose "undisturbed" days, specifying the disturbance either from the potential-gradient records themselves or meteorologically, but ISRAËL and LAHMAYER (1948) proposed that every observation should be used unless precipitation was actually falling at the time, giving a quite definite criterion.

5.42. Diurnal Variation of Potential Gradient

Results for the diurnal variation of potential gradient during fine weather vary considerably from one part of the world to another. As discussed in § 2.24., the main significance of such variations often lies in showing the variation of the resistance of the lowest layers of the atmosphere.

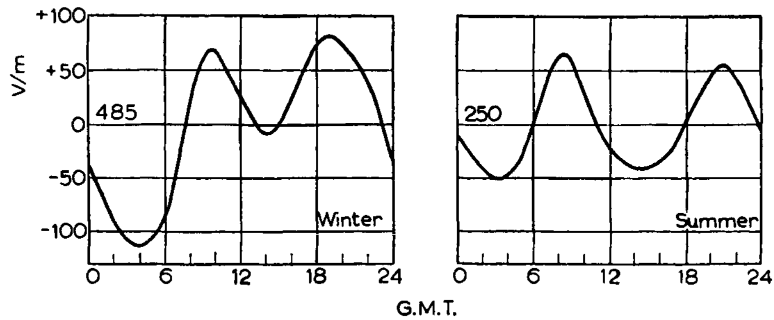


FIG. 24. Diurnal variation of potential gradient at Kew, winter and summer. (From SCRASE, 1934, Fig. 11 (lower 2 curves), p. 14.)

Over the oceans, in polar regions and in a few non-polar continental regions well removed from sources of pollution, there is a single period with a simultaneous world-wide maximum.

At most land stations the times of maxima and minima depend on local time and show that local influences predominate. In many cases, there occurs a double oscillation of potential gradient, with minima from 4 to 6 a.m. and from midday to 4 p.m., and maxima from 7 to 10 a.m. and from 7 to 9 p.m. In many cases the morning maximum disappears during the winter months. Some typical examples of the diurnal variation are shown in Figs. 24 and 25. The amplitude of the diurnal variation often amounts to over 50 per cent of the mean value.

5.43. Diurnal Variation in Undisturbed Conditions

Observations at Karasjok in Lapland (SIMPSON, 1905) and at Cape Evans in the Antarctic (SIMPSON, 1919) showed the variation of potential gradient to be simple in both cases with a single maximum, but with different times for the maximum. In both cases, there is no local source of pollution, so the effect of large ions in the atmosphere is small. HOFFMANN (1923) first pointed out that the times of the maxima in these two cases and in some other measurements are really simultaneous, though they differ in local time.

Shortly afterwards, MAUCHLY (1923), in analysing the results obtained during the cruise of the *Carnegie*, pointed out that the same result holds for the potential gradient over the sea, and more

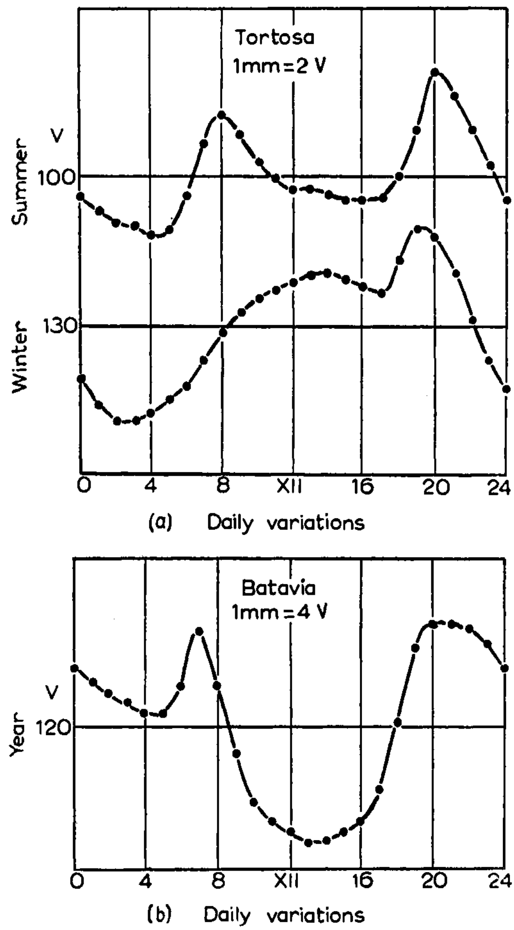


FIG. 25. Diurnal variation of potential gradient at Tortosa, summer and winter, and at Batavia, year. (From CHAUVEAU, 1925, Fig. B (lower 2 curves), p. 126 and Fig. C (lowest curve), p. 127.)

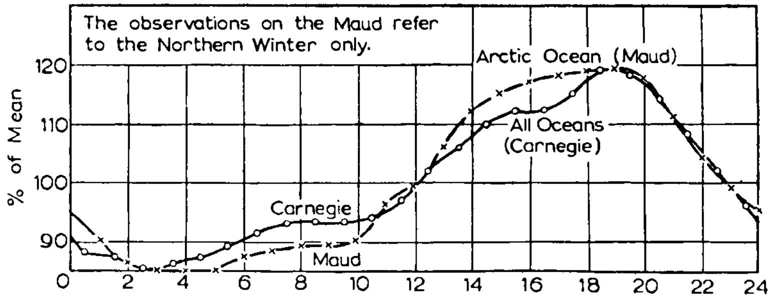


FIG. 26. Diurnal variation of potential gradient over oceans. (From WHIPPLE and SCRASE, 1936, Fig. 9A, p. 19.)

recent results have confirmed that, wherever there are no local sources of pollution, the potential gradient shows a maximum at about 19 hr Greenwich Mean Time, independent of the local time. Apart from polar and ocean measurements, this maximum has been found by SHERMAN (1937) in Alaska and by SHERMAN and GISH (1937) in South Dakota.

Referring to § 2.24., if λ and R are unaffected by time of day, then F shows the variations of V, and the maximum of potential gradient shows the maximum of the potential difference between the electrosphere and the earth; this has been correlated with the maximum of thunderstorm activity over the earth by WHIPPLE (1929a) (see § 11.7.).

PARAMONOFF (1950) eliminated local effects by taking 60 continental non-polar stations and arranging them in longitudinal regions; he then took the average, over all regions, for the potential gradient for each time of day (by G.M.T.) and obtained a curve in extremely good agreement with those for oceans and polar regions.

GISH (1942) pointed out that, other things being equal, the potential gradient should be less in low latitudes than in high, because of the difference in cosmic-ray intensity; in low latitudes, fewer ions are produced and so the columnar resistance is greater and F is decreased.

5.44. Local-time Effects

As stated in § 2.24., the potential gradient at the earth's surface F is given by

$$F = V/\lambda R$$

where V is the potential of the electrosphere, λ the conductivity of the air near the surface, and R the columnar resistance. In the cases considered in the last section, λ and R remain constant and variations in F reflect variations in V.

But at places where λ and R do not remain the same at all hours of the day, the variation of F depends on the variation of all three, λ, R and V, and of these it is often λ that is subject to the greatest variation, a variation which usually depends, in fine weather, on the time of day.

The conductivity of the air is due almost entirely to the small ions and depends upon the number of small ions present. As discussed in Chapter 4, the number of small ions depends mainly on the number of nuclei and the conductivity is a minimum when the number of nuclei is a maximum, thus giving a maximum of the potential gradient if the variations in V and R can be neglected in comparison with the variations in λ.

Very marked diurnal effects are found, depending on local time, especially near large cities where there will be the greatest introduction of nuclei into the atmosphere. Support is given to the idea that the potential gradient is greater when the number of nuclei is greater by observations that when the visibility is exceptionally good, the potential gradient is remarkably low, both being associated with the absence of nuclei; fog, on the other hand, reduces the number of small ions and, if no other effects intervene, produces high potential gradients.

WHIPPLE (1929b) verified the fact that the number of nuclei, and hence the potential gradient, depends upon the pollution entering the air. He found a change in the time of the summer

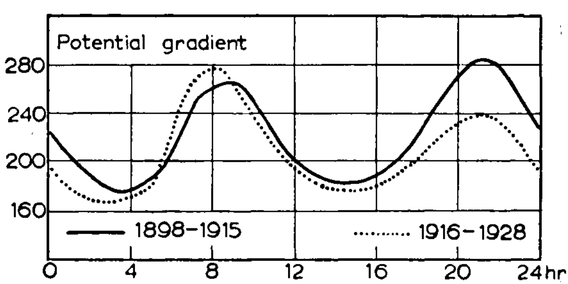


FIG. 27. Effect of "summer time" on the morning maximum of potential gradient. (From SCHONLAND, 1953a, Fig. 5 (upper), p. 26.)

morning maximum (in G.M.T.) as between the years before and after the introduction of "summer time" and ascribed this to the alteration of the habits of the people as to the times (in G.M.T.) of the lighting of fires and of the production of other forms of pollution (see Fig. 27).

Perhaps the most striking proof of the effect of pollution and its causes is shown by the measurements of SAPSFORD (1937) who found, in Samoa, a marked difference in the diurnal variation of potential gradient between Sundays and weekdays, caused by the natives having special fires on Sunday mornings and none on Sunday evenings (see Fig. 28). MÜHLEISEN (1953) also found a difference between Sundays and weekdays in an industrial area in Germany.

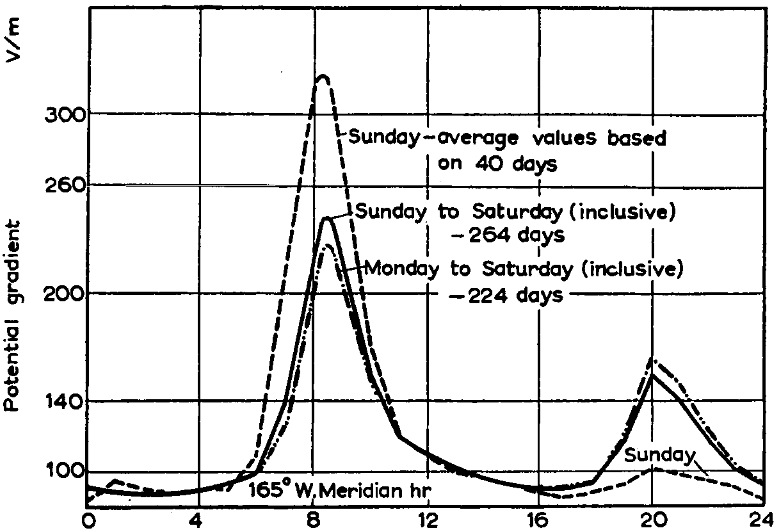


FIG. 28. Diurnal variation of potential gradient in Samoa. (From SAPSFORD, 1937, Fig. 1, p. 157.)

5.45. Origin of Local Diurnal Variation

The total number of nuclei present, and hence the conductivity, is affected not only by the production of the nuclei but also by the way in which they are dissipated; this dissipation is effected mainly by *austausch* (see § 2.32.). The production of nuclei is at a minimum in the early hours of the morning and their dissipation by *austausch* is at a maximum in the early afternoon; at these times, therefore, the conductivity is greatest near the ground and so, from the results of § 2.24., the potential gradient will show a minimum; thus the two daily minima can be given a general explanation. In addition, the facts that convection is more pronounced in summer than in winter and that domestic heating produces fewer nuclei in summer than in winter can account for the absence of the second minimum in the winter curves at some stations.

BROWN (1935) discussed the local effects by subtracting the world-wide variation, termed by him the "unitary variation" from the whole variation actually measured; he then found that the local effect could be considered to consist of a 24 hr wave with a maximum in the afternoon, together with a "depression" in the afternoon. He attributed the effects to convection of space charge and he found (BROWN, 1936) that the local effect is much reduced when there is continuous wind, which would prevent convection of space charge, and also (BROWN, 1937) that the local effect is reduced by cloud cover, without rain, as this decreases the heating of the earth's surface, thus reducing the convection.

ISRAËL (1949, 1953b) pointed out that, in the relation $F = V/\lambda R$, variations in V can be eliminated by using $H = F/F_0$, where F_0 is the potential gradient over the oceans at the same time as F is that at the point in question. Since F_0 is proportional to V, $H \propto 1/\lambda R$ and the variation, with time of day, of H gives the variation of $1/\lambda R$. From simultaneous measurements of λ, ISRAËL (1949) was able to determine how R varies during the day at different places.

ISRAËL (1953b) considered that BROWN's 24 hr wave with a maximum in the afternoon represents the variation of R, the columnar resistance, which is largely determined by the total number of nuclei in the air, irrespective of their height. BROWN's "depression", on the other hand, ISRAËL explained as a change in λ, the local conductivity close to the surface, due to the removal of nuclei to higher levels by convection.

ISRAËL (1953b) quoted two special cases in support of these ideas. At Huancayo, in Peru, the results of TORRESON and WAIT (1948) show that H and $1/\lambda$ vary very closely together, so that R remains nearly constant; in this case λ is varying only in a thin layer, so that R remains nearly constant. On the other hand, at the Jungfraujoch (ISRAËL, KASEMIR and WIENERT, 1951) in the autumn λ is unaltered for most of the day, but R varies through convection bringing up nuclei from below to levels above the station.

ISRAËL (1953b) also found that the potential gradient shows a similar type of diurnal variation to the water-vapour content, though of greater percentage amplitude, and concluded that there must be a similar cause. This is explained in terms of the introduction into the atmosphere of nuclei on the one hand and water vapour on the other, and their dispersal by convection; it may be compared with the relation between large-ion content and humidity (see § 4.13.).

KAWANO (1958a) considered the matter in greater detail, taking into account not only convection but also ionic recombination and the production of ions, particularly by the radioactive substances in the atmosphere, the distribution of which will be affected by convection. He was able to give a detailed account of the distribution of resistivity with height and an explanation of the local diurnal variation of potential gradient.

5.46. Sunrise Effect

NICHOLS (1916) first observed an increase in potential gradient at the time of sunrise. This was rediscovered by HOLZER (1955) and KASEMIR (1956) and discussed also by ISRAËL (1957). The peculiarity of the phenomenon is that there is a simultaneous increase in the air–earth conduction current, with little change in the conductivity; the effect is more pronounced in the summer than in the winter and is less noticeable at mountain tops than at lower levels.

Various suggestions have been made to account for the phenomenon. KASEMIR (1956) suggested an "*austausch*-generator" but CHALMERS (1957a) showed that this could not give a complete account of the phenomenon (see § 8.23.). CHALMERS's suggestion, that the explanation lies in MÜHLEISEN's (1956) discovery of the emission of positively charged nuclei into the atmosphere from industrial processes, does not fit in with the existence of the sunrise effect at mountain stations. The results of MÜHLEISEN (1958) on the effects of moisture may be relevant, or there may be some photoelectric effect of the sun's rays.

LAW (1963) related the sunrise effect to a change in sign of space charge and convection current and so to a difference of conduction current (see § 8.23.).

5.47. Annual Variation

From the same arguments as have been used in the discussion of the diurnal variation, it would be expected that the potential gradient would be greater in the local winter than in the local summer, since the amount of pollution is greater. This is found to be the case in most observations near large towns and other sources of pollution. In general, also as would be expected, the effects of local variations are less marked in summer than in winter and in some cases the diurnal maxima become indistinct in summer.

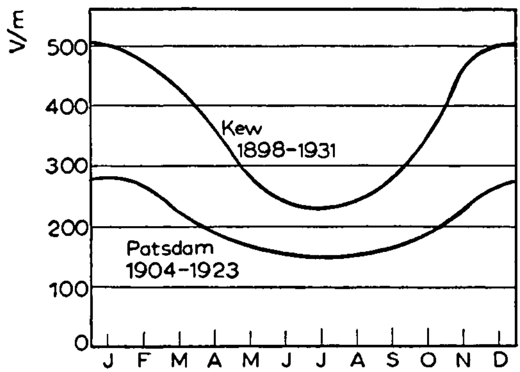


FIG. 29. Annual variation of potential gradient at Kew and Potsdam. (From SCRASE, 1934, Fig. 14 (top and third curves), p. 16.)

In places where there is no pollution, i.e. where the variation of potential gradient depends upon variations of V rather than upon variations of λ, it is found that the potential gradient is greater in the southern summer than in the northern summer, the maxima agreeing, as expected, in time for both hemispheres; the difference between the annual maximum and minimum is rather less than for the diurnal variation. These results have been discussed by WHIPPLE (1929a) and FISCHER (1962).

5.48. Secular Variation

PIERCE (1957) found that the annual mean value of the potential gradient at Eskdalemuir in Scotland had decreased quite markedly in the preceding 10 years and suggested that this might be ascribed to an increased conductivity, caused by radioactive fall-out from nuclear explosions. He found no comparable change in the measurements at Lerwick in the Shetland Islands; this difference may be due to the fact that fall-out does not remain on the surface of the sea and Lerwick has sea surfaces close at hand while Eskdalemuir is inland.

Reiter (1959) made extensive measurements at various heights, both of natural and artificial radioactivity and of potential gradient and air–earth current, and came to the conclusion that, for the amounts of artificial radioactivity found up to then in West Germany, there was no evidence for any effect on the atmospheric-electrical elements.

Pierce (1959) made calculations on the effects to be expected from the known fall-out of products of nuclear explosions and concluded that the effects of Eskdalemuir were of the order of 50 times as great; he suggested that, at Eskdalemuir, there were effects to be ascribed to radioactive leakage from atomic stations to the south-west.

Stewart (1960) found a decrease of potential gradient at Eskdalemuir, Lerwick, Lisbon and Oporto, and an increase of conduction current and of conductivity, but little change of potential gradient, at Kew. He found that the results could be ascribed to increased β-ray ionization from the top few centimetres of the ground, due to fall-out of fission products from nuclear explosions.

5.49. Local Effects on Mean Value

The actual mean value of the potential gradient varies quite widely from one place of observation to another. As has been discussed, this must be ascribed mainly to differences between the mean values of the local conductivity near the earth. A large part of such differences must be due to differences in the mean nucleus content, which controls the small-ion content and hence the conductivity.

But in some places the potential gradient is abnormally low, corresponding to high small-ion content, caused by greater ionization from radioactive substances present to a more than usual amount in the neighbourhood. An example occurs in the measurements of Norinder, Israël and Siksna (1954).

5.50. Mountain Measurements

Israël, Kasemir and Wienert (1951, 1955) measured the potential gradient at the Jungfraujoch (3472 m) and found that in the summer the diurnal variation of potential gradient is similar to that of other land stations but in winter the variation is that found over oceans or in polar regions. These results can be readily understood in terms of local conductivity and pollution. In summer, *austausch* (see § 2.32.) extends up to the height of the station and brings with it the disturbances due to pollution which depend on local time; but in winter, *austausch* does not extend to this height and so there are no effects of pollution.

ISRAËL (1957) commented on the fact that the mean fair-weather value of the potential gradient at the Jungfraujoch, after correction for the exposure factor, was 149 V/m, whereas at the same level in the free atmosphere it would be very much less. As ISRAËL pointed out, the lowest part of the columnar resistance is absent at the mountain station and so the conduction current is increased (see § 8.27.) as compared with a complete column with the same potential drop. Thus, if the conductivity is the same at the mountain top as in the free air, the potential gradient must be greater.

5.51. Space Charges

In air which is horizontally stratified from the electrical point of view, the equation of § 2.8. gives

$$\varepsilon_0 d^2 V/dx^2 = \varepsilon_0 dF/dx = -\varrho,$$

where ϱ is the space-charge density.

Thus, if there is a space charge, the potential gradient alters with altitude, and vice versa. The methods of measuring space charge will be considered in Chapter 6.

Localized space charges at small heights in the atmosphere can produce effects on the potential gradient at the earth's surface (see §§ 5.57., 5.68.).

5.52. Correlation of Potential Gradient with Nature of Air Mass

GHERZI (1961) observed very high values of potential gradient during periods of anti-cyclones of polar air at Montreal. He considered these results to show that the polar air mass must contain within it a positive space charge; this has suggested that the measurement of potential gradient might be a useful accessory to other meteorological measurements, and also that the term "fine weather" is not a sufficiently distinctive description.

5.53. Effect of Relaxation Time

KASEMIR (1950) pointed out that changes in potential gradient, and also in current and space charge, are governed by the "relaxation time" (see § 2.27.). When we are considering changes over periods which are long compared with the relaxation time, then the changes closely follow whatever causes them. Under such conditions it is correct, and much more convenient, to consider the current as the primary phenomenon; the potential gradient can then be thought of as the potential drop across the resistance of the corresponding part of the atmosphere, and the space charge as arising from the alteration of conductivity with height (see § 2.25.). It is possible to adopt the principle of the quasi-static state (see § 2.23.) and to consider slow alterations of this state.

On the other hand, if we are concerned with changes taking place in periods which are not long compared with the relaxation time, it becomes necessary to deal with the problem in terms of electrostatic effects, rather than current effects. The actual charges, rather than their rate of movement, have now to be considered as fundamental.

Thus there is a clear distinction to be drawn between the annual and diurnal variations of potential gradient on one hand, and short-period variations of less than an hour on the other.

5.54. Effects during Eclipses of the Sun

A number of measurements have been made of the effect of an eclipse of the sun on the potential gradient and other phenomena of atmospheric electricity. The results are very confusing, some showing an increase in potential gradient, others a decrease and yet others no noticeable effect at all. CHAUVEAU (1922) discussed the matter and, in particular, his own observations on 17 April 1912, using the same apparatus as had been in use for 20 years, and he found this day to show no distinctive features. On the other hand, JONES and GIESECKE (1944) made measurements on 25 January 1944, at Huancayo, again with apparatus in constant use, and found a definite lowering of potential gradient and changes in the other elements, as compared with a normal day. GISH (1944a), discussing these results, suggested that the effect was to be explained by the interruption of the heating of the earth by the sun, and depending on local circumstances. A more recent example of measurements was that of KOENIGSFELD (1953) on 25 February 1952, in the Belgian Congo; during the eclipse the potential gradient at the ground was found to be much lower than normally and to become negative; measurements with a radio-sonde with equalizers (see § 5.37.) showed that the effect extends to above 5000 m. The effect is ascribable to the change of convection resulting from temperature changes. MÜHLEISEN (1955) found a distinct drop in potential gradient at and just after totality in the eclipse of 30 June 1954, and discussed the possible causes. Other results and references have been given by CHAPMAN (1956) and ISRAËL and FRIES (1956).

It must be realized that if there are any effects due to particle, rather than radiative, effects from the sun, than the time and place of totality will not be that of the visual eclipse, and this must affect the conclusions.

5.55. Effect of Volcanic Eruptions on Potential Gradient

Japanese observers have investigated the effect of volcanic eruptions on the potential gradient observed at various distances to leeward of the eruption. HATEKEYAMA (1949) observed the effect of the eruption of Yake-yama on the potential gradient at Kakioka, about 250 km away, finding a negative potential gradient reaching beyond −1200 V/m; the shape of the potential gradient/time curve agreed with a spindle-shaped distribution of charge moving with the wind.

For the volcano Aso, HATEKEYAMA and UCHIKAWA (1951) made measurements close to the eruption and the results agreed with the theory that the larger particles of volcanic ash obtain positive charges and the smaller negative charges; laboratory measurements on the frictional effects of this ash agreed with this conclusion. On the other hand, for the volcano Asama, both outdoor and laboratory measurements gave results of the opposite signs, ascribable to the different kind of ash.

BLANCHARD (1965) discussed the electrical effects produced by the volcanic island of Surtsey and concluded that these were caused by the very rapid heating of sea water coming into contact with the volcanic magma (see § 3.21.).

5.56. Effects of Aurora

Various observers have reported a decrease in potential gradient during an auroral display (ANDREE, 1890, SHEPPARD, 1933; SHERMAN, 1934; SCHOLTZ, 1935; FREIER, 1961). On the other hand, KOENIGSFELD and VAN DER SCHUEREN (1963) found an increase.

Since the visual aurora is always above the base of the electrosphere, which should act as a perfect electrostatic screen, these results are surprising.

DOLEZALEK (1964) discussed the phenomena and various possible explanations, without giving preference to any one, and suggested further investigations.

5.57. Short-period Fluctuations of Potential Gradient

Fluctuations of the potential gradient can be detected over whatever period of time is considered. Long-period fluct uations have already been discussed in connection with diurnal and annual variations and can be ascribed in part to meteorological effects, including with these the nucleus content of the air, and in part to changes in the potential difference between the electrosphere and the earth.

When fluctuations of shorter period are considered, it follows, as ISRAËL (1959a) has pointed out, that in times which are not long compared with the relaxation time of the atmosphere, the fluctuations cannot represent effects of the electrosphere, but must correspond to something much more local. ISRAËL (1959a) defined fluctuations of periods between 1–2 min and 1 hr as the "agitation" and has given a detailed study of this.

Some of the effects may be ascribed to air masses of conductivity differing from that of the surroundings, but in many cases the phenomena must be caused by space charges at fairly low levels; the results will be discussed in §§ 6.13., 6.15.

All deviations from the normal values of the potential gradient can be ascribed to space charges somewhere, but, for the purpose of the present book, it is convenient to treat comparatively widespread and long-lasting effects in the present chapter, and to consider more localized and temporary effects in the chapter on space charge (Chapter 6).

5.58. Horizontal Potential Gradients

Apart from irregularities due to sloping ground, the potential gradient at the earth's surface must be vertical, since the earth is a conductor. But, above the earth, there may be horizontal components of the potential gradient, particularly when there are pockets of space charge close to the place of measurement. The matter has been discussed by IMIANITOV (1949a) who pointed that it is only charges close to the ground that could give rise to a horizontal component of potential gradient at a small height above the earth. He has made some measurements which indicate surprisingly large horizontal components, and it would be desirable to repeat measurements of this type.

5.59. Potential-gradient Effects from High-tension Cables

CHALMERS (1952a) found negative potential gradients to the leeward of high-tension cables (133,000 V) during periods of mist and fog; these must be explained by the production of an excess of negative ions where insulation is partially breaking down, the negative ions then travelling in the wind; under suitable conditions, effects were observed at 5 km from the cables. MÜHLEISEN (1953) made similar observations to as far as 7 km. It is clear that caution is necessary before a negative potential gradient can be definitely associated with a process of charge separation in clouds.

MÜHLEISEN (1953) also observed effects in fine weather from cables whose diameters were too small to be suitable for the voltage carried; in such cases, positive space charges were usually found, often giving potential gradients several times the normal value. It is probable that such effects would not occur with cables of suitable diameter for the voltage carried.

NORINDER, ISRAËL and SIKSNA (1954) found further examples of similar effects, and discussed them in relation to laboratory measurements of corona discharge.

5.60. Effects of Föhn Winds

REITER (1958) has distinguished between two different results of Föhn winds, which are, in the neighbourhood of his observations, warm dry southerly winds; in one type, the effects can be ascribed to drifting snow on mountain peaks and will not be further discussed. The more usual effects show a decrease of potential gradient, accompanied by no change or a slight increase in conduction current; these effects can be ascribed to an increase in the concentration of small ions, brought about by increased radioactivity in the air. These results are peculiar to the particular situation of REITER's observations and would not necessarily hold in other cases of Föhn winds.

5.61. Potential-gradient Variations under Clouds

WHITLOCK and CHALMERS (1956) found that under non-raining clouds, in particular stratus and strato-cumulus clouds, the variation of potential gradient is closely connected with the state of the sky overhead; the potential gradient shows minima under the thicker portions of the cloud, suggesting that negative charges reside in the lowest parts of such portions of the cloud. In some cases the effects were sufficient to make the potential gradient at the ground negative, and it does not appear that the effects can be accounted for merely in terms of a difference in conductivity, such as is described in § 3.4. For a discussion of the origin of these charges see § 15.42.

REITER (1955a) found effects with alto-cumulus clouds which also suggested a negative charge in the base of the cloud.

5.62. Effect of Dust Storms on Potential Gradient

In the violent dust storms of some tropical regions there are intense electrical effects. In general, these give rise to a negative potential gradient at the earth. A thorough investigation was made by RUDGE (1914), who found that the sign of the potential gradient produced depends on the chemical nature of the dust, acidic dusts producing a negative and basic dusts a positive potential gradient. The effect is due to a separation between, on the one hand, the larger particles of dust and, on the other, either smaller particles or the air; if a charge is produced on the air, this must mean the production of ions of some size.

FREIER (1960) found that the effect of a "dust devil" in the Sahara was explicable in terms of a dipole with negative charge above and positive below. CROZIER (1964) found similar effects in New Mexico.

LATHAM (1964) compared dust storms with snow storms and suggested that a similar type of mechanism might account for the electrical phenomena of both.

5.63. Effects at Coast

MÜHLEISEN (1959) investigated potential gradient and space charge at a station on the island of Sylt, particularly in relation to the differences in land and sea winds. He found that sea winds showed higher potential gradients, with more rapid changes in value, and positive space charges which must originate at the actual shore in the breaking of waves. The rate at which conditions typical of sea winds were set up, after rain and a change of wind, was slow, suggesting that the main effect might be the production of small positive ions, with subsequent attachment to nuclei.

5.64. Potential Gradients in Disturbed Weather

Considering the large number of continuous records that have been made of the potential gradient at various places, it is surprising how few results are available for periods of disturbed weather. The main reason for this is that the range of the recording instruments is that suitable for the fine-weather potential gradient, and in disturbed weather the values often get far outside this range; also some types of recording apparatus are affected by precipitation.

SIMPSON (1949) analysed records from a radioactive equalizer for values of potential gradient between ± 2000 V/m and also records of point-discharge current for values of potential gradient beyond ± 2000 V/m. He found that for rain caused by a warm front or other quiet conditions the potential gradient is usually negative and fairly steady, with values usually not beyond -1000 V/m. On the other hand, for rain associated with a cold front or other instability the potential gradient is often beyond ± 2000 V/m and undergoes rapid and frequent changes in sign. SIMPSON found that the character of the potential gradient record bears no relation to the rate of rainfall.

FUCHS (1955) also found the difference between the effects of warm and cold fronts and has given a composite diagram, in which, however, the effects of the cold fronts are less than those found by SIMPSON.

5.65. Potential Gradients in Quiet Rain

In quiet rain the potential gradient at the ground is usually found to be negative; ADKINS (1959a) found this to be true on a glacier when the cloud base was only little above the observer, so that secondary effects between the cloud and the ground could play no part.

KELVIN (1860b, c) found that the potential gradient sometimes remains positive at the top of a 30 m tower during rain, while it has become negative at the ground; CHAUVEAU (1900) found the same for the Eiffel Tower. Under these conditions, the surface charge at the ground is positive, while at the top of the tower it is negative; consideration of the lines of force shows that there must be a negative space charge in the levels of the air below the top of the tower. Various suggestions have been made as to the origin of this charge.

SMITH (1955) suggested that the negative potential gradient is produced by splashing of the raindrops at the ground, according to the effect found by LENARD (1892) and discussed in § 3.24. The results indicate that generation of charge by splashing occurs only if the rain is fairly heavy, so that splashing cannot be the only cause of the negative charge. But it may be that splashing plays some part. The charge produced by splashing depends on the ground surface and so it might be expected that any potential gradient effects due to splashing would depend on the surface. A particular case is that of sea water, where splashing is found to release positive charge, in contrast to normal splashing, releasing negative charge; the observational results for the potential gradient during quiet rain at sea are contradictory; SIMPSON and WRIGHT (1911) found a negative potential gradient as on land, but SWANN (1915) found that it usually remains positive.

ADKINS (1959a) suggested that the negative charge below the top of a tower could arise from charge released if the precipitation melts in this region; there is as yet no evidence that this in fact occurs. The results of REITER (1955a) showing a change in sign of potential gradient with a change from rain to snow, both in space and in time, agree with the suggestion that melting releases negative charge.

The results for towers might also be explained if raindrops break up in gusts of wind near the ground, again giving negative charge to the air.

5.66. Potential Gradients in Snow

In steady snowfall, SIMPSON (1949) found the potential gradient to be usually positive, in contrast to the usually negative value in steady rain. Polar measurements have also shown positive potential gradients in light snow when there is no drifting (SIMPSON, 1919; SVERDRUP, 1927; SHEPPARD, 1931).

REITER (1955a) made simultaneous measurements of potential gradient at different stations not very far apart but at different levels; he found, during precipitation, that the potential gradient was positive where dry snow flakes fell but negative where there was rain, mixed rain and snow or partially melted snow flakes. ADKINS (1959a) found the potential gradient to be positive during snow from a cloud quite close above the observer.

If it can be assumed that the potential gradient is positive in the free air where the precipitation is snow and negative where it is rain, then there must be a negative space charge in the region of melting. However, it may be that either the snow or the rain (or both) produces an effect on reaching the ground, and it is this effect which gives rise to the potential gradient observed, by providing a space charge. To distinguish between these possibilities, it would be necessary either to make potential-gradient measurements in the free air, or to detect the space charge.

When there are blizzards or drifting snow, the potential gradients are large and usually positive (SIMPSON, 1919). In snow showers, potential gradients are large and vary rapidly.

5.67. Potential Gradients in Fog

The effects of fogs on the potential gradient are consequences of the change of conductivity as between fogs and clean air, and are dealt with in § 7.11.

For a widespread fog, there is a considerable increase in potential gradient, but for a mountain fog of limited extent the potential gradient is decreased only slightly as compared with the fine-weather value.

5.68. Effect of Point-discharge Ions on Potential Gradient

When point discharge is occurring, ions are liberated and travel downwind from the point. WHITLOCK and CHALMERS (1956), by the use of two field mills, showed that the potential gradient to leeward of a discharging point is often very different from that to windward and suggested that this may be the origin of some of the more complex "field patterns" (§ 5.69.) found by SIMPSON (1949). The motion of the ions, and probably also their number, depends on the wind speed, and since the wind is often very gusty during the conditions which give rise to point discharge, it follows that there may well be violent variations in the space charge near to a field mill to leeward of a discharging point, and so corresponding variations in the measured potential gradient. When there are a number of discharging points in very disturbed weather the measured potential gradient is a complex phenomenon and may give very little information about charges in the cloud or in the precipitation.

In more steady conditions, with a single point, the potential gradient to leeward of a point will be less than that to windward by an amount corresponding to the effect of the point-discharge ions. DAVIS and STANDRING (1947) calculated the effect to be expected if the ions are assumed to travel horizontally in a straight line from the point.

WHITLOCK and CHALMERS (1956) gave results which appeared to indicate an effect only about one-third of that expected, but a more accurate investigation by MAUND and CHALMERS (1960) has shown that, in fact, the measured effects agree very closely with the prediction; MAUND and CHALMERS have shown, both theoretically and experimentally, that the turbulent diffusion of the ions from the straight-line path does not appreciably affect the difference of measured potential gradients to windward and leeward, provided the distance from the point is not too great.

5.69. Potential-gradient Patterns

SIMPSON (1949) found that in disturbed weather remarkable patterns are sometimes found in the records of potential gradient or of point-discharge current; he was able to divide these into wave patterns and symmetrical patterns.

Wave patterns, approximating to simple harmonic waves, sometimes with a regular change in amplitude, were found with up to 5 complete waves, with ranges from 700 V/m to 23,000 V/m and with periods from 4 to 106 min; the centre of the wave was always close to zero potential gradient. The most remarkable, which occurred during light snowfall, is reproduced in Fig. 30.

Symmetrical patterns of 5 different types were found, the most striking feature being the symmetry in time, as can be seen in the example shown in Fig. 31. The centre of a pattern often occurs at the time of the passage of some kind of discontinuity, e.g. a cold front, so that the two halves of the pattern occur with different air masses present.

Patterns of similar kinds have also been found by SIVARAMAKRISHNAN (1953), NORINDER, ISRAËL and SIKSNA (1954) and WHITLOCK and CHALMERS (1956).

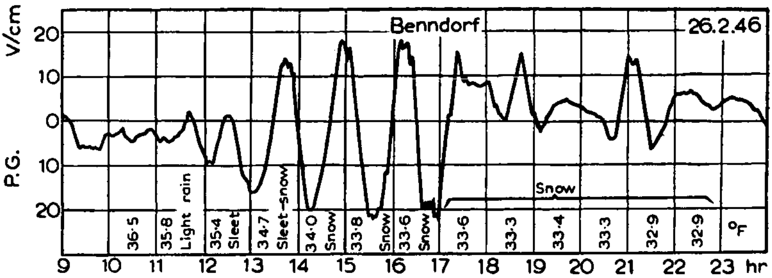


FIG. 30. Wave pattern of potential gradient during snow. (From SIMPSON, 1949, Plate VIII, Fig. A.)

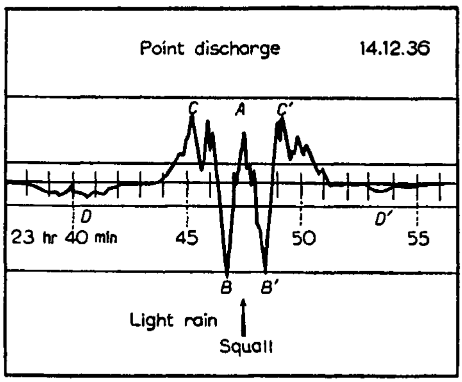


FIG. 31. Symmetrical pattern with squall. (From SIMPSON, 1949, Plate XII, Fig. A.)

Patterns might be accounted for by a suitable arrangement of charges, in the same vertical line for some symmetrical patterns and with horizontal separations for wave patterns and for other symmetrical patterns, such as the "N" pattern. The changes of potential gradient could then be produced either by the horizontal motion of the whole system over the observer or else by vertical motion of charges in the cloud. By the use of two field mills and measurement of time intervals (see § 5.36.), WHITLOCK and CHALMERS (1956) have shown, in some cases, that the patterns can be ascribed to charges moving horizontally with speeds somewhat greater than the surface wind speed but less than the estimated geostrophic wind speed. Because of the effects of point-discharge ions (see § 5.68.), this method is not available when the potential gradient is large.

5.70. Effects of Cosmic-ray Showers on Potential Gradient

WILSON (1957) suggested that very high energy cosmic-ray showers would affect the electrical state of the lower atmosphere by providing a highly conducting path over a fairly narrow region and so altering the lines of force and lines of current flow. He suggested that the investigation of simultaneous sudden changes of potential gradient at a network of observing stations might prove to be a simple method of detecting and measuring these showers. The effect to be expected depends upon the fact that the conductivity under normal conditions increases with height, whereas the conductivity in the column of the cosmic-ray shower would be roughly uniform.

Attempts to make exact calculations of the effect of such a column on potential gradients at various points have not been successful, but it appears that it would be possible by this method to detect showers of energy 10^{20} eV and perhaps 10^{19} eV. However, it is showers of lower energy, 10^{17} eV and 10^{18} eV, which are of particular interest to cosmic-ray physicists, and showers of higher energy are so rare that the method is not of practical use.

5.71. Jet Stream and Potential Gradient

FALCONER (1953) and SCHAEFER (1955) reported that exceptionally high values of the fine-weather potential gradient, as measured by the current from a radioactive equalizer (see § 5.14.), occurred when a jet stream was close to the place of observation (Albany, N.Y.). BENT (1955), on the other hand, using similar apparatus, found no regular relation between the potential gradient and the distance of the nearest jet stream at Mount Washington, U.S.A., but he did find several occasions when there was an unusually high current with a well-developed jet stream close.

GREYSTONE (1954) made a statistical analysis of records at Lerwick (Shetland Islands) of potential gradient, measured by a radioactive equalizer used in the normal way, and jet-stream proximity and found no correlation.

ISRAËL (1959b) discussed the matter and pointed out the difficulties involved in explaining any effect.

It may be important that the effect has been found when measuring the current through an equalizer and it is probable that this current depends not only on the potential gradient but also on the wind speed. If, therefore, a close jet stream causes higher winds at the ground, the observations could be explained without the need for assuming any electrical effects of the jet stream.

5.72. Potential Gradients in the Upper Atmosphere

Because of the very high electrical conductivity in the ionosphere (80–400 km), any potential gradients are necessarily very small. In consequence of this, the relative importance of magnetic effects is much greater than in the lower atmosphere and there are important interactions between electric and magnetic fields. The dynamo theory of geomagnetic effects leads to the idea of electrostatic fields and calculations show that their magnitude would be 10^{-2} to 10^{-3} V/m.

Attempts to measure these potential gradients in the upper atmosphere involve two difficulties in addition to the small magnitude of the fields. In the first place, any vehicle from which the field is to be measured is certain to acquire a charge of its own, from the effects of various radiations and particles impinging on it; this charge produces its own field, which must be eliminated, and the problem is complicated by the high conductivity of the surrounding medium. Secondly, in addition to the field, there is also a conduction current to any measuring instrument; if the field is measured by any kind of field machine, what is measured is essentially the displacement current as the exposure alters, and there is at the same time a varying conduction current, with the same frequency but different phase. WILDMAN (1965) has attempted to distinguish the two effects by using two different frequencies.

The techniques required for measuring electrostatic fields in the ionosphere have been discussed by SAGALYN (1965) and GDALEVICH (1965) has discussed specific rocket experiments.

CHAPTER 6

Space Charge

6.1. Free Space Charge

By the term "space charge" is meant the free, unbalanced charge in a volume of air, taking no account of the charges of both signs which balance one another. If, in 1 m^3 of air, there is positive charge amounting to $+Q_1C$ and negative charge amounting to $-Q_2C$, then the space charge is $(Q_1 - Q_2)$ C/m^3.

Space charges may be carried on small ions, large ions, dust particles, cloud or fog droplets, or precipitation particles, etc.

The M.K.S. unit of space charge is 1 C/m^3, but many workers have expressed their results in electronic charges per cm^3. One electronic charge per cm^3 is 0·16 $\mu\mu$ C/m^3.

6.2. Primary and Secondary Space Charges

Space charges can be divided, according to their origin, into "primary" and "secondary" space charges. Primary space charges consist of charges which come into the atmosphere at some place and are then transported by agencies which are mainly non-electrical to the place where they are measured; examples are the space charge due to precipitation and space charges in industrial smokes. Secondary space charges are those which arise from electrical conduction currents and, for example, include those which are consequences of variations in electrical conductivity (see § 2.26.) and those arising from point discharge. Wind, etc., may convert secondary space charges into primary.

In measurements of space charge, when it is the secondary space charges that are in question, it is important that the measuring apparatus shall alter the natural conditions in the atmosphere as little as possible.

6.3. Methods of Measurement

The methods of measurement of space charge can be divided into three main classes. In the first type, the space charge is drawn into an earthed cage and the potential at a point in the cage is measured. In the second type, the whole of the space charge is collected in some way and measured. In the third type, POISSON's law is used and the space charge is deduced from the change of potential gradient with height (see § 2.8.).

A very full account of the various methods of determining space charge has been given by VONNEGUT and MOORE (1958b), and the reader is referred to that for more details than are given here.

6.4. Cage Method

The cage method was suggested by KELVIN (1862) and appears to have been first used by CHAUVEAU (1902); KÄHLER (1927) first made comprehensive measurements with this method. The air whose space charge is to be determined blows or is drawn into a wire cage which is earthed; the potential at a point inside the cage differs from earth potential if there is a space charge within the cage and this potential can be measured; KÄHLER used a water-dropper for the purpose.

If the cage carries an induced charge in the earth's field, ions in the air stream approaching the cage will be affected by this induced charge, the ions of the same sign as the charge being repelled and those of opposite sign being attracted. An increase in the grid size of the mesh of the cage would decrease this effect, but might allow the external field to penetrate to the collector.

VONNEGUT and MOORE (1958b) constructed a cubical cage of side 12 ft with galvanized steel wire mesh. To measure the potential or potential gradient inside the cage they used a radioactive equalizer and various types of field machine at the centre of the cube and found agreement; they finally preferred a polonium equalizer for simplicity. They adapted the calculations of DODSON (1950) to give the space-charge density N (in electronic charges/cm^3) as

$$N = 780\ V/l^2$$

where V is the potential (in V) at the centre of the cube, of side l m.

MÜHLEISEN and HOLL (1952) used a variant of KÄHLER's method in which they measured, not the potential acquired by the water-dropper, but the charges on drops falling from an earthed dropper inside the earthed cage.

6.5. Filtration Method

OBOLENSKY (1925) was the first to use a method in which the air is sucked through an absorber which picks up the charge and allows it to be recorded by an electrometer; he used cotton wool as his absorber. BROWN (1930) used steel wool and his apparatus is shown diagrammatically in Fig. 32. KINMAN (1954) made careful compari-

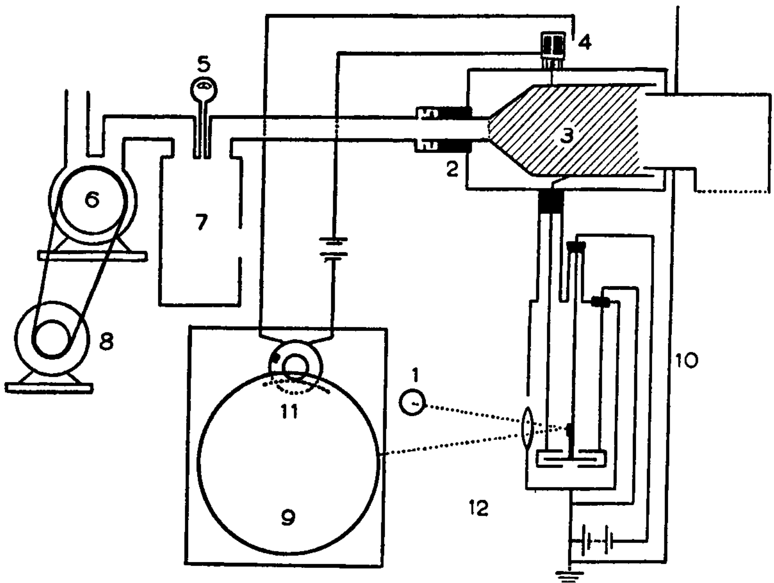


FIG. 32. Space charge apparatus. 1 = lamp; 2 = amber; 3 = steel wool; 4 = grounding key; 5 = pressure gauge; 6 = pump; 7 = surge chamber; 8 = motor; 9 = paper drum; 10 = metal plate in window; 11 = telechron motor; 12 = electrometer (From BROWN, 1930, Fig. 2, p. 6.)

sons between this type of apparatus and the cage type (§ 6.4.) and found good agreement.

VONNEGUT and MOORE (1958b) gave it as their opinion that the steel-wool fibres would be too large to trap all the condensation nuclei and some would escape; they suggested glass wool as more suitable and pointed out that the fact that glass wool is an insulator would be no disadvantage since a conducting case surrounding the glass wool would act as a "Faraday cylinder"; however, the presence of a charge on the fibres might repel other charges of the same sign and reduce the efficiency of collection. BENT (1964) has shown that this apparatus gives a very high efficiency of collection of charge.

Care must be taken with this type of apparatus to avoid effects of induced charges near the inlet. If the apparatus is set up in the open air, to obtain the natural space charge at a certain level, the cover of the apparatus must be earthed or there will be effects of displacement currents (see § 8.3.). In the natural field there will then be induced charges on the cover, of opposite sign to the potential gradient and so ions of the same sign as the potential gradient may be attracted to the cover near the inlet, thus failing to enter the apparatus. If it is hoped to remedy this by increasing the speed of suction, it is likely that the air sucked in is coming from levels other than that where the filtration apparatus is situated.

It may be possible to avoid the effect of induced charges near the inlet by putting the whole apparatus inside an earthed cage, but there arises the possibility of induced charges where the air enters the cage. Also, conditions inside the cage cannot be the same as those in the free air and the results then have less meaning.

SMIDDY and CHALMERS (1959) compared results obtained with this type of apparatus and those obtained with two double field mills (see §§ 5.32., 6.7.) and found good agreement except when there was reason to expect especially large numbers of small ions; under these conditions the results were as would be expected if induced charges had some effect.

In order to avoid effects of induced charges and to avoid the disturbance of the natural electrical state of the atmosphere by an earthed body, it would be desirable to maintain the cover of the filtration apparatus at the potential of its surroundings; it would then be necessary to have an earthed container between the cover and the collector, or there would be displacement currents. To bring the cover to the potential of its surroundings, a double field mill could be used, as it is undesirable to use a potential equalizer which produces ions (a radioactive source) or even droplets (a water-dropper).

6.6. Electrostatic Precipitation Methods

Of similar type to the filtration method are the electrostatic precipitation methods suggested by VONNEGUT and MOORE (1958b). The first of these is really an ion-counter (see § 4.11.) arranged to count ions of both signs by applying a field of either sign alternately or to alternate plates, and arranged to collect ions of very low mobility. An alternative method is to render the air so conducting that charges would move to the walls of the collecting tube however small might be their mobility; VONNEGUT and MOORE (1958b) used X-rays to make the air conducting, and also α-particles; a difficulty occurred with a zero reading of a negative charge which could not be eliminated.

6.7. Change of Potential Gradient

In order to produce a change in potential gradient of 1 V/m in 1 m there must be a space charge of $8{\cdot}8 \times 10^{-12}$ C/m^3, which is about 55 electronic charges/cm^3. Although it is not possible to measure potential gradients to an accuracy of 1 V/m with existing apparatus, it is possible to increase the vertical separation and to measure space charges of the order of 2×10^{-11} C/m^3.

DAUNDERER (1907, 1909) placed 3 lamps, in which flames acted as equalizers, at heights of 2 m, 1 m and ground level; in order to avoid any effects of the equalizers themselves, he measured only the differences between the potentials acquired by the collectors. NORINDER (1921) and SCRASE (1935b) used stretched wires with equalizers at their mid-points; NORINDER measured simultaneously the potentials attained by the wires at different heights, while SCRASE measured the difference in potential between two wires 1 m apart at different mean heights. BRASEFIELD (1959) used a kite-balloon at 43 m to support radioactive equalizers at 33 m and 21 m, and found no need to correct for the presence of the balloon; he also used a long pole to support an equalizer at 8 m, requiring a correction for the effect of the pole. Whereas the earlier workers used electrostatic instruments to measure potentials, BRASEFIELD used a current-measuring instrument in series with a very high resistance between the equalizer and earth, giving a method of measurement comparable with that of BREWER (see § 5.13.).

When space charges are being measured, it is clearly undesirable to have ions produced by the measuring device, since these ions give variations of conductivity and hence (see § 3.26.) space charges.

The only other method yet suggested to measure potentials or potential gradients at points above the earth's surface, without much distortion of the field, appears to be the double field mill (SMIDDY and CHALMERS, 1958) (see § 5.32.); since the double field mill measures both the potential and the potential gradient, one double field mill can be used to obtain the total space charge between the the ground and the level of the mill. With two double field mills the space charge between them can be determined.

It should be noted that this method includes any space charge on precipitation particles and it is probable that the filtration type of apparatus would not draw in such particles.

6.8. Values of Space Charge in Fine Weather

KÄHLER (1927) found values ranging from $+60$ to $+300$ $\mu\mu$ C/m^3 with a mean value of $+193$ $\mu\mu$ C/m^3. OBOLENSKY's (1925) results gave negative values in the summer and averaged, over a year, $+1{\cdot}2$ $\mu\mu$ C/m^3. BROWN's (1930) average for a year of continuous recording was $+28$ $\mu\mu$ C/m^3.

Using the change of potential gradient with height, DAUNDERER (1907, 1909) found a negative space charge in the lowest 2 m in winter and positive in the rest of the year; his average for the whole year was $+38$ $\mu\mu$ C/m^3. NORINDER (1921) found, for 0–3 m, an average value of -40 $\mu\mu$ C/m^3, but a positive charge at 8–9 m. SCRASE (1935b) divided his results into conditions of turbulent and still air; for turbulent air, the space charge was always positive, but, for still air, the value in the lowest 5 m was negative, and positive higher; his mean value was about $+10$ $\mu\mu$ C/m^3.

SMIDDY and CHALMERS (1959) found a mean value of -20 $\mu\mu$C/m^3 from 1–3 m, after neglecting a case where industrial smoke may have produced a positive space charge. They found the value of the space charge to depend upon the potential gradient, the higher the potential gradient, the more negative the space charge. An explanation was found in terms of the effects of radioactive rays from the ground (see § 8.20.).

LAW (1963) found values of $+10$ $\mu\mu$ C/m^3 by day and -20 $\mu\mu$ C/m^3 by night at a height of 50 cm. CROZIER (1963) found values of up to $+500$ $\mu\mu$ C/m^3 very close to the ground in very still night-time conditions. He connected this with the electrode effect (§ 8.19.). BENT and HUTCHINSON (1966) found values nearly always within the limits ± 60 $\mu\mu$ C/m^3.

MOORE *et al.* (1962) found, by measurements in aeroplanes, that from just above the ground up to 600 m the space-charge concentration seldom exceeds $0{\cdot}16\,\mu\mu\,C/m^3$ (1 electronic charge/cm^3),but these results were not concerned with measurements in the first few metres; the few cases of higher space charge were all associated with haze layers.

6.9. Space Charge and Ion Densities

A space charge of $+10\,\mu\mu\ C/m^3$ corresponds to a positive ion excess of about 6×10^7 ions/m^3. At Kew, where SCRASE made his measurements, the average number of small ions is $2 \times 10^8/m^3$ so that if the excess of positive ions is entirely composed of small ions, n_1/n_2 would be 1·42; actually n_1/n_2 is measured to be 1·35, and thus there would be a positive large-ion excess of about $10^7/m^3$. N_1 is about $7{\cdot}5 \times 10^9$ ions/m^3, so that the ratio N_1/N_2 still remains close to unity.

The largest space charge found by SCRASE was $+500\,\mu\mu\ C/m^3$ in still air at 5–7 m, corresponding to an excess of positive ions of $3 \times 10^9/m^3$. Under such conditions N_1 is about $1{\cdot}5 \times 10^{10}/m^3$, so that N_1/N_2 would reach only about 1·2 if all the excess were in the form of large ions.

6.10. Space Charges and Moisture

MÜHLEISEN (1958), working in an enclosed room, found that air in an enclosed space acquired a negative space charge in conditions of high humidity and a positive charge when the humidity was reduced. It does not seem clear what happens to the particles of opposite sign to those which form the space charge.

These results, if confirmed for the free air, would help to elucidate the phenomena at sunrise (see §§ 5.46., 8.23.).

6.11. Space Charges during Rain

ADKINS (1959b) measured the space charge during rain and found that the values were small unless the rain became heavy; in heavy rain, the results agreed with those of ion counting (see § 3.24.), giving a space charge opposite in sign to the potential gradient.

SMIDDY and CHALMERS (1959) found, also, that space charges appear only in heavy rain. They found the space charge produced always to be negative.

6.12. Space Charges at Cloud Levels

VONNEGUT, MOORE and BLUME (1957) used the measurement of potential gradient from an aircraft to obtain the space charge at various levels; they used radioactive probes on the wing tips and made measurements during banking at 20°. Their results showed the expected space charge at a discontinuity of conductivity (see § 2.26.), particularly at the top of the *austausch* region. But they also found some cases of negative space charges at cloud levels in the absence of clouds; similar results were obtained from KOENIGSFELD's (1953) measurements with radio-sondes. Since these negative charges were obtained only in flights over land it was suggested that they might be due to higher conductivity at a lower than a higher level, arising from radioactivity in the atmosphere.

6.13. Space-charge Packets

Measurements of changes in potential gradient over short periods of time (see § 5.57.) have shown effects which must be ascribed to space charges. In some cases, the space charges can be related to clouds, but in many cases they must be much lower.

One of the most noticeable of such effects is that caused by steam or smoke from a locomotive, always giving rise to a positive space charge. This was first noticed by KELVIN (1860a), commented on

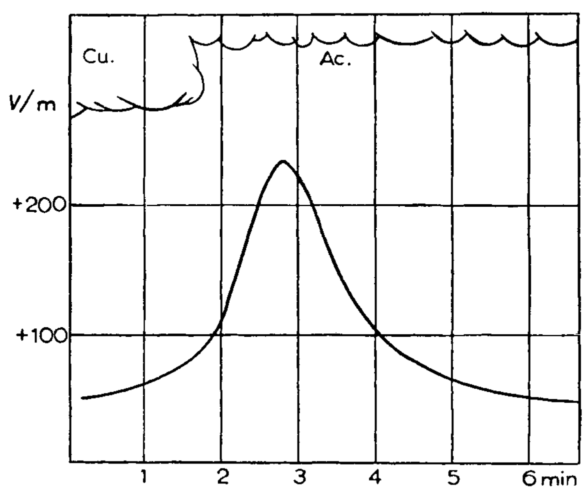


FIG. 33. Pulse due to space charge.

by ISRAËL (1950b) and others, and investigated in more detail by MÜHLEISEN (1953). WHITLOCK and CHALMERS (1955) investigated the rate of travel of packets of space charge believed to originate from locomotives, by using two field mills at a separation of 100 m in the direction of the wind and found them to move with the speed of the wind at a height of some tens of metres; they also found that the space charges could still be identified when the steam and smoke were no longer visible, at a distance of over 1 km from the source.

MÜHLEISEN (1953) also found positive space charges to be produced by diesel oil motors and other burning processes, but negative space charges from chemical laboratory and gasworks fumes. He found that the conductivity shows an increase close to the source of the space charge, but not at greater distances, indicating the initial production of small ions, which later become large ions and then have little effect on conductivity, though still affecting the potential gradient. MÜHLEISEN (1956) has discussed these space charges in relation to diurnal variations of potential gradient (see § 5.45.).

BRASEFIELD (1958), ISRAËL (1959a) and PIERCE (1964) have provided further examples of the effects of space-charge packets.

A different example of space-charge packets is found in the results of MOORE *et al.* (1962) who found positive space charges downwind of a television tower; the explanation appears to be that there is an electrode effect (see § 8.19.) at the top of the tower and this gives positive charge which is blown away by the wind.

6.14. Space Charges Produced by Natural Phenomena

One natural phenomenon which produces space charge is the electrode effect. CROZIER (1965) measured the natural space charge close to the ground, by measuring potentials with passive antennae at different levels (see § 5.17.), confirmed by some direct measurements. At night, with wind speeds less than 1 m/sec, he found positive space charges up to a height of 15–40 cm, with values of around 240 $\mu\mu C/m^3$. From this level up to at least 3 m, negative space charges were found with average values of the order of $-50\ \mu\mu C/m^3$. For short intervals, both the positive and negative charges reached values over three times those quoted. At night when the wind speed was greater than 1 m/sec and for all wind speeds during the day, the negative space charge disappeared but positive space charges up to about 80 $\mu\mu C/m^3$ were found close to the ground. An explanation of the seresults was given in terms of the assumption of a high rate of ion production close to the ground, particularly in low winds when radon can collect near the ground.

BENT *et al.* (1965) found space charges to be ascribed to point discharge from trees (see § 9.14.). BENT and HUTCHINSON (1965) found space charges associated with the melting of snow on the ground (see § 3.19.).

6.15. Potential-gradient Variations in Fine Weather

WHITLOCK and CHALMERS (1956) investigated short-period potential-gradient changes in times of the order of minutes. In fine-weather conditions, particularly with an inversion near the surface, so that there is no convection, the potential gradient is remarkably steady, with no more than 2 per cent change per minute.

Under less stable conditions, such as those which give rise to fine-weather cumulus clouds, or in cloudless weather with appreciable convection, the potential-gradient changes are greater, reaching 15 per cent change per minute. The potential gradient frequently shows cusp-like variations with a repetition rate which suggests, with the surface wind, a separation of aro und 2 km. Although cumulus clouds may be present, the cusps show no direct correlation with the clouds. It is suggested that the cusps may be related to convection cells, the upward moving part of the cell having an excess of positive charge from low levels; the sizes reported for convection cells are of the right order to agree with this theory.

BENT and HUTCHINSON (1966) found, by direct measurement of space charge, similar effects to those of WHITLOCK and CHALMERS and confirmed the identification of these effects with convection cells by finding, at the same time as the space charge reached a maximum, maxima also in absolute humidity and in temperature and a less well marked minimum in horizontal wind speed; they found that the effects extend to at least as high as 19 m.

ISRAËL (1959a) found that the amount of the "agitation" is greater the greater is the actual value of the potential gradient. BRASEFIELD (1958) found a similar relation between potential gradient and the frequency of its variation from space-charge effects.

6.16. Artificial Space Charges

The space charges produced by locomotives, etc. (see § 6.13.), can be classed as "artificial" and have already been discussed.

Deliberate attempts to create space charges have been made by VONNEGUT and MOORE (1958a), VONNEGUT *et al.* (1961) and VONNEGUT *et al.* (1962) and they have shown that the space charges so produced can be detected at a considerable distance from the source, and, by the aid of convection currents, can produce potentials much greater than those from which they originate.

BENT and HUTCHINSON (1966) reported two kinds of space charge produced, unintentionally, from man-made origins, but not so directly as those discussed in § 6.13. They found effects of negative space charges in mist which agreed exactly with the effects on potential gradients described by CHALMERS (1952a) and ascribable to breakdown of insulation at overhead high tension power transmission lines; in some cases the space charge had travelled 17 km into a region without mist. They also found effects of three different kinds which they were able to explain in terms of an electrode effect at an earthed metal lattice mast.

CHAPTER 7

The Conductivity of the Air

7.1. Ionic Conductivity

If i_1 is the vertical conduction current per unit cross-sectional area carried by positive ions, and i_2 for negative ions, and if F is the vertical potential gradient, then

$$\lambda_1 = i_1/F \quad \text{and} \quad \lambda_2 = i_2/F$$

denote the specific polar conductivities of the two signs.

If n_1, n_2 are the numbers of positive and negative ions, with mobilities w_1, w_2 and charges $\pm e$, then

$$\lambda_1 = n_1 e w_1, \qquad \lambda_2 = n_2 e w_2.$$

If there are ions of different mobilities, we can generalize these to

$$\lambda_1 = \Sigma\, n_1 e w_1 \quad \text{and} \quad \lambda_2 = \Sigma\, n_2 e w_2.$$

Large ions have small mobilities and their numbers are seldom, if ever, great enough relative to those of the small ions for them to play any significant part in the conductivity; the same is probably also true for the intermediate ions.

7.2. Leakage from Conductor

The conductivity can be measured by the leakage from a conductor, using the result of § 2.27., that:

$$Q = Q_0 \exp\left(-\lambda t/\varepsilon_0\right),$$

where λ is the negative conductivity if the conductor is positively charged and vice versa. This result was first derived by RIECKE

(1903) for a spherical conductor, but was extended to apply to any shape by SWANN (1914a).

There are certain restrictions as to the conditions in which RIECKE's law is applicable; in the first place, there must not be a potential gradient sufficiently large to give conditions outside the range of OHM's law, or, in other words, the current discharging the conductor must not appreciably alter the ionic content of the air near the conductor. Next, the conductor must not be in a field large enough to give induced charges which would cause different parts of the conductor to carry charges of different signs, or the current to the conductor would involve both signs of ion. Then, the law is vitiated if space charges are present near the conductor, and these might arise in still air by the electrode effect (see § 2.31.).

To avoid the conditions of non-adherence to OHM's law and also the setting-up of the electrode effect, it can be arranged that there is a flow of air past the conductor; however, it is necessary to ensure that the lines of air flow are parallel to the surface of the conductor, or else there may be divergences from RIECKE's law.

To avoid the effects of induced charges, the conductor should be shielded from the earth's field; this can be done by placing it inside a shield and ensuring a flow of air past it, or the shield can consist of a wire cage which allows a free flow of air. In either case, some care must be taken to avoid any effect of induced charges on the shield giving discrimination among the ions of different sign entering the shield.

The earliest apparatus for the measurement of conductivity was that of ELSTER and GEITEL (1899), but this was unsatisfactory, because the conductor was enclosed, and OHM's law did not hold. More satisfactory methods have been devised by GERDIEN (1905b) with a cylindrical condenser, and by SCHERING (1906) with a cage. Recently, KASEMIR and RUHNKE (1958) used a stretched wire, at earth potential, in the open air, with a charged grid nearby to provide a field at the surface of the wire; they considered that wind currents should be sufficient to overcome the effects of the distortion of the natural field by the wire and grid.

7.3. The Cylindrical-condenser Method

GERDIEN (1905b) used a cylindrical condenser, as discussed in § 2.54., with a sufficiently large air velocity to ensure that OHM's law holds. He used an outer tube of length 56 cm and radius 8 cm,

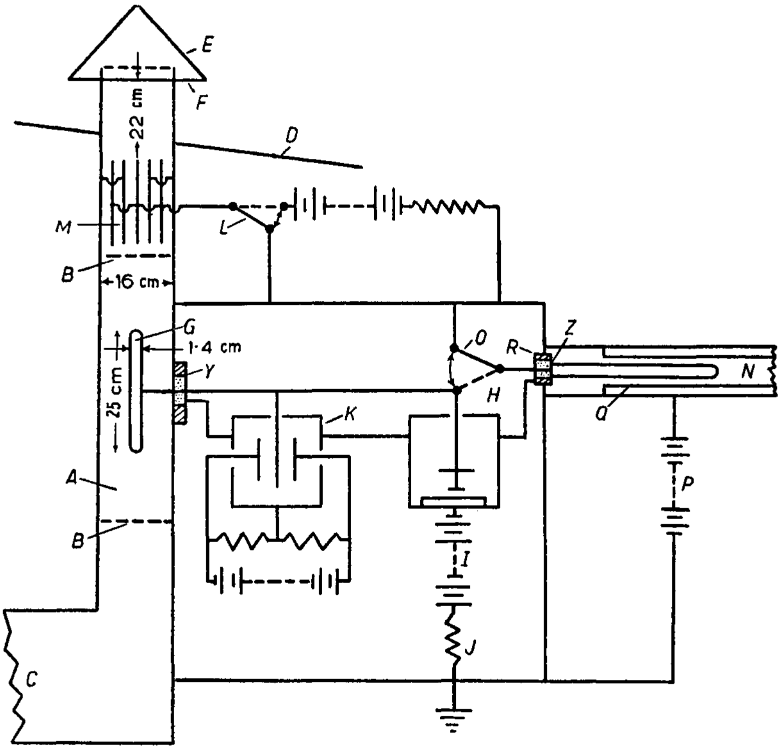


FIG. 34. Apparatus for measurement of conductivity. (From FLEMING, 1939, Fig. 13, p. 250.)
A: conductivity condenser, *B*: protecting screens, *C*: to fan, *D*: roof of building, *E*: cover, *F*: air inflow, *G*: central cylinder, *H*: high resistance radioactive cell, *I*: battery, *J*: resistance, *K*: electrometer, *L*: switch for zero determination, *M*: condenser for zero determination, *N*: condenser for calibration, *O*: switch for calibration, *P*: battery for calibration, *Q*: movable part of calibrating condenser, *R*: *Y*= guard rings, *Z*: insulator

with an inner electrode of length 25 cm and radius 0·75 cm. A potential difference of 100 V was applied between the electrodes, and the rate of air flow was made greater than the value L/t given by § 2.54.

The theory can be discussed either in relation to the results in § 2.54. or in relation to RIECKE's formula. From § 2.54., the number of ions which reach the central cylinder in unit time is the number which enter within a radius R, where

$$u(R^2 - b^2) \log_e a/b = 2w_1 VL.$$

Now if there are n_1 ions per unit volume, the number entering within this radius in unit time is

$$u\pi(R^2 - b^2)\,n_1,$$

which is, from above,

$$\frac{\pi n_1 2 w_1 VL}{\log_e a/b},$$

Thus the charge arriving at the central cylinder in unit time is

$$\frac{2\pi LVn_1\,ew_1}{\log_e a/b} = \frac{2\pi LV\lambda_1}{\log_e a/b}.$$

Alternatively, from RIECKE's formula, the current is $dQ/dt = -Q\lambda/\varepsilon_0$.

If V is the potential difference between the electrodes and C the capacitance, $Q = CV$, so that the current is:

$$CV\lambda/\varepsilon_0.$$

But the capacitance of a cylindrical condenser is $2\pi\varepsilon_0 L/(\log_e a/b)$, so that the same result is reached.

In practice, some ions are collected by the supports of the central electrode and also the cylindrical condenser, not being of infinite length, has a capacitance differing from its theoretical value. It is not easy to calculate the true capacitance, but SMITH (1953) devised an experimental method for its determination and if the value so obtained is used in $CV\lambda/\varepsilon_0$, the divergence from the theoretical formula is covered.

In GERDIEN's original method, an electrometer was connected to the central electrode, which was charged to a potential V_1; after a time t, the electrometer registered a potential V_2, so that, from RIECKE's formula $Q_2/Q_1 = V_2/V_1 = \exp(-\lambda t/\varepsilon_0)$, or

$$\lambda = \varepsilon_0/t \cdot \log_e V_1/V_2.$$

NOLAN (1940) adapted the apparatus for use as a null method, by altering the potential of the outer electrode in such a way as to maintain the central electrode at a constant potential, the change in potential produced by the charge arriving being compensated for by a change in the potential difference between the electrodes. The formula given above holds more accurately than before because there are no effects of stray capacitances. Also there is no need for calibration of an electrometer, since the ratio V_1/V_2 is that of two resistances on a potential divider.

Modern practice uses a constant potential difference between the electrodes and measures the current electronically, so that the capacitance is required. Adaptations of the apparatus for use with a radio-sonde in a balloon have been made by CORONITI *et al.* (1954), KOENIGSFELD (1955a), VENKITESHWARAN, GUPTA and HUDDAR (1953), LUGEON and BOHNENBLUST (1957), HATAKEYAMA *et al.* (1958) and JONES, MADDEVER and SANDERS (1959a). A design for use in aircraft was described by CORONITI (1960).

7.4. Wire-cage Method

SCHERING (1906) set up a charged conductor inside an earthed wire cage, and measured the rate of loss of charge. The use of the cage prevented the conductor from carrying any induced charge, and it was hoped that the normal air currents would cause the air near the conductor to be a fair sample of the natural air. It would therefore be necessary to avoid any effects of charges induced by the external fields on the wire cage, since such charges would alter the ionic composition of the air which enters the cage; SCHERING attempted to do this by setting up the cage under trees or under a roof or inside a larger cage, but the charges induced on these outer enclosures might have a similar effect to that which it was hoped to eliminate.

7.5. The Nolans' Method

NOLAN and NOLAN (1937), instead of measuring conductivity directly, measured n_1 and w_1 separately and so obtained λ_1 from $\lambda_1 = n_1 e w_1$. For the measurement of the mobility they used what amounts to a combination of the EBERT and GERDIEN methods for obtaining n_1 and λ_1. For this, two large identical cylinders were used, in one of which there was applied a sufficiently large field to collect all the ions of mobility above a certain value, thus giving $n_1 e$. In the second cylinder, smaller potentials were applied in succession, and the slope of the curve relating current and potential gave $\lambda_1 = n_1 e w_1$. By division, w_1 was obtained, the variable n_1 being eliminated. This method gives a kind of mean mobility suitable for use in the conductivity measurements, and more reliable than values of the mobility obtained otherwise. The value of $n_1 e$ was measured separately by an ion counter of the type described in § 4.11.

The advantage of this rather roundabout method over GERDIEN's seems to be that n_1 can be measured more accurately than can λ_1, and w_1 can be obtained in conditions similar to those in which n_1 is measured. However, in later work (NOLAN, 1940) the direct measurement of conductivity by an adaptation of GERDIEN's method was used (see § 7.3.).

7.6. Measurements Close to the Ground

HOGG (1939b) made measurements of the conductivity due to ions of each sign at various heights close to the ground, using an apparatus similar to GERDIEN's. If the tube is at earth potential, it will carry an induced charge which will tend to affect the distribution of ions in its neighbourhood; to avoid this, HOGG used a tube of cardboard, surrounded at its entry point by a metal ring whose potential was adjusted to be that of the air in its neighbourhood, found from potential-gradient measurements just before. The air after passing through the tube was then drawn into the conductivity apparatus.

HIGAZI and CHALMERS (1966) used similar apparatus, with the potential of the ring adjusted by means of a field mill continuously. In order to avoid loss of ions when the air stream went round a corner the cylindrical condenser was arranged vertically, immediately below the cardboard tube. Simultaneous measurements were made with two sets of apparatus, either measuring the conductivities for both signs, or else conductivities for the same sign at two heights.

HOGG's measurements were made in conditions of light wind, and he found that the conductivity due to positive ions was greatest close to the ground, falling to about half this value at 1 m and higher. At the same time, the negative-ion conductivity was very small close to the ground and rose to a value about equal to that for positive ions at 1 m. The total conductivity remained very nearly constant over the whole range of height concerned.

HIGAZI and CHALMERS found results agreeing with HOGG's only when there was very little wind, or when the potential gradient was higher than normal. For moderate winds, they found a ratio of positive to negative conductivity greater than unity at ground level, but averaging only 1·43; both conductivities, and their ratio, fell on going upwards from the ground.

The importance of these results will be discussed in § 8.15.

7.7. Indirect Determination of Conductivity

Since the current i, the potential gradient F, and the conductivity λ are related by the equation $i = \lambda F$, then λ can be obtained indirectly if i and F are measured. A direct measurement of i has been possible as yet only at the earth's surface and so the value of λ obtained has been that at the earth's surface; in a positive potential gradient this is, necessarily, due to positive ions alone unless negative ions emerge from the ground.

The measurements of NOLAN and NOLAN (1937), HOGG (1939b) and NOLAN (1940) all showed that λ measured indirectly at the ground was nearly the same as λ measured directly higher up, and the measurements of HOGG also showed that λ for positive ions gave the same results at the ground when measured directly and indirectly.

7.8. Measurements under Trees

O'DONNELL (1952) made measurements of conductivity at different levels, under conditions at first sight similar to those of HOGG (1939b) but found very different results. He did not find that the total conductivity at 1 m was equal to the positive conductivity close to the ground, nor did he find variations of the conductivities with height similar to those of HOGG.

CHALMERS (1953a) pointed out that the discrepancy is caused by the fact that HOGG made his measurements in the free air with no obstructions, whereas O'DONNELL, to avoid effects from induced charges, always set his apparatus up under the branches of large trees. Thus in O'DONNELL's measurements there was no potential gradient present and hence no reason to expect the absence of negative conductivity at ground level; the variation of conductivity with height is directly related to the presence of the potential gradient.

7.9. Cause of Variation of Conductivity with Height Close to the Surface

If the positive-ion conductivity close to the earth's surface is about twice that at one metre in calm conditions, and the potential gradient and mobility are the same, this must be caused by a greater number of positive ions at the lower level. HOGG (1939b) suggested that the reason for this might be a variation of the rate of production of ions with height, and he made some measurements to test this conclusion. CHALMERS (1946) showed that a suitable form of the variation of the rate of production of ions with height could lead to a quasi-static state in which the usual ionization equilibrium is modified by considering different currents entering and leaving the volume under consideration, and could give a small total space charge, with little alteration of potential gradient with height. HESS and O'DONNELL (1951) made measurements which confirm this idea and ascribed the effects to radioactivity in the ground and in the air (see § 8.20.), and PIERCE (1958) showed that the form of variation would be not far from that assumed by CHALMERS.

7.10. Variations of Conductivity

The fine-weather conductivity at many places shows a maximum in the early morning hours, with a fall soon after sunrise; this is probably accounted for by the formation of mist or by the increased pollution of the air.

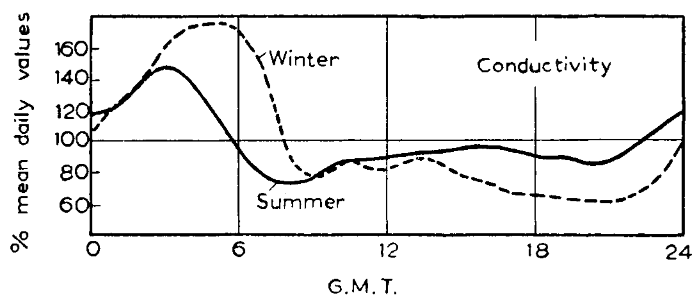


FIG. 35. Diurnal variation of conductivity at Kew. (From SCRASE, 1933, Fig. 6 (opp. p. 11) ("conductivity", "summer and "winter").)

As is to be expected from the discussion of the equilibrium of ionization and the presence of nuclei, any increase in the number of nuclei brings about an increase in the number of large ions and a corresponding decrease in the number of small ions, thus decreasing the conductivity. So we should expect, as is indeed always found, a minimum of conductivity in the winter, in places where pollution occurs.

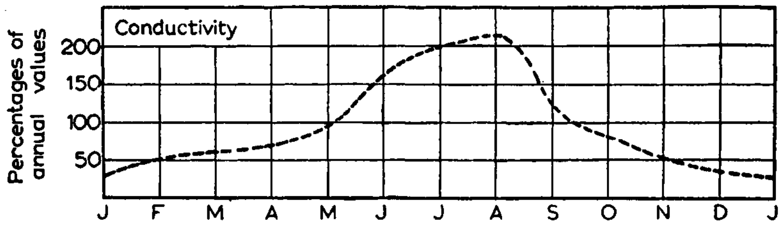


FIG. 36. Annual variation of conductivity at Kew. (From SCRASE, 1933, Fig. 4 ("conductivity" portions), p. 11.)

GOCKEL (1917) found a lowering of conductivity in mists and damp air; and many observers have found that the Föhn wind, a dry wind from high mountains, brings an increase of conductivity. MARKGRAF (1924) found that the conductivity is greater in a region of low pressure than in one of high. The effect of storms seems to depend on the locality; GOCKEL (1917) found an increase in conductivity in the Alps even before the arrival of a thunderstorm, but other observers have found no such effect of storms, and STARR (quoted by WHIPPLE and SCRASE, 1936) found no abnormality of conductivity due to thunderstorms except for slight changes associated with lightning flashes and lasting for a few seconds only. The question may be a local matter, depending on the height of the observer relative to the source of ions, e.g. point discharge; an observer high up might find an increase of conductivity caused by ions rising from below, while no similar effect would be found by an observer at lower levels.

Observations of conductivity at sea, where there are no local influences, show that there is a decrease in equatorial latitudes, which can be associated with the decrease in the intensity of cosmic radiation.

WRIGHT (1933) made measurements of conductivity and related them to the nuclei and smoke particles present. SCRASE (1935a) recalculated WRIGHT's results, using a correction for the distortion of the field, and also provided fresh measurements; he found that the influence of smoke particles on the conductivity is less than WRIGHT believed, so that it is justifiable to neglect the smoke particles in discussing ionization equilibrium.

7.11. Conductivity of Fogs and Clouds

In fogs, the conductivity is reduced by reason of the attachment of small ions to the fog particles, rendering the ions much less mobile. Measurements by ISRAËL and KASEMIR (1952) of the potential gradient and conduction current during widespread fog showed a considerable increase in potential gradient and a slight decrease in conduction current as compared with clear air, giving on the average a conductivity about one-third of its value in fine weather.

Mountain fogs give different results, since they are comparatively localized and really represent the interiors of non-raining clouds. ISRAËL and KASEMIR (1952) found a considerable decrease in vertical current but little change in the potential gradient, with the conductivity again having about one-third of its fair-weather value. KRASNOGORSKAYA (1961) found a larger change in λ_+ than in λ_- during the development of clouds on a mountain.

The difference between the two sets of results can be explained by the smaller extent of the mountain fogs, so the current can, so to say, find its way round the poorly conducting region, while the potential drop is little altered. On the other hand, for widespread fog, there is no way round and the whole current has to go through the fog, giving a greater potential gradient because of the reduced conductivity.

Attempts to measure the conductivity of a raining cloud have not been successful, because of shattering and splashing of the raindrops when air is drawn into the conductivity apparatus.

FREIER (1962) attempted to deduce the conductivity within a thunder cloud by a discussion of currents but CHALMERS (1964) pointed out that point-discharge currents could upset this argument (see § 12.14.).

7.12. Changes of Conductivity Before Onset and Dissipation of Fog

SERBU and TRENT (1958), at Argentia in Newfoundland, found that the decrease in conductivity and consequent increase in potential gradient became evident 1 to 2 hr before the onset of fog, and the reverse changes appeared ½ to 1½ hr before the dissipation of the fog. The same facts have been observed by a number of other workers in various places and a very thorough discussion has been given by DOLEZALEK (1962) who stated that the phenomena, while not 100 per cent reliable for forecasting, gave around 80 per cent reliability. It appears that the effect is more noticeable when there are few nuclei present.

DOLEZALEK (1962) discussed in detail the possible explanations, but did not find any one satisfactory, and he proposed further observations to elucidate the matter.

7.13. Balloon Measurements

Measurements of the variation of conductivity with height in balloon ascents have been made by WIGAND (1914), GISH and

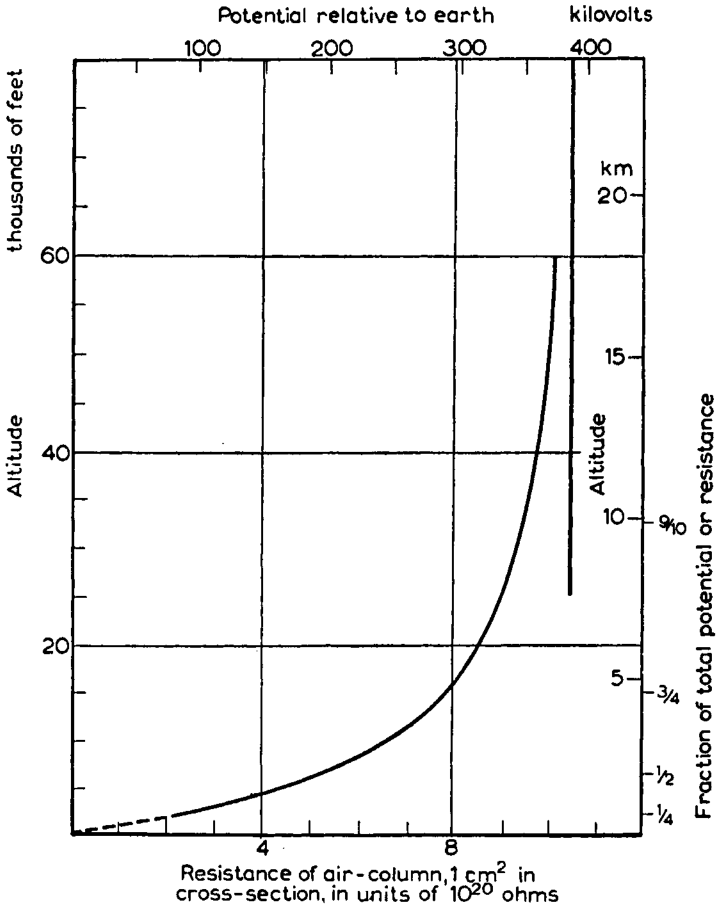


FIG. 37. Potential and resistance between the earth and points in the atmosphere. (From FLEMING, 1939, Fig. 10, p. 209.)

SHERMAN (1936) in *Explorer II*, STERGIS *et al.* (1955), KOENIGSFELD (1955a) and others; in the last two cases, radio-sonde methods were used to obtain the information. The results are more easily described in terms of resistance, rather than of conductivity, and it is convenient to use the term "columnar resistance" for the resistance of a column of air of unit cross-section. Figure 37 was obtained by GISH and SHERMAN (1936) from their conductivity results by integrating the columnar resistance up to different levels. Values for the potentials where obtained by assuming a constant vertical current density of 4×10^{-12} A/m². The results of STERGIS *et al.* agree in general very well with those of GISH and SHERMAN; they show greater regularity in the descent than in the ascent, which may be because in ascent the apparatus, below the balloon, is moving into air which may have been contaminated somewhat by the balloon, while the same is not true in descent.

STERGIS, REIN and KANGAS (1957a) carried out similar measurements with the apparatus of STERGIS *et al.* (1955) and found conclusively that there is no difference between the conductivity above a thunderstorm and that at the same level in fine weather.

WOESSNER, COBB and GUNN (1958) also used a balloon with radio-sonde methods to give positive and negative conductivities simultaneously. They found that their results could be represented by (converting to M.K.S. units):

$$\lambda_+ = 2{\cdot}70 \times 10^{-14} \exp(0{\cdot}254z - 0{\cdot}00309\, z^2)\ \text{mho/m}$$

and

$$\lambda_- = 4{\cdot}33 \times 10^{-14} \exp(0{\cdot}222\, z - 0{\cdot}00255\, z^2)\ \text{mho/m}$$

with z in km. The ratio of positive to negative conductivity was found to be greater than unity in the lower atmosphere, equal to unity at some height between 3 and 7 km, less than unity up to 18 km and then again about unity above that level.

GISH and SHERMAN (1936) found irregularities and a general decrease of conductivity at heights from 19 to 22 km but STERGIS *et al.* (1955) and WOESSNER, COBB and GUNN (1958) found the conductivity at these heights to follow the same increasing trend as at lower levels.

If it is assumed that ionization at great heights is due entirely to cosmic rays and also that large ions and nuclei can be neglected, so that small ions are removed only by recombination with other small ions, then it is possible to calculate, for any height, the equilibrium number of small ions and hence the conductivity, using the known variations with height of cosmic-ray intensity, recombination coefficient and mobility. GISH and SHERMAN (1936) showed that the calculated values agreed with those measured from a few km up to 18 km, and STERGIS *et al.* (1955) have made similar, more elaborate, calculations.

GISH (1944b) gave a formula derived from the *Explorer II* and other results which is (converted to M.K.S. units):

$$\gamma = 2.94 \times 10^{13} \exp(-0{\cdot}452z) + 1{\cdot}39 \times 10^{13} \exp(-0{\cdot}375z) + 0{\cdot}369 \times 10^{13} \exp(-0{\cdot}121z),$$

where γ is the resistivity in Ωm and z is the height in km.

PALTRIDGE (1965) made measurements up to 26 km with an ascending balloon and at 30·5 km with a constant-level balloon. His results showed that aerodynamic conditions were not such as to give agreement with theory and a correction was needed. Even so, the conductivity was less than expected by a factor of 3 at 30·5 km. The positive and negative conductivities were nearly equal from about 8 km.

7.14. Measurements in Aircraft

GISH and WAIT (1950) made measurements of conductivity in an aircraft, mainly flying over thunderstorms. They used a method similar to that of GERDIEN but with a direct-current amplifier in place of the electrometer; the inlet tube was mounted in the nose of the aircraft. Their results for the variation of conductivity with height agreed fairly well with the balloon measurements (§ 7.13.) and were fitted to a formula $\lambda = \lambda_0 + Az^2$. Expressing the conductivity λ in mho/m and the height z in km, their results gave for positive ions

$$\lambda_0 = 1{\cdot}7 \times 10^{-14}, \quad A = 0{\cdot}21 \times 10^{-14},$$

and for negative ions

$$\lambda_0 = 1{\cdot}1 \times 10^{-14}, \quad A = 0{\cdot}23 \times 10^{-14}.$$

For comparison, the balloon results of GISH and SHERMAN (1936), when fitted to the same formula, gave for positive ions

$$\lambda_0 = 1{\cdot}1 \times 10^{-14}, \quad A = 0{\cdot}28 \times 10^{-14}.$$

Although the measurements were made mainly over thunderstorms, there were no abnormal values to be attributed to the storms.

A more extensive set of measurements with similar apparatus was made by CALLAHAN *et al.* (1951), not confined to flying over thunderstorms. They also found results in general agreement with the balloon results and showed that, at heights above a few km, the conductivity is independent of time, season or place, except when haze is present (no flights were made in clouds) and agrees with that to be expected if ionization is produced by cosmic rays alone, with large ions and nuclei negligible. They found no significant difference between the conductivities due to positive and negative ions.

KRAAKEVIK (1958) made measurements of conductivities of both signs in an aircraft over Greenland, going up to 6 km, and found that the conductivity above 0·2 km could be represented by

$$\lambda = 2{\cdot}72 \times 10^{-14} \exp (z - 6)/3{\cdot}5 \text{ mho/m},$$

where z is in km. He found the ratio of the positive to negative conductivities to vary from 1·11 to 0·78 with an average value of 0·90.

CORONITI *et al.* (1952) investigated the effect of the charge of the aircraft on the measurement of conductivity, both in flight and under laboratory conditions. Their results may account for the anomalous values of negative conductivity found by GISH and WAIT (1950) at low altitudes.

ROSSMANN (1950) made measurements in a towed glider and found results similar to those in aircraft, but with abnormally high values of conductivity under special conditions. CORONITI and HEATON (1953) made measurements in an aircraft at various levels up to 6·4 km over the Pacific Ocean and found values of conductivity agreeing with those expected from cosmic-ray data.

SAGALYN and FAUCHER (1954) measured conductivity as well as temperature, humidity and large-ion content in the range 700 to 15,000 ft (about $\frac{1}{5}$ km to 5 km) and showed that there is a very marked change in all these parameters on passing through the upper boundary of the *austausch* region (see § 2.32.), the positive conductivity changing by a factor between 1·5 to 6·0; the level at which this occurs may be from 1000 to 10,000 ft (0·3 km to 3 km) and varies according to meteorological conditions. These measurements are in agreement with the potential-gradient measurements of LECOLAZET (1948a) (see § 5.38.). Within the *austausch* region, the conductivity varies considerably both in time and in place.

The measurements of CORONITI (1961) show how the conductivity depends on the presence of nuclei, as would be expected, since λ depends on n and n depends on N.

7.15. Rocket Measurements

Measurements of the conductivity at heights from 35 km to 80 km were made by BOURDEAU, WHIPPLE and CLARK (1959), with conductivity apparatus mounted in the nose of a "Viking" rocket, the results being telemetered down. Above 80 km, the apparatus would no longer function correctly, as the mean free path becomes of the same order as the dimensions of the apparatus. In considering the results it was necessary to take account of various possible disturbing effects, such as that of a charge on the rocket and those of shock waves.

The conductivity to be expected can be calculated from known values of the variations of cosmic-ray ionization, mobility and recombination coefficient with pressure and temperature, and so with height.

The results showed agreement with expected values up to about 50 km, but at greater heights the conductivity was less than expected, particularly for positive ions. Additional sources of ionization, such as collisions or photoelectric effects would give an increased conductivity and it was suggested that the decrease in conductivity could be ascribed to particulate matter such as is needed to account for noctilucent clouds (LUDLAM, 1957).

7.16. Conductivity over the Sea

SAGALYN (1958a) measured conductivity and ion and nucleus content over the Atlantic Ocean at various heights and at various distances from land. She found that the *austausch* region persists over the ocean and at distances of over a few hundred miles from the land, in fine weather, this region is very stable, with little change of thickness with time of day, and with thorough mixing of the air within the region. At the upper boundary of the *austausch* region, the conductivity increases suddenly by a factor of around 4; above the *austausch* region, the conductivity is closely the same over the sea as over the land. The nuclei present within the *austausch* region at large distances out to sea appear to have larger radii than those over the land and a smaller concentration, the factors in both cases being about 4.

7.17. The Columnar Resistance

The columnar resistance is defined as the resistance of a column of air of unit cross-section, 1 m^2 in the M.K.S. system, from the earth up to the electrosphere; bearing in mind the unit area, the columnar resistance can be expressed in Ω.

The columnar resistance can be obtained up to various heights from the results of conductivity measurements. From the *Explorer II* results, GISH and SHERMAN (1936) obtained a value of $10^{17}\,\Omega$ up to 18 km. If the conductivity continues to rise as expected from cosmic-ray data, then the total columnar resistance up to the electrosphere is less than 10 per cent greater than that up to 18 km. By the comparison of the vertical currents at places where there are large and small resistances in the lowest layers, it can be shown that the column above 18 km cannot have a resistance as much as 50 per cent of that below.

Using GISH's (1944b) formula (see § 7.13.) ISRAËL (1953a) found a columnar resistance of $4 \times 10^{16}\,\Omega$ up to 3 km, $1 \times 10^{16}\,\Omega$ from 3 km to 7 km and $2{\cdot}4 \times 10^{16}\,\Omega$ above 7 km, making a total $7{\cdot}4 \times 10^{16}\,\Omega$.

KRAAKEVIK (1958) over Greenland, found the columnar resistance up to 6 km to be $6{\cdot}0 \times 10^{16}\,\Omega$ and suggested a total columnar resistance of $8{\cdot}0 \times 10^{16}\,\Omega$.

ISRAËL (1949) was able to determine how the columnar resistance varies with time of day for a number of stations (see § 5.45.).

SAGALYN and FAUCHER (1954) pointed out that the *austausch* region contributes from 40 to 73 per cent of the total columnar resistance, and as this percentage varies from day to day, and within a single day, as well as from place to place, any attempt to calculate a columnar resistance from the conductivity measurements is bound to give irregular results.

If a columnar resistance of $10^{17}\,\Omega$ assumed to hold for the whole of the earth's surface, the total resistance of the atmosphere amounts to about $200\,\Omega$; with KRAAKEVIK's value, it is $160\,\Omega$.

7.18. Ratio of Polar Conductivities

Various measurements have been made of the ratio of negative to positive polar conductivities in the atmosphere; these can be divided into those in pure air, with very few nuclei present, and those in polluted air, where nuclei are important.

Measurements at high altitudes were first made by GISH and SHERMAN (1936) and gave a value of about 1·28, but more recent results have all given a lower value, e.g. CALLAHAN *et al.* (1951), KRAAKEVIK (1958a), SAGALYN (1958a) and CURTIS and HYLAND (1958), and PALTRIDGE (1965), giving values of around 1·06; SAGALYN (1958b) has pointed out that the value of 1·28 given by GISH and SHERMAN is not a direct measurement. In their measurements in a 3000 m^3 spherical chamber (see § 4.15.), PHILLIPS *et al.* (1955) found in pure air that the ratio of the conductivities is equal to the ratio of the mobilities, as would be expected with equal numbers of ions of both signs. At atmospheric pressure this value is closer to 1·28 than to 1·06, so SAGALYN (1958b) suggested that the value of 1·06 means that the ratio of the mobilities of the ions at high levels is closer to unity than near the ground; but the fact that the conductivity increases with height leads to a positive space charge (see § 2.26.) and this would reduce the ratio below the ratio of the mobilities.

At lower levels, where pollution is important, the ratio is found to be close to unity, although, very close to the ground, this is not true owing to effects related to the electrode effect (see § 7.9.). PHILLIPS *et al.* (1955) in the 3000 m^3 chamber found that the two conductivities were nearly equal when nuclei were present. This has been explained by SAGALYN (1958b) as follows: as discussed in § 4.28.:

$$\frac{n_2}{n_1} = \left(\frac{N_2}{N_1}\right)^{\frac{1}{2}} \left(\frac{\eta_{12}\,\eta_{10}}{\eta_{20}\,\eta_{21}}\right).$$

From the theory of BRICARD (1949), this gives

$$\frac{n_2}{n_1} = \left(\frac{N_2}{N_1}\right)^{\frac{1}{2}} \frac{D_1}{D_2},$$

and from EINSTEIN's (1905) thory (see § 2.14.)

$$\frac{n_2}{n_1} = \left(\frac{N_2}{N_1}\right)^{\frac{1}{2}} \frac{w_1}{w_2}.$$

But $\lambda = new$, hence

$$\frac{\lambda_2}{\lambda_1} = \left(\frac{N_2}{N_1}\right)^{\frac{1}{2}},$$

which is usually close to unity.

VOGLER (1959) has pointed out that the deposit of radioactive substances on the measuring electrodes will alter the apparent ratio of the polar conductivities because of the emission of α-particles.

7.19. Calculations of Conductivity and Potential Gradient

LECOLAZET and PLUVINAGE (1948) and LECOLAZET (1948b) made calculations on values of potential gradient to be expected at various points in a system which consists of a lower volume of conductivitiy λ, an upper volume of conductivity $10\,\lambda$, separated from the lower by a horizontal plane, and a sphere of conductivity $\lambda/3$ with a diametral plane in the plane of separation of the other two volumes. This is a representation of the system of a cumulus cloud at the top of the *austausch* region, as investigated with glider measurements by LECOLAZET (1948a) (see § 5.39.). LECOLAZET (1948b) showed that, except within a few metres of the boundaries, the results for the air, with ionic conduction, are exactly the same as for a system of three metallic conductors with the same relative conductivities and boundaries. On this basis, LECOLAZET and PLUVINAGE (1948) calculated the values of potential gradient to be expected, and found that the agreement with the experimental measurements of LECOLAZET (1948a) were as good as could be hoped for. Essentially these calculations amount to a detailed working out of what we have called the "traffic-jam effect" (see § 3.4.) and give quasi-static space charges at the boundaries between the media of different conductivities.

7.20. Conductivity in Clouds

Direct measurement of the conductivity of clouds is difficult by the ordinary methods, though PLUVINAGE (1946) has carried some measurements in mountain fogs at the Puy-de-Dome. Indirect determinations have been made for non-raining clouds by considering their effects on the potential gradient and on vertical current; two quite independent estimates by LECOLAZET and PLUVINAGE (1948) and by ISRAËL and KASEMIR (1952) suggest that the conductivity inside such a cloud is about one-third of the normal conductivity near the earth's surface. It is clear that such methods are not applicable to clouds which are giving precipitation or even to clouds, if any, in which there may be a process of charge separation without precipitation.

In ordinary air the conductivity is due almost entirely to the small ions and the large ions have their effect by reducing the number of small ions by combination. In a cloud, the cloud droplets also play a part, removing some of the small ions. PLUVINAGE (1946) calculated that the conductivity of a cloud, in the absence of appreciable external fields, should be given by: $qew/2\pi NaD$, where q is the rate of production of ions, e the electronic charge, w the ionic mobility, D the ionic diffusion coefficient, and N the number of droplets, of average radius a, per unit volume. PLUVINAGE's measurements give good agreement with the theory. The number of condensation nuclei and large ions in a cloud was shown by PLUVINAGE and ROCHÉ (1947) to be less than in the free air.

Conductivity in thunder clouds has been deduced by FREIER (1962) from measurements of recovery after lightning flashes (see § 12.14.) but CHALMERS (1964) criticized the argument.

7.21. Conductivity Measurements on Mountains

SCHILLING (1955) measured the conductivity of the air at various levels on mountains up to 4000 m and found surprisingly good agreement with aeroplane and balloon measurements at the same heights in the free air.

On the other hand, SAGALYN and FAUCHER (1954) found that, in summer, the upper limit of the *austausch* region over mountains is raised above that over nearby plains, though the whole layer is not so thick over mountains; thus, at a given height, the mountain measurements may be within the *austausch* region and would then give much lower results for conductivity than at the same height in the free air which is above the *austausch* region. In winter, however, the *austausch* region may not reach the top of the mountain and then the conductivity would be the same on the mountain as in the free air; SCHILLING's measurements must have been made under "winter" rather than "summer" conditions. Exactly similar conclusions as to the difference between summer and winter effects have been obtained from potential-gradient measurements (see § 5.50.).

PLUVINAGE and STAHL (1953) measured the conductivity in Greenland and found that the positive conductivity is often much greater than the negative, the factor even reaching 10 at a height of 1·5 m above the earth's surface in the centre of Greenland, where there is an ice-cap about 3000 m thick. They explained their results in terms of the "electrode effect" (§ 2.31.). The ice-cap presumably cuts off all effects of radioactivity in the earth's crust, so that there is not the variation of ionization with height that usually occurs. It may then be that the electrode effect is undisturbed by other factors and becomes fully evident (see § 8.20.). Similar results were found by RUHNKE (1962), who, however, found an increase of negative conductivity near the ground when the wind speed increased and was not able to give any satisfactory explanation.

7.22. Effect of Nuclear Explosions on Conductivity

HARRIS (1955) investigated the effects on the atmospheric conductivity produced by radioactive fall-out after nuclear explosions. His conclusion was that there are β-ray emitters deposited on the surface of the earth and these produce increased ionization and conductivity in the lowest few metres of the atmosphere, at those places where there is large fall-out; but changes in conductivity of similar and greater magnitude can occur from quite other causes. The corresponding effect on potential gradient was found to be small but detectable in favourable cases.

7.23. Conductivity inside Buildings

SCHILLING and CARSON (1953) reported the remarkable fact that there are simultaneous changes in the electrical conductivity inside buildings and outside; they found no definite explanation of the result, but suggested infiltration of air through cracks, etc., and this was verified by SCHILLING and HOLZER (1954). Similar results for ion counting had been found by KÄHLER (1934).

7.24. Conductivity and Physiological Phenomena

After their discovery that conductivity changes inside buildings are simultaneous with those outside, SCHILLING and CARSON (1953) suggested that this might be the factor relating meteorological and physiological phenomena (see § 2.37.). This was further discussed by SCHILLING and HOLZER (1954) who showed that it was a possible hypothesis, which needs more extensive study.

Earlier work by YAGLOU, BENJAMIN and CHOATE (1931), YAGLOU and BENJAMIN (1934) and WAIT and TORRESON (1935) had investigated the effect of the number of ions on physiological phenomena without reaching any definite conclusions.

CHAPTER 8

The Air–Earth Conduction Current

8.1. The Fair-weather Conduction Current

In this chapter we shall be dealing almost entirely with conditions in fine or fair weather, in which it can be assumed that no processes of charge separation are taking place in the atmosphere, and in which it can be assumed that the electrical phenomena are reasonably steady, so that the principle of the quasi-static state (§ 2.23.) can be used. The currents concerned are all vertical currents and are measured in A/m^2; strictly, therefore, the description should be "current density" rather than "current". A current carrying positive charge down is considered a positive current (see § 2.9.), and so there is a positive conduction current in a positive potential gradient.

The conduction current, i, in the atmosphere is related to the local potential gradient, F, and the local conductivity, λ, by $i = \lambda F$; this holds at or close to the ground, so that $i_0 = \lambda_0 F_0$, the suffix denoting values at the ground.

If there is no current other than the conduction current, then, under quasi-static conditions, this must be the same at all levels from the ground to the electrosphere and thus $i = V/R$, where V is the potential of the electrosphere and R the columnar resistance (see § 7.17.). There has tended to be a neglect of the proviso that there are no other currents and $i = V/R$ has been considered to be more universally true than is actually the case. KRAAKEVIK and CLARK (1958) pointed out the importance of convection currents, particularly in regions of high pollution, and this matter will be discussed in more detail in § 11.2.

In conditions where convection currents can be neglected, the formula $i = V/R$ holds and hence $F_0 = V/\lambda_0 R$. It follows that local effects, which influence λ_0 more than R, will be of more importance in measurements of F_0 than of i, and hence that for the determination of the way in which V varies with time it is better to use measurements of i than of F_0. In agreement with this, it is found that, where there are effects of local pollution, i shows less variation with time of day than does F_0.

8.2. Methods of Measurement

The air–earth current can be measured by two distinct methods and it is necessary to consider whether these two methods measure the same quantity, a matter which can be dealt with both theoretically and experimentally.

In the first place, it is possible to measure the actual charge reaching an isolated portion of the earth's surface in a given time. Provided that the area concerned is a fair sample of the earth's surface, this method certainly gives a direct measure of the actual air–earth current, but it is not easy to discuss how far it is a conduction current and how far a convection current. If there is a space charge in the lowest centimetres of the atmosphere, then any process of diffusion, particularly eddy diffusion, will bring this charge into contact with the ground and presumably some of it will remain; at the same time it is possible that some of the "bound" charge on the surface might be carried away by the diffusing air.

Alternatively, it is possible to measure, independently, the local potential gradient and the local conductivity, the product giving the air–earth conduction current. These measurements can be made, without difficulty, at heights above the earth's surface, and so give values of the conduction current at different levels.

If the results for the current obtained by the two methods do not agree, then it can be inferred that the conduction current, measured by the indirect method, does not comprise the whole current. Since the assumptions of a quasi-static state and of horizontal stratification give a current density which is the same at all levels, any divergence of the results by the indirect method from constancy with height would lead to the conclusion that there must be some other current, presumably a convection current, also not constant with height. Conclusions based on similar measurements at different heights may be less liable to error than those based on the comparison of results from different measuring methods.

8.3. Displacement Currents

When the direct method is used to measure the air–earth current, there is, in addition to the true air–earth current, also an effect due to the change of potential gradient, giving what can be called a displacement current.

From § 2.8., unit area of the earth's surface carries a charge $Q = -\varepsilon_0 F$; if F alters, there will be a displacement current $dQ/dt = -\varepsilon_0 dF/dt$ and this is not directly distinguishable from a true conduction current.

To give an example of the magnitude of this effect, let us suppose that the potential gradient alters by 300 V/m in an hour. The apparent current is then $0{\cdot}74 \times 10^{-12}$ A/m^2, to be compared with true currents of from 1 to 4×10^{-12} A/m^2. Since the potential gradient often changes much more rapidly than this, it follows that the instantaneous measurement of the apparent current cannot be expected to give an accurate value for the true conduction current, and some method of eliminating the displacement current is required.

8.4. Direct Method of Measurement

For the direct method of measurement it is necessary to isolate a portion of the earth's surface and measure the charge reaching it in a given time. In order that conditions shall not be different from normal, it is necessary for the collecting surface to be placed in the plane of the earth's surface and to be kept at, or very close to, the potential of the earth, for otherwise the lines of force will be considerably distorted and it will not be the true air–earth current that is measured.

As described in the last section, there is a displacement current as well as the true current if the potential gradient is changing, and this has to be eliminated or taken into account.

The first attempt at a direct measurement of the air–earth current was that of EBERT (1902) whose plate was not in the plane of the earth, nor was it maintained at earth potential, so that his results could not be accurate.

8.5. Wilson's Measurements

WILSON (1906, 1916) used the Universal Portable Electrometer (§ 5.21.) and the Capillary Electrometer (§ 5.23.) to make measurements of the air–earth current. The effects of displacement currents are eliminated since the plate is covered at the beginning and end of each reading. In order to keep the plate at earth potential, and so to avoid distortion of the field, the compensating condenser is used with the Universal Portable Electrometer, while the Capillary Electrometer effects this automatically. GOTO (1957) (see § 8.11.) adapted WILSON's method to continuous recording.

SMITH (1959) found that if WILSON's method is used without the compensating condenser and with a measuring instrument of small capacitance, there appears to be no current. The explanation is as follows: on the removal of the cover, the plate acquires a charge $-Q = -A\varepsilon_0 F$ on its upper side, where A is the area and F the potential gradient. If the electroscope or other measuring instrument is of small capacitance, then the corresponding charge $+Q$ must reside on the lower surface of the plate. Now, if the conductivity of the air is the same above and below the plate, these two charges $-Q$ and $+Q$ will be dissipated at the same rate and there will be no apparent current.

8.6. Simpson's Measurements

SIMPSON (1910) used a collecting surface of 17 m^2, placed in the plane of the earth's surface, and connected it to the reservoir of a water-dropping system, arranged so that the drops were formed at a point where no field existed, inside an earthed copper cylinder; the effect of the dropper was to keep the collecting area at the potential of the earth. The drops thus carried away with them all the charge reaching the area, and this charge was measured by an electrometer; the deflection of the electrometer was registered every 2 min and thus the air–earth current could be measured continuously. In order to take account of the effect of potential-gradient changes, SIMPSON had a continuous record of the potential gradient and was able to correct for the changes.

CHALMERS and LITTLE (1947) collected the charge arriving at an area of 1 m^2 in 10 min and then discharged it through a ballistic galvanometer. A large condenser decreased the leak and also prevented the collecting surface from acquiring a potential much different from that of the earth. This method took no account of the effect of potential-gradient changes, so that individual readings have little value owing to the continual fluctuations of the potential gradient and hence of the bound charge, but such effects average out over longer times and an accurate result is obtained for a long-period average.

8.7. Scrase's Method

In order to be able to make continuous measurements of the air–earth current by the direct method, it is necessary either to eliminate or to compensate for the displacement current; SCRASE (1933) used a quadrant electrometer for compensation. In the normal use of a quadrant electrometer for the measurement of currents, one pair of quadrants is connected to the current collector while the other pair is earthed; in SCRASE's apparatus the second pair is not earthed but, instead, is connected through a condenser to a potential equalizer. If, now, both pairs of quadrants are first earthed and then released, one reaches a potential determined by the charge arriving at the current collector, including both the air–earth current and the displacement current; the other quadrant receives a charge depending on the change of charge on the con-

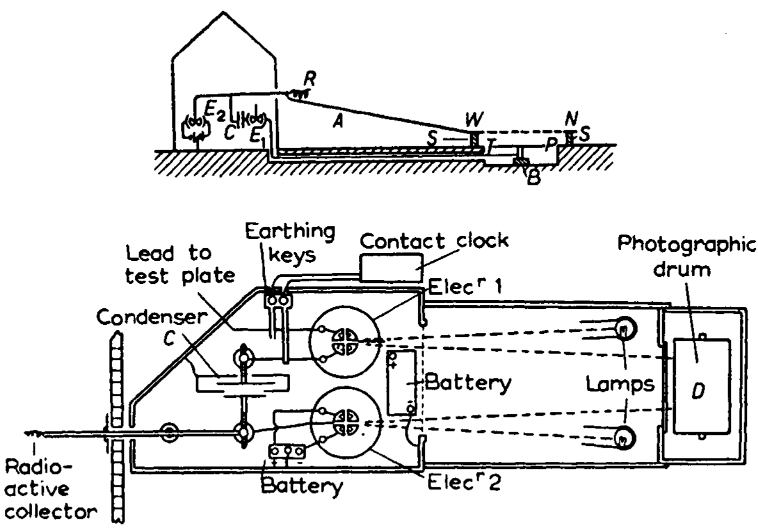


FIG. 38. Scrase's apparatus for air–earth current. (From SCRASE, 1933, Figs. 1 and 2, p. 4.)

denser and hence on the change of potential gradient. But the displacement current depends on the change of potential gradient, so a suitable adjustment of the capacitance of the condenser renders the effect of a potential-gradient change the same on both pairs of quadrants and thus the differential effect in the electrometer leaves just the air–earth current.

Unfortunately, SCRASE found that the polonium source that he used as a potential equalizer was too sluggish in its action for the method to be used as intended; a rapid change of potential gradient would affect the quadrants connected to the current collector immediately and those connected to the condenser only much more slowly. An increase in the strength of source would have had too great an effect on the conductivity of the air so SCRASE had to use another device.

He placed a wire net, connected to the potential equalizer, over the plate which acted as the current collector, at such a height that, in steady conditions, the potential gradient over the plate would not be affected by the presence of the net; in other words, the net lay along the same equipotential as contained the equalizer. Thus, instead of the plate being subjected to the natural displacement current, it is only subjected to the more sluggish changes as registered by the equalizer. In these conditions, the compensation for displacement current can be made exact. The capacitance of the collecting system was so large that its potential never differed much from that of the earth.

It can, perhaps, be argued that SCRASE's method does not give a true direct measurement of the air–earth current, and that what is really measured is the unipolar conductivity of the air between the wire net and the plate under conditions not far from the normal. The effect of connecting the wire net to the equalizer, rather than to any other potential, is to bring conditions close to the natural. When the net is not at the potential of its surroundings, owing to the sluggishness of the collector, the current to the plate is not quite the natural air–earth current.

Although SCRASE's original method did not succeed, it should be practicable with a more rapid form of collector and GOTO (1957) has used nearly this method with a form of agrimeter (see § 5.29.) replacing the equalizer. MÜHLEISEN (1953) described a method similar to that of SCRASE and avoided the difficulty by increasing the time constant of the system to 10 min.

8.8. Electronic Methods

KAY (1950) devised an electronic method of measuring small direct currents and CHALMERS (1953e) suggested that the output of an agrimeter should be connected through a condenser to give compensation for the displacement current, so that KAY's amplifier could be used for the direct measurement of air–earth current. However, the agrimeter output was not sufficiently steady.

KASEMIR (1951a) was interested in the diurnal variation of the air–earth current and arranged apparatus with a time constant of one hour, which was sufficient to even out any effects of displacement currents. He used an electronic amplifier and arranged the apparatus to be portable.

ADAMSON (1960) used a field mill to provide compensation in an arrangement similar to SCRASE's but with a double electrometer valve in place of the quadrant electrometer. An attempt to get compensation for all rates of change of potential gradient by equalizing the time constants was not successful completely as the time constants were not independent. The apparatus was actually used for the measurement of rain current to an open surface, but the compensation for the displacement current is as important also in the case of the air–earth current.

8.9. Matching Circuit

KASEMIR (1955) showed that, with a suitable circuit, the effect of potential-gradient changes can be eliminated. His argument, somewhat simplified, is as follows.

Let A be the area of the collecting plate, i the air–earth current density, F the potential gradient at the ground and λ_1 the positive conductivity near the ground, so that $i = \lambda_1 F$. The surface density of bound charge σ is $-\varepsilon_0 F = -\varepsilon_0 i/\lambda_1$. The total current at the collector, including displacement current, is

$$Ai - A\frac{d\sigma}{dt} = Ai + \frac{A\varepsilon_0}{\lambda_1}\frac{di}{dt}$$

if it can be assumed that λ_1 remains constant while i alters.

If the collector is connected to earth through a resistance R in parallel with a capacitance C, then, when V is the potential of the collector with respect to earth, the current j through R is V/R; there is a charge $Q = CV$ on the capacitance and if V alters, there

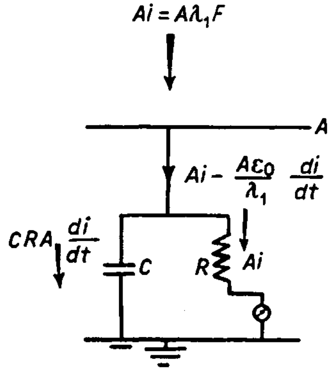


FIG. 39. Diagram of KASEMIR's method.

is a current $dQ/dt = CdV/dt$ through C. The total current from the collector must then be $j + dQ/dt = j + CRdj/dt$.

Since the collector is not accumulating charge,

$$Ai + \frac{A\varepsilon_0}{\lambda_1}\frac{di}{dt} = j + CR\frac{dj}{dt}.$$

If CR is chosen to be equal to ε_0/λ_1, then $j = Ai$, however i varies, that is to say the current through R is the true air–earth current and the displacement current is by-passed through C.

In practice, the difficulty would appear to be that λ_1 is not constant and so the condition $CR = \varepsilon_0/\lambda_1$ would not always hold unless C or R is continually altered. However if CR is approximately equal to ε_0/λ_1, then the effect of displacement currents is greatly reduced.

The method amounts to making the time constant of the collector circuit equal to the relaxation time (see § 2.27.) of the lower atmosphere. A more detailed discussion has been given by ISRAËL (1955a). The theory assumes that the conductivity remains constant and any change of potential gradient which is caused by a change of conductivity would not be by-passed and would be registered as an air–earth current.

The errors introduced by the incorrectness of matching have been discussed by RUHNCKE (1961 a).

8.10. Von Kilinski's Method

In order to avoid the difficulties of direct-current amplification, VON KILINSKI (1952) used the principle of the field mill to obtain, from the direct-current input of the air–earth current, an alternating current suitable for amplification. Figure 40 shows the method.

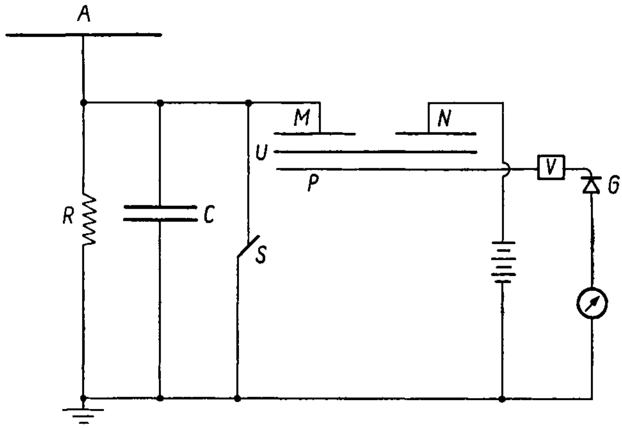


FIG. 40. von Kilinski's apparatus. (Copied and simplified from VON KILINSKI, 1952, Fig. 1.)

The current to A puts a potential on M; a fixed potential is put on N and the rotating sectors, U, alternately expose P to M and N and shield it. The amplifier V and the rectifier G give an output, which if the current is zero corresponds to the effect of N alone; if the current is negative, the output is less than that due to N alone. A switch S periodically shorts out the current and gives the zero, due to N.

The capacitance of C is made large so as to avoid effects caused by the displacement current, but it would seem that a field machine could be connected, through a condenser, to N and provide compensation for the displacement current.

An improvement to the method (VON KILINSKI, 1958) was obtained first by a slow rotation of the collecting plate, to break spiders' webs, and second by the use of the matching circuit (see § 8.9.).

8.11. Goto's Method

GOTO (1957) adapted WILSON's method (see § 8.5.) for continuous recording. If a plate is earthed when shielded from the earth's field, then exposed to the field and then shielded again, the net charge acquired is just that from the air–earth current. GOTO used three such plates, rotating beneath a hole in an earthed cover, somewhat similar in principle to the agrimeter (§ 5.29.), but without the earthing contact when the plate is exposed; he used the output to charge an electrometer in much the same way as was used in the early forms of the agrimeter.

The simple method, as described above, suffers from the disadvantage that when the plate is charged, on leaving the shielded region, it acquires a potential different from that of the earth and so distorts the lines of force and of current flow. In WILSON's original method this effect was eliminated by the use of the compensating condenser, applying a fixed potential to a suitable capacitance. To carry this out automatically, GOTO used a fixed capacitance and applied a suitable potential, derived from a water-dropper and then inverted. It would seem that a field machine could be used instead of the water-dropper.

8.12. Electronic Adaptation of Wilson's Method

CHALMERS (1962a) used WILSON's method with an electronic amplifier to measure the charge. Instead of adding together the charges from uncovering, collection and covering before measurement, as did GOTO (see § 8.11.), CHALMERS measured all three separately and then added them. The results showed that, in fact, the assumption of a constant conductivity, as used with a matching circuit (see § 8.9.) is not correct, and this method is therefore superior. In the form used, it could not give continuous readings.

8.13. Indirect Method of Measurement

The indirect method of measuring the air–earth current consists in measuring, separately, the potential gradient and the conductivity. In many of the measurements near the ground, the potential gradient has been measured by an equalizer, usually radioactive. The various methods of measuring conductivity (see Chapter 7) have been used, and it is necessary to include both signs of conductivity, since at a height of 1 m, ions of both signs play a nearly equal part.

The current $i = F(\lambda_1 + \lambda_2)$. But, very close to the earth's surface, the current, under normal calm fine-weather conditions, can be carried only by positive ions, so that, if primes represent values close to the earth, $i' = F'\lambda_1'$. If convection and diffusion currents can be neglected, $i = i'$; measurement shows that F and F' are very nearly the same. Thus $\lambda_1' = \lambda_1 + \lambda_2$, which is the result found by HOGG (1939b) (see § 7.6.).

8.14. Current to a Wire

KASEMIR and RUHNKE (1958) measured the current to an earth-connected wire at a height of 1 m and used the results in conjunction with measurements of the potential gradient by another wire, also at 1 m, carrying a radioactive equalizer.

If the wire, collecting the current, has a capacitance to earth of C and if the undisturbed potential at its position, as measured by the other wire with equalizer, is V, then the first wire when connected to earth must carry a charge Q such that

$$Q/C + V = 0.$$

If, on unit area of the wire, there is a charge dQ, the potential gradient at the surface of the wire is:

$$X = -dQ/\varepsilon_0.$$

Then, if the conductivity due to ions of opposite sign to Q, is λ, the current to unit area is

$$dj = \lambda X = -\lambda\, dQ/\varepsilon_0.$$

Integrating, for the whole wire,

$$j = -\lambda Q/\varepsilon_0 = \lambda CV/\varepsilon_0.$$

Since V is the potential at a height of 1 m, V gives the average potential gradient $\boldsymbol{F}$ over the first metre; thus j measures a multiple of $\lambda \boldsymbol{F}$.

It is clear that this method suffers under the usual difficulty of the displacement current, but KASEMIR and RUHNKE avoided this by the device of matching (see § 8.9.).

The method does not give a true direct measurement of the air–earth current, since all it really measures is the product $\lambda \boldsymbol{F}$; this is not even the unipolar conduction current at 1 m unless the potential gradient remains constant over the first metre of the atmosphere. Any effects of convection currents are not included in this measurement, and it therefore seems that this method should be best considered as an indirect rather than a direct method, in spite of the fact that it is a current that is measured. Perhaps it might be taken as a method to determine λ, since λF and F are measured with the two wires.

8.15. Measurements near the Ground

Early measurements of the air–earth current by the two methods appeared to indicate that the indirect method gave a result about twice that given by the direct method, and this agreed with theories current then (see § 8.20.). For example, LUTZ (1939) compared measurements made by the two methods at the same place, but at times separated by a period of years, so that local conditions may have altered. And ISRAËL's (1954) collection of results obtained at different places by the two methods shows a higher average for the indirect method, though this may have been due to the particular places of measurement.

Of importance in this connection is the variation of potential gradient with height in the lowest regions of the atmosphere. Simple consideration of the electrode effect (see § 8.19.) would lead to predictions of a quite appreciable difference between values at ground level and at 1 or 2 m. The results of WATSON (1929), SCRASE (1935b) and HOGG (1939b) have shown that, at Kew, the variation of potential gradient with height is very small, and SMIDDY and CHALMERS (1960) found the same for Durham. On the other hand, quite appreciable effects have been found over the Greenland ice-cap by PLUVINAGE and STAHL (1953) and RUHNKE (1962) (see § 7.21.) and over the sea by MÜHLEISEN (1961a).

NOLAN and NOLAN (1937) and NOLAN (1940) made simultaneous measurements of the air–earth current by the two methods at Glencree in Ireland, and found that the difference was only about 10 per cent. HOGG (1939b) (see § 7.1.) found that the positive and negative conductivities altered with height in the first metre, giving a total conductivity nearly independent of height.

LAW (1963) made measurements at Cambridge of ion densities and of space charge, and came to the conclusion that the conductivity varies with the height and so that the air–earth conduction current cannot be the same at all levels, thus necessitating a convection current also varying with height. The results of HIGAZI and CHALMERS (1966) lead to the same conclusion.

DOLEZALEK (1960d) discussed a number of measurements of potential gradient, air–earth current and conductivity. With OHM's law, it is assumed that there is no convection current; then deviations from OHM's law would appear, if a convection current actually exists; DOLEZALEK discussed the various other factors that would give apparent deviations from OHM's law and came to the conclusion that a comparison of results from the two methods of determining the air–earth current is not simply sufficient to give the convection current, if any.

8.16. Two Components of Conduction Current

A simple direct measurement of the conduction current at a level above the earth's surface is not possible, since a plate set up horizontally in fine-weather conditions would receive positive ions from above and negative ions from below and these two currents would tend to cancel out.

If, however, two plates close together are separated by an insulator, then each would receive one component of the current, and these could be added to give the total current. Compensation for the displacement currents due to field changes must be arranged.

Attempts to carry out the above suggestion are, at the time of writing, being made at Durham.

8.17. Measurements in Aircraft

A test of the importance or otherwise of convection currents can be made if the conduction current is measured by the indirect method at different levels. If there are no convection currents, the conduction current will be the same at all levels. The earliest results, by EVERLING and WIGAND (1921), showed the conduction current to decrease with altitude.

Recent work has been carried out by KRAAKEVIK at a number of different places. Over Greenland, KRAAKEVIK (1958a) found that the conduction current is practically constant with height, with a value of about $3{\cdot}7 \times 10^{-12}$ A/m^2. Over Chesapeake Bay, KRAAKEVIK and CLARK (1958) found the conduction current not to be the same at all levels. Above the *austausch* region (see § 2.32.), there is a constant value of about $1{\cdot}1 \times 10^{-12}$ A/m^2, but within the *austausch* regions (two separate regions were identified on the occasion in question), the conduction current was found to be considerably higher, about $1{\cdot}5 \times 10^{-12}$ A/m^2. If it can be assumed that quasi-static conditions existed, then it must follow that there is an additional current of about $0{\cdot}4 \times 10^{-12}$ A/m^2, either positive downwards above the *austausch* regions or negative downwards within the regions.

Since it is known that convection occurs within the *austausch* regions, KRAAKEVIK and CLARK considered that this might carry positive charge upwards, and the measurements of potential gradient showed space charges of about the correct magnitude.

KRAAKEVIK (1958b, 1961) reported other measurements of a similar type, in which the convection current could be deduced; in some cases the convection current was found to vary with height within the *austausch* region. In some cases, the convection current, taking positive current upwards, is more than half the conduction current, taking positive current downwards.

The existence of a continuous convection current in the *austausch* region would require a supply of positive charge for upwards convection and KRAAKEVIK and CLARK suggested that the source of this charge might be positively charged nuclei entering the atmosphere from combustion processes, as discussed by MÜHLEISEN (1956). Another source of positive charge might be positive charges from the sea (BLANCHARD, 1961) (see § 4.25.).

The average value obtained for the conduction current above the *austausch* region from various flights over the oceans was $2{\cdot}73 \times 10^{-12}$ A/m^2; if this can be taken as an average for the whole surface of the earth, then the total conduction current to the whole earth would be about 1400 A, or about 86 C/km^2/year.

The consequences of the fact that the convection current is appreciable and the conduction current not constant with height will be discussed in § 11.2.

8.18. Measurements with Radio-sonde

KASEMIR (1960) made measurements of the air–earth current in the free atmosphere with a balloon carrying a radio-sonde and using an adaptation of the method with a wire (see § 8.14.). As for the wire near the ground, what is measured is the product λF and not the total vertical current.

KASEMIR considered the effective area of the wire, the effects of initial charges on the balloon and the elimination of displacement currents with the matching circuit (§ 8.9.).

Results showed that, as expected, when above the *austausch* region, the air–earth conduction-current density remains the same at all heights. Its value over Greenland was found to be between 3 and 4 times its value over the Eastern United States, presumably because of the additional resistance of the polluted atmosphere near the ground.

Within the *austausch* region, the results were not reliable, since the matching condition could not be achieved.

UCHIKAWA (1961) measured the conduction current by the indirect method and found that its value in the *austausch* region was about 1·3 times that above the region.

8.19. The Simple Electrode Effect

In § 2.31. the electrode effect was discussed with the assumptions of uniform ionization and no convection or diffusion. Under these conditions, F would not be equal to F', but, instead $F'\lambda_1' = F(\lambda_1 + \lambda_2)$. Also λ_1' would not be equal to λ_1, since the number of small ions present depends on the recombination processes. A complete discussion must also include the effects of condensation nuclei, as well as effects of convection and diffusion and any variation of the rate of ionization with height.

The simplest case, that of air with no convection or diffusion, no nuclei, and uniform ionization, was investigated by VON SCHWEIDLER (1908), BEHACKER (1910) and SWANN (1913). A complete analytic solution is not possible in the general case, but the ratio of F' to F_∞ is 2·35, using modern values of the constants, and this is independent of the value of the current or of that of the rate of ionization.

The case where nuclei are present in such numbers that large ions predominate over small ions has been investigated by VON SCHWEIDLER (1929) and SCHOLZ (1931 b). SCHOLZ found that the equatons would be integrated only if $\eta = 2\eta_0$ (see § 4.28.). He then had to assume a value of η before he could obtain a result. If the result of WHIPPLE (§ 4.33.) is taken, that $\eta - \eta_0 = ew/\varepsilon_0$, so that $\eta = 2ew/\varepsilon_0$, then $F'/F_\infty = 2$, again independent of other parameters; if, however, as suggested by KEEFE, NOLAN and RICH (see §§ 4.31., 4.33.), $\eta - \eta_0 = ew/2\varepsilon_0$, then $F'/F_\infty = e = 2{\cdot}72$ and if one were to take $\eta - \eta_0 = 2ew/\varepsilon_0$, then $F'/F = (4)^{1}/_{3} = 1{\cdot}586$.

For other values than 2 for η/η_0, it is not possible to obtain analytical results, but it is hoped to use computer methods.

If, even in the simplest cases, any variation of the rate of ionization with height is introduced, the equations become intractable, though, again, computer methods might be applicable.

8.20. Convection Currents

WHIPPLE (1932) suggested that a way out of the difficulty of the absence of the electrode effect might be found in the existence of convection currents carrying positive currents upwards by a process of eddy diffusion. Thus the total vertical current at a height of 1 m would be less than $F(\lambda_1 + \lambda_2)$ by an amount equal to the current carried by the eddy diffusion and could then be equal to $F'\lambda_1$, the total current at the earth's surface, with F and F' nearly equal, as found by observation; such a process of diffusion would keep the density of positive ions nearly the same at all levels, so that λ_1 and λ_1' would not differ much, and thus the eddy diffusion current, or convection current, would be approximately equal to $F\lambda_2$, and so about equal to the resultant total current, but opposite in direction, giving a ratio of about 2:1 for the results by the two methods.

This theory cannot be quite satisfactory because, as mentioned in § 2·15., the current of eddy diffusion depends on d^3V/dh^3 or d^2F/dh^2, and, to give a value of this current equal to about λ_2, the value of d^2F/dh^2 would be such that F would alter appreciably in 1 m, contrary to the measurements. This argument can be stated thus: because the velocity of eddy diffusion is never very large, the space charge involved must be appreciable and this, in turn, requires a larger change of F than is actually found. HOGG's results show that λ_1 and λ_1' are not equal as is assumed in WHIPPLE's theory. LETTAU (1941) extended WHIPPLE's theory to include an eddy diffusion coefficient varying with height, but could not produce any better agreement.

If F does not alter with height, there is no space charge and no convection current, so that the conduction current must be the same at all levels, giving $\lambda_1' = (\lambda_1 + \lambda_2)$. The question then arises as to how λ_1' can be greater than λ_1, and it was first suggested by HOGG (1939b) and worked out by CHALMERS (1946) that the rate of production of ionization might be greater close to the ground than higher up, due to the effects of radioactivity in the ground. CHALMERS (1946) showed that it is possible to assume a not unreasonable rate of variation of ionization with height and to get conditions of a quasi-static state with ions of both signs, large and small, to give rough agreement with the observational results for potential gradient and conductivity, retaining small space charge (see § 7.9.). PIERCE (1958) showed that the actual variation of ionization with height is not far from that assumed by CHALMERS.

Where there is no radioactivity in the ground to provide this variation of ionization with height, the electrode effect would become evident, as in the cases quoted in § 8.15., for the Greenland ice-cap and the sea.

The results in aircraft (§ 8.17.) show that the assumption that convection is completely absent is not true, and LAW's results (§ 8.15.) confirm this. LAW (1963) has been able to give an account of his own results, involving both a convection current and a variation of the rate of ionization with height, and thus falling between the theories of WHIPPLE and of CHALMERS.

It appears probable that the extent to which the convection current is important may vary from one place to another.

8.21. Diurnal Variation

The diurnal variation of the air–earth current at Kew (SCRASE, 1933) shows different forms at different times of year. In the winter, the maximum occurs at about 6 hr with a minimum at about 21 hr, but in summer there is a much less marked maximum at about midnight and a minimum at about 13 hr. In general, there is a rough

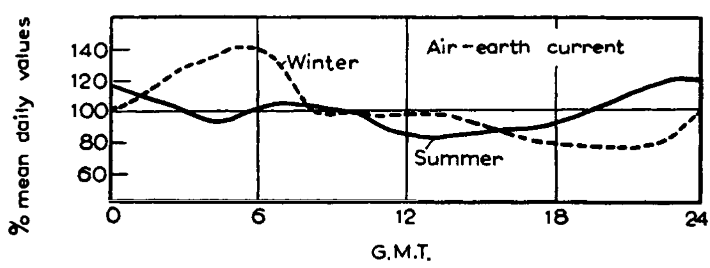


FIG. 41. Diurnal variation of air-earth current at Kew. (From SCRASE, 1933, Fig. 6 (opp p. 11) ("air–earth current", "summer" and "winter").)

inverse relation between the current and the potential gradient, though the current shows a less wide range of variation. This will be discussed in connection with the annual variation (§ 8.24.).

The results at Potsdam (KÄHLER, 1912), Davos, (DORNO, 1911) and Pavlovsk (OBOLENSKY, 1926) show a general measure of agreement with the results at Kew.

Over the oceans, where local effects on conductivity are absent, the diurnal variation of the air–earth current is the same as that of the potential gradient.

8.22. Variation of Air – Earth Current

The fundamental formula of fine-weather phenomena may be stated in the form:

$$i = V/R,$$

and so, by taking logarithms and differentiating:

$$\frac{1}{i}\frac{di}{dt} = \frac{1}{V}\frac{dV}{dt} - \frac{1}{R}\frac{dR}{dt}.$$

Thus the variation of i can be expressed partly as a world-wide variation concerned with the variation of V, the potential of the electrosphere, and partly as a local variation, concerned with the variation R, the local columnar resistance.

By considering the known variation of V with universal time and assuming the variation of R with local time to be the same everywhere as found by SAGALYN and FAUCHER (1956), OGAWA (1960a) constructed curves to give the diurnal variation of air–earth current at different longitudes. In view of the fact that the columnar resistance actually varies with time differently in different places, the agreement of OGAWA's calculated curves with the observational results, e.g. those collected by ISRAËL (1954), is quite good.

8.23. Sunrise Effect

At, or soon after, sunrise there is found an increase in the air–earth conduction current, whether measured by the direct or indirect method, as well as an increase in the potential gradient (see § 5.46.). CHALMERS (1957a) showed that the "*austausch*-generator" idea of KASEMIR (1956) would account for an increase in the current as measured indirectly, but not for an increase in the direct measurement. The explanation in terms of the injection of charged nuclei into the atmosphere from industrial processes (MÜHLEISEN, 1956) does not account for the existence of the sunrise effect at mountain stations; there may be effects due to moisture (MÜHLEISEN, 1958).

LAW (1963) suggested that the change of conduction current is related to the change in sign of the convection current, a larger conduction current being needed to keep the total vertical current constant.

8.24. Annual Variation of the Air–Earth Current

The annual variation of the air–earth current at different places showes quite different results. In some cases, there is a summer maximum, i.e. opposite to the maximum of the potential gradient; this occurs, for example, at Kew (SCRASE, 1933), at Val Joyeux (MAURAIN, HOMERY and GIBAULT, 1930), at Fort Rae (SHEPPARD, 1933) and at Canberra (HOGG, 1950). On the other hand, there is a winter maximum, agreeing with that of the potential gradient, at Potsdam (KÄHLER, 1912), Munich (LUTZ, 1911), Davos (DORNO,

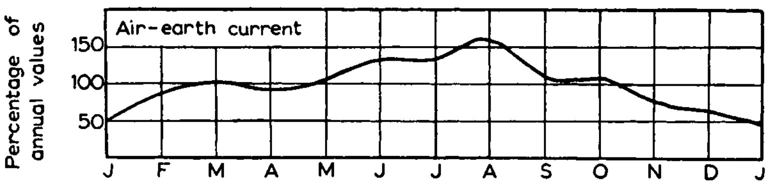


FIG. 42. Annual variation of air–earth current at Kew. (From SCRASE, 1933, Fig. 4 ("air–earth current" portions) p. 11.)

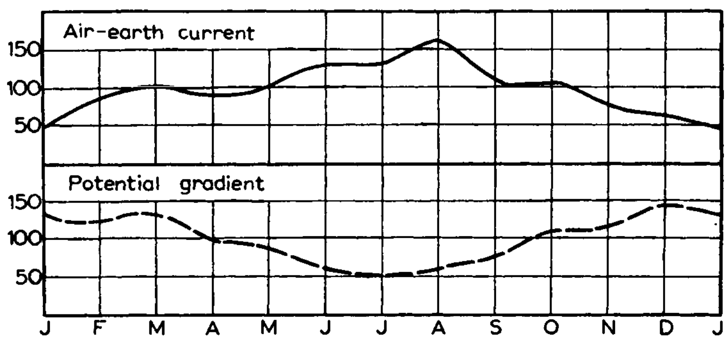


FIG. 43. Relation between air–earth current and potential gradient at Kew. (From SCRASE, 1933, Fig. 4 (top and bottom curves), p. 11.)

1911) and Huancayo (TORRESON, 1939). In other cases, the variations with time were irregular; a complete list was given by HOGG (1950). Some of these observations were taken at one time of day only and thus may be subject to effects of the alteration of the diurnal variation with time of year; for example, SCRASE (1935a) pointed out that the Kew observations taken only at 15 hr showed maxima in spring and autumn, while the continuous measurements (SCRASE, 1933) showed a summer maximum.

Over the oceans (MAUCHLY, 1926) or at high altitudes (e. g. the Zugspitze, LAUTNER, 1926), the annual variation is small.

Since $i = \lambda F$ and also $i = V/R$, it can be seen that, as V and R vary less than λ and F, i remains more nearly constant than λ or F. But, as R varies in the opposite sense to λ, but with less amplitude, it would be expected that i should follow the changes in λ and so should give a curve the inverse of that for F. Figure 43 shows that this is approximately true for the annual variations at Kew.

8.25. Relation between Air – Earth Current and Conductivity

HOGG (1950, 1955) has made two attempts to make more precise the relation between the air–earth current and the local conductivity at the earth's surface. From the relation of § 2.24., $i = V/R$, the conductivity affects i only in so far as it affects R. HOGG (1950) defined $\Lambda = 1/R$, the "columnar conductivity" and has divided this into an upper part Λ_c, effective at high levels and unaffected by local changes near the ground, and a lower part Λ_r effective near the ground and dependent on local conditions there; Λ_r is then some function of the measured specific conductivity, λ, near the ground.

Thus

$$i = \Lambda V = \frac{\Lambda_c \Lambda_r}{\Lambda_c + \Lambda_r} V = \frac{\Lambda_r}{1 + \Lambda_r/\Lambda_c} V = \frac{\mathrm{f}(\lambda)}{1 + \mathrm{f}(\lambda)/\Lambda_c} V.$$

If average values are taken over periods long enough to eliminate diurnal and annual variations of V, the comparison of i and λ for a number of different sets of observations should give $f(\lambda)$.

HOGG found that most of the observations could be fitted on a single curve which could be represented either by

$$i = 4{\cdot}2(\lambda + 70)^{0{\cdot}7}$$

or by

$$i = 0{\cdot}6\lambda^{2/3},$$

where i is in $A/\mathrm{m}^2 \times 10^{-14}$ and λ in $\Omega^{-1}/\mathrm{m} \times 10^{-16}$.

HOGG (1955) used the empirical result of GISH and WAIT (1950) (see § 7.14.) that the variation of conductivity with height is given by

$$\lambda = \lambda_0 + Ah^2.$$

Assuming that this relation holds down to the ground, where the conductivity would be λ_0, HOGG showed that the columnar resistance R is

$$\frac{\pi}{2(A\lambda_0)^{\frac{1}{2}}}$$

and so

$$i = \frac{2}{\pi}(A\lambda_0)^{\frac{1}{2}}\, V.$$

Comparison with measured values showed quite good agreement and mean values gave a result for V of $3{\cdot}4 \times 10^5$ V, close to accepted values (see § 11.2.). Since $i/\lambda_0^{\frac{1}{2}}$ should depend on V alone, HOGG used mean values over a period of many years in an attempt to establish whether V is affected by sunspot numbers. Although the apparent variation of V with time was similar for three stations, this did not agree with the sunspot variation.

The obvious weakness of these methods of relating air–earth current and conductivity is that they cannot take into account the abrupt change in conductivity at the upper boundary of the *austausch* region, as found by SAGALYN and FAUCHER (1954) (see § 7.13.), and the variability of the height of this boundary; thus, in HOGG's first method, A_r depends not only on λ, but also on this height; and in HOGG's second method, λ_0 is not the measured conductivity at the ground.

8.26. Synoptic use of Measurements

When simultaneous measurements are made of any two of the three quantities, F, i and λ, it is possible to make useful comparisons of the results at different places and to distinguish local from world-wide effects. As ISRAËL (1955b) has pointed out, a change in potential gradient may be produced by a nearby chimney or by a change in thunderstorm activity thousands of miles away, but simultaneous measurement of air–earth current or of conductivity would easily distinguish between these.

It is clear that comparative measurements, such as those of WHITLOCK and CHALMERS (1956) (see § 5.36.), would be of greater value if i or λ, as well as F, could have been measured simultaneously.

8.27. Mountain Measurements

HOLZER (1955) measured the air–earth conduction current at a number of different stations, in particular on mountains up to 4000 m. He found that the variation with time of day follows that of the potential gradient in undisturbed conditions (see § 5.43.) except that (a) the current is greater immediately after sunrise, and (b) the current is decreased during the middle of the day. Both effects may be due to convection currents altering conductivities, or having their own effects (see § 8.20.).

ISRAËL (1957) found that the mean value of the fair-weather conduction current at the Jungfraujoch was $13{\cdot}3 \times 10^{-12}$ A/m^2, to be compared with normal values of 2–4×10^{-12} A/m^2. On the idea that the conduction current should be the same at all levels, this appeared anomalous, but is, in fact, explained, as discussed in § 5.50., by the absence of part of the columnar resistance.

8.28. Measurements over Oceans

RUTTENBERG and HOLZER (1955) measured the air–earth current by the indirect method on the *Capricorn* expedition. The results of diurnal variation were in good agreement with that of potential gradient found on the *Carnegie* cruises (see § 5.43.), but statistical treatment of both sets of results suggests a small increase at local sunrise.

8.29. Air – Earth Current and Terrestrial Magnetism

From the general relation between magnetism and electricity, expressed as AMPÈRE's law, it follows that the line integral of the magnetic field strength round an area is equal to the total current through the area in a direction at right-angles.

Attempts have been made to verify this law in respect of the horizontal component of the earth's magnetic field. Results by SCHMIDT (1924) and BAUER (1920) appeared to show that the values of the current density necessary to satisfy the relation were of the order of 10^{-7} A/m^2, to be compared with the fine-weather current of the order of 10^{-12} A/m^2, but there was some doubt whether the accuracy of the magnetic measurements was sufficient for such comparisons to be made. More recently, SCHMIDT (1939) found the magnetic results gave no evidence for currents of this order.

Another method of dealing with this problem is that of PETERS (1923), who started at one definite point and drew paths running eastward, keeping everywhere at right-angles to the earth's horizontal field. If there are no vertical currents, these paths should close exactly on going right round the earth, and if the currents are of the order of 10^{-12} A/m^2, the paths should finish a few metres away from the start. But PETERS found that five paths in the northern hemisphere ended, on the average, 34 km south of their starting-points, while five paths in the southern hemisphere gave an average end-point 96 km north. Such results would require currents of about 10^{-7} A/m^2. Again, there is the doubt whether the magnetic measurements are sufficiently accurate.

Another phenomenon of a somewhat similar nature is that of vertical earth currents. If a telegraph line extending up a mountain side is connected to earth at two different levels, a current flows up the line from the lower to the higher level. If it can be considered that a current is flowing in the earth in the absence of the line, there must be currents down mountains on all sides, so that the circuit would have to be completed by a current through the air; however, such a current would have to be about 10^4 times the normal air–earth current. It has also been suggested that, in some way, the mountain acts as a battery which produces current only when the circuit is completed by the telegraph line, and in the absence of the line there is no current flowing but merely a difference of potential.

If currents of the order of 10^{-7}A/m^2 are found to be required to account for the magnetic results, these currents would have to be such that they are not measured by the direct method, and so could be composed only of particles which can traverse the collector without leaving any charge.

CHAPTER 9

Point-discharge Currents

9.1. Point Discharge

Under normal circumstances, the ionic current is carried by ions produced, as explained in Chapter 4, mainly by radioactive substances and cosmic rays. Each small ion so produced travels in the electric field until, as discussed in Chapter 4, it undergoes combination and ceases to be a small ion. The production of fresh small ions does not normally depend upon the small ions already present.

But if the electric field is sufficiently strong, there is a possibility of ionization by those already present. When this occurs over a long distance, a lightning discharge takes place, as will be discussed in detail in Chapter 14. When the process of ionization by collision is confined to the small volume near a point, where the field is enhanced by the concentration of lines of force ending on the point, there is the phenomenon of point discharge. When occurring at mast-heads, etc., this is known as St. Elmo's fire and when investigated in the laboratory it is termed corona discharge.

9.2. Ionization by Collision

The initial process of ionization consists of the removal of an electron from a molecule, leaving a positively charged ion. Under normal circumstances, the electron attaches itself to a neutral molecule, forming a negative ion, and then both positive and negative ions attach to themselves other molecules, as discussed in § 4.2.

However, if there is a sufficiently large electric field, an electron can acquire an appreciable amount of energy in the interval between its production and its first collision with a molecule; if this energy is sufficient, instead of combining with the molecule, the electron may be able to ionize the molecule, producing a fresh electron and ion. And this process can continue, so that, from one original electron a considerable number of ions may be produced.

It is possible that not only electrons but also the positive ions could produce fresh ionization by collision, but since an electron is smaller than an ion it is likely to travel farther before a collision and so acquire a greater energy. (All particles of the same charge gain the same energy in the same distance with the same field.) Thus, as the electric field increases, we may expect electrons to produce ionization by collision at a lower field than for positive ions.

9.3. Simplified Theory

The general discussion of point discharge in the atmosphere is simplified by the consideration of an ideal case. Imagine a horizontal cloud giving an electric potential gradient between it and the earth, so that the lines of force are vertical over a wide area. Now imagine a pointed conductor set up on the earth's surface; it will be seen that, near the conductor, the lines of force will be deflected from their vertical direction and will tend to concentrate on the point. Thus, at and near the point, the surface density of charge on the conductor and the field strength in the air will have values considerably greater than those at the earth's surface, the actual extent of this increase depending on the geometrical conditions. Thus, in circumstances which ordinarily would not give ionization by collision, the local field near a conducting point may reach the value necessary for ionization by collision in a limited region around the point.

For any point in a given position there is a minimum value of the potential gradient at the earth's surface nearby at which the point-discharge current becomes appreciable. For points at the heights of trees, this potential gradient is from 600 to 1000 V/m, and for trees themselves somewhat greater. When an isolated point is raised to 30 or 40 m, point discharge can occur in fine-weather potential gradients of under 200 V/m.

When the point is positively charged, i.e. in a negative potential gradient in the atmosphere, electrons move towards the point and positive ions away; when the point is negative, there are positive ions moving towards it, and the electrons moving away will attach themselves to molecules to form negative ions where they reach a region where they are no longer able to produce ionization by collision.

9.4. Pulsed Nature of Discharge

The process of ionization by collision produces positive ions or electrons which move into the point and ions of opposite sign which move away from it; these latter form a space charge which has the effect of reducing the potential gradient near the point, since lines of force end on these ions instead of on the point. This may go on until the potential gradient is so much reduced that ionization by collision no longer occurs and the discharge ceases. Then, as the ions move off, they no longer reduce the potential gradient near the point so much and ionization by collision starts again, the whole process now repeating itself. Thus, under suitable conditions, the discharge may occur in a series of pulses.

The pulsed nature of the discharge has been investigated in laboratory experiments by TRICHEL (1938), who found that it occurs very regularly when the point is the negative electrode; each pulse carries the same quantity of charge, of the order of 10^{-10} C and the potential gradient determines the frequency of the pulses and thus the average current. But occasionally there is a change in the quantity of charge per pulse, associated with a visible change in the position of the discharging spot on the point. LARGE and PIERCE (1955) have found exactly the same results for natural point discharge, in a positive potential gradient, from a point 7 m above the ground, in disturbed weather. CHALMERS (1965b) found that a wind in the direction of the main current flow caused an increase in the current, the interval between pulses and the charge per pulse, the last effect being the greatest. The subject has been studied in more detail by PIERCE, NADILE and MCKINNON (1960).

When the point is positive, corresponding to a negative potential gradient in the atmosphere, the effects are different; in laboratory experiments, when the potential gradient is only just sufficient for discharge, there are "inductive kicks" with, sometimes, small and irregular pulses; for higher potential gradients, the current becomes more steady. Entirely similar results were obtained in the atmosphere by LARGE and PIERCE (1955).

9.5. Laboratory and Atmospheric Conditions

In laboratory experiments on point discharge, the volume in which ionization by collision occurs may be of dimensions comparable with the distance between the electrodes; but for atmospheric point discharge, the "electrodes" are the point, at earth potential, and the cloud. Thus in laboratory experiments the ions that have moved away from the point move quickly to the other electrode and have little effect, whereas in the atmosphere they can act as a space charge and affect matters largely.

Also, in laboratory measurements there is usually no wind that can remove the ions from the neighbourhood of the point, but in the atmosphere wind cannot be avoided.

Thus, in general, the results of laboratory experiments cannot be carried over to atmospheric phenomena, and many of the investigations of atmospheric point discharge can be carried out only in the atmosphere, where conditions are not very much under control.

9.6. Early Measurements

Measurements of the current through a measuring instrument connected between an exposed point and the earth appear to have been made first by COLLADON (1826), who used points at the top of a pole some 10 m high. PELTIER (1840) fixed points to a kite and measured the current and WEBER (1886–92) used both kites and captive balloons. Other early workers who measured point-discharge currents included QUETELET (1849), LAMONT (1852, 1862), LEMSTRÖM (1882–4, 1898, 1900), WEISE (1904) and DIECKMANN (1912). These observers found that appreciable currents occur, even from low points, in the strong potential gradients of storms and showers, but that a considerable height is needed in fine weather. WEBER, who measured currents up to 6μA continuously, with larger values around the time of a lightning flash, seems to have been the first to suggest that point discharge from natural points, e.g. trees, might be of importance in questions of the electrical condition of the atmosphere. WILSON (1920) returned to this idea and suggested that point discharge might be a very important factor in the transfer of charge between clouds and the earth; this led directly to the work of WORMELL (1927, 1930) and SCHONLAND (1928b).

9.7. Wormell's Microvoltameter

In order to test the part played by point discharge in the transfer of charge between the electrosphere and the earth, WORMELL (1927) devised a method of measuring the total amount of charge brought down to the earth by point discharge over a considerable period

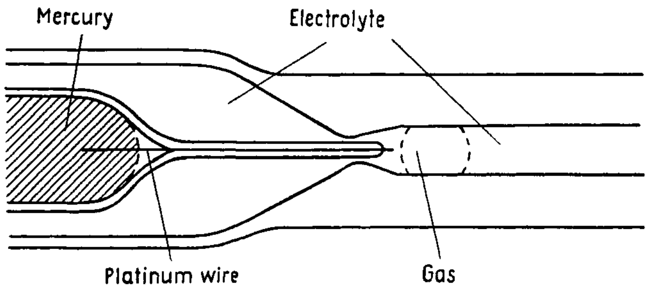


FIG. 44. Wormell's microvoltameter (one electrode). (From WORMELL, 1927, Fig. 3, p. 445.)

of time. He uesd a metal point, mounted on a pole and raised to about the height of the trees in the neighbourhood. To measure the quantity of electricity passing, WORMELL devised a "microvoltameter" consisting of a tube containing dilute sulphuric acid, with two platinum wires to form electrodes; he arranged that the bubbles of gas produced by electrolysis should go into capillary tubes where they could be measured by a microscope. It is known that the volume of hydrogen produced at one electrode is twice that of the oxygen produced at the other. If 1 C of charge produces a volume x of oxygen and a volume $2x$ of hydrogen and if q_1 C pass in one direction and q_2 C in the other, the volumes of gas produced at the two electrodes are $v_1 = q_1x + 2q_2x$ and $v_2 = q_2x + 2q_1x$. If v_1 and v_2 are measured and x is known, q_1 and q_2 can be obtained.

Over a period of 3 years, WORMELL found that twice as much negative as positive charge came to earth through his point. In order to estimate the current density brought to the earth by point discharge and so to get an idea of the importance of point discharge in what he called "the electrical balance sheet of the earth", WORMELL had to know, or guess, how many points exist in a given area with the same effectiveness as the one used. WORMELL counted the number of trees per km^2 higher than this point and assumed that this would give the number of points similar to the one used, with the idea that the effect of those higher would roughly balance the effect of some being very close together. His results gave a net negative charge of 0·13 C per annum through his point and about 800 similar points per km^2, so that 1 km^2 receives about 100 C of negative charge per annum from point discharge.

9.8. Schonland's Measurements

SCHONLAND (1928b) took an actual tree, which he supported on insulators and connected to the earth through a galvanometer. The total charge passing could be estimated by the integration of the galvanometer readings. Schonland was interested in the effects during thunderstorms, for it is then that the point-discharge currents are greatest; therefore his results are not complete for any periods of time, but rather represent particular results for a series of thunderstorms. He was concerned mainly in estimating the relative effects of the various ways in which charge is transferred to or from the earth during a thunderstorm. He was able to estimate the charge brought down to earth per unit area by lightning flashes and he measured the charge brought down by rain and conduction currents (see § 11.9.).

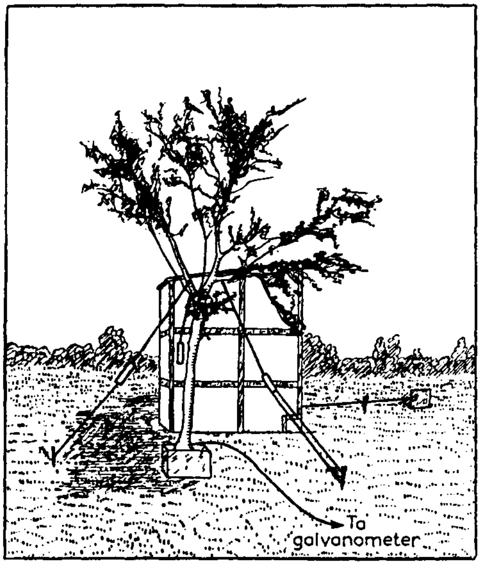


FIG. 45. Schonland's tree. (From SCHONLAND, 1928, Plate 6 (upper), facing p. 256.)

To complete the knowledge of the transfer of charge during storms, SCHONLAND had only to convert the current brought down through his single tree into the current brought down over an area. In this he seemed to have an easier task than WORMELL, for the trees in the neighbourhood of his observations (Somerset West in South Africa) are all similar to the one he used and are fairly evenly spaced; also, since SCHONLAND used a typical tree, there was no question as to the relative efficiency of a point and a tree; SCHONLAND attempted to maintain natural conditions by tying leafy branches to his tree. But a glance at the photograph in SCHONLAND's paper suggests that the tree used is not in natural surroundings and that the nearest trees are considerably more distant than the average separation of 5 m that is stated. Thus, if the amount of point-discharge current from a tree depends on its situation relative to other trees, SCHONLAND's was not a fair sample.

Assuming his value of the separation of trees, SCHONLAND found that, on the average, point discharge brings down 20 times as much charge as lightning flashes in the same storm, both being almost always negative, while the amount of rain charge is negligible. He also obtained a value of 2·1 A as the average total current below a thunderstorm. These figures may well require correction if the tree actually used is better exposed than the typical trees of the neighbourhood. If it can be assumed that the same current flows through a South African thunderstorm as through one in America, then it is possible to use the results of GISH and WAIT (1950) (see § 11.17.) that the average current is 0·5 A and deduce that the tree SCHONLAND used is about 4 times as effective as the average tree; with the value of 1·3 A from the results of STERGIS, REIN and KANGAS (1957b), the ratio is only 1·6. WORMELL (1953a), by considering the maximum point-discharge current density beneath a storm, found a still greater factor. But the conclusion remains that point discharge plays the most important part in the transfer of charge between thunder clouds and earth.

9.9. Point-discharge Ratio

WORMELL (1927, 1930) found the ratio of the quantities of negative to positive charge brought down to the earth to be 2·0:1. WHIPPLE and SCRASE (1936) made comprehensive measurements of the total current due to point discharge at Kew over a period of 2 years, using a point of height 8·4 m, similar to that of WORMELL, but recording the currents through a galvanometer by means of a drum camera; the integration of the records gave the total quantity of charge passing during any interval of time. Their ratio was 1·7:1, rather smaller than that of WORMELL, but they were able to measure smaller quantities of charge and so may have obtained a more accurate value. Other workers have used apparatus of the same type, with the exception of ALLSOPP (1952), who collected the charge on a condenser, with a spark gap in parallel to discharge the condenser when the charge reached a certain value; the total charge was then obtained by counting the sparks produced, and the direction of the current was noted.

TABLE 1. Point-discharge Ratio

Observer	Place	Ratio of −ve to +ve	Duration
WORMELL (1927, 1930)	Cambridge (England)	2·0	3 years
WHIPPLE and SCRASE (1936)	Kew (England)	1·7	2 years
IMMELMANN (1938)	Pretoria (S. Africa)	2·8	2 years
YOKOUTI (1939)	Japan	2·1	1 year
GERASHIMOVA (1939)	Elbrus (Caucasus)	1·6	14 years
CHIPLONKAR (1940)	Colaba (India)	2·9	1 year
PERRY, WEBSTER and BAGULEY (1942)	Nigeria	2·86	9 months
LUTZ (1944)	Munich (Germany)	2·0	3 years
CHALMERS and LITTLE (1947)	Durham (England)	1·36	8 months
ALLSOPP (1952)	South Africa	2·2	
MICHNOWSKI (1957)	Poland	1·5	6 storms
SIVARAMAKRISHNAN (1957)	Poona (India)	2·0	1 year

Results are tabulated in Table 1 and suggest that perhaps the ratio of negative to positive charge is greater in tropical and subtropical than in temperate regions.

9.10. Time Lags

LUTZ (1941) and HUTCHINSON (1951) observed that when the potential gradient is changing sign, the point-discharge current often changes sign after the potential gradient. This is explained in terms of space charge in the region below the point but above the apparatus for measuring potential gradient, a space charge produced largely by points in this region. When the potential gradient is positive, the point discharge fron the measuring point and also from lower points produces negative ions, giving a negative space charge. If the effect of the cloud now decreases, a time will come when this negative space charge can produce a greater effect at the potential-gradient measuring apparatus than the still positive effect of the cloud; so the potential gradient at the ground changes sign, while the potential gradient at the point, and therefore the point-discharge current, remains positive. LUTZ (1941) pointed out that it is the point-discharge current, rather than the potential gradient near the ground which better indicates the effects due to the clouds. HUTCHINSON (1951) used similar ideas of space charge to account for deviations, in the form of "humps", from the law of WHIPPLE and SCRASE (1936). If all the points concerned in point discharge were in exactly similar situations, then time lags and humps would not occur, and the relation between point-discharge current and potential gradient would presumably hold more closely.

9.11. Discharge from Points Attached to Kites and Balloons

Measurements have been made of currents flowing down wires attached to kites or balloons; in some cases, special points were provided for point discharge but in others discharge occurred at existing points or at the surface of the wire. In stormy weather, the currents are similar to those of trees and lower points, but of greater magnitude. In fine weather, also, point discharge can occur and this was first found, if not recognized as such, by DE ROMAS (1783) with a kite; of the early workers mentioned in § 9.6., PELTIER (1840) also used a kite and WEBER (1886–92) used both kites and balloons.

NUKIYAMA and NAKATA (1926) measured the current in a wire attached to a kite-balloon at various heights, and RANGS (1942) measured the currents from kites at various heights, together with simultaneous measurements of the potential gradient at the ground.

DAVIS and STANDRING (1947) used a captive kite-balloon at heights up to 2400 m and found a general increase of current with height; they also found a general increase of current with the potential gradient at the ground and with the wind speed; they discussed the effects on the potential gradient, measured at the ground below the balloon, caused by the induced charge on the balloon cable and by the ionic charge released by point discharge and travelling in the wind. WICHMANN (1951) also measured currents from a balloon in fine weather and used these to give relative values of the potential gradient at different times of day, neglecting any effect of wind, and assuming the formula of WHIPPLE and SCRASE (1936) (see § 9.15.).

CHALMERS and MAPLESON (1955) measured currents from a point attached to a captive balloon, at lower heights than those used by DAVIS and STANDRING (1947), going up only to 200 m; they found the current to increase with wind speed, with height and with the potential gradient at the ground.

9.12. Point Discharge from Isolated Point

KIRKMAN and CHALMERS (1956) used a point at about 30 m, considerably higher than the neighbouring trees; they found point discharge to occur for potential gradients down to 250 V/m, i.e. in some conditions of fine weather, and they found the current to depend on the wind speed. For potential gradients over 1000 V/m, point discharge occurred at nearby trees and the ions so produced interfered with measurements. The results will be discussed in § 9.16.

9.13. Multiple Points

CHIPLONKAR (1940) made measurements with a group of 4 points and found, rather surprisingly, that the total current through the group is less, with the same potential gradient, than the current through a single point. On the other hand, SIVARAMAKRISHNAN (1957) found, under somewhat similar conditions, a greater current from 4 points than from one.

BELIN (1948), in laboratory experiments, found that each point in a group of 7 or 13 gave as much current, under similar conditions, as a single point.

CHALMERS and MAPLESON (1955), with a captive balloon in fine weather, compared the currents through a group of 8 points and through a single point, and found that, in the same potential gradient and wind, the 8 points together gave about half the total current by one point.

Clearly, the discrepancy must be ascribed to the separations of the points relative to the lines of force; certainly SIVARAMAKRISHNAN's points were more widely separated than CHALMERS and MAPLESON's and would more nearly act independently, and in the laboratory conditions of BELIN the points would be quite independent.

In the laboratory, JHAWAR and CHALMERS (1965) used 7 points, 6 in a hexagon with the other at the centre, and varied the length of the side of the hexagon, keeping the rest of the apparatus the same. They found that their results could be expressed by stating that the "starting voltage" is increased as the points become closer, and that, for any one separation of points, if one allows for the difference of starting voltage, the current is a multiple of that for a single point, the multiple being greater the greater the separation of the points.

There is scope for a systematic investigation of the effect of multiple points in natural point discharge in the free atmosphere, to be carried out in a place where point discharge is frequent, comparing the current through a single point with currents through sets of points with different separations.

9.14. Discharge from Trees

SCHONLAND'S (1928b) measurements were made with a tree cut down and supported on insulators, but the other measurements described have all been made with metal points. From the point of view of the effective separation of points (see § 9.24.), it is important to compare the artificial point with the natural tree.

In § 5.68., the effect of point-discharge ions on the potential gradient to leeward of the point has been discussed, and MAUND and CHALMERS (1960), have attempted to use this method to determine, indirectly, the point-discharge current through a living tree. The results obtained showed that a sycamore tree in full leaf gives no measurable effect even when the potential gradient reaches 5000 V/m; for ash and poplar trees not in leaf, discharge appears to commence at about 1000 V/m, not far different from the value for a single point at the same height.

By-passing through a galvanometer most of the current down a tree, by leads into it at two different heights, MILNER and CHALMERS (1961) showed that the current through a lime tree is comparable with that through a point at similar height, but starts at a higher potential gradient. ETTE (1966b) found that the by-passed current is only a fraction of the whole current through the tree.

CHALMERS (1962c) found that, when the potential gradient is varying rapidly, such as after a close lightning stroke, the currents through a tree and through a metal point do not correspond. ETTE (1966a) has discussed this in detail.

Ions produced at trees from point discharge have been detected as space charge by BENT *et al.* (1965).

9.15. Relation between Point-discharge Current and Potential Gradient

Measurements under laboratory conditions have shown that the relation between point-discharge current and potential gradient is of the form:

$$I = a(F^2 - M^2),$$

where I is the current, F the potential gradient and a and M are constants. It has therefore been usual to attempt to fit measurements of atmospheric point discharge to a similar formula.

When measurements are made of the point-discharge current in natural conditions, it is difficult to obtain accurate results because conditions are usually changing very rapidly; hence there is always a wide scatter of the results and the agreement with any formula is never very close.

WHIPPLE and SCRASE (1936) measured the point-discharge current and, simultaneously, the potential gradient at the ground, and fitted their results to the above formula, finding, for positive currents down through the point, $M = 780$ V/m and $a = 8 \times 10^{-14}$ A/(V/m)^2; for negative currents, $M = 860$ V/m and $a = 10 \times 10^{-14}$ A/(V/m)^2.

SCHONLAND'S (1928b) results show the same general form as those of WHIPPLE and SCRASE, but do not fit the formula exactly.

CHIPLONKAR (1940), CHAPMAN (1952), YRIBERRY (1954) and SIVARAMAKRISHNAN (1957) have also fitted their results to the same formula. HUTCHINSON (1951), who measured the potential gradient by a bound charge method, found "humps" in the curve and showed that these could also be seen in the results of WHIPPLE and SCRASE; these have already been discussed (§ 9.10).

The results here quoted so far are those for point discharge from a point which is at about the same height as natural points in the neighbourhood and no account has been taken of wind speed.

An analysis by KIRKMAN and CHALMERS (1956) of the results of WHIPPLE and SCRASE (1936) and of HUTCHINSON (1951) to fit a formula

$$I = A(F - M)^n$$

showed that the best value of n is 1·1. This is closer to the formula of CHAPMAN (1956) (see § 9.16.) which suggests that perhaps wind speed is an important factor. The difference between this value of 1·1 and a value of 1 is caused by a few observations during large values of F, and might be accounted for by the fact that larger values of F are associated with more highly disturbed conditions, during which the wind speed is likely to be greater.

9.16. Effect of Wind Speed

The first measurements to show that wind speed affects the natural point-discharge current were probably those of DAVIS and STANDRING (1947), which showed an increase of current with wind speed, but no general formula was obtained. CHALMERS and MAPLESON (1955), with a captive balloon, made measurements of currents with wind speed, W, up to 22 km/hr and potential difference, V, between the point and its surroundings up to 22 kV, and found an empirical formula

$$I = KW^{1/4}V^{7/4}.$$

For an earthed point at height h in a potential gradient F, $V = Fh$, if space charge in the height h can be neglected.

KIRKMAN and CHALMERS (1956), with an isolated point at 30 m, found

$$I = k(F - M)(W + b).$$

CHAPMAN (1956), on general considerations, suggested that the correct form of the relation between point-discharge current and potential should be

$$I = a(V - V_0)\,v,$$

where v is the velocity with which ions are removed from the neighbourhood of the point.

When there is little wind, v is proportional to F, since the ions are moved away by the field, and so

$$I = A(F - F_0)\,F,$$

quite similar to the formula of WHIPPLE and SCRASE (1936).

On the other hand, if there is much wind, v is equal to the wind speed W and the formula becomes

$$I = a(V - V_0)\,W = ah(F - F_0)W,$$

which is not far from the formula of KIRKMAN and CHALMERS (1956).

CHAPMAN (1956) tested his formula in three ways: by point-to-plane discharge in still air, by point discharge in a wind tunnel and by measurements with a point on an aircraft, and found good agreement. LARGE and PIERCE (1957) measured the current from a point raised to a high potential and exposed to the natural winds; at the suggestion of CHALMERS, they considered v as the vector sum of the wind speed and a term involving V and obtained

$$I = a(V - V_0)(W^2 + c^2V^2)^{\frac{1}{2}}.$$

CHALMERS (1957b) showed that the results of CHALMERS and MAPLESON (1955) and of KIRKMAN and CHALMERS (1956) agreed as closely with this formula as with the empirical formulae previously given, but that the constants a, V_0 and c differ for the different sets of observations, as might be expected since different points were used. The approximate theory of § 9.19. suggests that the constant a should be $2\pi\varepsilon_0$ for all points.

MILNER and CHALMERS (1961) found that, if wind speed is included, the current from a point at tree-top height agrees with the formula given, so that the formula of WHIPPLE and SCRASE (1936) is not correct, even for points at such heights. MILNER and CHALMERS also found that the current through a tree, as measured by the by-passing technique (see § 9.14.), is related linearly to the potential gradient, for a constant wind speed.

9.17. General Space-charge Theory

Considering times long compared with the intervals between pulses, in a steady field the conditions are stationary or quasi-static, and it is possible to enquire theoretically into the relation between the point-discharge current and the potential difference between the point and its surroundings. At first sight it would appear that we should need to know the details of the processes of ionization by collision in order to discuss the current close to the point; however, this turns out not to be necessary, for just the same reasons as apply in the rather analogous case of the space-charge theory of thermionic emission.

The process of point discharge produces a unidirectional ionic current away from the point and this acts as a space charge to reduce the field close to the point. Since the current depends on the field close to the point, there must be a situation of balance when there is just enough current and so space charge to produce the field near the point that supplies the current.

It turns out that it is not possible to work out the situation in detail, even with no wind, except in special cases. CHALMERS (1952b) has shown that it is possible to obtain solutions which lead to the formula of WHIPPLE and SCRASE (1936) in the absence of wind, for a point which is one of a rectangular array. CHALMERS and MAPLESON (1955) obtained a formula for a single point which agreed with their empirical formula, with a suitable choice of a parameter; with a different choice, agreement is also found with the formula of KIRKMAN and CHALMERS (1956).

9.18. Widespread Point Discharge

The relation between current and potential difference for a region of widespread point-discharge currents was first given by WILSON (1925) and was discussed in detail by WHIPPLE and SCRASE (1936).

If we consider a level at height h where the potential gradient is X and where there are n ions, each carrying a charge e, per unit volume, then if w is the ionic mobility, the current density is

$$i = newX.$$

From POISSON's law,

$$dX/dh = ne/\varepsilon_0.$$

If conditions are quasi-static, i must be the same at all levels and

$$i = \varepsilon_0 wX\, dX/dh$$

giving:

$$[X^2] = 2i[h]/\varepsilon_0 w.$$

This holds throughout the volume in which the ionic current is uniform, but clearly is not accurate close to the ground, where the point-discharge current comes from discrete points.

The equation can be expressed either as

$$X^2 - X_0^2 = 2ih/\varepsilon_0 w,$$

or as

$$X^2 = 2i(h + c)/\varepsilon_0 w,$$

where X_0 represents the value of X at the ground if the equation continued to hold so far, and c represents the distance below the ground at which X would then be zero.

It would need further discussion to decide how X_0 or c depends on i, since i has been taken as constant in the integration, but both X_0 and c are small.

Using the second form of the equation, the potential at a height h is

$$V = (8i/9\varepsilon_0 w)^{\frac{1}{2}} \, [(h + c)^{\frac{3}{2}} - c^{\frac{3}{2}}].$$

WHIPPLE and SCRASE (1936) pointed out that these results show a very considerable increase of potential gradient with height.

The theory, as given, does not take into account the fact that ions, produced in the point-discharge process, are small ions and only continue to exist as such for a limited time; then they will become large ions by attachment to condensation nuclei, and will have a much lower mobility. It is, however, not profitable to discuss conditions of large ions, since the quasi-static state does not persist for long enough to allow large ions to travel far. The theory, also, takes no account of wind; for a sufficiently widespread region, horizontal wind would have no effect, but it the region is limited in extent, wind cannot be neglected. A vertical wind would also have an effect, as discussed by CHALMERS (1944).

9.19. Approximate Theory for Isolated Point

A single isolated point can be discussed approximately if we neglect its connections with earth and consider a point at zero potential inserted into a region at potential V, taken as negative.

Close to the point, the ions moving away from the point are under the influence of the electric field; far from the point, they are much more under the influence of the wind. And at some intermediate distance, the velocity of the ions in the field is just equal to the wind speed, in magnitude. As an approximation, it is assumed that the volume round the point is divided into two regions by a sphere at which the ion velocity and wind speed are equal. In the inner region, (A), it is assumed that wind can be neglected; and in the outer region (B), it is assumed that wind removes the ions immediately they enter the region from (A).

As discussed in § 9.17., it is not necessary to consider the details of the process of ionization by collision, but we can imagine that, within a radius a from the point, the field is greater than the minimum value X_0 for ionization by collision. Let X be the field strength at any point, W the wind speed and w the mobility of the positive ions.

Then if b is the "intermediate distance" mentioned above, dividing the regions (A) and (B), $X_b = W/w$.

In the region (A), for any sphere of radius r, if there are n ions carrying charge e per unit volume, the outward current through the sphere is

$$I = 4\pi r^2 newX.$$

For two neighbouring spheres, GAUSS's theorem gives

$$4\pi r^2 ne = \varepsilon_0 \frac{d}{dr}(4\pi r^2 X).$$

Thus

$$\frac{I}{wX} = 4\pi\varepsilon_0 \frac{d}{dr}(r^2 X)$$

which integrates to

$$\left[\frac{Ir^3}{12\pi w\varepsilon_0}\right] = \left[\frac{r^4 X^2}{2}\right],$$

with the lower limit of integration at $r = a$, $X = X_0$.

Since a is small, we can neglect the values at the lower limit and get

$$\frac{I}{6\pi w\varepsilon_0} = rX_r^2.$$

Putting

$$\frac{I}{6\pi w\varepsilon_0} = K^2, \quad X_r = Kr^{-\frac{1}{2}}.$$

Thus the potential at b is

$$V_b = \int_a^b - X_r\, dr = -2K(b^{\frac{1}{2}} - a^{\frac{1}{2}}),$$

or, approximately,

$$V_b = -2Kb^{\frac{1}{2}}.$$

But

$$X_b = Kb^{-\frac{1}{2}} = W/w$$

$$V_b = \frac{-2K^2 w}{W}.$$

In region (B), there is assumed to be no space charge and so GAUSS's theorem gives

$$X_r = \frac{b^2}{r^2} X_b = \frac{b^2 W}{r^2 w}.$$

Then

$$V_r - V_b = -\int_b^r X_r \, dr = \left[\frac{b^2 W}{rw}\right]_b^r.$$

If

$$r \to \infty \qquad V_\infty - V_b = -\frac{Wb}{w} = -\frac{K^2 w}{W}.$$

Thus

$$V_\infty = -\frac{3K^2 w}{W} = -\frac{I}{W 2\pi\varepsilon_0}.$$

So

$$I = -2\pi\varepsilon_0 WV.$$

This formula, derived by CHALMERS (1962b), can be improved by replacing V by $(V - V_0)$ where V_0 is the minimum potential difference for point discharge to occur; this takes care of the neglect of a in the above deduction.

9.20. The Alti-electrograph

Since the point-discharge current through a point depends upon the potential gradient, this can give a method of measuring potential gradients in the atmosphere. SIMPSON and SCRASE (1937) and SIMPSON and ROBINSON (1940) have used this method with a balloon. As only small weights can be carried by the balloon, it

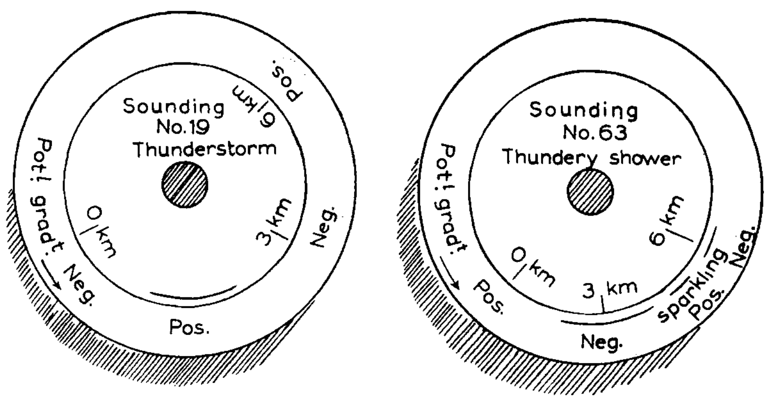


FIG. 47. Alti-electrograph record. (From SIMPSON and SCRASE, 1937, Plate 12, following p. 314.)

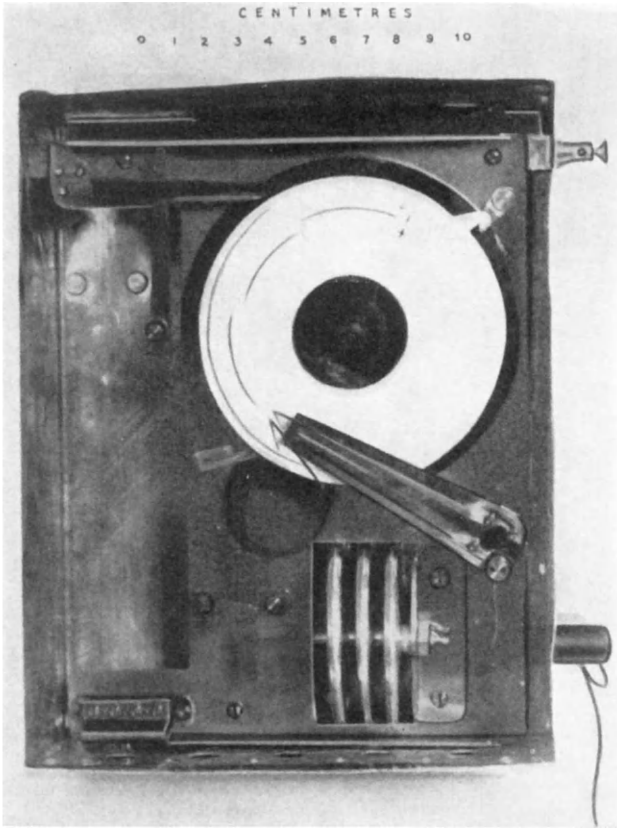


FIG. 46. The alti-electrograph. (From SIMPSON and SCRASE, 1937, Plate 10, Fig. 1, following p. 314.)

was not possible to make direct measurements of the magnitude of the point-discharge currents, but only of their direction.

They used "pole-finding paper", which is paper impregnated with potassium ferrocyanide and ammonium nitrate and used with iron electrodes, to indicate the direction of the current flowing between 4 points just below the balloon and a point 20 m lower on a wire trailing below the balloon. In addition, the balloon carried a hydrograph to give the humidity and an aneroid to give barometric pressure and hence height. Further, it carried an aneroid release, so that the apparatus would fall by parachute after reaching a predetermined height, and the records were recovered.

The instrument gave definite results for the sign of the potential gradient at different heights in thunder clouds and storm clouds, but was not sufficiently sensitive to give records in continuous rain except occasionally. The results will be discussed with other observations on the electrical structure of thunder clouds in Chapter 12. The apparatus was called the "alti-electrograph".

9.21. Width of Alti-electrograph Traces

The alti-electrograph was designed merely to determine the direction of the vertical potential gradient in and near clouds; but it was found that the width of the trace produced could be related to the current flowing through the apparatus and this raised the possibility of using the records to determine the magnitude as well as the direction of the potential gradient, if the point-discharge current can be related to the potential gradient.

SIMPSON and SCRASE (1937) assumed that the points on the alti-electrograph act similarly to a single discharging point near the ground and obey the law of § 9.15., $I = a(F^2 - M^2)$, found by WHIPPLE and SCRASE (1936), but with a correction for pressure as determined by TAMM (1901) in laboratory experiments. To obtain the constant a, they made measurements of the width of the trace and the potential gradient at the ground when the ascent of the balloon commenced.

The results obtained gave values of the potential gradient within thunder clouds which were considerably lower than anticipated and lower than found more directly by GUNN (1948). This has given rise to considerable controversy as to the interpretation of the alti-electrograph results, in particular as to whether the WHIPPLE–SCRASE law can be applied. Since the balloon is carried very largely by the wind and has only a small velocity relative to the air, it would seem that, in CHAPMAN's formula (see § 9.16.), the term v would be proportional to F and an F^2 formula is correct. CHAPMAN pointed out that the application of his formula to his own measurements (CHAPMAN, 1953), with a radio-sonde, would lead to a correction giving potential gradients within the cloud more closely in accord with expected values; but a similar correction does not apply to the alti-electrograph, unless there was an unexpected error in the original calibration to give a; it is conceivable that the relation between the current in the alti-electrograph and the potential gradient would be different close to the ground from that at a higher level, or there might be a sufficiently great increase in potential gradient between the ground and the level at which the first measurements were made, so that the calibration would have an error.

It should, however, be emphasized that, whatever conclusion is reached about the correctness or otherwise of the interpretation of the widths of the alti-electrograph traces, this cannot detract from the value of the alti-electrograph for its original purpose of determining the sign of the vertical potential gradient in and near thunder clouds.

9.22. Radio-sonde Method

An adaptation of the alti-electrograph has been described by KREIELSHEIMER (1947) and CHAPMAN (1952, 1953). In this, instead of using the pole-finding paper, the current through the points is used to give one of the indications of a radio-sonde, transmitting a signal depending on the current. The only results by KREIELSHEIMER's method, published by KREIELSHEIMER and BELIN (1946), indicate a polarity of the cloud, in New Zealand, opposite to that found by the alti-electrograph in England; the signals at times were too large for the apparatus, so no results are available as to the maximum potential gradients reached. CHAPMAN (1953), in America, reported results similar to those from the alti-electrograph and a calibration of potential gradient, using a field machine, suggested values of the same order as obtained from the alti-electrograph; his maximum value of potential gradient was 21,000 V/m. But this has been corrected as discussed in § 9.21.

9.23. Potential Gradients below Clouds

The theory of § 9.18. shows that for widespread point discharge there should be a considerable space charge in the region below the cloud and hence a considerably greater potential gradient just below the cloud than at the earth's surface. The exact extent of this increase depends on the point-discharge current density and so on the effective separation of discharging points similar to the one used (see § 9.24.).

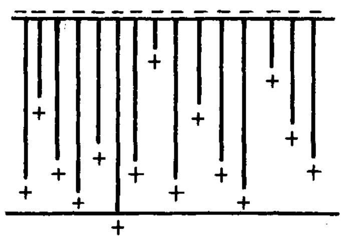


FIG. 48. Lines of force and space charges below thunder cloud.

But the measurement of the widths of the alti-electrograph records show little, if any, increase of potential gradient with height below clouds, and certainly much less than would be expected with any reasonable estimate of the effective separation of the points. In a negative potential gradient, there is certainly a point-discharge current taking negative charge down to earth and liberating positive ions; if these ions do not rise up towards the cloud in the form of a space charge, giving an increase of potential gradient, then the question arises as to where else they could go. Below a thunder cloud, they might diffuse or be blown sideways, but point discharge is sometimes found below clouds giving widespread continuous rain (e.g. CHALMERS and LITTLE, 1940) and the theory of § 9.18. is then more closely applicable than below thunder clouds. In such cases the increase of potential gradient should be very clearly shown by the alti-electrograph, but, in fact, in similar conditions, no such effect has been found.

No satisfactory solution of the problem has yet been put forward. CHALMERS (1939) considered the effect of negative ions coming from the cloud to remove the space charge, and also (1944) the capture of ions by falling drops or their removal by upward air currents, but none of these is sufficient to account for the absence of the expected increase of potential gradient with height.

The remaining possibility is that there has been a wrong interpretation of the widths of the alti-electrograph records in terms of potential gradients. A mere divergence of the relation between point-discharge current and potential gradient from that of WHIPPLE and SCRASE (1936) could not account for the fact that the majority of the alti-electrograph records show little or no increase of current with height, unless the true form of the relation somehow involves the height of the balloon. It has been suggested that perhaps the current through the alti-electrograph actually measures, not the potential gradient, but the vertical ionic current density, which would be reasonably constant below a cloud. But, up to the present, there has been no theoretical support for this suggestion.

The idea that the potential gradient does increase with height and that the interpretation of the widths of the alti-electrograph records is incorrect is supported by two other lines of evidence. The results on the relation between rain current and point-discharge current are interpreted (see § 16.20.) in terms of the picture of § 9.18. and do not appear capable of interpretation otherwise. Also the change of potential gradient due to a lightning flash can be explained in terms of a space-charge blanket below the cloud (see § 12.10.) but not otherwise.

9.24. Effective Separation of Discharging Points

As already discussed, the use of the results of currents from a single point to give the effect of point discharge in the balance of currents to and from the earth requires a factor to convert from the current through a point to the current density over an area. This is provided if we know the effective separation of other points, assumed in rectangular array, equivalent to the point used for measurement. The discussion of the relation between rain current and point-discharge current also involves the effective separation.

Early estimates of the effective separation were merely based on counting trees, etc., except for the case of SCHONLAND (1928b), discussed already (see § 9.8.). SIMPSON (1949), from his measurements of rain current, obtained a value for Kew of 22 m to compare with WHIPPLE and SCRASE's (1936) rough estimate of 25 m (see § 10.20.).

CHALMERS (1951), using SIMPSON's rain results in more detail, obtained a value for Kew of 14·5 m, and using the results of HUTCHINSON and CHALMERS (1951) for Durham a value there of 6·1 m (see § 10.20.). SMITH (1951 a) at Cambridge used the results of WORMELL (1939) for potential-gradient changes due to near lightning flashes and was able to deduce the current density and thus the effective separation for his point (not the same one as used by WORMELL (see § 12.12.). CHALMERS (1953 c) used the electrical model of a thunder cloud, given by SIMPSON and ROBINSON (1940), to account for potential gradients at the earth's surface, and the relation of WHIPPLE and SCRASE (1936) between point-discharge current and potential gradient, so as to obtain the total point-discharge current below a thunder cloud in terms of the effective separation of points. If it is now assumed that this current is equal to the average current above a thunder cloud, found by GISH and WAIT (1950) to be about 0·5 A a value of 11 m is found for Kew. Using the value of 1·3 A found by STERGIS, REIN and KANGAS (1957 b), the figure is about 7 m.

If the natural discharging points were actually all of the same height and in rectangular array, the term "effective separation" would have a well-defined meaning, but since, in fact, there are different points, at different heights, which become effective as dischargers at different potential gradients, it is probable that the ratio of the current through a single point to the total current over an area is dependent on the value of the current, and so on the potential gradient, so that the idea of a fixed "effective separation" is only approximately true.

9.25. Total Current below Cloud

It has often been suggested that the total point-discharge current below a cloud should depend very considerably upon the number and nature of the points at the earth's surface below the cloud. For example, WICHMANN (1951) has objected to the inclusion of sea thunderstorms in considering the transfer of charge between the clouds and the earth, because there can be little point discharge.

If a cloud produced a definite potential gradient at the earth, irrespective of the point-discharge current, then this conclusion would be justified; but this neglects the effect of the space charge produced by the point discharge. A better approach is to consider that the cloud produces a definite rate of separation of charge and so, if conditions become quasi-static, a definite current is required to dissipate these charges. The particular conditions below the cloud will not alter greatly the total dissipating current below the cloud, but the actual potential gradient at the earth's surface is merely incidental; it is that potential gradient which is necessary, under the existing conditions, to produce the point-discharge current required. Just as we have seen in § 2.25. for fine weather it is much simpler and more fundamental to discuss matters in terms of currents rather than of potential gradients and space charges, so here, where space charges become more important since currents are unidirectional, it is again simpler and more fundamental to discuss currents.

The total current beneath a cloud is very nearly the same whatever the conditions at the earth's surface, as has been shown in more detail by CHALMERS (1952c). This, incidentally, shows that a lightning conductor is almost entirely without effect in dissipating the charge in the cloud (see § 14.51.).

CHAPTER 10

Precipitation Currents

10.1. Importance of Rain Electricity

KELVIN (1860b) appears to have been the first to suggest the measurement of the electric charge on rain, but he was not successful with his "electropluviometer".

The subject first became important when the suggestion arose that rain and other forms of precipitation might be important in the maintenance of the negative charge on the earth.

Another point of interest in regard to the charges on precipitation is the question of their origin. It is generally recognized that water drops and ice particles, which either fall as snow or hail or melt to give rain, may play an important and possibly fundamental part in the separation of electricity in clouds; therefore measurements which give information about the charges on rain and snow may help in elucidating the processes at work in the separation of charges. The charges on raindrops and snow flakes must originate either in the clouds or in the air below, and any fresh knowledge about these charges and their relation to other factors can be expected to contribute to the discussion of the general problem of the production of electric charges in the clouds.

When conditions during rainfall are fairly steady, it is possible to apply the principle of the quasi-static state (see § 2.23.), so that the total vertical current will be the same at all levels, provided that horizontal currents can be neglected. At the earth's surface, the total current comprises the conduction current, the precipitation current and perhaps a point-discharge current and also lightning. In estimating the total, a knowledge of the precipitation current is needed.

10.2. Measurements of Rain Charge by Electrometer

Most of the measurements of rain charges have been made with an insulated collector connected to an electrometer. The collector is shielded from the earth's field, to avoid any effect due to the change of bound charge when the potential gradient alters. Care must be taken that there shall be no effects due to splashing, since LENARD (1892) found that the splashing of water gives rise to electric charges, both on the water drops and in the air; so it is necessary to ensure that drops falling on the shielding vessel do not splash into the collector and also that drops falling into the collector do not splash out.

The deflection of the electrometer is recorded or noted at regular intervals and returned to zero, a fresh measurement then beginning. Since an electrometer has only a certain range of readings between the smallest which is measurable and the largest which remains on the scale, currents only within a certain range of values can be measured; when the measurements are made by eye, rather than by automatic recording, the range of sensitivities can, in some cases, be adjusted according to the conditions, or the time between zeros may be varied.

Measurements of the type discussed were first made by ELSTER and GEITEL (1888), who exposed their collector to the rain for periods varying between 5 sec and 2 min, depending on the conditions; eye observations were used; KÄHLER (1908) and SCHINDELHAUER (1913) used a similar method, noting the electrometer readings every minute. SIMPSON (1909) recorded and earthed the electrometer every 2 min. BENNDORF (1910) used two collectors which were exposed alternately for 10 min and gave automatic recording of the potential acquired in that time. BALDIT (1911–12) used observations by eye every 15 sec, criticizing those who had used longer periods, since the current often changes sign many times in 2 min; he allowed his electrometer to charge up, observing every 15 sec until the deflection became too large, when he earthed the electrometer. BERNDT (1921) measured the charge collected in 5 min. BANERJI (1932, 1938) measured the rain charge for a few thunderstorms, using SIMPSON's (1909) method. SIMPSON (1949) used continuous recording, earthing every minute.

Since the greatest rain current tends to occur with the greatest rate of rainfall, MCCLELLAND and NOLAN (1912), followed by MCCLELLAND and GILMOUR (1920), MARWICK (1930) and SCRASE

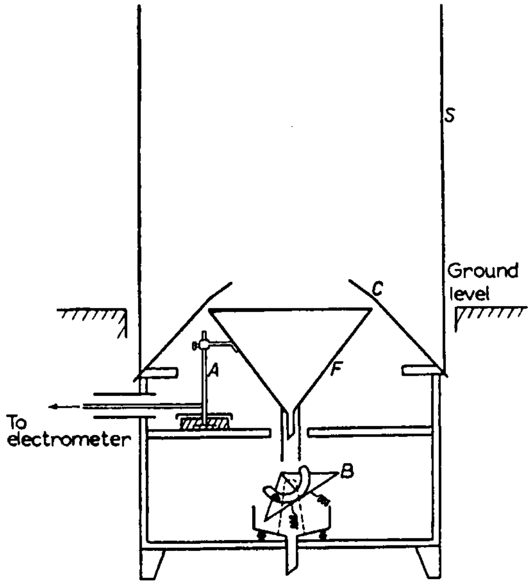


FIG. 49. Scrase's apparatus for rain current. (From SCRASE, 1938, Fig. 1, p. 4.)

(1938) preferred to measure the charge carried by a definite quantity of rain, rather than that received in a definite time. In this type of arrangement, the rain falls into a bucket, shielded from the earth's field and protected from the effects of splashing; when the bucket is full, it tips up and the electrometer is earthed, observation being made of the charges received between two earthings.

SIMPSON (1909) and SCHINDELHAUER (1913), alone of the earlier workers, made continuous measurements, all the other workers measuring only at selected times. More recently, SCRASE (1938) and SIMPSON (1949) made continuous measurements by photographic recording and so had records not only of the average currents over a period, but also of the instantaneous currents.

10.3. Direct Measurement of Current

GERDIEN (1903) measured the actual current density, connecting his collector to an electrometer with a shunting resistance of $10^{12}\Omega$, so that the electrometer deflection never becomes large.

BENNDORF (1912) criticized this arrangement, since accidental resistance changes and leakages would be of the same order as this resistance and the method could not be relied on.

Modern insulators and resistors have now made the same type of method available and STOCKILL (1954) used a d.c. amplifier of high stability developed by KAY (1950), with a resistor of $10^{11}\,\Omega$ and a negative feed back circuit to give a direct measure of the rain current. RAMSAY and CHALMERS (1960) used a vibrating-reed electrometer for a similar purpose, and JOLIVET (1959) a d.c. amplifier.

10.4. Completely Exposed Receiver

If the main point of interest in the precipitation current measurements is the question of the part played by rain in the maintenance of the earth's charge, then it is clear that the shielded receiver used in the methods described above is not very satisfactory, for what is measured is the charge on the actual raindrops (and not all of them), whereas in natural conditions there may be other processes associated with the rain which contribute to the total transfer of charge to the earth; if, for example, the conduction current during rainy weather differs widely from that in fine weather, this current might be of magnitude comparable with the rain current but would not be registered by a shielded receiver. Also there may be effects due to splashing which would be excluded by a shielded receiver, though existing in natural conditions. The actual transfer of charge to the earth can be measured only by isolating a portion of the earth's surface and determining the charge received by it in conditions as close to the natural as possible.

An important disadvantage of the shielded receiver is that the results may not represent a fair sample of the rain, for driving rain would not enter the receiver. SCRASE (1938) found that his apparatus received only half the amount of rain caught by a standard rain gauge. This difference would become serious if the charges on drops depend on the size, since the smallest drops are more easily removed by the wind and would be excluded from a shielded receiver.

The use of a completely exposed receiver involves a "displacement current" as discussed in § 7.5., when there are potential-gradient changes and hence some compensation for these must be used to obtain the true current.

The first use of a completely exposed receiver was probably that of WEISS (1906), who used a wire brush to catch the raindrops, but his results were vitiated by the collection of a charge, either by point discharge or by displacement current, even when there was no rain. KOHLRAUSCH (1909) avoided this by shielding the brush from the atmospheric field and thus his method comes into the previous class.

HERATH (1914) used a large area of cloth (24 m^2), suspended by insulators and connected to a galvanometer, the other side of which was earthed; the deflection of the galvanometer was recorded photographically on a moving drum. This method suffers under the disadvantage that the conditions are not quite the natural ones in regard to effects of splashing, etc.; no attempt was made to compensate for potential-gradient changes.

WILSON (1916) suggested that conditions would approach the natural, and so give only the natural effects of splashing, etc., if there were a guard ring round the collecting plate and if the plate and guard ring were covered with grass or soil. WILSON himself made a few observations with such an arrangement and it was used by SCHONLAND (1928b) for observations associated with thunderstorms; he used a capillary electrometer and photographic recording. CHALMERS and LITTLE (1940, 1947) used the same method for simulating natural conditions, but collected the charge received for 10 min and then discharged it through a galvanometer, with photographic recording. In none of these methods was any attempt made to compensate for potential-gradient changes.

CHALMERS (1956) used a similar method, with a simple electronic amplifier, to measure the charge received in 4½ min; in this case, compensation for potential-gradient changes was provided by simultaneous measurements of the potential gradient, as in SIMPSON's method for the air–earth current (see § 8.6.). ADAMSON (1960) used a d.c. amplifier to measure the current flowing, with compensation from a field mill as for the air–earth current (see § 8.7.) and the same apparatus was used by RAMSAY and CHALMERS (1960).

SIMPSON's (1949) measurements were made with a collector which received as much rain as a standard rain gauge and so was not completely shielded from potential-gradient changes; this method thus comes into a category between those of this section and § 10.2.

Compensation for both displacement and conduction currents could be obtained by comparing the current to an exposed receiver with that to an inverted plate nearby; if the height of the inverted plate were adjusted so that its reduction factor (see § 5.18.) was unity, the same areas would be used for both receivers and the effects of local space charges near the ground would be the same. If, in addition, a field machine were used to measure field changes alone, the conduction current near the ground during rain could be measured.

10.5. Single Drops

If the main object of the investigation of rainfall electricity is considered to be the question of the origin of the charges on the raindrops, then it is probable that more information can be obtained by measuring the charges on the individual drops, rather than by making observations which give only the average of a large number of drops. In addition to the drop charge, there are other parameters which are measured, in particular the size of the drop and the potential gradient.

Single-drop charges were first measured by GSCHWEND (1922), who used a collector connected to a sensitive electrometer, as did BANERJI and LELE (1952). CHALMERS and PASQUILL (1938), HUTCH-

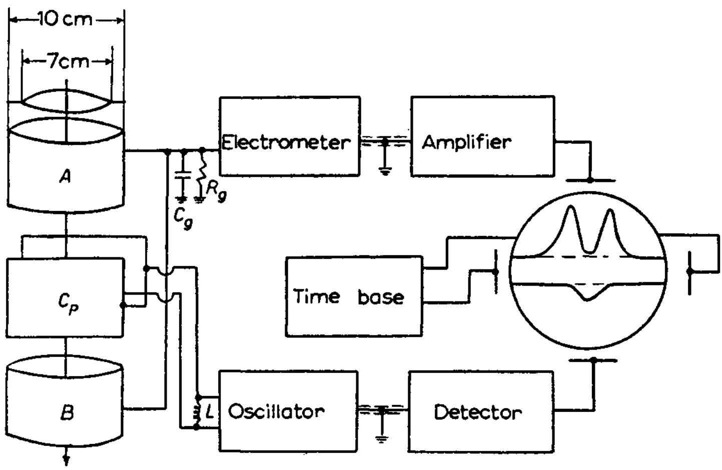


FIG. 50. Apparatus for measuring single-drop charges. (From SMITH, 1955, Fig. 1, p. 24.)

INSON and CHALMERS (1951), GUNN and DEVIN (1953), MAGONO, ORIKASA and OKABE (1957) and ARABADJI (1959) used electronic amplifiers in place of the electrometer. Instead of catching the drops in a collector and obtaining a single pulse, GUNN (1949), FEDEROV (1951), SMITH (1955) and JOLIVET (1959) allowed them to pass through a ring on which charges became induced, and thus obtained a double pulse, which was amplified and displayed on a cathode-ray oscillograph. SMITH (1955) tilted his receiver into the wind, so as to collect a fair sample of the drops.

To measure the sizes of the drops, GSCHWEND (1922) and HUTCHINSON and CHALMERS (1951) used the method of WEISSNER (1895), collecting the drops on prepared paper and measuring the stain; GUNN (1949) used a somewhat similar method with photographic fixing of the stain. BANERJI and LELE (1952) allowed the drops to fall into the receiver of a fine-bore manometer. GUNN (1949) and SMITH (1955) used two induction rings and found the time interval between the two pulses, giving the rate of fall and hence the size of the drop; JOLIVET (1959) used a wide induction ring and measured the breadth of the pulse to give the speed and hence the size. For larger drops, for which the rate of fall does not vary much with the size of the drop, SMITH (1955) used the variation in the capacity of a parallel-plate condenser through which the drop fell. ARABADJI (1959) used the impulse of the fall of the drop on a piezoelectric crystal.

GSCHWEND (1922), GUNN and DEVIN (1951), BANERJI and LELE (1952), SMITH (1955) and MAGONO, ORIKASA and OKABE (1957) all measured the potential gradient at the same time with independent apparatus. HUTCHINSON and CHALMERS (1951) had their receiver somewhat exposed to the atmospheric field and measured the potential gradient from the bound charge on covering and uncovering the receiver; they also measured the point-discharge current independently.

BANERJI and LELE (1952) and, for short periods, SMITH (1955) made observations, at the same time as with single drops, on the rain current to a shielded receiver.

GUNN (1947, 1950) carried his apparatus on an aircraft, using a single ring to measure charges; he was not able to measure sizes of drops. FLUEGGE and PILIÈ (1965) used similar apparatus in aircraft and measured drop size by measuring the light scattered by the drop.

10.6. Need for Continuous Observations

Even more, perhaps, than in any other measurements of atmospheric electricity, it is necessary to have continuous observations of rainfall electricity over a considerable period of time before it is possible to reach any definite conclusions in regard to the resultant transfer of charge to the earth by rain. SCRASE (1938) pointed out that the great range of the currents must mean that the results of measurements on rain can be valid for general conclusions only if great care is taken to avoid missing any part of the rain during the period investigated, for he found that as much as half the total negative charge measured was contributed by as little as 2 per cent of the total rain collected. Thus any eye observations might easily miss an important part of the total charge. And it is in conditions of very heavy rain that insulation might break down most readily and cause defective records. This conclusion is reinforced by the result of CHALMERS and LITTLE (1947), who found that in one 10 min period of soft hail there was an average current of -7×10^{-8} A/m^2, this being larger, by a factor of some thousands, than the normal results for rain. CHALMERS and LITTLE (1947) also pointed out that a few days in a period contributed a quite disproportionate amount to the total charge collected, and if these had been accidentally missed, the whole conclusion as to the balance of charge brought down might have been altered. SCRASE's (1938) results show that a long period of time is needed for general conclusions, for in two successive years, when little rain was missed, the ratios of positive to negative charges were 0·7 : 1 and 1·4 : 1 respectively.

It is clear that local conditions are likely to play an important part in the electrical effects of rain, and measurements are needed in many parts of the world, with widely differing conditions, before it will be possible to state with any degree of certainty what part the charges on rain and other forms of precipitation play in the maintenance of the earth's charge.

10.7. Classification of Rain Results

Before the general results are discussed, it is desirable to warn the reader that they are obtained from observations made over relatively short periods of time and in few parts of the world, so that future results may show that the results to date do not form a representative selection. In particular, it should be realized that few observations have been made in the very intense rains of tropical regions and few, too, in the particular conditions of mountainous country, while none at all appear to have been made in polar regions.

It should also be pointed out that different observers have all used different areas of receiver, with different arrangements of shielding, and different times of recording, so that their results are not strictly comparable.

Many observers have divided rain into three distinct types: (a) "continuous rain", "ordinary rain" or "non-stormy rain"—German *Landregen*; (b) showers or squalls (without thunder or lightning)—German *Boenregen*; and (c) storm or thunder rain—German *Gewitter*.

10.8. Ratio of Positive to Negative Charge

The measured ratio of positive to negative charge depends very much on the conditions of measurement. The only truly correct measurement of this ratio is with single drops as, in any other method, each observation is likely to include drops of both signs. The larger the number of drops included in the measurement, the more probable it is that the drop sign which is generally most common will predominate in each measurement. SIMPSON (1942) has taken an example in which the true ratio of the signs of the charges of single drops is 2 : 1, but where grouping of the drops in fours gives a ratio of 3·7 : 1 and in eights 4·2 : 1.

For single drops, GSCHWEND (1922) found a ratio of 1·5 : 1 (positive : negative); CHALMERS and PASQUILL (1937) found 1·30 : 1, or omitting snow, sleet and hail 1·23 : 1. In stormy rain, ARABADJI (1959) found 1 : 2, and in thunderstorms, GUNN (1949) found 1·6 : 1 and GUNN and DEVIN (1953) 1·2 : 1.

For measurements of many drops at once, all the observers except ELSTER and GEITEL (1888), GERDIEN (1903), BANERJI (1932, 1938) and JOLIVET (1959) have found an excess of positive charge, but the actual ratio has depended on the conditions of measurement, varying from 1·10 : 1 (SCRASE, 1938) and 1·36 : 1 (BALDIT, 1911, 1912) to 15 : 1 (HERATH, 1914) and 30 : 1 (SCHONLAND, 1928b, in a few thunderstorms).

In nearly all cases in which steady rain has provided most of the observations, a positive excess of charge has been found and one may conclude that it is probable that the total electricity of precipitation is positive. But the places and conditions of measurement may not be a fair sample for the whole world.

Results for continuous rain show an excess of positive charge usually more marked than in the other types of rain. There are many occasions when continuous rain is found to be positively charged over a long period of time; one of these has been discussed in detail by Chalmers and Little (1940).

Squall or shower rain has usually shown an excess of negative charge, and thunderstorm rain often, but not always, a positive excess, though not in each individual storm.

10.9. Times of Rainfall and Quantities of Water

In general, it has been found that the periods during which there is positive rain amount to a longer time than those during which there is negative rain, and often the ratio of these has been found to be greater than the ratio of the total quantities of charge, showing that periods of negative charge tend to bring more highly charged rain than periods of positive charge. As in the case of the total quantities of charge, the actual values have little meaning, since they depend on the conditions of measurement. There is some evidence that the ratio of the periods of positive and negative rainfall depends on the rate of rainfall; for example, Simpson (1909) found that the ratio increases as the rate of rainfall increases.

The ratio of the positive to negative quantities of water is greater than the ratio of the total quantities of charge, showing that negative drops tend to have a greater density of charge than positive. Scrase (1938) found that the ratio of positive to negative quantities of water depends on the rate of rainfall, being larger for both high and low rates of rainfall than for intermediate rates.

The charges per unit volume of water have been measured by most of the observers who have measured the rate of rainfall. The results by different observers and for different kinds of rain vary from 3×10^{-6} C/m^3 to 7×10^{-3} C/m^3. Average values for continuous rain are of the order of 10^{-4} C/m^3, while in showers and storms the averages reach nearly 10^{-3} C/m^3. As for the previous factors discussed, the mixture of positive and negative charges caused the values to be less than for single drops and to depend upon the number of drops included in each measurement.

10.10. Numbers of Single Drops

Those who have measured single drops have given the ratio of positive to negative drops; often this ratio is greater than the ratio of the total charges, showing again that those drops which carry negative charge are, on the average, more highly charged than those carrying positive charge. Results are shown in Table 2 (drops with zero charge, when detected, are not included in the ratio).

TABLE 2

	Total number of drops	+ve/−ve
GCHWEND (1927)	1,537	1·77
CHALMERS and PASQUILL (1938)	17,582	1·71
GUNN (1949)	159	0·77
HUTCHINSON and CHALMERS (1951)	913	0·95
FEDEROV (1951)	5,863	1·31
BANERJI and LELE (1952)	3,748	0·80
GUNN and DEVIN (1953)	7,279	1·00
ARABADJI (1959)	982	0·83
JOLIVET (1959)	18,000	2·5
KRASNOGORSKAYA (1961) (including snow)	83,743	1·007

10.11. Vertical Rain Currents

The vertical rain current is another factor which depends upon the number of drops included in the observation; the fewer the drops, the larger are the measured currents likely to be. The measurement of rain current does not require the knowledge of the rate of rainfall.

In general, it has been found that the currents in continuous rain vary from about 10^{-12} A/m^2 up to nearly 10^{-10} A/m^2, while in squall, showers and storms the currents become as high as 10^{-8} A/m^2. The highest recorded currents are one of $2{\cdot}0 \times 10^{-7}$ A/m^2 given as doubtful by SCHONLAND (1928b) and one of $-7{\cdot}3 \times 10^{-8}$ A/m^2 in soft hail observed by CHALMERS and LITTLE (1947).

In all cases where shielding excludes some of the rainfall, the rain currents measured would be less than the true rain current to an exposed area of the earth, if the rain collected is a fair sample.

10.12. Charges on Single Drops

The values of the charges on single drops, as measured by various observers, show a very wide range, from $\pm 3 \times 10^{-15}$ C, the smallest charge measurable by FEDEROV (1951), up to over -10^{-10} C, measured by BANERJI and LELE (1952)

Probably because of different lower limits, different observers have found very different mean values, FEDEROV's being about 10^{-13} C and BANERJI and LELE's 10^{-11} C; that the lower limit must be important is shown from the results of HUTCHINSON and CHALMERS (1951), whose lower limit was 3×10^{-14} C and who found 561 drops out of 1476 to have charges below this limit.

Within clouds, GUNN (1947, 1950) found charges up to 5×10^{-11} C, considerably greater than values found with similar apparatus by GUNN and DEVIN (1953) on the ground during thunderstorms.

One of the most striking features of the results for single drops is the fact that there are almost always drops of similar sizes, arriving at nearly the same time, with charges of different magnitudes and even of different signs, although they must be presumed to have been very close to one another throughout their fall; this is most noticeable in the measurements of SMITH (1955), who measured a considerable number of drops in a short period of time (see § 10.17.).

With a range of over 10^4 in the values of the charges measured at the ground, it is clear that it is not possible to use the same apparatus for the measurement of charges of all magnitudes, unless some range-changing device is available. Within one period of rainfall, the range of charges is much less wide, but may still be too great to be covered adequately by a single measuring device.

10.13. Relation between Precipitation Charge and Potential Gradient

Many observers have discussed the question of a relation between rain charge and potential gradient. Among the earlier workers, ELSTER and GEITEL (1888) and BENNDORF (1910) found that in many cases there is a definite inverse relation between the potential gradient/time and precipitation current/time curves, and SIMPSON (1909) found a general excess of positive rain charge and of negative potential gradient during rain. However, many later workers found no evidence for any such relation; for example, SCRASE (1938) found that both for storm rain, mainly positively charged, and for shower rain, mainly negative, there was about the same proportion of time during which the potential gradient was negative; and GUNN and DEVIN (1953) measured the charges on single drops in two thunderstorms and found no relation between the sign of the charge and the sign of the potential gradient.

On the other hand, SCRASE (1938) and CHALMERS and LITTLE (1940) noticed cases of long-continued positive rain and negative potential gradient, and in one of SCRASE's instances both the rain current and the potential gradient changed sign for the same short period. NOTO (1939) found opposite signs for rain charge and potential gradient in 530 intervals out of 622; HUTCHINSON and CHALMERS (1951) and MAGONO, ORIKASA and OKABE (1957) obtained similar results with single drops; mountain measurements (see § 10.19.) also show the same.

SIMPSON (1949) and SIVARAMAKRISHNAN (1961) obtained conclusive evidence of an inverse relation between rain current and potential gradient, when conditions are reasonably steady, particularly with high potential gradients, and they found that the two often change sign in opposite directions at the same time (the "mirror-image" effect, see § 10.27.). For lower potential gradients, SIMPSON (1949) still found that there is a relation between rain charge and potential gradient, but the line does not go through the origin, so that there are instances when both rain current and potential gradient are positive. RAMSAY and CHALMERS (1960) found that the relation between rain current and potential gradient is much more evident in winter than in summer.

KRASNOGORSKAYA (1961) expressed similar results in terms of an inverse relation between potential gradient and space charge of precipitation.

10.14. General Interpretation of Relation between Charge and Potential Gradient

There are two main ways in which the relation between rain charge and potential gradient can be interpreted and, as we shall see, there is some reason to believe that one is correct in the case of high values of potential gradient and the other for low values. In the first place, it can be considered that there is some process which gives a charge, usually positive, to the rain, the opposite charge remaining in the cloud or possibly in the air below; this latter charge would then give rise to a potential gradient at the ground of sign opposite to that of the rain. This is the interpretation preferred by SCRASE (1938) and assumed by BANERJI and LELE (1952); as we shall see in § 10.22., this may be correct for steady rain with low potential gradients.

On the other hand, the potential gradient may be set up first and the raindrops then acquire their charges by the capture of ions moving in this potential gradient, according to the theory put forward by WILSON (1929) (see § 3.11.). This is the theory used by SIMPSON (1949), CHALMERS (1951), SMITH (1955) and SIVARAMAKRISHNAN (1961).

Provided it is assumed that there is only one process of charge separation at work, it is possible to conclude that, if conditions are reasonably steady and we can apply the principle of the quasi-static state (§ 2.23.), the first interpretation leads to a current which is usually positive downwards at all levels, i. e. of the same sign as the rain. On the other hand, if the rain charge is derived from the ionic current, the total rain current can never be larger than the ionic current, so that, when the rain is positive, the total current is negative downwards in this case. A determination of the total current can thus distinguish between the two interpretations, as was pointed out by CHALMERS and LITTLE (1940).

10.15. Space Charge of Rain

In any discussion of the relation between rain charge and potential gradient, it may be necessary to take into account the effect upon the potential gradient at the ground of the actual charges falling in the rain.

To simplify the discussion, let us assume a widespread cloud from which there falls a rain current of density J A/m^2, the drops falling with a velocity of v m/ecs. The space charge density of the rain is then J/v C/m^3 and so the rate of change of potential gradient with height dF/dh is $-J/\varepsilon_0 v$.

Taking a value of 10^{-11} A/m^2 for J and 4 m/sec for v, dF/dh is about 1/3·5 V/m per m. Thus if the rain carries the same charge down through a height of 1 km and if there is no ionic space charge, the potential gradient at 1 km differs from that at the ground by nearly 300 V/m.

MAGONO and ORIKASA (1961) have gone into the matter in more detail and showed that the space charge of the rain may account for some divergences from the "mirror image effect" (see § 10.27.).

10.16. Relation between Rain Current and Point Discharge

During the periods when the potential gradient was large enough to give point discharge, SIMPSON (1949) was able to compare the rain current with the point-discharge current through a single point. He found that, for any one rate of rainfall, the rain current is proportional to the point-discharge current and of opposite sign, the factor of proportionality increasing with the rate of rainfall. SIMPSON's results could be expressed by any of the following:

$$J/I = -2{\cdot}0 \times 10^{-4}(R')^{0{\cdot}57} \{-9{\cdot}734 \times 10^{-8}(R')^{0{\cdot}655}\},$$

$$J/I = -\frac{1}{5{\cdot}5 \times 10^2}(1 - e^{-0{\cdot}058R'}) \{-0{\cdot}8 \times 10^{-6}(1 - e^{-0{\cdot}093R'})\},$$

or

$$J/I = -\frac{1}{400}\left(\frac{R'}{R' + 20}\right)\left\{-1{\cdot}61 \times 10^{-6}\left(\frac{R'}{R' + 6}\right)\right\},$$

where J is the rain current in A/m^2, I the point-discharge current in A and R' the rate of rainfall in mm/hr. SIVARAMAKRISHNAN (1961) obtained similar results but with different numerical values, partly because of a different point; his values are given in brackets above.

HUTCHINSON and CHALMERS (1951) compared the charges on single drops with the point-discharge current. They found that, for any one size of drop, the charge is proportional to the point-discharge current and of opposite sign, the factor of proportionality increasing with the size of the drop. If it is assumed that the drop is of the average size for the particular rate of rainfall, then the relations of BEST (1943, 1950a) between these can be used to compare the results of SIMPSON and those of HUTCHINSON and CHALMERS; they are found to give good agreement provided it is realized that what is significant is the point-discharge current density, not the current through a single point and that the effective separations differ in the two cases.

10.17. Smith's Single-drop Observations

SMITH (1955) measured the charges of a large number of drops during a short interval of time, and obtained a distribution curve of charge with mass; Figure 51 is a typical example of the charges observed during a period of 2 min, in which the potential gradient had a large negative value which remained fairly constant. These results show the smaller drops to be of charge opposite in sign to

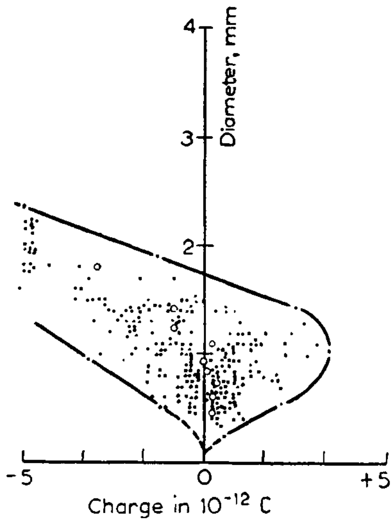


FIG. 51. Charges on single drops. (From SMITH, 1955, Fig. 9A (No. 24A), p. 31.)

the potential gradient, while the larger drops have the same sign of charge as the potential gradient.

SMITH found that, on adding together all the charges measured during a period, he usually obtained a total with a sign the same as the potential gradient. This is in contradiction to the results of SIMPSON (1949) who found opposite signs for rain charges and potential gradient, and SMITH found that when he used apparatus similar to that of SIMPSON he obtained results similar to those of SIMPSON. SMITH explained the discrepancy by pointing out that he was unable to measure the charges on the smallest drops and these would be expected to carry a charge of opposite sign to the potential gradient; his limit of charge measurement was about 3×10^{-13} C, so that a considerable number of drops might have been missed and would yet contribute to the total current.

These results may explain why the inverse relation between rain current and potential gradient has been found in some measurements and does not appear in others. SMITH's results suggest that this inverse relation arises mainly from the smallest drops, and if there is an appreciable wind these drops would be most likely to be caught by the shield of the rain collector and so fail to reach the collector itself. Thus the shielding that is used to avoid effects of potential-gradient changes may prevent the appearance of the inverse relation between charge and potential gradient. It may be significant that the inverse relation has been most noticeable in cases without shielding (CHALMERS and LITTLE, 1947; CHALMERS, 1956) or where the shielding is not very effective (SIMPSON, 1949). On the other hand, RAMSAY and CHALMERS (1960) found little difference between the currents to shielded and unshielded collectors.

There appears to be a serious discrepancy between the results of SMITH and those of HUTCHINSON and CHALMERS (1951). In the measurements of the latter which were comparable with those of SMITH, namely those in which the potential gradient was large enough to give point discharge, HUTCHINSON and CHALMERS found that 105 drops had charges of opposite sign to the potential gradient, as against 37 of the same sign, and this preponderance persisted even to the largest drops, contrary to what would be expected from Fig. 51. An explanation lies in the fact that in the period represented by Fig. 51, HUTCHINSON and CHALMERS would have measured at most 6 drops and probably only 2 or 3; these drops would probably have been of near average size, and, remembering that Fig. 51 does not include drops too small for SMITH to measure, it can be seen that the few drops measured by HUTCHINSON and CHALMERS might well be expected to correspond to the lower part of Fig. 51, on the right-hand side of the zero, i. e. of sign opposite to the potential gradient. Larger drops measured by HUTCHINSON and CHALMERS have occurred during heavier rain when the whole scale of the diagram would be enlarged as compared with Fig. 51. It may also be pointed out that exceptionally large drops entering the apparatus of HUTCHINSON and CHALMERS might have hit the rim and would then have splashed and been rejected. The apparent discrepancy may not be a real one, but may be due to poorer sampling in the measurements of HUTCHINSON and CHALMERS than in SMITH's.

It is possible that, had GUNN and DEVIN (1953) also measured drop size, they might have found some correlation between drop charge and potential gradient, but depending on drop size, as in SMITH's results.

10.18. Raindrop Charges in Clouds

GUNN (1947, 1950) used a single inducing ring, carried on an aircraft, to measure the charges on the individual drops in clouds; the method used on the ground to obtain the masses was not applicable to these measurements.

The first measurements were made in a cloud of a weak cold front exhibiting no thundery activity, with the freezing level at 11,000 ft (3350 m). Negative charges averaging $1{\cdot}3 \times 10^{-11}$ C were found from the ground up to 20,000 ft and positive charges averaging $1{\cdot}1 \times 10^{-11}$ C from 10,000 ft to 26,000 ft.

A further set of measurements were made in a thunderstorm at 7 different levels up to 20,000 ft. The freezing level was at 14,000 ft and the maximum electrification, amounting to 9×10^{-11} C per drop, both positive and negative, was found at 7500 ft, where the temperature was +10 °C. Charges of both signs occurred at nearly all levels.

In both sets of measurements it was found that the resultant charges of the drops would give space charges of appreciable magnitude and hence potential gradients much different from those actually measured; thus there must be charges on cloud droplets or in the air, so as to neutralize more closely the drop charges.

PHILLIPS and KINZER (1958) also measured the charges on individual drops in thunder clouds and found large charges of both signs.

GUNN (1952) attempted measurements of the charges on cloud droplets, by the method described in § 13.7., in clouds that were precipitating or about to precipitate; while the results were not so definite as for non-precipitating clouds, there appeared to be a tendency for the larger drops to carry negative charges. In mountain clouds, WEBB and GUNN (1955) found that, when there were traces of rain present, the charges became larger and more variable than without rain, but no definite predominance of either sign could be distinguished.

10.19. Precipitation Charges at Zugspitze

KUETTNER (1950) made measurements of the charges on precipitation at the Zugspitze (10,000 ft or 3000 m approximately) in the lower portions of thunder clouds. He found that the electric charges on solid precipitation particles were nearly always of sign opposite to the potential gradient when temperatures were above −5° C, while at lower temperatures the relation was not so marked. High potential gradients appeared to be associated particularly with graupel and the central lightning area was usually identical with the area of greatest precipitation.

REYNOLDS (1954) also found that, on mountain tops, the precipitation, whether solid or liquid, carried a charge of sign opposite to the potential gradient.

10.20. Explanation of Relation with Point Discharge

SIMPSON (1949) suggested that the relation between rain current and point-discharge current could be explained in general terms if the raindrops acquire their charges by capturing the ions which have been produced by point discharge and which are moving upwards. The theory of WILSON (1929), developed in detail by WHIPPLE and CHALMERS (1944) (see § 3.11.), shows how such ion capture can be calculated, giving charges on the drops depending on the potential gradient. SIMPSON pointed out that this theory could account for the actual rain charges only if the potential gradient is considerably greater at a higher level than that measured close to the earth's surface; such an increase is to be expected from the space charge from point discharge (§ 9.18.), but is not found from the alti-electrograph results (§ 9.23.).

SIMPSON also pointed out that the second and third forms of his relation (§ 10.16.) lead to a limiting value of the ratio of rain current to point-discharge current for high rates of rainfall; this, he suggested, might be accounted for if the heaviest rain could capture all the point-discharge ions moving upwards. In such a case, the rain-current density would be equal to the point-discharge current density, thus giving a method of obtaining the latter, and hence the effective separation of discharging points (see § 9.24.). The value obtained in this way is in reasonable agreement with earlier estimates of the separation.

BEST (1953) showed that if raindrops obtain a charge proportional to the square of their radius, other things being equal, as is predicted by WILSON's theory, then it is possible to deduce a relation between rain current and rate of rainfall similar to SIMPSON's first relation (§ 10.16.), making use of results for the distribution of drop sizes in rainfall of various rates (BEST, 1950a). This was carried further by SIVARAMAKRISHNAN (1960), who showed theoretically that the exponent of R' should be 0·75, in fair agreement with his own measurements (see § 10.16.).

CHALMERS (1951) gave a detailed explanation of the results of both SIMPSON (1949) and HUTCHINSON and CHALMERS (1951) in terms of ion capture. Assuming that the drops start with charges which themselves are the cause of the field below, and taking into account the space charges of ions and of drops, he obtained, by numerical calculation, the drop charges under different conditions and found that both sets of results can be explained if the drops start at a height of around 0·8 km and if the effective separation of discharging points is somewhat less than previous estimates. In this work, only the average drops were considered and deviations from the average were neglected. The height of 0·8 km is rather less than might be expected, but, apart from this, the very much idealized picture gives a reasonable account of the results and this lends support to the idea that this is the correct explanation. SIVARAMAKRISHNAN (1961) carried out somewhat similar calculations but did not reach so satisfactory a conclusion.

SMITH (1955) also made calculations on much the same basis, but dealt with drops of different sizes under similar conditions; he showed that the charges to be expected agreed reasonably well with his observational results, the larger drops retaining their original sign of charge, while the smaller drops change sign, as do the average drops on CHALMERS's theory.

OGAWA (1960) considered the matter rather differently. He assumed that, in the "charging region", the drops would acquire their maximum charge of $-12\pi\varepsilon_0 X a^2$ (see § 3.11.); then he used the results of MARSHALL and PALMER (1948) for the relation between the rate of rainfall and the numbers of drops of different sizes, and the results of BEST (1950b) for the terminal velocity of raindrops, and so deduced

$$J = -2{\cdot}12 \times 10^{-6}\, P_h\, (R')^{0{\cdot}48} + 0{\cdot}54 \times 10^{-6}\, P_h\, (R')^{1{\cdot}05},$$

where P_h is the potential gradient in the charging region.

He then compared his values of P_h with the potential gradient P at the ground and obtained $P_h = 3P^{1{\cdot}3}$. This gave values of P_h which were of the same order as expected from the space-charge theory (§ 9.18.). He next attempted to compare these results with those of SIMPSON (1949) but this is vitiated by the fact that he assumed that point-discharge current is proportional to the square of the potential gradient. In the absence of measurements of wind speed, it is not possible to use the correct formula (see Chapter 9).

While these theories give a general explanation of the relation between rain current, point-discharge current and rate of rainfall or drop size, there remains the fact that it is still necessary to assume that there exists the increase of potential gradient with height which is predicted by the space charge theory (§ 9.17.), but not found by the alti-electrograph (§ 9.23.). Also, on these theories, the rain acquired its charge in accordance with the potential gradient at a height of some hundreds of metres and so at a time of the order of minutes before it reaches the ground; one would therefore expect a delay time of this order between changes of potential gradient and changes of rain current. But results for the "mirror-image effect" (see § 10.27.) do not show such a delay consistently.

10.21. Rain Current with Low Potential Gradients

SIMPSON (1949) also made measurements of the rain current with low potential gradients and found that his results could be expressed by

$$q = -4{\cdot}8 \times 10^{-8} (P - 400),$$

where q is the charge in C/m^3 and P is the potential gradient in V/m.

If expressed in terms of the rain current, J, rather than the charge per unit volume, then J is proportional to the rate of rainfall, R.

SIMPSON suggested that the "400" might represent the normal, fine-weather, potential gradient, so that q is proportional to the deviation from this value.

RAMSAY and CHALMERS (1960) found that the variation of J with R is probably not linear.

SIVARAMAKRISHNAN (1960) found results similar to those of SIMPSON (1949), but q is proportional to $(P - 100)$; his results show that q increases somewhat with rate of rainfall.

REITER (1965) also found the current to depend on $(P - C)$ and in his case C was 40 V/m, agreeing with the local fine-weather potential gradient.

10.22. Total Current with Low Potential Gradient

When the potential gradient is low, the total current close to the earth's surface consists of the precipitation current and the conduction current together with any possible effects of splashing; in these conditions there is no lightning and no point discharge. The total current can be obtained from the measurement of the charge acquired by a completely exposed receiver, as used by CHALMERS and LITTLE (1947) and by CHALMERS (1956), with a correction for the displacement current (see § 8.3.).

CHALMERS (1956) found that, in the great majority of cases of continuous rain with potential gradients between ± 800 V/m, the total current is positive downwards; the potential gradient at the same time was usually negative, most frequently around -100 V/m to -200 V/m. These results show that the rain current is usually greater than the conduction current and therefore, from the arguments of § 10.14, it seems likely that the process of charge separation is one which gives a positive charge to the rain and a negative charge accumulates in the cloud.

The variation of total current with potential gradient was found to be represented by

$$K = -1{\cdot}18 \times 10^{-14}(P - 150),$$

where K is the total current in A/m^2 and P is the potential gradient at the earth's surface in V/m.

RAMSAY and CHALMERS (1960) made similar measurements and found similar results with the constant depending on the rate of rainfall.

From SIMPSON's (1949) results, also, it can be concluded that the total current is usually positive downwards; when the potential gradient is $+400$ V/m, the rain current is zero, and so whatever may be the conductivity, the conduction current is positive downwards. When the potential gradient is zero, the rain current is positive and so in both these cases the total current is positive downwards. Unless the conductivity during rain is very considerably greater than in fine weather, the conclusion can be drawn that the total downward current is positive over the whole range covered by SIMPSON's formula (§ 10.21.).

10.23. Charges on Snow, Sleet and Hail

SIMPSON (1909) found that the charge on snow is more often positive than negative, and that the vertical currents and charges per unit mass are greater for snow than for rain. Agreement with these results has been found by WEISS (1906), GSCHWEND (1927), MARWICK (1930) and CHALMERS and PASQUILL (1938). But, on the other hand, ELSTER and GEITEL (1888), KÄHLER (1908), SCHINDEL-

HAUER (1913), MCCLELLAND and NOLAN (1912) and MCCLELLAND and GILMOUR (1920) all found an excess of negative charge.

The divergence between the different results is probably due to differences in the conditions under which the measurements were made; SIMPSON (1942) and CHALMERS and LITTLE (1947) found that quietly falling snow is usually negatively charged, while snow under turbulent conditions is more often positive.

NAKAYA and TERADA (1934) measured the signs of the charges on snow flakes, though not their magnitudes, by deflection in an electric field. They found a preponderance of negatively charged flakes in dry snow, but when the flakes had water drops attached, they were often positively charged. MAGONO, ORIKASA and OKABE (1957) found the charges on individual snow flakes to be usually opposite in sign to the potential gradient.

SIMPSON (1949) found the same relation between precipitation current and point-discharge current occurred with snow as with rain, but the snow behaved as though the rate of precipitation were greater than the true value. This suggests that snow acquires charge from the point-discharge current more readily than does rain, as might be expected from the larger cross-sectional area of the snow flake. For low values of potential gradient, with steady snowfall, SIMPSON found the potential gradient to be usually between 0 and +400 V/m, and the charge on the snow to be negative, not positive as it would be for rain in the same potential gradient. SIMPSON did not have sufficient records to obtain a relation between snow charge and potential gradient.

REITER (1965) found that, for steady snowfall, the relation between current and potential gradient is similar to that for steady rainfall, but that snow is usually negatively charged and rain positively; the value for zero charge was, as for rain, 40 V/m.

MAGONO and SAKURAI (1963) measured indirectly charges on snow pellets during drifting snow and deduced a large positive space charge close to the surface, a negative space charge at a height of about 80 cm and a small positive space charge at higher levels.

10.24. Total Current in Continuous Snow

Using the same apparatus as for continuous rain (see § 10.22.), CHALMERS (1956) measured the total current in continuous snowfall, with potential gradients between ± 800 V/m. The results show that, in contrast to rain, the downward current is usually negative, with potential gradients of both signs.

The variation of total current with potential gradient was found to be represented by

$$K = -0{\cdot}92 \times 10^{-14} (P + 425),$$

where K is the total current in A/m^2 and P is the potential gradient at the earth's surface in V/m. Comparing these results with those for rain (see § 10.22.), it is seen that the slopes of the K: P curves are not very different, but there is a great difference in the intercepts; if the slope depends on the rate of precipitation, then these results suggest that the average effective rate of precipitation, in rain and snow is nearly the same.

10.25. Origin of Charges on Continuous Rain and Snow

The general results, that during continuous rain the current is positive downwards, while during continuous snow it is negative, can be interpreted in two main ways, by considering that the precipitation obtains its charge either in the cloud or at or near the ground.

Since most precipitation either originates as ice or at least is in the form of ice for some part of its history, it must be presumed that any process which gives charge to snow in the cloud will also be operating in the rain cloud, but in the latter case there is a second process also at work.

If the charge-separation process operates only close to the ground, the direction of the separation may be different for rain and snow. It must then be imagined that the charge of the sign opposite to that on the precipitation is carried upwards by the up-draught that is associated with continuous precipitation and becomes attached to cloud droplets.

One line of evidence which agrees with the process occurring near the ground is that discussed in § 5.65., leading to the conclusion that, during quiet rain there is sometimes an appreciable negative space charge in the lowest 30 m or more of the atmosphere.

Any explanation of the charges on rain and snow must be able to account for the relation between the total current and the potential gradient. In particular, some explanation must be available for the value of the potential gradient for which the total current is zero (+150 V/m for rain and −425 V/m for snow), or for the values of potential gradient for which the precipitation current is zero. This difference was not, however, found by REITER (1965). When there is no charge on the precipitation one might assume that there is no separation of charge taking place and then one would expect that conditions are not far removed from those in fine weather, giving a value of the potential gradient not far from that found for rain, but very different from that found for snow. It may be significant that the slopes of the current/potential gradient lines for rain and snow are not very different. Up to the present there has been no satisfactory explanation of these results; a complete account would have to consider the space charge of the precipitation and also ion capture by the falling particles. A more detailed discussion of the problem is given in §§ 13.2., 13.3., 13.4.

10.26. Splashing at the Ground

SMITH (1955) attempted to explain the negative potential gradient and positive rain charge in continuous rain with low potential gradient by the effects of splashing giving, as LENARD (1892) had found, a positive charge to the splashed water and a negative charge to the air. SMITH supposed that this negative charge moves upwards in the air by turbulent diffusion or in the up-draught until there is created a sufficient potential gradient to produce point discharge, after which stage, in steady conditions, there can be a quasi-static state in which the negative charge liberated by the splashing returns to earth through the points. SMITH thus deduced a relation between the resulting potential gradient and the rate of rainfall, but this would give larger negative potential gradients than are usually found during continuous rain.

SMITH considered that instruments measuring the rain current are also subject to this splashing effect, presumably by the negative charge diffusing out of the collector, leaving inside the positive charge measured. Even if this occurred, SMITH's theory would not account for the relation between rain charge and potential gradient, and there could certainly be no such effect in the measurements of HUTCHINSON and CHALMERS (1951), where the effect of each drop was measured as it arrived, and where 230 drops during continuous rain with a small negative potential gradient showed a positive charge as against 134 with a negative charge.

COLLIN and RAISBECK (1964) found some evidence for effects of splashing particularly in high potential gradients when the shielding of the collector is not adequate.

10.27. The "Mirror-image Effect"

SIMPSON (1949) found not only that the rain current and the potential gradient are usually opposite in sign during steady conditions, but that, when conditions were changing not too violently, the two changed sign nearly simultaneously, so that the time variation of the rain current often appeared as a mirror image of the time variation of the potential gradient or of the point-discharge current (e. g. Fig. 52). SIVARAMAKRISHNAN (1957) found that the

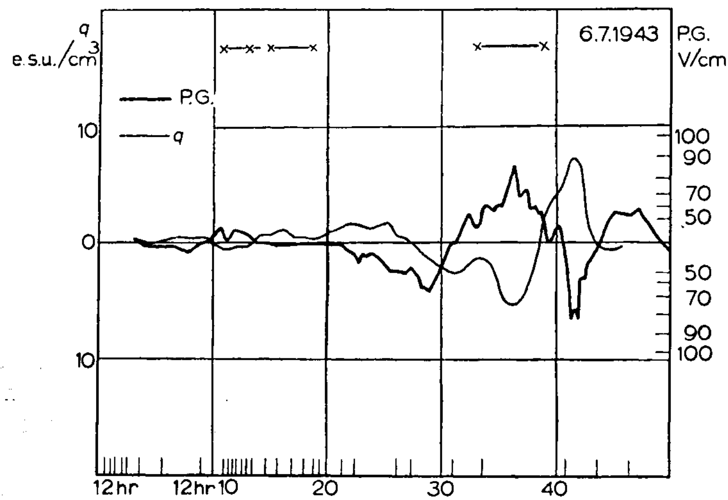


FIG. 52. The mirror-image effect. (From SIMPSON, 1949, Fig. B, Plate XVI.)

change of sign of the precipitation current often synchronizes more closely with the change of sign of the potential gradient at the ground than with that of the point-discharge current. The early work was largely concerned with conditions in which point discharge occurs, and the time scale was such that time differences of less than a few minutes could not be detected. Later work particularly that with low potential gradients, e. g. that of RAMSAY and CHALMERS (1960), showed that there are appreciable time differences, sometimes in one direction and sometimes in the other, between the two zero values; not only do the crossings of the zero correspond, but also, and more easily identified, the maxima and minima. OGAWA (1960) found examples of the mirror-image effect in showers; in fewer than 20 percent of the cases, he found a time difference with, more frequently, the rain current changing before the potential gradient; the time difference in some cases reached up to 8 min.

In general, the same factors as give the inverse relation must operate in the mirror-image effect also, that is to say, the capture of ions during point discharge, as discussed in § 10.20., or the leaving behind, to produce the potential gradient, of a charge opposite to that on the precipitation.

But, whereas the inverse relation can be discussed in terms of a quasi-static state (§ 2.23.), the mirror-image effect involves changing conditions, and cannot be as simple.

We can think of two ways in which the conditions may change; either there could be a portion of cloud with more (or less) electrical violence than the rest, and this approaches the observer and then recedes from him; or else there could be a fairly widespread increased (or decreased) development of electrical effects.

Some observations of OWOLABI and CHALMERS (1965) on precipitation currents at separated receivers suggest that both these types of change may occur. REITER (1957) found that changes of sign of potential gradient occur simultaneously at stations at different levels and separated horizontally by as much as 12 km, showing that there were developing, rather than approaching, clouds of different activity.

CHALMERS (1965a) considered the different conditions and concluded that, if point discharge is occurring, the maximum of potential gradient should appear before the maximum of precipitation current, in the conditions of both approaching and developing clouds. With low potential gradients, the same would occur for an approaching cloud, but for a developing cloud the maximum of the precipitation current would appear before that of the potential gradient. MAGONO and ORIKASA (1960) considered that the space charge of the precipitation would affect the mirror-image effect.

RAMSAY and CHALMERS (1960) found that any relation between precipitation current and potential gradient is much less marked in summer than in winter.

10.28. Shower Clouds

Relations between precipitation current and potential gradient are much less evident for shower clouds than for continuous precipitation and Reiter (1965) suggested that there is another parameter involved, namely the turbulence. He found that the greater the instability of the atmosphere, the greater the frequency of sign changes of potential gradient and ascribed these to the cells of turbulence and convection.

10.29. Charge on Wires in Snowfall

Barré (1953) investigated the charges acquired by stretched steel wires in blizzards in Greenland; in suitable conditions, a wire 50 m long received currents of over 10^{-5} A. At temperatures above -10 °C, small or zero effects were found; from -10°C to -15°C there was a tendency to a positive excess, but below -15°C a negative excess. During an actual fall of snow the charging was almost invariably positive and increased as the height of the wire increased. On the other hand, when the blizzard was formed by drifting snow, the results could be explained in terms of two types of charged particles, the negative being higher than the positive, so that in a light wind only negative charge reaches the wire while in a very strong wind it acquires positive charge. While the wire might be acquiring the charges on the snow particles, there is also the possibility that charges are separated by the impact of the snow on the wire, and it is therefore dangerous to draw definite conclusions from such results as to the charges on the snow.

Magono and Takahashi (1959) also made measurements on the charging of a wire during snowfall, making simultaneous observations of potential gradient and precipitation current; they found the charge to be proportional to the potential gradient. Various explanations were offered and rejected, including that of the wire acquiring the charge on the snow and that of frictional electrification by the impact of the snow on the wire. They found that they could explain the results if the snow acted as a potential equalizer, removing the charges induced on the wire in the electric field.

10.30. "Precipitation Static"

This term was originally applied to effects on radio communications on aircraft flying through clouds and precipitation; this was traced to the charging produced on the aircraft by frictional effects of snow and ice and by the effect of glazing by supercooled cloud droplets; the aircraft becomes negatively charged with the corresponding positive charge carried into the air. The potential difference between the aircraft and its surroundings quickly reaches a value sufficient to produce point discharge at projections on the aircraft and it is the radiation from this discharge which is picked up by the radio-receiving aerial. The effects can be reduced if it can be arranged that point discharge occurs most readily at special points situated such that radiation from them is at a minimum at the receiving aerial. For further details the reader is referred to the report of the American investigation (GUNN *et al.*, 1946).

More recently, the term "precipitation static" has been applied to similar interference with the reception of broadcast television signals during thundery conditions; there are here two effects, one of point discharge at the receiving aerial itself or at points on the supporting mast nearby, and the other of charged precipitation falling on the aerial. These effects have been reduced (PAGE, 1961) in the first case by placing an earthed point at a higher level than the aerial, so that the discharge occurs mainly at this point and so that the field close to the aerial is diminished, and then placing tuned rods to reduce the signal from the point travelling to the aerial; to reduce effects of actual precipitation the aerials were encased in insulating tubes.

CHAPTER 11

The Transfer of Charge

11.1. The Maintenance of the Charge on the Earth

During fair-weather conditions there is normally a positive potential gradient close to the ground, giving a negative bound charge on the surface of the earth. But, at the same time, the atmosphere is a conductor and there is a current of positive electricity coming to the earth, tending to neutralize the bound charge. In spite of this current, the bound charge on the earth's surface remains practically the same from day to day in fine weather; although there are changes with time during the day, these are comparatively small. LINSS (1887) was the first to realize that the conduction current would neutralize the bound charge in a short period of time, if the latter were not maintained in some way. The results of SCRASE (1933) can be used to determine the time in which the positive charge reaching an area of the earth's surface is equal to the negative charge present on this area. SCRASE found the mean value of the potential gradient at Kew to be 365 V/m; the charge per m^2 is then $-\varepsilon_0 F = -3{\cdot}23 \times 10^{-9}$ C. SCRASE's mean value for the air–earth current was $1{\cdot}12 \times 10^{-12}$ A/m^2, so that the charge would be neutralized in $3{\cdot}23/1{\cdot}12 \times 10^3$ sec, i.e. about 48 min. (The simplicity of this calculation using M.K.S. units may be compared with the calculation using E.S.U., where 365 V/m has to be converted into E.S.U. before applying $F = -4\pi\sigma$, and then $1{\cdot}12 \times 10^{-12}$ A/m^2 has also to be converted into E.S.U.) Over other parts of the world, where the potential gradient is smaller and the air–earth current larger, the results of the calculation give shorter times than for Kew; over the oceans, the time is about 6 min.

The problem set by these results is to explain how it is that, at Kew, at the end of the 48 min, the earth's surface still carries a charge of $-3{\cdot}23 \times 10^{-9}$ c/m^2, although in that time a charge of $+3{\cdot}23 \times 10^{-9}$ C/m^2 has arrived by conduction. The answer to this must be that the earth is a conductor and the corresponding negative charge has arrived at some other of the earth. It was first suggested by WILSON (1920), and has since been much confirmed that this negative charge reaches the ground in those parts of the world which are experiencing stormy weather.

The only known processes by which charge can reach the earth are:

(a) Air–earth conduction and convection currents.
(b) Point-discharge currents.
(c) Precipitation currents.
(d) Lightning discharges.

Point discharge is, of course, a form of conduction current, but it is convenient to consider it separately.

Convection currents may be conveniently considered together with conduction currents since both are measured together when the air–earth current is measured by the direct method.

The maintenance of the charge on the earth must correspond with the maintenance of the potential of the electrosphere and we shall discuss this first.

11.2. Methods of Determining the Potential of the Electrosphere

There are two different procedures by which the potential of the electrosphere may be deduced and, although at first sight they appear to be equivalent, in fact there is a distinction.

From § 2.24., under quasi-static conditions, $V = iR$, where V is the potential of the electrosphere, i the air–earth conduction current density and R the columnar resistance (see § 7.17.). As described in § 7.17., the columnar resistance can be calculated from the results of conductivity measurements at various levels. If i can be determined also, then V is at once obtained. This we shall call the columnar resistance method.

If the potential gradient, F, is measured at different heights, then clearly $V = \int F dh$ gives the potential of the electrosphere.

If conditions are truly quasi-static then the two methods might be expected to yield the same result. But this is true only if the value of i is the same at all levels, i. e. if there is no vertical current other than the conduction current. If there are convection, or other, currents then $V = iR$ is no longer correct and, instead, we should use $V = \int idR$ and the simple method of the columnar resistance fails. Before it is justifiable to use the columnar resistance method, it is essential to verify that i is the same at all levels; for this purpose it is necessary to measure F as well as the conductivity at different levels, since no direct method of measuring the conduction current in the atmosphere has yet been devised. And when F is measured at different levels it appears simplest to obtain V by direct integration without needing to consider R at all.

Any convection current is likely to be confined to the *austausch* region, so that the conduction current may be different in the *austausch* region from its value above. Therefore, even if a direct measurement of the air–earth current at the earth's surface were to measure the total current, this would give the value of i above the *austausch* region and not necessarily the value within the region, and so could not be used in the columnar resistance method. It would not even be correct to assume that the convection current within the *austausch* region remains constant with change of level, and so the conduction current may also vary with height, in contradiction to the simple ideas of the quasi-static state used in § 2.24. The recognition that the two methods do not necessarily give the same result is due to KRAAKEVIK and CLARK (1958).

11.3. Results from Columnar Resistance Method

The results quoted will be values for the mean potential of the electrosphere, with adjustment where necessary for the variation with time of day.

GISH and SHERMAN (1936) used the balloon results for conductivity (§ 7.17.) and a value of $3{\cdot}6 \times 10^{-12}$ A/m^2 for the conduction current, and obtained a value for V of 4×10^5 V. GISH (1951) used other results and gave a value of $3{\cdot}6 \times 10^5$ V, and from ISRAËL's (1953a) figures the value would be $2{\cdot}7 \times 10^5$ V.

CLARK (1957) measured the potential gradient and KRAAKEVIK (1958) simultaneously the conductivity during aeroplane flights over Greenland. Unlike the results over Chesapeake Bay (see § 8.17.), these results showed a constant conduction current density of $3{\cdot}7 \times 10^{-12}$ A/m^2 at all levels. With KRAAKEVIK's estimate of $8{\cdot}0 \times 10^{16}\,\Omega$ for R, this gave a value of $2{\cdot}9 \times 10^5$ V for V and is probably the most reliable value, since there is little pollution in Greenland to affect the conductivity measurements at low levels or to provide space charges and electric convection currents; further, meteorological conditions were unfavourable to convection at the time of the measurements.

11.4. Results from Integration Method

The measurements by CLARK (1957) of the potential gradient at various levels above Greenland yield an integrated potential of $2{\cdot}2 \times 10^5$ V up to 6 km and an estimated value of $2{\cdot}83 \times 10^5$ V for the potential of the electrosphere.

From the measurements of KRAAKEVIK and CLARK (1958) (see § 8.17.) over Chesapeake Bay, the integrated value of the potential was $3{\cdot}41 \times 10^5$ V at about the time of the diurnal maximum, which corresponds to a value of $2{\cdot}84 \times 10^5$ V at the time of the mean value.

CLARK (1958) considered various other results and came to the conclusions that the best value would be $2{\cdot}9 \pm 0{\cdot}3 \times 10^5$ V.

FISCHER (1962) used the same method from several individual balloon ascents and came to very similar conclusions to those quoted above. He also found agreement with the diurnal and

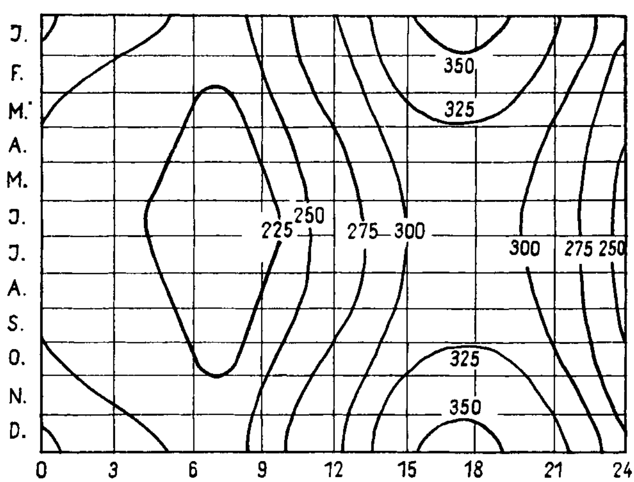


FIG. 53. The variation of potential of electrosphere (in kV) with time of day and of year. (From FISCHER, 1962.)

annual variations deduced from potential-gradient measurements (see §§ 5.43., 5.47.). FISCHER's results for the variation of the potential of the electrosphere with time of day and time of year are given in Fig. 53.

SWANN (1955) made the interesting calculation that a single discharge of 30C would change the potential difference between the electrosphere and the earth by about 200 V, i.e. by one part in about 10^3.

11.5. The Charge on the Electrosphere

The conductivity of the atmosphere increases with height and above heights of the order of 40–60 km it is so great that the potential differences within it are negligible and one is in the electrosphere (see § 2.21.). For most purposes of atmospheric electricity it is possible to consider the electrosphere as equivalent to a metallic conductor and it is then legitimate to enquire as to the charge on its inner surface.

Measurements within the complete conducting shell of the electrosphere can give no means of obtaining any information about any charges outside this shell. Since the electrosphere must be impervious to lines of force, the sum of all the charges inside it must be zero, these comprising the charges on the inside of the electrosphere, those in the atmosphere and clouds and those on the surface of the earth.

CHALMERS (1953b) showed that the total charge on the inner surface of the electrosphere is zero. GISH and WAIT (1950) confirmed by STERGIS, REIN and KANGAS (1957a), have shown that the conductivity of the air above a thunder cloud is no different from that at the same place in fine weather.

Consider a surface surrounding the earth above all thunder clouds, such that the surface is composed of points at all of which the specific conductivity, λ, has the same value; if λ depends only on height, the surface is a sphere concentric with the earth, but if there is any dependence on latitude, it is no longer spherical. Consider unit area of this surface and let the lines of force crossing this area have a charge σ at their upper ends; the potential gradient across this area is then σ/ε_0 and the current across it $\lambda\sigma/\varepsilon_0$. Since the current flows along the lines of force, the current $\lambda\sigma/\varepsilon_0$ must be $-d\sigma/dt$. Now, adding up for all the areas making up the surface considered, any effects of higher horizontal currents will be eliminated, so

$$\Sigma d\sigma/dt = -\Sigma\lambda\sigma/\varepsilon_0.$$

If Q is the total charge above the surface, $\Sigma\sigma = Q$ and $\Sigma d\sigma/dt = dQ/dt$ and so, since λ is constant at the surface,

$$dQ/dt = -\lambda Q/\varepsilon_0,$$

showing that Q rapidly approaches zero and remains zero thereafter unless there is a transfer of charge across the surface other than by conduction.

If we now take the surface concerned to be just below the lower boundary of the electrosphere, however we may choose to define this, then it follows that the total charge on the inner surface of the electrosphere is zero. Even if there are occasional discharges from cloud tops to the electrosphere, the results above show that the charge is quickly lost.

It also follows that the sum of the charges in the atmosphere, in the clouds and on the surface of the earth must be zero.

11.6. Correlation with Thunderstorm Activity

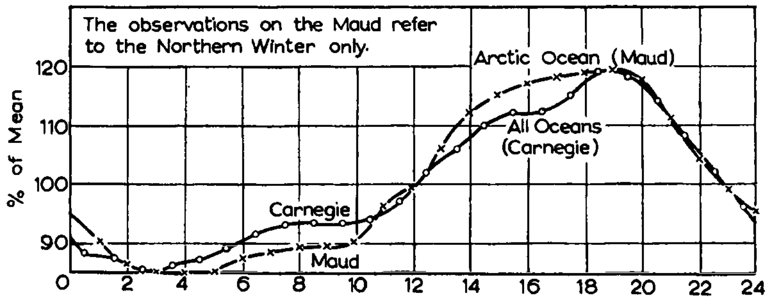
If negative charge is transferred to the earth in periods of storm, then it would be expected that the total negative charge on the earth would be at its maximum at times when the storms over the earth are at a maximum. The greater negative charge involves a greater fine-weather potential gradient and, if there is no change in columnar resistance, therefore a greater potential difference between electrosphere and earth.

The matter may be regarded very simply from the result of § 11.5. If each thunderstorm provides a definite current to the electrosphere (about 0.5 A according to GISH and WAIT (1950) or about 1·3 A according to STERGIS, REIN and KANGAS (1957b)), then the more storms the larger the positive current from the electrosphere, always 0·5 A or 1·3 A per storm. If the columnar resistance remains constant, the electrosphere–earth potential difference must be proportional to the fine-weather current density, and so, since the area of the earth's surface concerned does not vary appreciably, the potential of the electrosphere is proportional to the number of storms in action.

By considering the transfer of charge to the earth from storms and away from it by the fine-weather current, WHIPPLE (1929a) reached a similar conclusion. CHALMERS (1953d) pointed out that WHIPPLE had missed out the effect of charges induced on the earth by storm clouds, but that this effect was probably small compared with the effect of actual charges arriving at the earth; PIERCE (1958) worked out this effect in detail and showed that it would amount to only 1/20 of the total effect.

11.7. Correlation between Potential Gradient and Storms

It was first suggested by APPLETON (1925) that the variation of potential gradient in undisturbed conditions (§ 5.43.) might be the same as the variation in the total number of storms over the whole



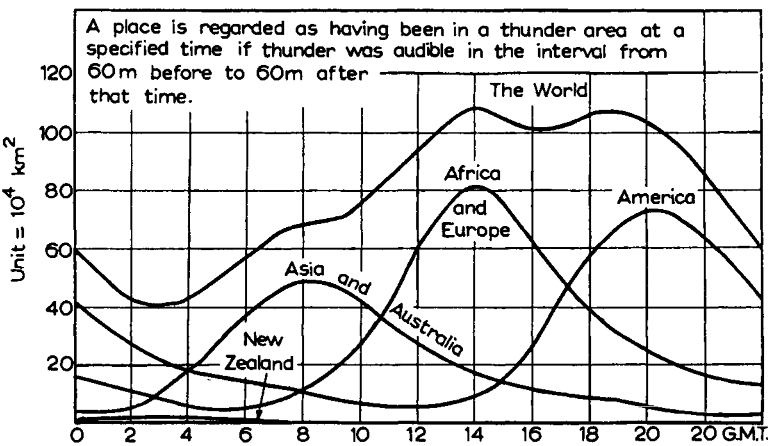


FIG. 54. Diurnal variation of land thunder areas and of potential gradient in unperturbed conditions. (From WHIPPLE and SECRASE, 1936, Figs. 9A and B, p. 19.)

earth. WHIPPLE (1929a) made use of the thunderstorm data of BROOKS (1925) to find the diurnal variation of thunderstorm activity all over the earth. The most intense thunderstorms occur in tropical Africa and in tropical South America, in each case during the afternoon hours, so that the world thunderstorm activity reaches a maximum in the late afternoon G.M.T. WHIPPLE and SCRASE (1936) plotted in detail the thunderstorm activity in the different continents and found maxima at 14 and 19 hr G.M.T. (see Fig. 54 which contains a reproduction of Fig. 26 for comparison). A slight adjustment in favour of America relative to Africa and Europe would give exact agreement with the diurnal variation of potential gradient in undisturbed conditions, where the maximum is found at 19 hr. To give the correct variation about the mean, it is necessary to add to the results in Fig. 54 a constant number for sea thunderstorms, constant because there is no evidence for a variation with time of day. WICHMANN (1951) objected to the inclusion of sea storms, but this is nullified as discussed in § 9.25. WHIPPLE pointed out that not only does the correlation with storm activity show agreement whith the undisturbed potential gradient as regards the times of maxima and minima, but both show a gradual rise to the maximum and a rapid fall afterwards. The agreement of these results with the expectations of § 11.6. confirm the idea that it is in storm conditions that the earth receives its negative charge to balance the positive charge received in fine weather.

FISCHER (1962) found evidence that a tropical thunderstorm may transfer more charge to the earth than a storm in temperate latitudes; this could explain the maximum of electrosphere potential in the southern summer although there is then a minimum of total storms.

11.8. Correlation between Current and Storms

HOGG (1950) pointed out that measurements of the air–earth current might show better correlation with storms than measurements of the potential gradient. From the fundamental formulae for fine weather (§ 2.24.),

$$i = V/R \quad \text{and} \quad F = Vr/R,$$

so the use of i, rather than F, for determination of variations in V, eliminates effects of variations in r. This idea is reinforced by the arguments of § 11.5. about the charge on the ionosphere. HOGG found the expected correlation.

HOGG also suggested a correlation between the seasonal variations of air–earth current and thunderstorm activity. He was faced with the problem as to whether, over the oceans, the seasonal variation of potential gradient reflects a seasonal variation of current or of conductivity. He used results from various parts of the earth and combined them to obtain a value for the total current to the earth. This gave quite good correlation with seasonal thunderstorm variation if it is assumed that the conductivity over the ocean remains constant, as found by TORRESON *et al.* (1946), but not so good if the conductivity has the same seasonal variation as the potential gradient.

11.9. The Electrical Balance Sheet

If the negative charge on the earth is to be maintained, there must be, for the earth as a whole, a balance between the different processes bringing charge to the earth. WORMELL (1930) first discussed the possibility of working out such a balance for a small area of the earth, and used available data to give results for Cambridge. Since then, more accurate data have become available for some places, and the problem can be discussed in somewhat more detail, considering each process in turn. Attempts have also been made to determine the balance sheet for the earth as a whole.

11.10. Conduction Current

Measurement of the air–earth current at various places has given values from 1×10^{-12} A/m^2 to 4×10^{-12} A/m^2. For Cambridge, WORMELL (1930) assumed a value of about 2×10^{-12} A/m^2 and so obtained a result of +60C reaching 1 km^2 in a year.

At Kew, the current is less, and the results of SCRASE (1933) give +35 C/km^2/year. For the oceans, the current is about +100 C/km^2/year, and overland, where there is no disturbance by pollution, the value may be still greater.

For Durham, CHALMERS and LITTLE (1947) found a value agreeing approximately with WORMELL's estimate of 2×10^{-12} A/m^2 and so giving +60 C/km^2/year.

For the world as a whole, KRAAKEVIK's (1961) figures give about 86 C/km^2/year (see § 8.17.).

11.11. Convection Current

If the conduction current is measured above the *austausch* region, as discussed in § 8.17., then there is no convection current to complicate the matter. If, however, the conduction current is measured, by the indirect method, within the *austausch* region, then there must be an effect of the convection current and the magnitude of this is difficult to measure.

If the conduction current is measured by the direct method, there will be no convection current as such, but there remains the question as to whether the area used is a fair sample of the earth's surface, in regard to the production of ions and charged particles at and close to the earth's surface.

11.12. Point Discharge

Wormell (1930) measured the point-discharge current from a single point over a period of about 4 years and obtained an average annual resultant charge passing of about −0·12 C; he made the guess that there would be about the equivalent of 800 such points per km^2, giving a contribution to the maintenance of the negative charge on the earth of about −100 C/km^2/year. Later, Wormell (1953a) gave reasons for believing that this estimate was too low, and suggested an increase to about −170 C/km^2/year.

Whipple and Scrase (1936) also measured the total point-discharge current over a period and chose an effective separation of points at Kew which gave, by accident or design, also a value of −100 C/km^2/year. The more recent estimates of the separation of discharging points at Kew (see § 9.24.) suggest that this is too low by a factor of from 2 to 4.

Chalmers and Little (1947) found a total point-discharge current of 0·037 C in 8 months. Making a guess of an effective separation of similar points of 25 m, this gave −90 C/km^2/year, but a more modern estimate would give a higher value. The more recent estimates of the separation of points at Durham have been concerned with a point in a different situation and so cannot be applied to the above figures.

The conclusion that, in England, the total charge reaching 1 km^2 per year is of the order of −200 C, may be subject to modification if the result of Maund and Chalmers (1960) is confirmed that trees in leaf give point discharge less readily than has been expected (see § 9.14.).

11.13. Precipitation Currents

WORMELL (1930) originally assumed a value for the total precipitation current intermediate between those found by SIMPSON (1909) and by SCHINDELHAUER (1913), giving a total of about -20 C/km^2/year.

The later work of SCRASE (1938) should give a more reliable value, at any rate for Kew. If we take the two years of SCRASE's measurements, the value is $+7$ C/km^2/year, but if we take only the second, more reliable, year, the value is $+22$ C.

CHALMERS and LITTLE (1947) obtained a total vertical current of $+100$ C/km^2/year, of which $+60$ C was ascribed to fine weather, leaving $+40$ C for precipitation.

It should be realized that measurements of rain currents with a shielded receiver do not include all the rain—SCRASE (1938) collected only half the rain that is caught by a standard rain collector. This may affect appreciably the figures quoted above for Kew.

In view of the paucity of results for tropical regions, an estimate for the whole world is bound to be very tentative.

11.14. Lightning Discharges

The total effect of lightning discharges in bringing charge to 1 km^2 of the earth in a year must depend on (1) the number of discharges and (2) the average charge brought down by each.

For Cambridge, WORMELL (1930) originally made a rough estimate of -20 C. From his own lightning observations, he has more recently (WORMELL, 1953a) made a more reliable estimate. His observations show that the net transfer of charge to earth is equivalent to 28 per cent only of the charge concerned in each discharge; potential-gradient change measurements (WORMELL 1939) led to a conclusion of 1 flash per year per km^2, but only 0·4 flashes to earth, the remainder being cloud flashes; but some discharges bring positive charge, so that the effective number is only 0·28. The average charge concerned in a flash is about 20 C, so that the total for 1 km^2 per year at Cambridge is $-5{\cdot}6$ C.

GOLDE (1945a) estimated the number of flashes to earth from a survey of "near misses" and obtained values considerably greater than those of WORMELL. For Cambridge, his value would be about 2 per km^2 per year, giving a total charge of -28 C.

For Kew, the frequency of discharges is rather greater than for Cambridge, and for Durham rather less. Using WORMELL's figures, the charges to reach the earth become about −6·5 C for Kew and −5 C for Durham and, using GOLDE's figures, −32 C for Kew and −25 C for Durham.

For the whole world, ISRAËL (1953a) used the value of BROOKS (1925) of a total of 1800 storms in existence at one time, and took a value of 60 lightning flashes to ground per hour for each storm; from the accepted value of 20 C per flash and an estimate that 80 per cent of the flashes bring negative and 20 per cent positive charges, he reached an average lightning current of $-0{\cdot}67 \times 10^{-12}$ A/m^2, corresponding to about −20 C/km^2/year.

11.15. Cosmic Rays

It is known that the incoming primary cosmic rays are positively charged nuclei, so that at the top of the atmosphere there is a positive current travelling towards the earth. At lower levels, when the primary particles have produced secondaries, this positive excess will be maintained, since the secondary particles will be of both signs equally; it is only when a positive particle is actually stopped that there is no longer an excess current towards the earth.

The suggestion has been made that the contribution of cosmic rays to vertical currents in the atmosphere ought not to be neglected; however, the magnitude of the effect is far too small to be appreciable. REYNOLDS (1956) gave an estimate of about 0·15 A as the positive excess arriving at the top of the atmosphere over the whole earth. This gives an average of 10^{-2} C/km^2/year, which is very much smaller than the other factors involved. In addition, a very large proportion of this positive excess is stopped before reaching the earth and the actual positive current into the earth is only a few per cent of the figure given.

11.16. Total Transfer

Using the results discussed in the preceding sections, it is possible to make estimates of the total transfer of charge for the various places where the individual measurements have been made. It is to be regretted that similar measurements are not available for other places where conditions differ from those in England. In the table below, in the estimates for Durham and Kew, values for lightning are those based on the figures of WORMELL (1953a), rather than on those of GOLDE (1945a).

WAIT (1950) made an estimate for the whole world, but it should be pointed out that his estimate for point discharge is based on the

TABLE 3

Place	Author	Fine weather conduction	Point discharge	Precipitation	Lightning	Total
Cambridge	WORMELL (1930)	+60	−100	+20	−20	−40
Cambridge	WORMELL (1953a)	+60	−170	+20	−5·6	−96
Durham	CHALMERS and LITTLE (1947)	+60	−90	+40	−35	−25
Durham	Revised (1957)	+60	−180	+40	−5	−85
Kew	CHALMERS (1949)	+35	−125	+22	−45	−113
Kew	Revised (1957)	+35	−300	+22	− 6	−249
World	WAIT (1950)	+100	−30	+20	−20	+ 70
World	ISRAËL (1953 a)	+90	−100	+30	−20	0

idea that the oceans and other flat portions of the earth would have no point discharge. However, CHALMERS (1952c) has concluded that the total point-discharge current below a cloud is largely independent of the nature of the surface, so that oceans would play their part; this would increase WAIT's estimate for point discharge.

ISRAËL (1953a) used more or less reliable estimates for some of the factors and adjusted the less certain so as to give the total balance for the whole of the earth that is to be expected.

Since the results in England give an excess of negative charge passing to the earth and it is to be predicted that a similar, and even greater, excess of negative charge would arrive at tropical regions, it is pertinent to enquire where there would be the corresponding positive excess, and the answer is probably to be found in the deserts and perhaps also in the polar regions.

In view of the progress made in various branches of atmospheric electricity, it is surprising that there are no additions to be made to this table since the previous edition of this book.

11.17. Currents above Clouds

GISH and WAIT (1950) pointed out that, on the average, the total current from the top of a thunder cloud to the electrosphere must be equal to the total upward current within the cloud and also equal and opposite to the total current down to the earth from the cloud. Of these the current above the cloud is the simplest, since there are no precipitation or conduction currents and probably no discharges. Accordingly, GISH and WAIT measured the total current above thunderstorms by making traverses in an aircraft over the storm and measuring both the potential gradient and the conductivities at various points. The flights were made at about 40,000 ft (12 km) and, except in one case, were well above the storm cloud. Any effect of a charge on the aircraft itself was eliminated by using two field mills, on the top and bottom, to measure the potential gradient, (see § 5.38.).

Integrating over the whole area above the storm, the results for the total current varied from 0 to +1·4 A, with the exception of one case when the aircraft passed through the top of the cloud and a value of +6·5 A was obtained. Neglecting this case, when there may have been other currents, the average current was +0·5 A upwards.

Similar measurements, using a balloon which went over the tops of thunder clouds, were made by STERGIS, REIN and KANGAS

(1957b). From 20 storms, giving values ranging from +0·6 A to +4·3 A, the average current was +1·3 A. All these storms investigated were in the Central Florida area and the result might not be typical for all thunderstorms.

11.18. Horizontal Currents

There are currents taking positive charge vertically upwards above storms and currents taking positive charge downwards in fine weather, so that somewhere there must be horizontal currents to link up the circuit. At the lower ends they must be in the earth and at the upper ends they can be stated to be "in the electrosphere".

In order to make this more precise, Israël and Kasemir (1949) considered the variation with height of the ionization due to cosmic rays and calculated the level at which the horizontal currents would reach their greatest intensity. Depending on the particular assumption made, the results gave a height between 50 and 65 km. Since this is considerably below the level at which the ionosphere, as identified by radio-wave reflection, is taken to commence, the distinction between the electrosphere and the ionosphere is required.

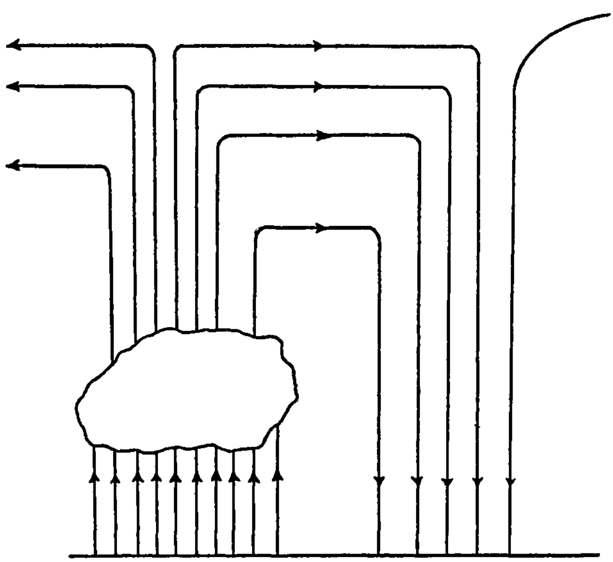


Fig. 55. Horizontal currents.

11.19. Number of Thunderstorms

Meteorological observational stations record "thunderstorm days", namely, those days on which thunder is heard, and the number of such days per year defines the "isoceraunic level" of the station. This may be a rather poor guide to the amount of thunderstorm activity at a station, since there is no distinction between days on which there are several storms of long duration and those on which there is a single short storm.

Lightning counters (see § 14.37.) give a much more accurate picture than thunderstorm days but are not yet sufficiently standardized or sufficiently widespread in use. However, PIERCE, ARNOLD and DENNIS (1962) quote results which appear to show that lightning counter measurements are more closely correlated with isoceraunic level than might have been expected. They gave a result of $0{\cdot}40 \pm 0{\cdot}20$ as the number of lightning flashes per km^2 per annum per thunderstorm day.

BROOKS (1925) estimated, from thunderstorm-day observations, that the total number of storms in existence at any one time is about 1800. Since this did not take into account any cases where more than one storm occured in one day, his result is probably an underestimate and the total number may well be nearly 3000.

For an air–earth current density of $3{\cdot}5 \times 10^{-12}$ A/m^2 as found over the oceans by AULT and MAUCHLY (1926) and by TORRESON *et al.* (1946), the total fine-weather current to the whole earth would be 1800 A and BROOKS's statistics show that the area affected by storms is too small to alter this estimate.

With the result of GISH and WAIT (1950) of 0·5 A per storm, the greater part of this 1800 A can be balanced by effects during storms, and with the figure of 1·3 A obtained by STERGIS, REIN and KANGAS (1957b), 1400 storms would be sufficient to balance the fine-weather current. A contribution might be expected from non-stormy weather and WORMELL (1927, 1930) has shown that there is considerable point discharge during showers which never reach the stage of lightning and which, therefore, would not be included in BROOKS's statistics.

11.20. Charge Transfer in Non-stormy Rain and Snow

The conditions in non-stormy rain and snow can be divided into two groups, according to whether there is point discharge or not. When point discharge occurs, the relation between the charge on the rain and the point-discharge current can be explained, as discussed in § 10.20., if the rain acquires its charge from the point-discharge ions; this, therefore, seems to require that the point-discharge current is greater than the rain current, so that the resultant current is in the same direction as the point-discharge current, thus usually conveying a negative charge downwards. Therefore, when there is point discharge, mainly in showers and heavy continuous rain, the transfer of charge is in the same direction as in thunderstorms.

On the other hand, when there is no point discharge, the direct measurement of the total current (CHALMERS, 1956; see § 10.22.) and its indirect estimation from SIMPSON's (1949) results show that the current usually brings positive charge down. With continuous snow (see § 10.47.), the results show a total negative charge to earth; if the same is true for most snowfall, there may be a considerable alteration in the figures for precipitation current over the whole world.

CHAPTER 12

The Thunder Cloud

12.1. Polarity of Clouds

The phenomena of lightning and the other electrical effects of thunder clouds show with certainty that the thunder cloud must be the seat of intense electrical charges; and the frequency of discharges inside the cloud, especially in tropical regions, suggests that there are charges of opposite signs in the cloud, rather than that there is just a single sign of charge. The most important problems of the thunder cloud are the determination of the arrangement of the charges in the cloud and their magnitude. It is generally considered that there is an arrangement of the charges with the concentration of one sign higher up than that of the other sign. Such a cloud is called "bipolar" and it is termed of positive polarity if it has the positive charge uppermost, and of negative polarity in the reverse case.

There has been considerable debate as to the polarity of the typical thunder cloud; SIMPSON (1927) for a long time believed that thunder clouds are of negative polarity, while WILSON (1925) held the opposite view. The alti-electrograph results (see § 9.16.) have shown that the main distribution of charge is that of an upper positive and lower negative charge, and that often, perhaps always, there is a concentration of positive charge in a limited region below the negative charge, in the base of the cloud. These results were obtained in England and the earlier South African work (e.g. SCHONLAND, 1928a) seemed to show a simple bipolar cloud of positive charge without the additional positive charge in the base, but later work has shown that this occurs also in South African storms (MALAN, 1952).

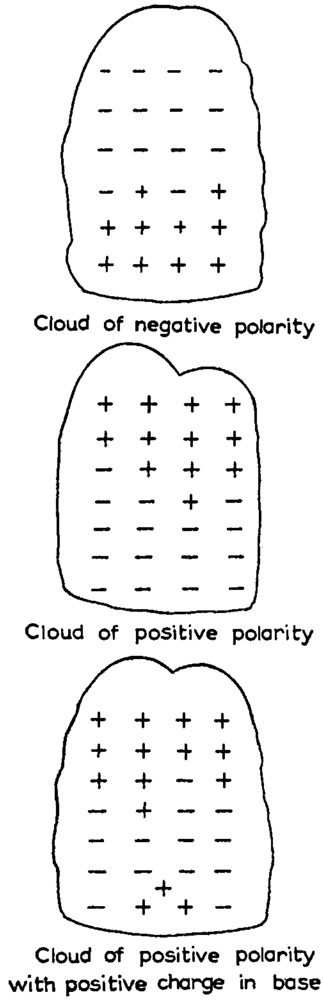


FIG. 56. Polarity of clouds.

12.2. Methods of Investigation

One of the methods of obtaining information about the electricity of thunder clouds is to measure the potential gradient below and near clouds; often these values of potential gradient are considerably larger than those of fine weather and so cannot be measured by the same instruments as are used in fine weather unless a change of sensitivity can be arranged. More definite information is afforded by the measurement of potential-gradient changes produced by lightning flashes, which correspond to the transfer of a charge from one portion of the cloud to another or from the cloud to the earth. The alti-electrograph, which yields information as to the potential gradients in the atmosphere below, in and above clouds, gives even more certain results as to the distribution of electricity in clouds.

It follows from electrostatic considerations (see § 2.10.) that the electrical structure of a cloud can never be obtained unambiguously from measurements carried out at the earth's surface, even if the cloud remained still and of the same electrical structure and if an unlimited number of observations could be taken. If only a few measurements of potential gradient can be made at only one station for a cloud which is in motion and is altering in electrical structure, the information must give little certain knowledge and at best can merely eliminate some suggested cloud structures. For more definite results, it is necessary to go into the third dimension, e.g. by the use of the alti-electrograph.

12.3. Potential Gradient due to Thunder Cloud

For the sake of simplicity, we shall first consider a cloud consisting of two equal and opposite charges of electricity, imagined to be point charges, one vertically above the other. The vertical potential gradient at the ground must change sign as the cloud approaches from a distance to the overhead position, as can be seen from the following general argument; when the cloud is far distant, the two charges are practically equally distant from the point of measurement and so the magnitudes of the potential gradients due to each will be nearly equal; but the directions differ and so the vertical component of potential gradient is greater for the upper charge. Horizontal components cancel for each charge because of the effect of the electrical images (see § 2.19.). Thus, for a cloud of positive polarity the potential gradient when it is far distant is positive. But, when the cloud is nearly overhead, the directions of the two charges are almost the same, but the lower charge now produces the greater potential gradient, being closer to the point of measurement; for a close cloud of positive polarity, the potential gradient is negative. At some intermediate distance, the vertical potential gradient due to the cloud must be zero.

Considering the matter mathematically, let us suppose a cloud of positive polarity, with a charge $+E$ at a height H and a charge $-E$ at a height h; then if observations are made at a horizontal distance R from the cloud, the vertical potential gradients are

$$\frac{-2EH}{4\pi\varepsilon_0(H^2 + R^2)^{3/2}} \quad \text{and} \quad \frac{+2Eh}{4\pi\varepsilon_0(h^2 + R^2)^{3/2}}$$

the factors 2 arising from the electrical image charges.

The sum of these two is zero when

$$R^6 - 3h^2H^2R^2 - h^2H^2(h^2 + H^2) = 0$$

or

$$R^{\frac{2}{3}} = h^{\frac{2}{3}}H^{\frac{2}{3}}(h^{\frac{2}{3}} + H^{\frac{2}{3}}).$$

If h and H are considered as constants, this equation gives the value of R for which the potential gradient is zero, namely, the "reversal distance" at which the potential gradient changes sign. If h and H are nearly equal, the reversal distance is $2H$; if $H = 2h$, the value is about 1·2 h.

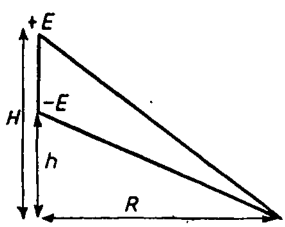


FIG. 57. Potential gradient due to thundercloud.

But, usually, the two charges in the cloud are not equal, are not in the same vertical line and are not point charges, so that the above theory is far too simple. However, unless the difference in the two charges is very great, or the line joining them is very much inclined, we can retain the conclusion of the simple theory that, for a cloud of positive polarity, the potential gradient is positive when the cloud is distant and negative when the cloud is near, the exact reversal distance depending on the magnitudes and positions of the charges. For a cloud of negative polarity the signs are reversed. It must, however, be remembered that, even if the potential gradient agrees exactly with that to be expected from a simple charge system, there are still other, more complex, charge systems which would give the same potential-gradient distribution at the earth's surface.

So far, we have discussed only the sign of the potential gradient and not the magnitude; we can see that, in general, the potential gradient should be greater when the cloud is near than when it is distant; we then expect that a cloud of positive polarity will have its most intense potential gradients negative. If we think of all the lines of force from a cloud coming to the ground and none to the electrosphere and if we think of the cloud as having two equal and opposite charges, then the induced charges on the ground will just cancel out; however, some lines of force from the upper pole will go to the electrosphere not to the ground, and if calculations are made as discussed in § 12.22. involving the change of conductivity with height, the lower charge is found to be of greater numerical magnitude than the upper. These results mean that, for a cloud of positive polarity, a measurement of any effect depending on potential gradient below the cloud will show an excess effect of negative potential gradient; this will be enhanced if the measurement is, for example, of point discharge which does not occur at all if the potential gradient is near zero, as will be the case more for positive than for negative potential gradients.

If a cloud is found to give a fairly small positive potential gradient when at a distance, a larger negative potential gradient when it has approached close to the observer, and again a smaller positive potential gradient when it has receded, then there can be no escape from the conclusion that the structure is, at least, similar to that of a bipole of positive polarity. But it is seldom that the observations can be as certain as this, for there is usually more than one cloud in the sky and the effects of any particular cloud, when distant, may be masked by effects of other clouds which are not of interest at the time.

12.4. Measurements of Potential Gradients due to Thunder Clouds

The methods of measuring potential gradient with equalizers are not very suitable for the measurement of potential gradients due to thunder clouds. Apart from the fact, mentioned in § 12.3., that a different order of sensitivity is required from that in use in fine weather, an equalizer is often seriously affected by rain. Also the potential gradients due to thunder clouds change very rapidly and the usual equalizers are not capable of dealing with rapid alterations, so that results obtained may not be good samples of the actual potential gradients.

The same objections do not apply to the bound-charge methods of measuring potential gradient but these do not yet appear to have been much used for continuous recording of thunderstorm potential gradients, except by MALAN and SCHONLAND (1950) and by SMITH (1954).

SIMPSON and SCRASE (1937) and SIMPSON and ROBINSON (1940) used an equalizer to give rough values of the potential gradient, and in particular its sign; their results are given in Fig. 58, which is interpreted to mean that it is probable that thunder clouds are usually of positive polarity but that, at any rate sometimes, there is another region of positive charge in the base.

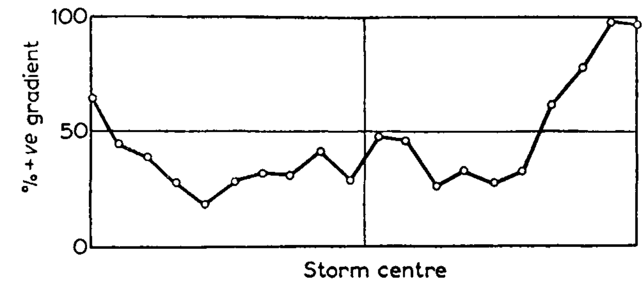


FIG. 58. Frequency of positive potential gradients at ground. (From SIMPSON and ROBINSON, 1940, Fig. 15, p. 324.)

A plate or sphere exposed to the field has often been in use in thunderstorms for measurements of potential-gradient changes (see § 5.35.). When the surface is covered over, the charge passing gives a measure of the bound charge and so of the potential gradient, but the covering, of course, puts the apparatus out of use for potential-gradient change measurements, so that this type of measurement has not often been made during storms and, when made, may not have given a fair sample of the potential gradients. SCHONLAND (1928a) has given results of this kind and he found the potential gradient to be negative in each of 13 storms passing over the observing station; in 10 out of 11 storms at a distance, the potential gradient was found to be positive and in 3 cases the actual reversal was found. These results leave little doubt that the thunder clouds observed by SCHONLAND in South Africa are predominantly of positive polarity.

WORMELL (1939) made similar investigations in England. He used, according to the value of the potential gradient, a sphere at 4·8 m or 1·5 m above the ground, a plate flush with the ground or an inverted test plate, shielded from above, which carried the bound charge due to lines of force which curve round the shield (see Fig. 59). In addition to the alterations of bound charge related to

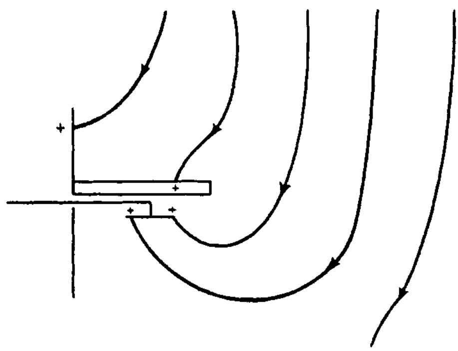


FIG. 59. The inverted test plate.

alterations of potential gradient, his conductors received charge by conduction and, except for the inverted plate, also by rain; during rainfall, the inverted plate was mainly used. WORMELL made his measurements with a capillary electrometer (see § 5.22.), which kept his conductor at earth potential. On covering or uncovering the conductor, the capillary electrometer registered the bound charge and hence the potential gradient; but the potential gradient could also be estimated fairly closely from the capillary electrometer records, since the zero altered only slightly on account of conduction currents. In his analysis of potential-gradient measurements, WORMELL considered only the fairly steady values before lightning flashes, not the rapidly varying values immediately after flashes. He related the potential gradients to the distance of the storm centre, found by the lightning–thunder time interval. Unlike SCHONLAND, he found potential gradients of both signs, both for distant and close storms, but he found that the potential gradients due to distant storms were more often positive, while those of close storms were more often negative. For the closest storms (nearer than 3 km), the negative preponderance was a little less than for storms at about 5 km, agreeing with the results of SIMPSON and ROBINSON (1940), shown in Fig. 58. Whereas SCHONLAND found the potential gradient due to close storms to be negative in every case, WORMELL found the preponderance of negative potential gradient to be only 2:1. This seems to indicate a quite definite difference between the English and South African storms.

12.5. Measurements by Point Discharge

Another method of measuring potential gradients below thunder clouds is by measurement of the point-discharge currents through a raised point. As discussed in Chapter 9, the current depends on the potential gradient, though perhaps not linearly. Point-discharge measurements have the advantage that they can be kept in operation throughout the storm without interfering with other observations; also they give rapid response and are unaffected by rain. But it must be realized that, as described in § 9.10., the point-discharge current changes sign after the potential gradient at the ground, and so does not give a complete measure of the potential gradient when this is altering; however, as LUTZ (1941) pointed out, the point-discharge current probably gives a better measure of the effect of the cloud than does the actual potential gradient at the ground. The fact that wind speed is also a factor in the point-discharge current ought to be taken into account, but it has not usually been measured. The general results, quoted in § 9.9., show that there is a definite preponderance of negative charge coming to the earth by point discharge for all the places of measurement. Now, point discharge currents do not occur for low potential gradients and are stronger for high potential gradients, so these results are in agreement with the idea that the strongest potential gradients below thunder clouds are negative, as we should expect immediately below a cloud of positive polarity.

12.6. Potential-gradient Changes due to Lightning

Measurement of potential-gradient changes due to lightning flashes can be used to give information in regard to the relative positions of the charges transferred in the flashes. Such measurements have one outstanding advantage over measurements of the actual potential gradients; since a potential-gradient change is due to one single lightning flash, and visual observation can often determine which cloud is concerned, there is not the masking by other clouds of the effect of the cloud under investigation, such as occurs in measurements of the actual potential gradients.

A difficulty that arises in the interpretation of lightning-flash potential-gradient changes is that the effects to be expected differ according to whether the discharge is within the cloud or from cloud to earth; this question can sometimes be settled by visual observation, or by the detailed examination of the potential-gradient change.

The mathematical discussion of potential-gradient changes can be simplified if it is realized that the lightning discharge is essentially a transfer of charge from one point to another, and that it is irrelevant to consider these charges which have not been transferred.

A lightning flash which transfers a charge $+E$ from a height H to a height h produces a potential-gradient change which is just the same as the actual potential gradient due to $-E$ at H and $+E$ at h. Thus, if the lightning flash is vertical, it is possible to use the theory of § 12.3., and it follows that, when a positive charge moves downwards, the potential-gradient change is the same as the potential gradient produced by a cloud of negative polarity. So, for a distant cloud in which a positive charge moves downwards, the potential-gradient change is negative, while for a near cloud it is positive; for a cloud in which a negative charge moves downwards, the reverse is true. There must be a reversal distance for potential-gradient changes due to lightning flashes within clouds, the actual distance depending on the heights of the two ends of the flash.

For a discharge to earth, the effect is that of the destruction of a charge in the cloud; now, a negative charge in the cloud would, alone, produce a negative potential gradient at the earth at all distances, so that a lightning flash bringing negative charge to earth would destroy this negative potential gradient and give a positive potential-gradient change. This is what is to be expected if a negative charge at the base of a cloud of positive polarity discharges to the earth.

Neglecting the possibility of discharges from the top of the cloud to earth and also of discharges to the electrosphere, we can form Table 4 to give the sign of the potential-gradient change to be expected in different cases.

TABLE 4

	Cloud of positive polarity		Cloud of negative polarity	
	Near	Far	Near	Far
Discharge in cloud	+	−	−	+
Discharge to earth	+	+	−	−

It will be seen that the polarity of the cloud can be distinguished by an examination of the potential-gradient changes due to near lightning flashes, but the effects of distant flashes give no certain indication of polarity unless it is known independently whether the flash is within the cloud or to earth.

The above results may not be correct if the lightning flash is far removed from the vertical, but the more distant the flash, the greater would be the inclination to the vertical that would be necessary to reverse the sign of the changes as given in Table 4.

These results have not considered the "branched discharge" suggested by SIMPSON (1927) in which the charge leaves the lower side of the cloud but does not reach the earth, either neutralizing a space charge already in the lower air or else providing such a space charge. From the point of view of the discussion above, such a discharge would behave like a discharge within the cloud and would give a reversal distance. But, since the process involves lowering the lower charge, the potential-gradient change will be just the opposite to that produced by a true cloud discharge in the same thunder cloud. Since the branched discharge is lower in the atmosphere than a cloud discharge, the reversal distance will be smaller.

12.7. Measurements of Potential-gradient Changes

Measurements of potential-gradient changes have been made by the alterations in bound charge, using mainly apparatus we have already discussed in Chapter 5. MALAN and SCHONLAND (1950) appear to be the only workers who have yet used a field machine for continuous measurement of potential gradients and potential-gradient changes, using a cathode-ray oscillograph for display; SMITH's (1954) apparatus is designed for the same purpose. APPLETON *et al.* (1926) investigated the potential-gradient changes produced by distant lightning flashes, using a radio-receiving aerial and various measuring instruments including a cathode-ray oscillograph.

WILSON (1916, 1920) made measurements mainly on lightning flashes nearer than 10 km, and his general conclusion was that there is an excess of positive potential-gradient changes, i.e. that the potential gradient changes from negative to positive, increases positively or decreases negatively. Since these are mainly "near" flashes, the results give evidence for the clouds being of positive polarity. SCHONLAND and CRAIB (1927) and SCHONLAND (1928a) found unmistakable evidence of a reversal distance for discharges, visual observations of which showed that they were within the cloud; for distant flashes, most of the potential-gradient changes were negative, and the few which gave positive potential-gradient changes were usually correlated with visual observations of a flash to ground. The results show that South African thunderstorms are practically all of positive polarity. These results were confirmed by those of HALLIDAY (1932), also in South Africa.

APPLETON *et al.* (1926) found that, in the vast majority of cases, the potential-gradient change was negative, i.e. of opposite sign to the majority of WILSON's results. This difference in sign is to be correlated with a difference in the average distance and suggests that this is the effect of reversal due to flashes within clouds. If this is so, then there can be no doubt that the clouds must be mainly of positive polarity. APPLETON *et al.* mention one particular case when there was only one single thunder cloud giving discharges, and the potential-gradient changes altered from negative to positive as the cloud approached the overhead position and then back to negative as the cloud receded; this could be nothing but a cloud of positive polarity with flashes within the cloud, and the reversal distance was between 8 and 10 km.

WORMELL (1939) made measurements of potential-gradient changes due to flashes at various distances, and this work was followed by PIERCE (1955a). Analysing the results of WILSON, WORMELL and himself, including observations during over 20 years between 1920 and 1949, according to the magnitude of the potential-gradient change and therefore approximately according to the distance, PIERCE found the ratio of numbers of positive and negative potential-gradient changes; for distances less than 6 km the ratio is about 4·4, but for greater distances it is less, falling to about 1·1 at more than 50 km. PIERCE did not confirm WORMELL's result that the ratio is smaller for the closest flashes. These results suggest that cloud and earth flashes are approximately equal in number, and that clouds are of positive polarity; the cloud flashes show a reversal distance of about 7 km.

WHIPPLE and SCRASE (1936) made measurements by point discharge of potential-gradient changes due to discharges at distances of less than 5 km and found a preponderance of positive changes of 3·5 : 1.

12.8. Multiple Station Recording of Potential Gradients and Potential-gradient Changes

If it is assumed that the charges in a cloud can be represented by two point charges at different points, then there are 8 unknown quantities, corresponding to the 3 coordinates and the magnitude of each charge. Hence, from the measurement of the potential gradient at 8 or more stations simultaneously it should be possible to obtain the 8 unknown quantities. This would seldom be feasible because there are usually several clouds all producing effects at the same time.

Instead of measuring the potential gradient, it is possible to measure the potential-gradient change produced by a lightning flash; then the number of unknowns is reduced to 7, since a charge moves from one point to another and the potential-gradient change depends on this one value of charge alone. The use of a number of simultaneous measurements to obtain the characteristics of a lightning flash was first carried out by WORKMAN, HOLZER and PELSOR (1942) and later by BARNARD (1951), HACKING (1954) and REYNOLDS and NEILL (1955). The results were in general agreement with those found by other methods.

FITZGERALD (1956) used the method of simultaneous measurements to investigate the charges in clouds and also gave some very detailed analysis of the errors to be expected and of the validity of the conclusions.

TAMURA (1958) used the same method both for cloud-ground discharges and for intracloud discharges. His conclusions were that there is a region in the thunder cloud from which a cloud-ground discharge carries down negative charge and another region, horizontally separated from the first, where intracloud discharges occur. If the lower positive charge is added, one reaches the double dipole as postulated by KUETTNER (1950) (see § 12.17.).

12.9. Electric Moment Destroyed

If a lightning discharge moves a charge Q through a distance d, then there is a change in "electric moment" of $M = 2Qd$, the factor 2 arising from the electrical images in the earth. If observations are made at a distance L, which is large compared with d, the change in potential gradient δF is M/L^3.

PIERCE (1955a) analysed the potential-gradient changes from a number of distant lightning flashes and found the inverse cube relation to hold both for positive and negative values of δF. The median value of M was found to be about 110 C km, agreeing with earlier estimates by WORMELL (1939). The division of M into parts to be associated with different stages of the lightning discharge will be discussed in § 14.29.

WANG (1963) at Singapore found the mean value of M to be 260 C km.

12.10. Pre-discharge Potential Gradients

A surprising fact noticed by WHIPPLE and SCRASE (1936) was that quite often (in nearly 30 per cent of the observations) the potential gradient before a lightning flash was insufficient to give point discharge, i. e. less than about 800 V/m. And in a number of cases (about 40 per cent) the potential gradient after the lightning flash was of opposite sign to that before the flash. WORMELL (1953a) found the same results and gave a reproduction of a record in which the potential gradient changes from -5500 V/m to $+37{,}000$ V/m for a discharge at 3 km.

These results can be easily accounted for if there is a space charge residing in the air below the cloud, as predicted by the theory of § 9.18., from point discharge, and as discussed in § 9.23. Then the low pre-discharge potential gradient is explained by the effect of the cloud being largely neutralized by the space charge, while the later potential gradient of opposite sign is due to the space charge itself. But the increase of potential gradient with height that is predicted from the space charge does not appear in the alti-electrograph results (see § 9.20.).

12.11. Recovery of Potential Gradient after Lightning

Many observers have noticed that after a lightning flash that is not less than 3 km away, the potential gradient returns to something very close to its pre-discharge value along a curve which is not

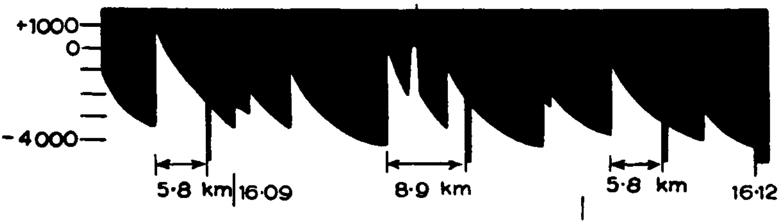


FIG. 60. Recovery of potential gradient after lightning. (From WORMELL, 1953, b, Fig. 1 (second line), p. 477.)

far from exponential, with a time constant of a few seconds (7 sec is given as a mean value). The exponential curve is just what one would expect if there are: (1) a constant rate of separation of charge, and (2) opposing dissipation of charge proportional to the extent to which charge has already been separated; this latter can be comprised of conduction and other currents within, above and below the cloud. WILSON (1920) calculated that the regenerating current in a cloud must be about 5A; WORMELL (1953a) gave a value of 3A.

TAMURA (1954) discussed the recovery curves, making use of an assumption similar to that of HOLZER and SAXON (1952) (see § 12.22.), that conductivity alters with height, and he found good agreement between observation and theory for discharges inside clouds, giving a curve closely exponential for distant discharges.

12.12. Recovery after Close Flashes

WORMELL (1939) found that, for flashes closer than 3 km, the potential-gradient changes follow a different form from that for more distant flashes, dying away at first much more rapidly. This can be accounted for by the rapid dispersal of the space charge, which is responsible for the high potential gradient immediately after the flash. SMITH (1951a) used these results to obtain a value for the relation between potential gradient and point-discharge current density. SMITH's argument, somewhat simplified by approximations, is as follows: suppose that, before the flash, the potential gradient is small, owing to a near balance between the effects of the negative charge in the base of the cloud and the positive space charge below; the discharge now removes the negative charge and leaves a potential gradient X due to the space charge; this potential gradient is related to a surface density of charge on the earth, $\sigma = X\varepsilon_0$.

Now the potential gradient X gives rise to a point-discharge current density J which is BX^2, if the formula of WHIPPLE and SCRASE (1936) (see § 9.15.) is assumed and if X is sufficiently large for M in the WHIPPLE-SCRASE formula to be neglected. The current J produces ions which neutralize the space charge, and at the same time itself neutralizes the surface charge.

So

$$J = -d\sigma/dt = -\varepsilon_0 \, dX/dt.$$

But $J = BX^2$, so that

$$dX/dt = -BX^2/\varepsilon_0.$$

Thus for close discharges which give a small pre-discharge potential gradient the recovery should proceed according to a law $dX/dt \propto X^2$. This is in agreement with WORMELL's results and allowed SMITH to obtain a value of B. The simple WHIPPLE–SCRASE formula involves the current from a single point, while B is concerned with the current per unit area, so that SMITH was able to deduce the effective separation of points (see § 9.24.).

TAMURA (1954) (see § 12.11.) calculated the recovery curves to be expected for cloud discharges at short distances, but without considering space charges from point discharge, and obtained results in agreement with the observations.

A complete description of recovery curves should take into account both the space charge from point discharge and the alteration of conductivity with height, and this may mean that the apparently exponential recovery curve is not as simple as it appears; the consequences of this in relation to the internal dissipation current have yet to be worked out.

12.13. Relaxation and Regeneration Times

GUNN (1954b) pointed out that there are two important time constants which arise in the building up of electric charge in a thunder cloud. In the first place, there is the ordinary relaxation time, ε_0/λ (see § 2.27.), which is concerned with conduction processes moving charges; the value of λ, and therefore of the relaxation time, for the inside of the cloud is not known, but GUNN assumed it to be similar to that at the same level outside the cloud. In addition, there is what GUNN termed the regeneration time, depending on the rate at which the separation of charge is taking place, a time which, in an active thunderstorm, is much less than the relaxation time.

GUNN produced evidence of two kinds of lightning discharges, one of which occurs inside clouds, with a frequency depending on the regeneration time (see § 14.32.), and the other, from cloud to earth or cloud to air, with a frequency 5 or 10 times less, and depending on the relaxation time.

12.14. Conductivity in Thunder Clouds

FREIER (1962) has interpreted the recovery curves of potential gradient after lightning flashes in terms of the conductivity inside the cloud, and reached the conclusion that the conductivity must be around 20 times that in the free air at the same level.

Although an exponential rate of recovery might appear to involve only one constant, that interpreted by WORMELL (see § 12.11.) as concerned with the rate at which charge is separated, in fact if there is a final state which is approached, this does involve the rate of dissipation of charge.

If FREIER's conclusions are correct, then the process of charge separation must involve some creation of small ions. But CHALMERS (1964) criticized FREIER's argument because point-discharge currents were not taken into account.

12.15. The Alti-electrograph Results

The general mode of action of the alti-electrograph has already been discussed (§ 9.20.). The results (SIMPSON and SCRASE, 1937; SIMPSON and ROBINSON, 1940) showed that the balloons went into regions with potential gradients of both signs, in most cases with a negative potential gradient at the greatest heights above the cloud. Since the clouds are moving during the observations and the balloons do not ascend vertically, there is difficulty in interpreting some of the results, but in most cases there is no doubt that there is an upper positive charge and a lower negative charge; out of 15 soundings, for 13 different storms, 7 gave direct evidence of a positive charge in the lowest part of the cloud. Figure 61 shows the situations of the centres of charge.

SIMPSON and ROBINSON (1940) showed that the upper positive charge is usually at a temperature below -20 °C and always below -10°C. The temperature at the negative charge, in 13 cases out of 15, is below 0 °C and is about 15°C higher than at the positive charge. The lower positive charge is at a temperature above the freezing point in 5 cases out of 7.

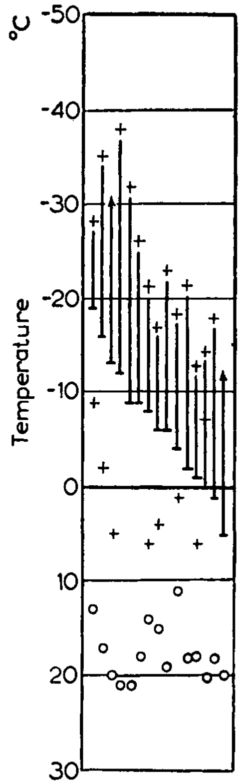


FIG. 61. Temperatures at charge centres and ground. (From SIMPSON and ROBINSON, 1940, Fig. 12, p. 321.)

The percentage of positive potential gradients varies with the height as shown in Fig. 62, giving quite definite evidence of three concentrations of charge, and a similar curve of the variation of the percentage of positive potential gradients with temperature (Fig. 63) is even more definite, the dotted curve showing what might be expected if there were only two charges.

In attempting to convert the qualitative results of the alti-electrograph into quantitative results from the widths of the traces, SIMPSON and SCRASE (1937) reached the conclusion that the potential gradients beneath the clouds were much less than would be expected on the theory of the space-charge blanket. Instead of showing the expected increase from the earth up to the cloud base, the alti-

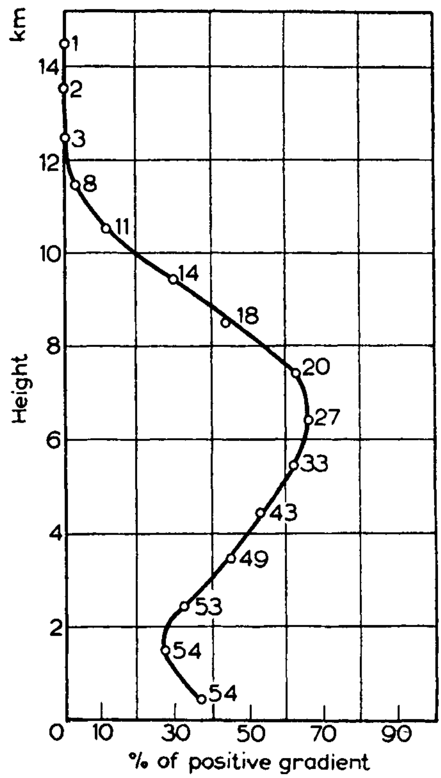


FIG. 62. Frequency of positive potential gradients according to height. (From SIMPSON and ROBINSON, 1940, Fig. 13, p. 321.)

electrograph records frequently remained of the same width, showing no effect of space charge. This problem has been discussed in §§ 9.23. and 12.10.

CHAPMAN (1953) with a radio-sonde found results essentially similar to those of the alti-electrograph.

12.16. Zugspitze Measurements

KUETTNER (1950) made a number of measurements at the Zugspitze (10,000 ft), which was usually above the base of a thunder cloud. Measurements made for any one storm can be considered to give a cross-section of the storm at a particular altitude, and, since the alti-electrograph results indicated that temperature is the

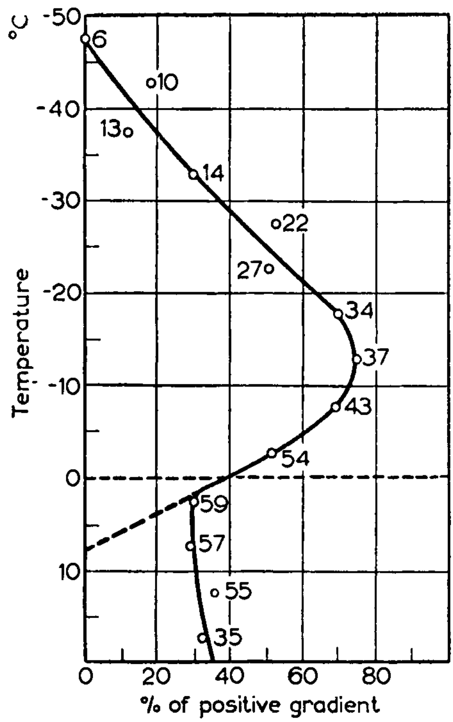


FIG. 63. Frequency of positive potential gradients according to temperature. (From SIMPSON and ROBINSON, 1940, Fig. 14, p. 323.)

main determining factor in the positions of electric charges in the cloud, measurements made under different temperature conditions should give cross-sections corresponding to different relative levels in the storm. While individual storms gave divergent results, the average results gave a picture of a thunder cloud quite similar to that from the alti-electrograph. Since the presence of the mountain obviously interferes with the actual potential gradient, KUETTNER merely measured the sign and obtained this from the point-discharge current. He found the centre of the lower positive charge to be at about the freezing-point level, rather than at +3°C as found by the alti-electrograph; the negative charge is at about −8°C and the upper positive charge higher, the actual level being always above the Zugspitze and so not directly measurable. The upper positive charge seemed to lag behind the lower charges, so that, behind the cloud, the potential gradient was more often positive than in front.

KUETTNER also recorded other facts which may be of importance in discussions of the origin of thunderstorm charges. The lower positive charge and the negative charge appear to be bound to the same vertical air column which shows heavy precipitation, strong down-draught and initiation of lightning. The main charges reside within the cloud and not either in the precipitation or the free air. There appears to be no distinction in kind, but only in degree, between shower clouds, without lightning, and true thunder clouds;

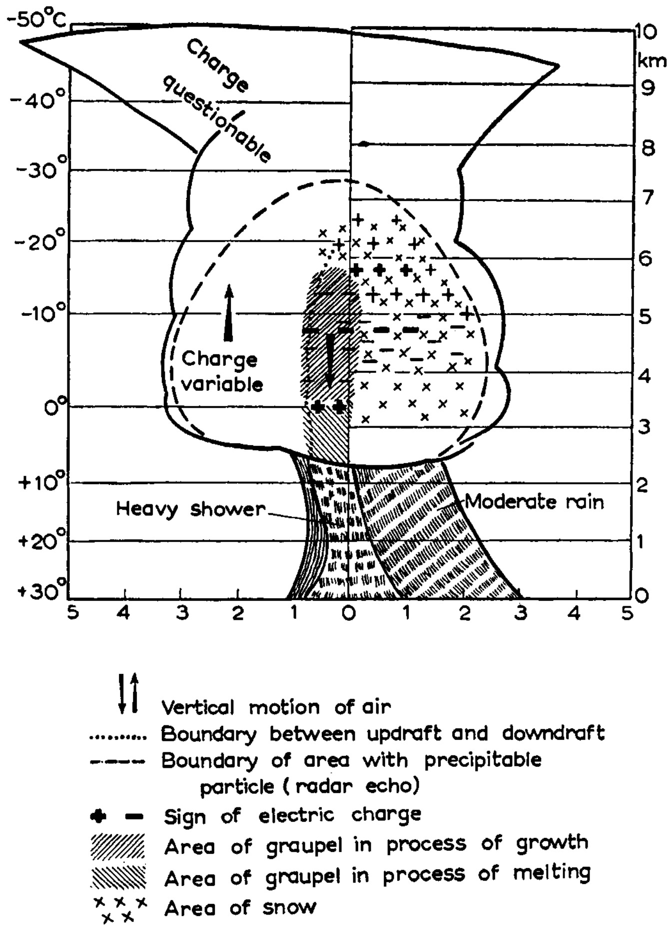


FIG. 64. KUETTNER's model of thunderstorm cell. (From KUETTNER, 1950, Fig. 12, p. 331.)

the same applies to snow-shower clouds except that the lowest portion of the structure is cut off by the cloud base or the ground. It seems a necessary condition for a thunderstorm that the cloud base should be lower than the freezing level. The potential gradient in the central area of the thunderstorm shows violent fluctuations quite apart from those connected with lightning. Large hail is by no means a necessary occurrence in a thunderstorm, but graupel, formed by the freezing of supercooled water droplets on to ice particles, seems to occur very frequently.

The agreement on most points between these results and those of the alti-electrograph shows that the thunderstorm develops at a particular temperature level, without reference to the presence of the earth's surface.

12.17. Double Dipole Theory

The results of the alti-electrograph and Zugspitze measurements show that the simple picture of a thunder cloud with two charges is incomplete and there must be added to it the lower positive charge. In addition to the main process of charge separation, leading to the two main charges, there may be a second, lower, process of charge separation giving the lower positive charge and adding to the main negative charge; alternatively, the lower positive charge may arise by some secondary process (see §§ 16.35., 16.36., 16.37.).

Kuettner (1950) considered two distinct processes of charge separation, occurring at different horizontal positions in the cloud, as well as at different levels. He considered that each process involves the upper charge remaining on the cloud particles and the lower charge falling out in the precipitation, so that the upper charge soon becomes larger than the lower charge from this process. Kuettner imagined the lower process to occur in the centre of activity of the cloud, in the region of strong down-draught, and he associated it with the process of graupel growth, the charges occurring at temperatures of -8°C and 0°C; these he termed the "graupel dipole". The second process of separation, in the opposite direction, is considered to occur in the snow region, mainly behind the down-draught region, giving charges at -20 °C and -8 °C; these he termed the "snow dipole". An important difference between this picture and earlier ideas is that the negative charge is here ascribed more to the lower than to the upper separation process, and is not in the same vertical line as the upper positive charge. This can account for the otherwise puzzling observation of WORKMAN, HOLZER and PELSOR (1942) that lightning flashes within clouds are often more nearly horizontal than vertical.

This picture would predict precipitation charges of different signs at different heights, but the signs expected are exactly opposite to those found by GUNN (1947) (see § 10.18.) for a cloud of a weak cold front.

12.18. Potential Gradients inside Clouds

Estimates of values of the potential gradient within clouds have been made from the widths of the alti-electrograph traces, but this method fails when the potential gradient exceeds a value estimated at about 10,000 V/m, as the current is then sufficient to give sparking at the electrodes; as stated earlier (§§. 9.23., 12.10.), there are reasons for doubting whether the estimates of potential gradient by this method are at all accurate. It has frequently been stated that the alti-electrograph seldom found potential gradients over 10,000 V/m and this has been taken as evidence that higher values are confined to very small volumes. But, in fact, when sparking occurred, this showed the existence of a potential gradient of greater magnitude and ROBINSON (1951) has drawn attention to the fact that 15 per cent of the useful ascents come to a violent end in a rapidly increasing potential gradient, probably with a flash of lightning through the balloon. The fact remains that the volume of the cloud, in which the potential gradient was sufficient for sparking, is smaller than would be predicted if the limit for sparking in the apparatus is 10,000 V/m.

ROSSMANN (1950) reported on one flight in a towed glider through the air just beneath a storm cloud and measured, directly, a maximum potential gradient of 9000 V/m; inside the cloud the potential gradient would probably have been considerably higher. CHAPMAN (1953), using the radio-sonde method, found values of potential gradient up to only 21,000 V/m in a thunder cloud and much lower in snowstorms.

GUNN (1948) made measurements in an aeroplane flying through a thunder cloud and found much larger values of the potential gradient; in one case there was a potential gradient of −340,000 V/m at the surface of the aeroplane, after correction for the effect of the aircraft's own charge, just before it was struck by lightning. GUNN found high values of the potential gradient to be much more extensive than had been found by the balloon measurements. The conducting aircraft would tend to concentrate lines of force in its neighbourhood, but this effect would not be sufficient to account for the discrepancy.

GUNN (1957) emphasized the fact, derived from his own observations, that there is an abrupt increase in potential gradient on entering a cloud; he related this to the change of conductivity, as discussed in § 2.26. The same has been found from tethered-balloon measurements by MOORE, VONNEGUT and BOTKA (1958).

12.19. Measurements in Aircraft

FITZGERALD and BYERS (1962) have used aircraft instrumentation, described in § 5.38., in flights near and in thunderstorms and clouds building up to storms. They found that there appear to be only small electric fields until precipitation begins; then fields develop corresponding to negative charge in the precipitation and positive remaining in the anvil, but there are also indications of lower positive charges. In one case a large hail shaft showed a negative charge.

12.20. Breakdown Potential Gradient

Normally in air at atmospheric pressure a potential gradient of about 3×10^6 V/m is required for breakdown with small sparks. At the pressures and temperatures in thunder clouds this would be reduced to about half. MACKY (1931) found that, in the presence of water drops, the potential gradient required is reduced, the exact value depending on the radii of the drops; for drops of 2 mm radius, the value is 1×10^6 V/m. It might be suggested that ice crystals, with their sharp points, should reduce the sparking potential gradient even more than do water drops, but LOEB (1953) has pointed out that the effect is very much limited by the high resistivity of ice. REYNOLDS (1954) suggested that sparking will commence in a cloud when the potential gradient is $1{\cdot}5 \times 10^6$ V/m.

For long sparks in air, NORINDER and SALKA (1951) found the breakdown potential gradient in air to be only about $4{\cdot}5 \times 10^5$ V/m at standard pressure and $2{\cdot}7 \times 10^5$ V/m at a height of 4 km, a representative height for a thunder cloud.

For a lightning flash to be initiated it is sufficient for the breakdown potential gradient to be reached in a small region, and the breakdown is then propagated. If, however, the breakdown potential gradient is reached at the surface of a conductor, there will be corona discharge or point discharge, but not necessarily lightning.

12.21. Quantities and Heights of Free Charges

There have been various estimates of the quantities of free charge in the different parts of an average thunder cloud. SIMPSON and ROBINSON (1940) found that the potential gradients below and near to their average thunder cloud would fit with charges of +24C at 6 km, −20C at 3 km and +4C at 1·5 km. As they did not include any space charges from point discharge, the actual lower cloud charges would be somewhat less.

GISH and WAIT (1950) found that the fields above a "less electrically active" storm cloud could be represented by +39C at 20,000 ft (about 9·5 km) and −39C at 10,000 ft.

MALAN (1952) found that his results for potential-gradient measurements at the ground could be represented by +40C at 10 km, −40C at 5 km and +10C at 2 km, again not considering space charges.

HOLZER and SAXON (1952) pointed out that the difference of conductivity at different levels would require different magnitudes for the upper and lower charges (see § 12.22.). KASEMIR (1965b) also took into account the change of conductivity on entering the cloud, and concluded that the true free charges should be much larger, giving a value of −340C for the negative charge. These calculations do not take into account the current of point-discharge ions moving upwards into the cloud, which must alter the effective conductivity. However, KASEMIR's basis of discussion in terms of currents, rather than of electrostatics, is certainly important.

FITZGERALD and CUNNINGHAM (1965) found negative charges at heights above 9·1 km, from measurements in actual flights through storms, and pointed out that measurements from outside the storms would have led to quite different conclusions.

It should be pointed out that the above results are conclusions derived from storms in a variety of different places and conditions, so that it would not be expected that the quantities or heights of the charges should be the same in all cases.

Apart from KASEMIR's, the figure given suggest that a lightning flash neutralizes a considerable proportion of the free charges in a cloud. But, as will be discussed in § 16.10., there may be much larger charges which have been separated on to different carriers but have not yet been segregated in space.

12.22. Distribution of Conduction Currents

In many cases, the calculations of charge distribution and currents near thunder clouds have been based on simple electrostatic principles, considering the earth as a conductor and the air as a very poor conductor of constant conductivity, while neglecting the electrosphere; the simple calculation of § 12.3 is an example. HOLZER and SAXON (1952) improved on this by assuming a simple bipolar cloud in a medium whose conductivity increases exponentially with height. They made the assumption that the upper charge has a value of 20C and found different values for the lower charge according to the separation of charges assumed; for a separation of 4 km the lower charge is 45C. The results for the currents agree with observations, as far as they can be tested, reasonably well; it is found that the current at 12 km, as measured by GISH and WAIT (1950), nearly all goes to the electrosphere and very little returns to earth. It is also found that currents to be expected from distant storms at 40 km are small compared with the normal air–earth current.

The assumption that the conductivity of the atmosphere continually increases with height is not in agreement with the *Explorer II* results (GISH and SHERMAN, 1936) which indicated a fall of conductivity at about 19 km and over; this would affect the conclusions above, and would result in appreciable currents from distant storms. However, the more recent balloon observations of STERGIS *et al.* (1955) show no such decrease in conductivity and the simpler theory appears adequate.

By assuming a conductivity at the base of the thunder cloud the same as in free air at the same level, HOLZER and SAXON are leaving out of consideration the quite considerable point-discharge current between cloud and ground.

12.23. Radar Observations of Thunderstorms

WORKMAN and REYNOLDS (1949a) observed a number of thunderstorm cells both visually and by radar reflections. Their results are presented, as an average for 12 storms, in Fig. 65. One of the most interesting of their observations is the very sudden appearance of the radar return, which must be due to the production of large

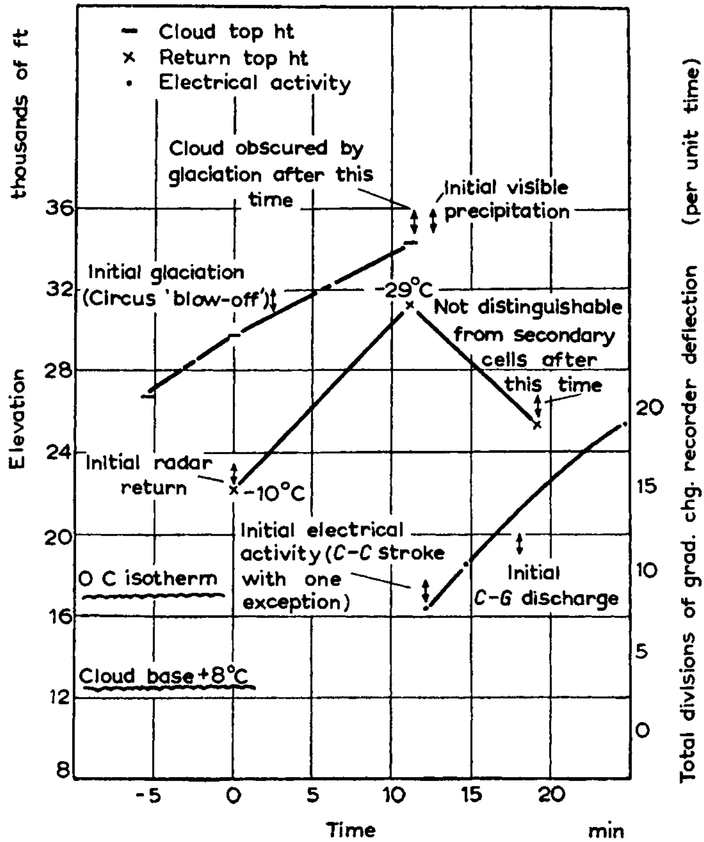


FIG. 65. Thunderstorm cell observations. (From WORKMAN and REYNOLDS, 1949a, Fig. 1, p. 143.)

enough particles in the cloud, though this occurs several minutes before precipitation becomes visible or before electrical activity is evident. After the initial radar return, the cloud development continues and the top of the "radar-cloud" rises, though remaining below the top of the actual cloud. Then, about the same time as electrical activity and visible precipitation begin, the top of the radar-cloud starts to descend.

REYNOLDS (1954) has drawn attention to the fact that clouds in New Mexico often develop to the height where the temperature is about $-20°C$ without producing lightning, while discharges never occur unless the cloud top reaches at least the height where the temperature is $-25°C$.

12.24. Relations between Electrification and Precipitation

On the basis of results similar to those described in § 12.23., REYNOLDS and BROOK (1956) discussed the relation between electrification, as evidenced by a significant change in potential gradient from its fair-weather value, and precipitation, as evidenced by 3 cm radar echoes. They formed the conclusion that precipitation is a necessary, but not sufficient, condition for the onset of thunderstorm electrification; a further condition appears to be rapid vertical development, and when this exists from the early stages, there appears to be little time interval between the first appearances of electrification and of precipitation echoes.

MOORE, VONNEGUT and BOTKA (1958) analysed similar data and pointed out that there might be electrification within the cloud considerably earlier than when it was detectable at the ground; by extrapolation backwards of the approximately exponential growth of electrical effects, they concluded that the potential gradient within the cloud might well be several times its normal value at the time when precipitation started. They obtained confirmation of these conclusions by measurements with a radio-sonde on a tethered balloon which entered the cloud. Their suggestion was that the electrification is due entirely to the rapid vertical development and that precipitation may be the consequence rather than the cause of electrification; they quoted an experiment by Lord RAYLEIGH (1879) in which electrification was shown to assist coalescence of drops. A theoretical treatment of the subject by SARTOR (1960) showed that coalescence should be much increased by the presence of an electrostatic field, even if the drops carry charges of the same sign. FREIER (1960b) produced further evidence for the same effect.

It is certainly necessary to obtain more evidence in regard to the precedence of electrification and precipitation, to decide the important matter as to which is to be regarded as cause and which as effect.

12.25. Gush of Rain after Lightning

It has many times been observed that there is often an abrupt heavy gush of rain shortly after a lightning flash. The question as to whether the heavy rain causes the lightning, or whether the lightning causes the gush, or whether both are consequences of some other cause, is very relevant to the problem of the relationship between electrification and precipitation.

The suggestion has been made that, in some way, the electric field holds up the precipitation particles, and when this is released by the lightning flash they can then fall.

Using radar observations from below a thunder cloud in the Bahamas, MOORE *et al.* (1962) found that a new echo, indicating large drops, often appeared shortly after a lightning flash, when there had previously been no such echo, and that heavy rain fell some minutes later. This indicates that probably lightning causes the heavy rain rather than the reverse and it was suggested that the electric field increases the probability of drops coalescing. Further evidence from observations of thunderstorms in New Mexico has been provided by MOORE *et al.* (1964), with similar conclusions in regard to radar echoes and confirmatory direct observations of rainfall; the gush may sometimes be of hail.

12.26. Warm-cloud Thunderstorms

While the alti-electrograph and Zugspitze results show that the main thunder-cloud charge centres are at temperatures below 0°C, there are several reports of storms giving lightning while completely at temperatures above 0°C (e. g. FOSTER (1950), APPLEMAN (1957), MOORE, VONNEGUT, STEIN and SURVILAS (1960), MICHNOWSKI (1963)). Unfortunately there are not yet any reports of such storms which have been investigated sufficiently to indicate the polarity of the cloud.

Cases of warm thunderstorms are necessarily rare, and in those reported the clouds either have grown later to reach above the 0°C level or have dissipated quickly. In order that a cloud can develop to thunderstorm conditions it must be of quite considerable vertical extent and this could be achieved below the 0°C level only if the base of the cloud is at a high temperature, i. e. only in tropical conditions. Therefore the lack of very many reports may be the consequence of the special conditions needed for warm thunderstorms rather than an indication that they do not exist or are "freaks". Further, such clouds would have only rather weak up-draughts and so, probably, rather weak electrical effects.

If it should turn out that these warm thunder clouds have the same polarity as the normal thunder cloud, it would be necessary, for these clouds at least, and perhaps by implication for all thunder clouds, to seek for a mechanism for the main separation of charge which does not require the presence of ice, since it is inconceivable that ice could occur in these clouds.

On the other hand, if warm thunder clouds should turn out to have the opposite polarity to that of normal thunder clouds, it could be suggested that these clouds represent a particularly intense manifestation of a separation process which produces the lower positive charge in the normal thunder cloud. One consequence of this would be that the lower positive charge would have to be the result of a distinct separation process and not merely the concentration of point-discharge ions produced lower down. A second consequence would be that the lower process of separation could not be connected with melting, as has been suggested.

A detailed investigation of the polarity of warm thunder clouds is most desirable. This could be obtained from measurements of potential gradients near such clouds, but it would be preferable to measure potential-gradient changes caused by lightning flashes, with simultaneous discrimination between cloud and ground flashes.

12.27. The Lower Positive Charge

WILLIAMS (1958) discussed the evidence in regard to the lower positive charge and came to the conclusion that its location is in the up-draught of the front of the cloud and close to the sharp boundary of the front of the precipitation zone, occupying a width of the order of 1 km in the direction of motion of the cloud.

If, as suggested by the results of CLARENCE and MALAN (1957) (see § 14.10.), the lower positive charge plays an essential part in a cloud–earth lightning discharge, then the pattern of hits to ground should yield information about the location of this charge, and WILLIAMS used the observations of FETERIS (1952) to obtain such results.

12.28. The "Physical Storm"

Many writers on thunderstorms, including particularly WALL (1948) and ISRAËL (1950a), have taken the view that there is no essential difference between the processes at work in a thunder cloud and in a non-thundery shower cloud; in a thunder cloud the potential gradient reaches the value necessary for a discharge, while in a shower cloud it does not. It is then possible to consider a "physical storm", which is a continuously acting shower cloud, in which the potential gradient never quite reaches the values necessary for lightning, but which otherwise has all the characteristics of the thunder cloud; in an actual storm the lightning can then be treated as an accident of secondary importance.

On this view, it is possible to consider the physical storm as a quasi-static state, in which the currents must be the same inside, above and below the cloud. The magnitude of such a current is found from the measurements of GISH and WAIT (1950) above the storm to be about 0·5 A on the average, or from those of STERGIS, REIN and KANGAS (1957b) to be about 1·3 A. Then there must be a process, or processes, within the cloud producing separation of charge at a rate of several amperes, most of which is dissipated by currents within the cloud; leaving a resultant current of about 0·5 A or 1·3 A to be dissipated by the external currents above and below the cloud.

12.29. Distinction between Storms and Showers

WICHMANN (1952b) differed emphatically from the idea of the "physical storm", with lightning as no more than an accident. Instead, he put forward the view that there is a difference, not only in degree, but also in kind, between the processes in the shower cloud and in the thunder cloud. He pointed out that the lower positive charge appears to be peculiar to the thunderstorm, and he suggested that it may be only when this charge exists that there can be a sufficiently strong potential gradient to give lightning, a suggestion previously made by SCRASE (1938). According to this view, there is some process common to showers and thunderstorms, which gives rise to the upper separation of charge; and, in addition, there is, in thunderstorms alone, a lower process which gives, at the same time, the positive charge and the intense potential gradient to produce a discharge.

Quite apart from the question of the second process of charge separation, WICHMANN also pointed out that the idea of considering thunderstorms to be represented by a quasi-static state must be incorrect, since each thunderstorm cell is a dynamical phenomenon, which passes through all its stages in a comparatively short time, a time which is in fact of the same order as the relaxation time of the lower atmosphere.

12.30. Mountain Thunderstorms

Writers of mountaineering literature have pointed out that there appear to be two types of mountain thunderstorm; one behaves quite similarly to an ordinary thunderstorm, giving lightning flashes both within the cloud and from cloud to earth; as might be expected, the earth flashes strike to mountain peaks preferentially. In addition, there are clouds which do not give rise to flashes when in the free air, but which show intense electrical effects when lying on mountain peaks, giving much point discharge particularly at metal points such as ice axes; this type of cloud gives rise to lightning flashes as it approaches or leaves a peak, particularly the latter; a very good (or should one say, striking?) example of this type was given by MURRAY (1951). A discussion of the subject was given by SMYTHE (1950).

It would appear that, in the first type of storm discussed, the process of separation of charge takes place at higher levels than the peaks, but, in the second type, charge separation occurs at mountain level and the local potential gradients become very large, though not large enough for a flash; as the cloud moves off, the charges become still more strongly separated, and a flash results.

12.31. Thunderstorms and Penetrating Radiation

WILSON (1925) pointed out that the potential gradients within thunder clouds might be sufficiently large for electrons to be accelerated in spite of the opposing forces in their passage through the air; he suggested such "runaway" electrons as the origin of cosmic rays but this was definitely disproved by the absence of much variation of cosmic-ray intensity with time. SCHONLAND (1930) attempted to find these accelerated electrons, but realized that the strongest potential gradients within thunder clouds would be such as to drive the electrons upwards; his failure to find such electrons below a thunder cloud was evidence that the negative potential gradient below the cloud is not sufficiently strong. SCHONLAND and VILJOEN (1933) used a counter and found effects due to electrons, occurring simultaneously with, and just before, lightning flashes from storms at distances greater than 30 km; they attributed these to electrons projected upwards, as suggested by WILSON and

SCHONLAND, and then bent in the earth's magnetic field. CAIRNS (1933) and APPLETON and BOWEN (1933) found similar effects, and HALLIDAY (1934), who used a cloud chamber actuated by atmospherics, found definite evidence of electrons moving from the north and particularly noticeable when the thunderstorm was in the west; this is in exact agreement with what is to be expected if the electrons are projected upwards and bent in the earth's magnetic field.

SCHONLAND (1930) found that there is a definite decrease in the intensity of the natural cosmic rays immediately below a thunderstorm; he ascribed this to a slowing down of the charged components of the cosmic rays in the intense fields in the thunder clouds.

12.32. Thunderstorms and "Cosmic Noise"

FLEISCHER and FALCONER (1960) found that thunderstorms could produce effects on the level of "cosmic noise" at 18 Mc/s, derived from radio-astronomical sources; some of these effects are very similar to those produced by solar flares, but their place of action must be much lower in the atmosphere.

12.33. Tornadoes

VONNEGUT (1956), VONNEGUT and MOORE (1958a) and VONNEGUT (1960) brought forward a suggestion that tornadoes, which cause so much damage in the United States, owe their concentrated force to electrical effects. Instead of the electrical energy being only a small part of the total energy, as is the case in a normal thunderstorm, VONNEGUT thinks that the energy of a tornado is largely electrical and that the dissipation of this energy in discharges gives rise to temperature differences sufficient to maintain the rotational energy of the storm. VONNEGUT (1960) gave several references to observations of intense electrical manifestations in tornadoes and quoted the observations of JONES (1955) that there are 10 to 20 discharges per second; if these involve as much energy each as a normal lightning discharge, there is ample electrical energy available and VONNEGUT considered the conversion of this first into heat and then into wind. VONNEGUT, MOORE and HARRIS (1956) showed, in a laboratory experiment, that an air vortex exercises a stabilizing influence on a high-voltage electrical discharge and it is suggested that this may be of importance in tornadoes.

Accepting the suggestion that tornadoes are powered by electrical effects, we would still have to enquire where the electrical effects originate and why all thunderstorms do not develop into tornadoes. Answering the second question first, it would seem that the distinguishing feature of a tornado is the concentration of the energy into a small volume.

On VONNEGUT's theory of the origin of thunderstorm electrification (see § 16.32.), the electrification is due to convection and hence the intense localized electrification of a tornado would then be caused by intense localized convection. On the more usual theories of thunderstorm electrification depending on precipitation phenomena, it would be precipitation which could be taken as the primary phenomenon.

It must remain for further investigations such as have been suggested by VONNEGUT (1960), to confirm, or otherwise, the suggestions above.

WILKINS (1964) used laboratory model vortex experiments and theoretical investigations, and came to the conclusion that the electrical effects were consequences of the phenomena of a tornado rather than causes of these phenomena.

12.34. Thunder

Although storms are known as "thunder" storms and the clouds concerned as "thunder" clouds, the audible phenomenon of thunder is probably less impressive than the visible phenomenon of lightning and has been much less extensively investigated.

It is clearly true that thunder is a secondary phenomenon, the sound waves being caused by pressure changes, caused in turn by the heating produced in the lightning stroke. A detailed discussion of the acoustical phenomena of thunder has been given by REMILLARD (1960).

Since sound waves travel much less rapidly than light, the time difference between the observations of light and sound give the distance from the observer of that part of the discharge from which thunder is first heard, i. e. for a vertical stroke to earth, from the point of fall.

12.35. Cosmic Thunderstorms

BRUCE (1959) suggested that the conditions for the separation of charge in the earth's atmosphere, leading to thunderstorms, might be paralleled in the "atmospheres" of stars and even galaxies, and he has used this idea of electric discharges to account for several of the unexplained phenomena of astrophysics, with time as well as space scales much larger. It is outside the scope of this book to discuss BRUCE's interesting speculations.

CHAPTER 13

Non-stormy Clouds

13.1. Electrification of Non-stormy Clouds

Non-stormy clouds are much more frequent in occurrence and much steadier in their conditions than storm clouds, and it might, therefore, be expected that electrical effects associated with the former would be more studied and better known. However, in fact, the much more intense effects of the thunder cloud have led to its more detailed study, and the electrical effects of non-stormy clouds have been somewhat neglected.

The clouds of most importance in regard to electrical effects, after storm clouds, are nimbo-stratus, or continuous rain or snow, clouds. Some electrical effects are also to be ascribed to fair-weather cumulus, alto-stratus, strato-cumulus and perhaps even cirrus-type clouds. In this chapter we shall also consider electrical effects in fog and mist.

13.2. Nimbo-stratus Clouds

In the absence of any measurements with instruments, such as the alti-electrograph, in nimbo-stratus clouds, the only information that is available is that which can be obtained from currents and potential gradients at the earth's surface.

Conditions during continuous precipitation from nimbo-stratus clouds are reasonably steady and it should therefore be possible to apply the principle of the quasi-static state, so that the total vertical current density is the same at all levels, and can therefore be obtained by measurements at the earth's surface.

The results of CHALMERS (1956) show that there is a distinct difference between the cases of clouds giving rain and snow (see §§ 10.22., 10.24., 10.25.). It is probable that, in England, any precipitation that reaches the ground as rain has been in the form of snow at an earlier stage in its history; seldom are there nimbo-stratus clouds wholly at temperatures above the freezing point and so, even if precipitation is initiated by the coalescence process, the particles are likely to have risen, during their growth, to levels where they will freeze.

Thus any process which occurs within the snow cloud can be presumed also to occur within the rain cloud, and so the snow cloud seems to give the simpler physical conditions and will be dealt with first.

13.3. The Nimbo-stratus Snow Cloud

The results of CHALMERS (1956) are that the total vertical current downwards to the ground from a nimbo-stratus cloud giving snow is negative (see § 8.11.). If it can be assumed that conditions are "quasi-static", then there must be a negative current downwards at all levels, so that the conduction current above the cloud brings down negative charge. Therefore the potential of the top of the cloud must be above that of the electrosphere, i. e. it must be greater than $+2{\cdot}9 \times 10^5$ V with respect to earth.

CHALMERS (1956) found that the potential gradient during snowfall from nimbo-stratus clouds was sometimes positive and sometimes negative, the mean value being slightly negative. Many other workers, e. g. REITER (1955a, 1965), have found the characteristic of snowfall to be a positive potential gradient at the ground.

If there is a process of charge separation at, or very close to, the ground, giving a positive charge to the air and a negative charge to the snow, then the vertical current in the region between the cloud and the ground must be carried by the positively charged air moving upwards; thus, if the snow falls from the cloud uncharged, the space charge in the region below the cloud must be the positive charge of this upward-moving air. If F is the potential gradient at the ground and h is the height of the cloud base, then the positive space charge reduces the potential at the cloud base to below the value Fh. If we assume h to be 1 km, then even if F is $+290$ V/m the cloud base has a potential below $+2{\cdot}9 \times 10^5$ V and, as discussed above, the top of the cloud has a potential above this. To maintain such a potential difference within the cloud, there would have to be some process of charge separation within the cloud. If F is less than $+290$ V/m, the separation within the cloud must be greater.

The very attractive idea, that the difference between the rain cloud and the snow cloud is to be ascribed to effects at or close to the ground, is not tenable, because the phenomena for the snow cloud cannot be accounted for by just considering the separation of charge to occur close to the ground.

Since there must be some separation of charge within the cloud, it seems reasonable to assume that this represents the only process of charge separation, unless we become forced to assume some other process.

The picture of the nimbo-stratus snow cloud, with positive charge above and negative below, is thus similar to that of the main charges in the thunder cloud.

ANDERSON (1965) reported measurements above snow clouds, giving currents bringing positive charges downwards, in agreement with the discussion above.

13.4. The Nimbo-stratus Rain Cloud

As discussed above, it is reasonable to suppose that rain drops falling from nimbo-stratus clouds will have been in the solid state for part of their previous history. It is therefore fairly certain that, for a cloud which gives rain, there occurs within the cloud the same process as for the snow cloud, which, as discussed in § 13.3., gives a separation with positive above and negative below. But, since the results show that, in the rain cloud, the total separation is in the opposite direction, it follows that for the rain cloud there must be some other process of charge separation giving negative above and positive below, and this process more than outbalances the snow process. Arguments similar to those for the snow cloud cannot give as definite a result because of the complicating effects of the simultaneous action of the snow separation of charge, but it seems probable that the rain separation does not occur at or near the ground. It therefore appears likely that the process that is specific to rain may be connected with the act of melting.

In the nimbo-stratus rain cloud there appear to be two processes at work, and we already know that there must be two processes at work also in the thunder cloud. Since, in both cases, the upper separation of charge gives positive above and the lower separation negative above, it is very tempting to suggest that the same processes act, on a different scale of intensity, in both types of cloud.

13.5. Trade-wind Cumulus Clouds

FITZGERALD (1956) investigated the electrical structure of small cumulus clouds, all at temperatures above the freezing point, near Puerto Rico, by means of aircraft flying in and near the clouds. The results showed a negative excess of the order of 10^{-3} C, with indications of a positive charge in the upper part of the cloud. He suggested that the charges might originate from the selective capture of ions, by WILSON's process (see § 3.11.).

FITZGERALD and BYERS (1958) extended the above results and showed that the negative charge is concentrated more in regions of high liquid-water content. As soon as solid precipitation was observed, a positive charge appeared, of greater magnitude than the negative charges in all-water clouds.

13.6. Electrification of Minor Shower Clouds

TAMURA (1956) investigated the electrical nature of minor shower clouds which appear in Japan in the cold season; these are of dimensions of the order of 2–3 km and give potential gradients up to only ± 2000 V/m, with no lightning.

In 7 out of 10 cases the polarity of the cloud was found to be negative and in 3 positive, so that more often the separation of charge is opposite in direction to that of the typical thunderstorm. It is also to be noted that the showers appear to be caused by convection and that the precipitation is often in the form of snow. It is clear that these clouds present a difficult problem in that they appear to be meteorologically similar to, but smaller than, thunder clouds, yet the electrical structure is often opposite in sign.

13.7. Charges on Droplets in Non-raining Clouds

GUNN (1952) investigated the charges in non-precipitating warm cumulus clouds (temperature about 20 °C), at a height of about 1400 m, by the use of apparatus carried in aircraft. The measurement of potential gradients at the boundary of and within the cloud showed values of less than 1000 V/m, from which it follows that there can be little free space charge within the cloud. By means of a centrifuge, he collected all droplets of diameter greater than about 10 μ and found charges on these of about $+1{\cdot}3 \times 10^{-9}$ C/m^3; he showed that such charges were not produced by the shattering of drops in the centrifuge, but must exist on the drops in the cloud. With a large-ion collector, he measured the charges on all ions and particles up to a diameter of about 0·01 μ and found charges of about $-2{\cdot}0 \times 10^{-9}$ C/m³. Droplets of diameters between 0·01 μ and 10 μ would be missed by both methods of measurement and, since the cloud as a whole is nearly electrically neutral, such droplets should contribute about $+0{\cdot}7 \times 10^{-9}$ C/m³.

WEBB and GUNN (1955) made similar measurements in clouds at two mountain stations (about 500 m and 2000 m high); their values for the charges were smaller than GUNN's for clouds in the free air by factors of from 3 to 6, but both for the larger drops and for ions they were of the same sign as GUNN's; however, for the mountain clouds, if the cloud as a whole is to be neutral, the droplets from 0·01 μ to 10 μ would have to be negatively charged, while for clouds in the free air the charge would be positive. In some cases, it was possible to count the total number of droplets of over 10 μ in diameter in a volume of cloud and, combining these results with those for charges, the average charge per droplet amounted to less than 1 electronic charge; since, however, this is the average, irrespective of sign, it may well be that some droplets carry charges amounting to several electronic charges, as indicated by theory (see § 13.8.).

TWOMEY (1956) measured the charges on individual cloud droplets at mountain stations by allowing them to fall through an alternating horizontal electric field; from the zig-zag motion he was able to derive the size and charge of the droplets. In clouds in which no ice existed he found almost always positive charges on the droplets, the charge per drop being approximately proportional to the surface area. When ice was present, negative charges also appeared and there was no correlation between charge and size.

PHILLIPS and KINZER (1958) used a method very similar to that of TWOMEY in conditions also quite similar and found quite different results. They found the charges on the individual drops to follow a Gaussian distribution about a mean value close to zero, instead of the definite preponderance of positive charges found by TWOMERY. When the drops were divided into groups of different sizes of drops, PHILLIPS and KINZER found that the mean charge was negative for the smallest drops and positive for the larger drops, in agreement with GUNN's results; the mean values were of the order of 1 electronic charge or less per drop; if the average charge, irrespective of sign, is obtained, this is from 4 to 8 electronic charges. ALLEE and PHILLIPS (1959) extended this work to the study of supercooled clouds and found similar results, in agreement with the theory of GUNN (1955a).

MAGONO and KIKUCHI (1961) measured the charges on large cloud drops (over 40 μ diameter) in clouds at the top of Mount Teine (1023 m) and found results in agreement with those of PHILLIPS and KINZER and of ALLEE and KINZER, rather than with those of TWOMEY, obtaining both positive and negative charges on drops at all temperatures; however, negative charges preponderated. They extended the work to ice particles, which they found to be more highly charged. The charges appeared in all cases to be approximately proportional to the cube of the diameter. They also found that, in general, there was a "mirror-image relation" (more properly, an inverse relation) between the charge and the existing potential gradient.

BARKLIE, WHITLOCK and HABERFIELD (1958) measured, in an aeroplane, the charge on air from a cloud sucked through a tube with sharp bends and found a preponderance of negative charge; since larger particles, calculated to be those of diameter greater than 1 μ under the conditions used, would hit the walls at the bends, this negative charge is that carried by particles less than 1 μ in diameter. They also suspended a cylinder below the aeroplane and calculated that it would capture particles larger than 10 μ in diameter, the smaller particles being carried round the cylinder in the air stream; they found that the cylinder acquired a positive charge, so that their results are in complete accord with those of GUNN. That the effects are not caused by shattering appears to be completely verified by the results obtained in laboratory experiments (see § 3.23.), demonstrating the effect of ions in the charging of the particles.

WHITLOCK and CHALMERS (1956) measured the potential gradient at the earth's surface below non-raining clouds, particularly stratus and strato-cumulus (see § 5.61.) and found effects to be attributed to negative charge in the cloud base, more pronounced in the thicker portions of the cloud. A similar result was discussed by ISRAËL (1959a).

Measurements of cloud-drop charges during precipitation have been discussed in § 10.18.

REITER (1965) has found negative charges in the bases of swelling, but non-precipitating, cumulus clouds and also in the bases of alto-stratus clouds transforming to nimbo-stratus. The same conclusion was reached for strato-cumulus clouds by KRASNOGORSKAYA (1961).

13.8. Theory of Cloud-droplet Electrification

PLUVINAGE (1946) discussed theoretically the capture of small ions by cloud droplets in a cloud which is substantially neutral, in which there is no appreciable potential gradient and in which ions of both signs have similar properties and are present in equal numbers. His results show that there will be, ultimately, a distribution of charges on the droplets, the values of the charges depending on the size of the cloud droplets. For droplets of average radius 2·5 μ, the average charges are $\pm 5e$, and for average radius 20 μ, $\pm 14e$, where e is the electronic charge. Calculations of conductivity based on this theory agree reasonably well with measurements in mountain fogs (see § 7.20.).

GUNN (1954c, 1955a) has worked out a theory on somewhat similar lines and extended it to the case where the polar conductivities of the two signs are not equal. The individual droplets carry charges which are distributed round a mean value which is zero when the conductivities are equal, but which is displaced towards the side of greater conductivity by an amount of the order of 1 electronic charge when the conductivities are not equal. The results of measurements in clouds (§ 13.7.) and in the laboratory (§ 3.23.) accord with the theory.

13.9. Charges in Mist and Fog

In wet mists, when the precipitation is too small to be recorded as such, it has often been found there are small currents to the earth, frequently of opposite sign to the potential gradient, so that they could not be conduction currents. SCRASE (1933) estimated that the negative current he found could be accounted for by each droplet carrying about 35 electronic charges. NOLAN (1940) and ISRAËL and KASEMIR (1952) found similar effects.

WIGAND (1926) and WIGAND and FRANKENBURGER (1930) obtained charges amounting to up to 2000 electronic charges per droplet in fogs, but PLUVINAGE (1945) pointed out that they had assumed all the charges to be on the droplets and had neglected the presence of ions.

WEBB and GUNN (1955) measured the net charges on the droplets and ions in a fog by the same method as for a cloud (see § 13.7.) and found an average charge on the larger droplets of -4×10^{-10} C/m^3, and $+14 \times 10^{-10}$, C/m^3 on the ions, corresponding to an average of less than one electronic charge per droplet.

CHAPTER 14

The Lightning Discharge

14.1. Investigation of Lightning

In this chapter we shall give an outline of the phenomena of lightning and the reader who desires further information is referred to the detailed accounts of MEEK and PERRY (1944–5), WORMELL (1953b), SCHONLAND (1956), MÜLLER-HILLEBRAND (1961) and to a comprehensive book by MALAN (1963), or to a book in more popular style by SCHONLAND (1964).

The lightning flash has been investigated photographically by the Boys camera in which the movement of lenses or film is used to separate effects which are visually included in the same flash, but which actually follow one another in quick succession. For photography of the flash, it is necessary to predict the direction of view and point the camera correctly; it is clear that photographic investigation can cover only that part of the discharge which occurs below the cloud base.

Lightning can also be investigated by the effects produced on the electrical potential gradient. Some confusion had been caused, first, by using apparatus designed to respond to the very rapid changes that are observed photographically and so not recognizing slower changes; later, too, there were further apparent discrepancies between the time scales of photographic and electrical processes arising from the fact that the first process occurs within the cloud and is not visible.

Direct investigation of the electric currents during lightning flashes is made difficult by the impossibility of predicting where lightning will strike; except in the case of very high buildings, such as the Empire State Building in New York, which are often struck, it is necessary to use simple apparatus which can easily be

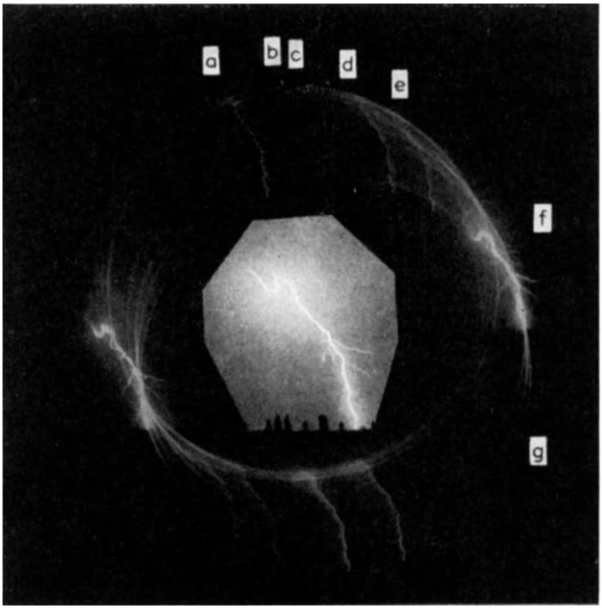


FIG. 66. Boys camera photograph. (From SCHONLAND and COLLENS, 1934, Plate 6, Fig. 6, between pp. 662 and 663.)

duplicated, so that a number can be installed in the hope that a few will give useful results. Measurements have been made of the lightning-discharge peak currents, rates of rise and fall of current and total charge passing.

Almost all of the most interesting results in regard to lightning have been concerned with flashes from cloud to earth rather than with flashes from cloud to cloud or within one cloud.

Now that the different features of a lightning flash have been fairly well established, it is possible to consider the whole flash in chronological order, rather than, as has been necessary in the past, to describe various features and point out discrepancies.

14.2. The Moving Camera

The first use of a moving camera to investigate the structure of a lightning flash was probably that of WALTER (1903), who found a single flash to be composed of a number of separate discharges.

BOYS (1926) first designed the camera which has been used for much of the work on the structure of the lightning flash. Essentially it consists of a fixed photographic plate, with two lenses revolving in a circle at opposite ends of a diameter. Each lens distorts the image of a non-instantaneous flash, but the two images are distorted oppositely, so that, from a comparison of the two pictures and a knowledge of the velocities of the lenses, it is possible to deduce the direction and speed of the visible discharge processes. Since the different processes at work have very different velocities, the maximum of information can be obtained only by the use of two moving cameras; in one form of such apparatus, one camera with two lenses gives about 50 rev/sec, while a single lens moves much more slowly giving only one revolution to 59 of the two-lens system. BOYS constructed his camera and carried it with him for several years in England without finding any opportunity to use it; however, when it was taken to South Africa, where storms are much more frequent and where it is easier to get all-round visibility, results were quickly obtained.

BOYS (1929) introduced an improved form of camera, still with two lenses, in which the lenses are fixed but the photographic film moves; this allows of easier interpretation.

MALAN (1957) has discussed theoretically the problems of photographing lightning and designed a camera in accordance with the principles derived, using a lens of long focal length and wide aperture; in this camera, the film moves as a belt between two pulleys.

14.3. General Results of Photographic Measurements

The structure of the visual lightning flash was elucidated using the Boys camera in a series of papers by SCHONLAND and his co-workers (SCHONLAND and COLLENS, 1934; SCHONLAND, MALAN and COLLENS, 1935, 1938; HODGES and COLLENS, 1938). The first main feature of the results is that what appears visually as a single lightning flash is usually not a single discharge but a number of separate processes following on one another. It is convenient to distinguish between the visual "flash" and its component "strokes" The general results show that each flash contains from 1 to 42 "main" strokes and that each main stroke is preceded by a "leader" stroke.

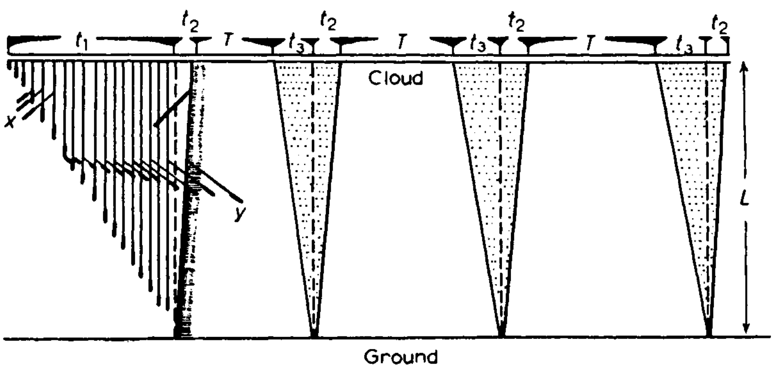


FIG. 67. Boys camera photograph of typical vertical discharge (not to scale). (From SCHONLAND, MALAN and COLLENS, 1935, Fig. 1, p. 598.)

Average values

$t_1 = 10^4\ \mu$sec $\quad t_2 = 40\ \mu$sec $\quad t_3 = 1000\ \mu$sec $\quad T = 3 \times 10^4\ \mu$sec

Leader strokes proceed downwards from the cloud; in many cases of the leader to the first stroke of a flash, the leader proceeds in steps and shows considerable branching.

When the leader stroke reaches a region from 5 to 50 m from the ground, a streamer from some point connected to the earth comes up to meet it and then the upward main stroke commences. This is much brighter than the leader stroke and follows it in its path.

After a time interval, there may be a second leader stroke, followed by a main stroke, and then, in many cases, a third and further strokes. The leaders to strokes after the first do not show steps except rarely at the lower end of the stroke, and they usually follow the same track as the first main stroke. Such unstepped leaders are termed "dart" leaders.

The idealized Boys-camera picture of a typical lightning flash is shown in Fig. 67.

14.4. Electrical Measurements

The electrical effects of lightning discharges are usually measured by the change produced in the potential gradient at a point which is nearly always on or close to the surface of the earth. There are two separate phenomena involved, the effect of the electrostatic potential and that of electromagnetic radiation; the first gives terms involving the inverse second and third powers of the distance, while the radiation effect is proportional to the inverse first power.

As we shall see in § 14.5., the electrostatic effect, proportional to the inverse cube, gives information about the change of position of the moving charge, and this is shown by measurements close to the discharge. On the other hand, the radiation effect depends on the acceleration of the charge and is more prominent at large distances.

Some of the methods used for the measurement of potential-gradient changes have already been described (§§ 5.34., 12.7.). The continuous recording of potential gradients, on a scale sufficiently open to show very rapid changes, is very wasteful of recording material even if recording is confined to times when storms are near, and any kind of triggering device would mean that the first stages of a discharge would be missed. To avoid these difficulties, CLARENCE and MALAN (1951) used a magnetic tape recorder; they "played back" the record, obtained permanent recordings of the interesting portions and then erased the tape for further use.

14.5. Electrostatic Field

The electrostatic effect of a lightning discharge is essentially that of the movement of a charge from one point or region to another. If we think only of the charge which moves and neglect any other charges present, then we must consider not only the charge concerned but also the effects of charges induced on the conducting earth; as has been discussed in § 2.19., these induced charges can be replaced by an "image" charge. So, if the charge which moves is $+Q$ at a height H, we must add an image charge $-Q$ at a height $-H$, and we can consider the two to form an electric dipole of electric moment $M = 2QH$. A discharge which takes the charge Q down to a height h alters the electric moment by $\delta M = 2Q(H - h)$ and for a discharge to earth $\delta M = 2QH$.

If the apparatus used has a very small resolving time, it is possible to measure not only δM for the whole flash, but also δM for the separate components.

The simple electrostatic potential V due to the dipole at a distance r, large compared with the length of the dipole, is given by the ordinary formula

$$V = \frac{M\cos\Theta}{4\pi\varepsilon_0 r^2}.$$

This leads to an electrostatic potential gradient at a point on the earth's surface of $M/4\pi\varepsilon_0 r^3$.

The change of potential gradient for any particular initial and final positions of a charge can be calculated quite easily; but if the flash is inclined, this has to be taken into account by using

$$\delta V = M_1/4\pi\varepsilon_0 r_1{}^3 - M_2/4\pi\varepsilon_0 r_2{}^3,$$

where r_1 and r_2 will no longer be the same.

14.6. The Inverse-square Term

In the literature on the effects of lightning on the electrical potential gradient, the term in $1/r^2$ is often referred to as an "induction" term, but this is incorrect.

In the derivation of the potential due to a dipole, the terms in $1/r$ cancel provided that the charges at the two ends of the dipole are equal and opposite. When we consider a dipole whose magnitude is altering, the potential at a point must be calculated, not from the values of the charges at the time concerned, but from their values and positions at a time r/c earlier, where r is the distance of the charge from the point and c is the velocity of light; then, except in the plane of symmetry, r will differ for the two ends of the dipole and so, effectively, the charges at the two ends are different. Thus the terms in $1/r$ do not cancel and we are left with a term in $1/r^2$ in the potential gradient. It is not difficult to show that this term is

$$(dM/dt)/4\pi\varepsilon_0 cr^2.$$

As can be seen from this argument, the term has nothing whatever to do with "induction" in any of the meanings of this word.

In actual measurements, this term is of little importance, since it is small compared with the electrostatic term for close discharges and small compared with the radiation term for distant discharges.

14.7. Radiation Field

It is shown in textbooks on electromagnetic theory that the radiation from an accelerated particle gives an electric field and also a magnetic field. The ratio of these is constant and their vector product gives the Poynting vector, the rate of flow of energy. Since this must follow an inverse-square law, the electric and magnetic fields must each follow a $1/r$ law.

The textbooks show that the electric field is $(d^2M/dt^2)/4\pi\varepsilon_0 c^2 r$. At great distances from the source, this term becomes much more important than the other terms.

14.8. Magnetic Measurements

A lightning flash is a current and therefore produces a magnetic field. This, if it can be measured, depends upon the current and so on dM/dt and varies with the inverse square of the distance. In addition, there is a magnetic component of the radiation and this, like the electrical component, depends on d^2M/dt^2 and so on the inverse first power of the distance.

The magnetic effect has been measured in two ways; NORINDER (1947) used a frame aerial, shielded to eliminate electrostatic effects, so that a changing magnetic field would produce an e.m.f., and he applied this to an integrating circuit (see § 2.49.). He was thus able to obtain a value for the magnetic field strength, provided he was sufficiently close to the discharge (less than 5 km) for the magnetic effect to be much larger than the radiation effect. With suitable assumptions he was then able to deduce the current.

WILLIAMS and BROOK (1963) (see § 14.23.) made direct measurements of the magnetic field with a fluxgate magnetometer.

14.9. General Results of Electrical Measurements

The early work on potential-gradient changes emphasized rapid, rather than slow, changes, partly because the measurements were mainly concerned with distant flashes, from which there is little radiation from slow changes, and partly because the apparatus was designed to respond to changes corresponding to the rapid effects discovered photographically. APPLETON and CHAPMAN (1937) found that the potential-gradient changes could be divided into three parts, corresponding to the stepped leader, the main stroke and a continuing discharge. SCHONLAND, HODGES and COLLENS (1938) found good correlation between electrical and photographic results for the same flash.

Measurements at Cambridge (WORMELL and PIERCE, 1948; PIERCE, 1952, 1955a, 1955b; WORMELL, 1953a, 1953b; PIERCE and WORMELL, 1953) were made using oscillographic equipment designed to show the effects of the slower processes in discharges at distances from 30 km to 150 km. The results show that the greater part of the total change of electric moment occurs, not with the rapidity of the visual flash, but much more slowly. KITAGAWA *et al.* (1953) obtained similar results using a field mill and a magnetic tape recorder.

The discrepancy has been resolved by CLARENCE and MALAN (1957), who were able to identify two preliminary processes occurring before the first visual process; it seems probable that the Cambridge workers were not able to recognize electrically the visual leader process. Both the preliminary processes and the visual leader process often show steps for the first stroke of a flash.

The time scale from electrical measurements is: changes before the main, return, stroke, from 0·05 sec to 0·4 sec; main stroke ("R") some msec; final change ("S") 0·1 sec.

14.10. Preliminary Processes

Clarence and Malan (1957) were able to resolve the discrepancy between the time scales in photographic and electrical observations of leader processes by the discovery that the processes before the main return stroke are more complex than had been realized previously. The first process is termed the "preliminary" or "B" potential-gradient change, followed by an "intermediate" or "I" change before the visual "L" change; the electrical measurements which have indicated a long time-scale have incorporated all these processes in the "leader" process.

The B process is identified as a discharge between the lower positive charge and the main negative charge, and results for the "reversal distance" (see § 12.3.) agree with this interpretation. At distances which are not too great, the B change, which lasts from 2 to 10 msec, sometimes shows steps.

The I change is interpreted as the charging up, with negative charge, of the conducting channel previously produced; this proceeds continuously and, usually, slowly, taking up to 0·4 sec.

It is not until the I process is complete that the true, visual, leader can proceed towards the ground, and this L process occupies the time as obtained from the photographic observations.

Takeuti, Ishikawa and Takagi (1960) found that, in about 50 per cent of the flashes, there is a "pre-preliminary" process, lasting even over 0·5 sec. No interpretation of this has yet been offered.

14.11. The Pilot Streamer

Schonland (1938) pointed out that the difference between the stepped leader to the first stroke and the dart leaders to subsequent strokes must lie in the fact that, for the latter, an ionized path already exists, whereas in the first stroke it is necessary for the insulation of the air to be broken down. Schonland suggested that the effective velocity of the stepped leader is, in fact, the velocity of the process of breakdown which is carried by what he called the "pilot streamer". He suggested that the usual process for the development of the pilot streamer is that there is a process of ionization in front of the tip of the streamer by means of electrons in the tip moving under the action of the field present. If the potential gradient is just sufficient for the ionization to occur, we get the minimum velocity of the streamer, namely, $1 \cdot 0 \times 10^5$ m/sec, and SCHONLAND found little branching in such cases. If, however, point discharge has produced a considerable space charge in the region below the cloud, the potential gradient becomes greater than the minimum, the pilot streamer becomes faster and extensive branching occurs.

SCHONLAND (1953b) considered the pilot streamer in detail and pointed out that the absence of steps in the electrostatic potential-gradient change measurements (CHAPMAN, 1939, etc.) shows that the pilot streamer and stepped leader cannot be explained by any sudden pulsations. He finds much evidence for identity between the pilot streamer of lightning and that of the long spark in laboratory investigations (ALLIBONE, 1948).

PIERCE (1955b) provided evidence from potential-gradient changes for lower pilot-streamer velocities, down to $1 \cdot 5 \times 10^4$ m/sec, and these might be barely possible on the theories. However, these changes might be due, not to pilot streamers of visual stepped leaders, but to the "preliminary" or "B" process, inside the cloud, where conditions are different.

There is no direct evidence for the existence of the pilot streamer, but it would be very difficult to account for the phenomena without it. The average current in the pilot streamer appears to be about 100 A.

14.12. The Stepped Leader

Photographic investigations have shown that stepped leaders can be divided into two types, called α and β. In type α leaders, the length of step is usually between 10 m and 200 m, with an average value of about 50 m. The average time taken by a step is about 1 μsec, so that the average velocity of traverse of the step is about 5×10^7 m/sec.

After a step, there is a pause of between 30 and 100 μsec before the next step starts, giving an effective velocity of propagation, which is the velocity of the pilot streamer, of the order of 10^5 to 10^6 m/sec.

In stepped leaders of type β the effective velocity at first is considerably higher, from 6×10^5 to 3×10^6 m/sec, but at some stage in the path the velocity drops to about 10^5 m/sec and sometimes the stroke never reaches the ground at all. The upper parts of type

β leaders are heavily branched and highly luminous with long steps; in the slower portion of the stroke it becomes faint with shorter steps.

Photographic observations show that about 65 per cent of the leader strokes are of type α, but some, which start within the cloud, might not show visually the rapid portion. Potential-gradient change measurements by MALAN and SCHONLAND (1951a) showed at least two-thirds of the initial leaders to be of type β, but some of these may be the "preliminary" discharges (see § 14.10.) and not the visual stepped leaders.

The pilot streamer must produce an ionized path, along which an electric current can flow, carrying charge with it. If the current flows more rapidly than the pilot streamer progresses, the current catches up to the streamer and then, for some reason, the stepped leader carrying the current is delayed while the pilot streamer goes on. Theories as to how this is brought about will be discussed later (§ 14.13.).

By comparing the rates of change of current in the step and return processes, HODGES (1954) has obtained values of the current surges in steps of types α and β as about 620 and 2600 A respectively.

14.13. Theories of Stepped Leaders

SCHONLAND (1938) put forward two possible explanations of the delay which causes the steps; according to one view, there must be a drift of electrons down the stem of the pilot streamer, but some electrons are captured by atoms and positive ions, so that time is required for a fresh supply of electrons to advance before there is a sufficient potential gradient to carry the current further. In the alternative view, the time is required for positive ions to gather below the tip of the leader and so to give a sufficient potential gradient at the tip. MEEK (1939) made calculations based on SCHONLAND's first idea, and, with not unreasonable assumptions, obtained agreement with the data.

More recently, SCHONLAND (1953b) attempted to explain the stepped nature of the leader in terms of a decrease of the potential gradient in front of the leader tip below the value at which it can advance in the form of a thermally ionized channel. Calculations based on this theory give good agreement with measured average values of the radius of the leader (2·5 m) and of the step length (20 m).

BRUCE (1944) put forward another alternative view, based on the transition from glow to arc discharge. He took into account the "corona discharge", similar to point discharge, from the leader stroke channel outwards, not only at its tip, but also along its sides, and he found reasons for believing that arc conditions will set in, for a time, after the streamer has progressed a certain distance, hence the steps. MEEK and PERRY (1944–5) dissented from this view on the grounds that experimental glow–arc transition data have been found for short gaps between metal electrodes and it is not legitimate to apply these to lightning.

PIERCE (1955b) made use of BRUCE's theory to give a general account of the leader process and the potential-gradient changes that occur; his value for the current is about 600 A. SCHONLAND (1962) has accepted BRUCE's theory and considers that the central "arc" core is probably only 2 mm in diameter at a temperature of about 10,000°C.

14.14. The Dart Leader

For a second and subsequent strokes, there is not a stepped leader but a "dart" leader. Dart leaders are similar to the steps of a stepped leader, but, since the ionized path exists already, there is no need for a pilot streamer and no steps. The ionization in the path of the dart leader is older than during the passage of the stepped leader, so that more recombination and diffusion have taken place and the remaining ionization is therefore less intense; in consequence, the velocity of the dart leader is less than the velocity in the individual steps, but greater than that of the pilot streamer. Values for the velocity of the dart leader are from 1×10^6 to 2×10^7 m/sec.

When there has been an exceptionally long interval between successive strokes in a flash, there may be little ionization left, particularly at the bottom of the path, and then there may be need for a new pilot streamer to ionize the air afresh, which explains the occasional cases of steps at the lower ends of leaders to discharges other than the first of a flash.

14.15. The Upward Streamer

As the leader stroke progresses downwards towards the earth, charge is being brought down in the channel and thus the potential gradient near the ground is being increased. The field strength necessary for local ionization by collision is about 30,000 V/cm and this has been reached close to points, trees, etc., even before the lightning flash starts, and gives point-discharge currents. But when the approaching leader causes a considerable increase in potential gradient, the region in which the value of 30,000 V/cm is reached becomes larger and upward streamers can progress from points, trees and even small projections on the ground. One of these upward streamers will ultimately join with the leader and give the "point of fall" of the lightning flash.

GOLDE (1945b) discussed the conditions necessary for the production of the upward streamer and came to the conclusion that the length would depend on the quantity of charge concerned in the leader and on the height of the point from which the upward streamer starts; the actual height of the point of junction might be from a few metres, for a weak stroke going to ground, up to over 50 m for a strong stroke to a point several metres up. GOLDE (1947) discussed some photographs which confirm these general conclusions.

14.16. The Main Stroke or Return Stroke

As soon as the leader, in one of its branches, reaches the earth, there will be an ionized path, acting as a good conductor, connecting the negative charge, which has become distributed through the branched leader, to earth. The charge in the leader then quickly travels to earth, being "drained" away, starting from the bottom of the leader; the main stroke thus appears to travel upwards, as though it were a stroke of positive charge, but, since the luminosity arises from the change in the motion of the charges, it is in fact the negative charges in motion which cause the effects. There are considerable charges distributed in the branches of the first leader and these, too, are drained back as soon as the main stroke has got as far as the branching-point; then the main stroke, which appears to be travelling *up* the main path, appears to travel *down* the branches, and catches up with the leader process still moving, much more slowly, down the branch.

Since the main stroke is removing charge along an ionized path, its velocity is greater than the effective velocity of the leader and has a value of from 2×10^7 m/sec up to $1{\cdot}5 \times 10^8$ m/sec; the maximum current is usually of the order of 2×10^4 A though 10^5 A is sometimes reached.

Since a dart leader moves more rapidly than a stepped leader, the ionization near the top of the leader is fresher for a dart leader than for a stepped leader at the time when the main stroke starts, and so the main stroke to a dart leader may be more rapid and powerful than that to a stepped leader.

14.17. Current in Return Stroke

BRUCE and GOLDE (1941) used results of measurements of currents to conductors that are struck, and expressed the results in the form

$$i_t = i_0[\exp(-\alpha t) - \exp(-\beta t)],$$

where i_t, i_0 are the currents at times t and 0 after the start of the return stroke. They gave average values of $i_0 = 28$ kA, $\alpha = 4{\cdot}4 \times 10^4$ and $\beta = 4{\cdot}6 \times 10^5$. PIERCE (1960) pointed out that there is likely to be a difference between effects during the return stroke to the first leader and return strokes to subsequent leaders, because of the difference in ages of the ionized column (see § 14.16.); for the first, stepped, leader, the ionization near the top of the column has had an appreciable time (of the order of tens of milliseconds) to dissipate; but for subsequent dart leaders, the time is much less (about 1 msec). He suggested that, for strokes after the first, the constant β should be altered from $4{\cdot}6 \times 10^5$ to 7×10^5.

BARLOW, FREY and NEWMAN (1954) suggested that there should be a third, more slowly decaying, term and HEPBURN (1957) showed this to be necessary to account for the intensities of atmospherics of frequencies less than 1 kc/s.

NORINDER (1951) obtained values of the currents in return strokes from magnetic measurements (see § 14.8.) and used the same formula as BRUCE and GOLDE; but he found the constants to be $i_0 =$ 20 kA, $\alpha = 7 \times 10^3$ and $\beta = 4 \times 10^4$. This is a very large discrepancy, since BRUCE and GOLDE's figures give a maximum of 20 kA at $t = 5{\cdot}6\,\mu$sec and a fall to half this value after 20 μsec, and NORINDER's a maximum of 12 kA at 53 μsec and a fall to half value after 200 μsec. PIERCE (1958) attempted to explain the discrepancy in terms of lateral corona current, without success. MALAN (1963) pointed out that NORINDER interpreted his results with the idea that the same current is flowing in the whole of the channel; since, in the early stages, the current is mainly at the lower end, the difference between the two sets of results can be partly explained.

14.18. Diameter of Channel

SCHONLAND (1937) measured photographs in which successive main strokes had been separated by the wind, so that each stroke could be observed separately and its diameter determined. He concluded that the average diameter of the main-stroke channel is 16 cm.

HILL (1963a) considered that the size of his "fulgamites" (see § 14.35.) gave evidence of a channel diameter of only a few mm.

UMAN (1964) measured the holes produced in fiberglass by lightning flashes and found some of diameters between 2 and 3·5 cm and others of diameters between 2 and 5 mm.

14.19. Multiple Strokes

After a single leader and main stroke have been completed, there is often, after an interval, another stroke which consists of a dart leader and main stroke. After this, there may be further similar strokes following at about the same intervals; in some cases there may be a continuing current with continuing luminosity (see § 14.23.).

SCHONLAND (1938) suggested that the subsequent strokes were to be ascribed to leader strokes starting from fresh charge centres and travelling to the original centre or to the track. BRUCE and GOLDE (1942) suggested that the process develops in the opposite direction, from the original channel to the fresh centre; the channel is at earth potential after the first return stroke and carries a surface positive charge induced by the negative charge in the cloud, and the potential gradient is likely to be greatest near the channel. Alternatively, PERRY (1944) and MCCANN (1944) suggested that the charging process within the cloud might be sufficiently rapid for fresh charges to develop at the original centre. The results for junction leaders (see § 14.25.) confirm that the idea of BRUCE and GOLDE (1942) is more probably correct.

The latter strokes usually follow the path of the first, but cases are known, e.g. WALTER (1936), where this is not so. KITAGAWA, BROOK and WORKMAN (1962) found that when the channel had lost luminosity for over about 90 msec it was no longer sufficiently conducting for another stroke to follow the same path; in such cases there must be a fresh stepped leader developing a new channel. In some cases, a later stroke follows the path of an earlier one for part only of the track and then diverges, perhaps along a branch which failed, in the original stroke, to reach the ground.

14.20. Number of Strokes

In discussing the number of strokes in a multiple flash it is necessary to distinguish between separate leader-return stroke combinations and minor changes of luminosity, such as those associated with K changes (see § 14.26.), and this is not always achieved by photographs.

Large numbers of separate luminosity maxima have been reported, e.g. 54 (WORKMAN, BROOK and KITAGAWA, 1960), 48 (MALAN, 1956) and 40 (LARSEN, 1905). But these are not all separate leader-return stroke combinations; in the case of WORKMAN, BROOK and KITAGAWA there were 26 and in that of MALAN 16; WORKMAN, BROOK and KITAGAWA (1960) have concluded that in LARSEN's flash there were no more than 16 separate leader-return stroke combinations.

The number of strokes in a flash must depend on the positions of the various charge centres. Both British (PIERCE, 1955a) and South African (SCHONLAND, 1956) observations show that frontal storms give a much higher proportion of multiple flashes than do heat storms, and this is consistent with the picture of a frontal storm being larger and more complex than a heat storm. The number of strokes per flash does not appear to show any marked dependence on latitude.

BRUCE and GOLDE (1942) analysed the relative frequency of different numbers of strokes in a complete flash, from the various observations available to them at the time, mainly from South Africa; they found that above 50 per cent of the flashes had at least 2 main strokes and 25 per cent at least 4. More recent results have shown a larger percentage of multiple strokes, values including 62 per cent (PIERCE, 1955a), 78 per cent (SCHONLAND, 1956), 81 per cent (NORINDER and KNUDSEN, 1961), 83 per cent (KITAGAWA and KOBAYASHI, 1958), 86 per cent (KITAGAWA, BROOK and WORKMAN, 1962) and 87 per cent (MALAN, 1956). Some of the data are shown in Fig. 68. A good average may be to regard 80 per cent or more of all flashes as multiple, with 3 as a median number of strokes per discharge.

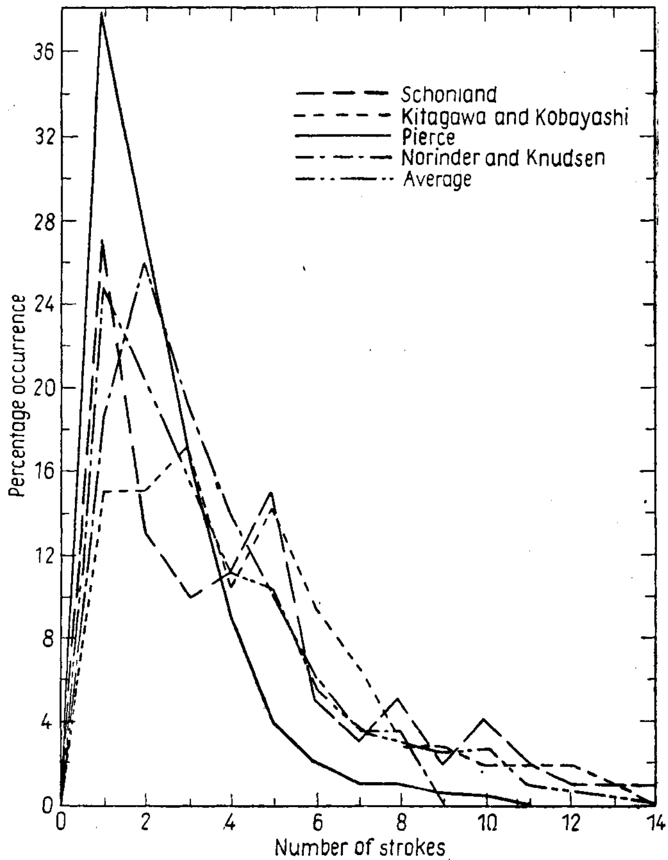


FIG. 68. Distribution of multiple flashes.

14.21. Intervals between Strokes

BRUCE and GOLDE (1942) found the average interval between strokes to be about 70 msec, but SCHONLAND's (1956) analysis gave a binomial with the most probable value about 45 msec and a long tail. PIERCE (1955a) obtained a median value of 65 msec; KITAGAWA and KOBAYASHI (1958) found some cases of long intervals for high-order strokes, but fewer than 1 per cent of the intervals are over 500 msec and none exceeding 1 sec has been recorded.

In § 14.19., it was stated that KITAGAWA, BROOK and WORKMAN (1962) had found that the ionized path would no longer support a fresh dart leader after 90 msec, so it might be expected that this time would represent the longest interval between strokes following the same path. However, if there is a continuing current (see § 14.23.), the ionization is maintained, and after the current has ceased a fresh dart leader can follow the path.

LARSEN's (1905) flash lasted 624 msec, MALAN's (1956) 1·5 sec and that of WORKMAN, BROOK and KITAGAWA (1960) created a record of 2 sec.

14.22. Duration and Frequency of Discharges

BROOK and KITAGAWA (1960a), from the analysis of records of isolated and widespread storms in New Mexico, put forward the following conclusions: for isolated storms, the mean duration of a single flash, probably proportional to the amount of charge involved, increases as the frequency of discharge increases. But for widespread storms, the average interval between discharges is an inverse function of the horizontal extent of the storm, as would be expected if the widespread storm consists of a number of independent cells.

BROOK and KITAGAWA (1960a) defined a "thunderstorm activity index" by measuring the cumulative duration of all discharges in a certain time.

14.23. Continuing Currents

Lightning flashes are usually considered to involve discontinuous processes, but MALAN (1954) found that there often occurred slow final potential-gradient changes, described as "S" changes; these occur most frequently after flashes with fewer than 4 strokes. These slow changes are considered to be caused by a continuous discharge from regions of the cloud higher than those tapped by the earlier separate strokes. MALAN suggested that the discharge is continuous, rather than intermittent, if the charges in the cloud are sufficient to maintain the ionization in the conducting channel below the cloud. This is supported by evidence that a continuing luminosity was observed photographically in about the same proportion of cases as the slow final potential-gradient change was observed electrically, though a direct correlation was not obtained. MALAN considered that the intermittent and continuous discharges remove charges from regions in the cloud up to the -41°C isothermal.

Further work was done by KITAGAWA, BROOK and WORKMAN (1962) and BROOK, KITAGAWA and WORKMAN (1962), who found that in about 50 per cent of multiple flashes there is at least one period of continuing current after a return stroke, lasting at least 0·04 sec and sometimes up to 0·5 sec; these continuing currents never occur after the first stroke of a flash and are sometimes, but not always, associated with the last stroke. The continuing current contributes an appreciable amount to the total charge brought down by a flash, and those flashes which include a continuing current lower about twice as much charge as those which do not. The actual values of the continuing current range from 38 A to 130 A. During the continuing current there are often detectable changes in current and luminosity, in some cases similar to the "K-changes" (see § 14.26.).

Continuing currents are also found in discharges to high buildings (e.g. the Empire State Building, see § 14.39.) and to towers in the Alps (BERGER, 1961), and show currents of the same order as those found in strokes to ground.

Another method of measuring the continuing currents has been used by WILLIAMS and BROOK (1963), who used a fluxgate magnetometer to measure the magnetic field produced by the continuing current; with measurements of the distance, from the lightning–thunder time interval or from an array of sound-ranging microphones, and the assumption of the length of the stroke, they were able to obtain values for the current, the average being 184 A, of the same order as found by other methods. The average charge brought down by continuing currents was found to be 31 C, representing an important contribution to the total charge in the flash.

14.24. Heights of Charge Centres

Heights of charge centres can be obtained from measurements of the electrical effects due to the clouds; various results have been obtained for the height of the negative charge centre, values being 3 km (SIMPSON and ROBINSON, 1940); 3·3 km (GISH and WAIT, 1950); 5 km (MALAN, 1952); 6 km (REYNOLDS and NEILL, 1955); 6–8 km (TAMURA, 1958) and 3–6 km (TAKAGI, 1961). These differences may well be caused by the locations of the storms investigated, the position of the charge depending on such factors as temperature.

Detailed investigations of the heights of the charge centres responsible for the successive strokes of a multiple flash were made by MALAN and SCHONLAND (1951b) who used 5 different methods mostly involving measurements of electrical potential-gradient change. Their results showed that, in most cases, successive strokes come from charge centres lying at increasing heights, probably in the same vertical column and therefore associated with the same cell. HACKING (1954) used simultaneous observations at three stations to verify that successive strokes come from regions of increasing height, but found that occasionally there are considerable

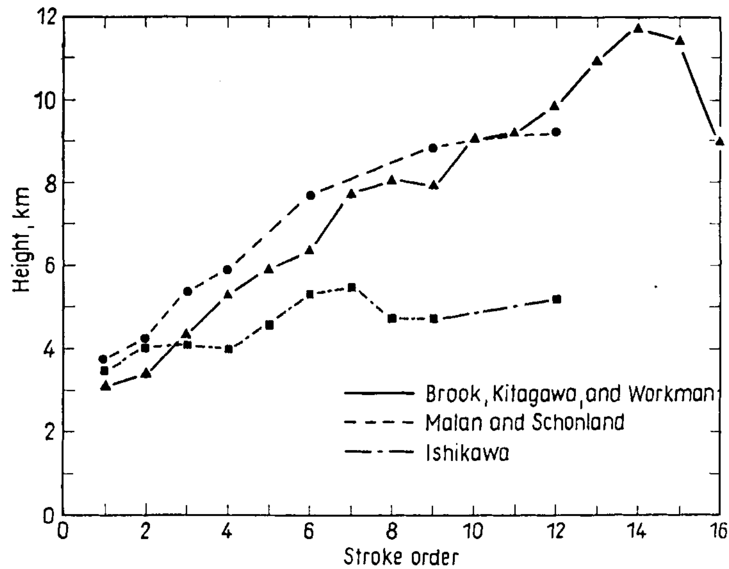


FIG. 69. Height of discharge vs. stroke order.

horizontal displacements. The results of BROOK, KITAGAWA and WORKMAN (1962) showed that, in New Mexico, there is an increase in height with order of stroke similar to that found in South Africa (see Fig. 69).

On the other hand, results in England are different; PIERCE (1955a, b) found that, in 71 per cent of the flashes showing more than one main stroke, the centres must have been at very nearly the same level, though separated horizontally, and in only 29 per cent were there centres at increasing heights.

Results on Japan appear contradictory; for summer thunderstorms, KITAGAWA and KOBAYASHI (1958) found results similar to those in South Africa. They also found that the density of negative charge decreases with height in the column and that the negative charge is non-uniformly distributed, being associated with a number of small cells of about 300 m diameter. ISHIKAWA (1961), however, found very little change of height with stroke order (see Fig. 69).

14.25. Junction Leaders

MALAN and SCHONLAND (1947) observed the potential-gradient changes during the intervals between the successive strokes of a flash and interpreted these as caused by the movement of charge involved in the joining-up of the different charge centres responsible for the separate strokes. In many cases, the junction potential-gradient change was found to be negative when close and positive when distant, corresponding to the motion of negative charge downwards or positive charge upwards in the cloud.

MALAN and SCHONLAND (1951 a) found seven cases where the junction potential-gradient change altered from positive to negative during the change and one where the alteration was in the opposite direction. For the 7 cases this is interpreted as positive charge moving upwards at just the right distance away to give the reversal, similar to SMITH's "loops" (see § 14.30.), and the estimates of distances and heights agree with this idea.

BROOK and VONNEGUT (1960) made visual observations of the junction process and found many of them to be more nearly horizontal than vertical, showing that the phenomena in New Mexico are more nearly similar to those in England than to those in South Africa (see § 14.24.).

TAKAGI (1961) found that the junction process, like the intra-cloud flash (see § 14.30.), consists of a slow positive streamer followed by a more rapid movement of negative charge; in a cloud-to-ground flash the positive streamer of the junction leader travels from the top of the main discharge, joining up with other negative centres of charge; these charges then flow to earth either as part of a continuing current or as the later parts of a multiple stroke.

14.26. Small Rapid Changes

KOBAYASHI *et al.* (1958) observed small rapid field changes, named "K" changes between and after flashes to ground. These K changes are recognized by luminosity variations and by the detection of high-frequency pulsations, and occur at irregular intervals of from 4 to 30 msec. These changes are interpreted as the neutralization of small concentrations of negative charge in the cloud.

14.27. Effects of Close Discharges

MALAN and SCHONLAND (1947) investigated the electrical and optical effects of discharges closer than 6 km. By a triggering device due to SCHONLAND and ELDER (1941), they were able to obtain photographs of daylight discharges. As the triggering device depends on sudden changes of potential gradient, WORMELL (1953b) pointed out that these measurements would tend to record more readily those strokes which have very prominent stepped leaders and to ignore those without stepped leaders, so that a representative sample would not be obtained.

The general interpretation of results from close discharges differs from that for more distant discharges, since electrostatic, rather than radiation, effects predominate. For close discharges the potential-gradient change produced by the leader process is of a "hook" shape and always corresponds to a lowering of negative charge. The time scale is longer than for visual flashes and more recent work interprets this in terms of the "preliminary" discharge (see § 14.10.).

The potential-gradient changes due to leaders of close discharges seldom show the "step" effects which might be expected from a stepped leader, although these do appear for more distant discharges. For close discharges, electrostatic effects predominate and so it must be concluded that the main transport of charge is in the continuously moving pilot streamer and not in the stepped leader. On the other hand, for distant discharges it is the radiation effect which is important and this depends upon acceleration of charges, so that small charges with high acceleration in the steps would give more marked effects at greater distances. Similar conclusions have been reached by CHAPMAN (1939) and PIERCE (1955a, b). Statistical results on the relative numbers of stepped and unstepped leaders are liable to error, since stepped leaders are more prominent visually, as well as more readily actuating a triggering device.

The recovery of potential gradient after close discharges has been dealt with in § 12.12.

14.28. Discharges Bringing Positive Charge to Earth

Discharges which show, for a distant flash, a negative change of potential gradient and, at the same time, the rapid "R" change which denotes a discharge to earth, are rather rare; BRUCE and GOLDE (1941) estimated that about 10 per cent of discharges to ground bring down positive charge and PIERCE (1955a, b) found that about 10 per cent of discharges to ground contain at least one positive stroke. But ISHIKAWA (1961) found the proportion of discharges bringing positive charge down to the ground to be less than 4 per cent of all discharges to earth and less than 1 per cent of all lightning flashes.

The changes of electric moment associated with such discharges are large and the leader and final stages take unusually long times. These facts suggest that these positive discharges are from the upper pole of the cloud rather than from the lower positive charge. BERGER, quoted by SCHONLAND (1962), found positive discharges in the later stages of thunderstorms, when negative charges will have passed to the ground in flashes and the remaining positive charge in the top of the cloud is subsiding; exceptionally large charges, up to 100°C, and currents, up to $1{\cdot}8 \times 10^5$ A, have been reported in such cases.

KITAGAWA *et al.* (1953) found a few cases of discharges bringing positive charge to earth in a single stroke and interpreted these as discharges from the lower positive charge.

Multiple positive discharges are very infrequent and it is more likely for a positive first stroke to be succeeded by a negative second stroke than a positive.

KASEMIR (1962) has found cases, in which there is a negative potential-gradient change for a ground flash, and which therefore have been interpreted as bringing down positive charge, but which, in fact, involve a large negative potential-gradient change in the "predischarge" or preliminary processes (§ 14.10.) and a smaller positive change in the main discharge, so there is still negative charge coming to ground.

14.29. Changes of Electric Moment

PIERCE (1955a) has given a table of the average change of electric moment in the various sections of a lightning flash. For flashes consisting of only one stroke, they are

$$L = -35\,\mathrm{C\,km}, \quad R = -29\,\mathrm{C\,km}, \quad S = -25\,\mathrm{C\,km}.$$

For flashes of two strokes

$$L = -31\,\mathrm{C\,km}, \qquad 1\mathrm{st}\,R = -28\,\mathrm{C\,km},$$

$$2\mathrm{nd}\,R = -34\,\mathrm{C\,km},$$

and S, when present, -19 C km. For the 29 per cent of the cases in which there is a J change, the average is -13 C km.

PIERCE's "L" includes the "B" and "I" changes (see § 14.10.).

14.30. Intracloud Flashes

PIERCE (1955a) has shown that the very rapid "R" potential-gradient change occurs only if the flash reaches the ground; for flashes within a cloud the potential-gradient changes are all slow and the total duration amounts to about 0·2 sec.

KITAGAWA and BROOK (1960) compared records of potential-gradient changes for ground and cloud discharges; they found that the initial portions of the records differ considerably. Both show pulses but, for the ground discharge, the mean repetition rate is about 80 μsec, while for the cloud discharge it is about 680 μsec. In the later portions of the records the cloud discharge shows exactly similar characteristics to the "J" or junction-leader changes between the strokes of a multiple flash to ground (see § 14.25.), and in both cases the K changes (§ 14.26.) are evident. They suggested that the difference between the initial portions might be ascribed to the different temperatures and hence to a difference between the droplets of water present which affect the breakdown of insulation (see § 12.20.); for the ground discharges the negative charge is, on the average, at a temperature of -6°C, but for cloud discharges the centre of the dipole is at about -18°C.

PIERCE (1955a) analysed records of lightning flashes at Cambridge, England, and found that the direction of motion of charge could be summarized as in Table 5.

SMITH (1957) investigated the slow potential-gradient changes due to cloud flashes, lasting several tenths of a second. Many of these

TABLE 5

	Ground flashes	Cloud flashes	Total
Negative charge down	35	40·5	75·5
Negative charge up	2·5	11	13·5
Charges in both directions	1·5	9·5	11
Total	39	61	100

show "loops", involving a change of potential gradient, first in one direction and then in the other. SMITH explained these in terms of the reversal distance (see § 12.3.); if one portion of the discharge is within the reversal distance and the other outside, then the potential-gradient changes will be in opposite directions. By analysis of the results, SMITH showed that, in the majority of the discharges with a single loop, there must be negative charge moving upwards, i.e. the discharge starts near the lower pole. In a number of cases, double loops occurred with two changes of sign of the potential-gradient change, and these are less easy to interpret; it can be shown that they cannot be explained by an inclined discharge entering and then leaving a "reversal zone". Either the discharge must change direction or else there must be two, nearly simultaneous, discharges, possibly one starting from the upper charge at about the same time as the other starts from the lower charge, and meeting at an intermediate point.

BROOK and KITAGAWA (1960a) suggested, from the model of HOLZER and SAXON (1952) (see § 12.22.), that the total number of return strokes should be approximately equal to the total number of cloud strokes in a storm which lasts long, or for all storms over a long period, and gave some results in agreement with this conclusion.

TAKAGI (1961), and OGAWA and BROOK (1964), in contrast to SMITH (1957), found that most intracloud discharges start with a streamer from the positive upper portion of the cloud, travelling with a velocity of about 10^6 cm/sec and that when this streamer, in one of its branches, reaches an accumulation of negative charge, there is a rapid upward movement of about 1 C of charge at a velocity of about 10^8 cm/sec over a distance of about 1 km. It may well be that SMITH was unable to recognize the positive streamer and that his results correspond to the upward negative streamer found by TAKAGI.

TAKAGI found that sometimes there is also a slow upward negative streamer at about the same time as the slow positive streamer and so a cloud discharge involves the two streamer processes joining.

TAKAGI estimated that the local concentrations of charge are roughly 100 m in diameter, agreeing with the extent of the thunderstorm "sub-cell" or "turret".

OGAWA and BROOK (1964) found that the "K" changes (see § 14.26.) were comparable with return strokes, and might well be thought of as currents which occur when the streamer reaches a region of concentrated charge.

14.31. Cloud and Ground Flashes

The results of PIERCE (1955a), that 39 per cent of all flashes come to ground, have already been given (§ 14.30.). More recent work, quoted by PIERCE, ARNOLD and DENNIS (1962), suggested to them that the higher the latitude the higher the proportion of flashes to ground, and they gave a formula for the fraction of ground flashes as

$$0{\cdot}1 + 0{\cdot}35 \sin \lambda,$$

where λ is the latitude.

SCHONLAND (1928a) obtained results suggesting that, in South Africa, the ground flashes are only 9 per cent of the whole and WANG (1963) at Singapore found 17 per cent.

The general explanation is that the lower charge centre is farther from the ground in tropical storms than in those farther from the equator, probably because the charge centres develop at definite temperatures (see § 12.15.).

PIERCE (1955a) found that, comparing "heat" and "frontal" storms, there appeared definite differences; in heat storms half the flashes reach the ground, while in frontal storms only 31 per cent do so. On the other hand, NORINDER and KNUDSEN (1961) concluded that a greater proportion of flashes come to ground for frontal than for heat storms.

14.32. Minor Discharges

GUNN (1954b) found small discontinuities in the potential-gradient records of close thunderstorms, always of opposite sign to that of the charging process, and frequently recurring quite regularly. He interpreted them as localized discharges within clouds, giving no visible lightning, and with a repetition frequency depending on the regeneration time (see § 12.13.). Similar minor discharges, this time in a cloud which gave no visible lightning flashes, have been found by WHITLOCK and CHALMERS (1956); it may be that these merely represent the effects of distant lightning. GUNN's discontinuities may be related to the K changes (see § 14.26.).

14.33. Runaway Electrons

It has been suggested, originally by WILSON (1923), that electrons with a sufficient energy in the high electric fields of thunderstorms might reach the "runaway" condition in which they would gain more energy from the field than they could lose, and would then continue to travel for quite long distances.

HILL (1963b) used photographic emulsions sensitive to electrons in an attempt to detect these electrons near the foot of lightning flashes, but he was unable to find any effect. In agreement with this conclusion, HILL was also unable to find any radioactivity, which might have been produced by fast electrons, in copper caps struck by lightning.

CLARENCE (1965) found evidence supporting the idea of upward runaway electrons above thunder clouds, and applied these to problems in connection with whistlers (§ 15.11.).

14.34. Potentials and Energies

SCHONLAND (1964, 1953b) estimated the potential differences involved in clouds or between cloud and earth just before a lightning flash as between 10^8 and 10^9 V. The quantity of charge concerned is on the average about 20 C, so that the energy in a flash is of the order of 10^{10} joules.

If we take the result of GOLDE (1945a) that there are about 2 flashes to earth per year per km^2 in England, and if we suppose that all the energy of every lightning flash to earth could be harnessed, then for each km^2 we should have about 2×10^{10} joules per year or a power of about 600 W continuously. So much for any idea of harnessing lightning to give power supplies! And actually a great part of this energy is used up in heat along the path of the lightning flash, so that the harnessing would have to take place at the cloud, rather than at the bottom end of the flash.

HOLZER and SCHILLING (1954) discussed further the possibility of utilizing the electric energy of the atmosphere and concluded that it is, at present at least, quite impracticable.

14.35. Fulgamites

HILL (1963a) placed copper caps over the ends of lightning conductors at the top of a 1000 ft mast, to investigate individual lightning strokes. He found that each lightning stroke produced a "fulgamite", a small pip or mound of copper, melted and raised above the slightly indented surface of the metal. The fulgamites were only a few mm across and this led to the conclusion that the lightning channel is much less broad than had been thought (see § 14.18.). Calculations of the quantities of charge involved gave results from 0·02 C to 15 C, mostly considerably less than those deduced from electrical measurements.

14.36. Warning of Lightning

In many instances, it would be very desirable to have a simple device which would indicate that a lightning flash is likely in a particular region.

One method of doing this would be to use a device which detects or counts a close lightning flash (see § 14.37.) and the suggestions of PIERCE (1956) and KREIG (1960) have been incorporated into a device by MOORE *et al.* (1962), using the electrostatic field change due to the lightning flash, with several levels of sensitivity.

Such a device would, however, not be effective in giving warning of the first lightning flash in a developing storm and so MOORE *et al.* (1962) suggested that the warning should be given when the point-discharge current through a certain exposed point reaches a determined value, and have developed devices of this kind.

14.37. Counting of Lightning Flashes

Attempts have been made to count the number of lightning flashes that are received at a particular station. One of the earliest was that of FORREST (1943), detecting the high-frequency radiation; GANE and SCHONLAND (1948), in the "ceraunometer" preferred to make use of electrostatic field changes. Other devices have been made by ITO, KATO and IWAI (1955), LOCK (1956), SULLIVAN and WELLS (1957), MÜLLER-HILLEBRAND (1959) and HORNER (1960). The problems have been analysed in detail by PIERCE (1956) and BROOK and KITAGAWA (1960b), with the idea of considering whether a counter can be devised which will detect most flashes within a certain distance and few beyond. The conclusion of BROOK and KITAGAWA (1960b) is that such a counter cannot give accurate results, and that it is even difficult to distinguish between intracloud and cloud-ground discharges unless a detailed examination of the field changes is undertaken. At night, a photoelectric photomultiplier counter, with suitable exposure, can be used to detect all flashes within visible range.

MALAN (1962) devised a counter which distinguishes between cloud and ground flashes by comparing the radiation on 5 kc/s and 100 kc/s, the former being much larger in ground flashes.

The results of HOLZER and DEAL (1956) on low-frequency atmospherics suggest that these can be used to count lightning flashes over a large part of the earth (see § 15.8.).

Methods which depend upon optical, rather than electrical signals from the lightning channel have been suggested by BROOK and KITAGAWA (1960), MACKERRAS (1963) and SALANAVE and BROOK (1965). In the method of the last of these, a filter allows into the counter only the $H-\alpha$ radiation, which is particularly prominent in lightning flashes; this method is available for daytime use and can be arranged to count only the cloud-to-ground flashes.

14.38. Discharges from Cloud to Electrosphere

It has been suggested that there might be discharges to the electrosphere from the upper pole of a thunder cloud either simultaneously with, or immediately after the main stroke to the ground from the lower pole. Visual evidence of such effects has been reported by BOYS (1926), MALAN (1937), WOOD (1951), REYNOLDS (1954) and HOFFMAN (1960), and there appeared to be some indication also from the electrical measurements of MALAN and SCHONLAND (1947), though MALAN later ascribed the effects to the slow final process inside the cloud (see § 14.23.).

GISH and WAIT (1950) and STERGIS, REIN and KANGAS (1957b) found no indication of any alteration in electrical conductivity above a thunder cloud, such as would be produced, at least momentarily, by such discharges; however, it must be realized that, at these levels, the relaxation time is short and any increased ionization would be quickly dissipated. The conduction currents above thunder clouds, as obtained by GISH and WAIT (1950) and STERGIS, REIN and KANGAS (1957b), appear adequate to account for the whole of the current through the cloud, so that if there are upward discharges from a thunder cloud, these do not contribute much to the total current.

Isted (1954) found evidence for abnormally strong spasmodic and periodic ionization at levels of about 70 km, while there was stormy weather in the neighbourhood below, sometimes simultaneously with lightning and at other times when no lightning was observed. This suggests that there might be something like point discharge in operation in the electrosphere when the potential gradient above the cloud becomes sufficiently large, and would imply that these currents penetrate well into the electrosphere. At the lower levels, where the conductivity has been measured, such currents could spread out to give no difference from ordinary conduction currents. Similar results have been reported by Rastogi (1962).

Rumi (1957) and Atlas (1958) have found radar reflections which they interpret as originating in the ionization from lightning flashes going up to 20 km, i. e. cloud-electrosphere discharges.

14.39. Discharges to Skyscraper

McEachron (1939, 1941) investigated lightning strokes to the Empire State Building (then 1250 ft, now 1410 ft) in New York, and his results show a distinctive difference from those previously discussed; the discharge is initiated by a stepped leader from the building upwards, not from the cloud downwards. The steps, as observed photographically, are very similar in length and speed to those in the normal downward leader, but after the stepped leader has reached the cloud there is no return main stroke, but instead a continuing flow with continuous luminosity. Strokes following the first are preceded by downward dart leaders.

McCann (1944) found that the critical height of a building for the distinction between the upward and downward leaders is about 600 ft. McEachron found that 84 per cent of the strokes brought down negative charge. From an analysis of the data, Bruce and Golde (1942) found that the ratio of negative to positive discharges is greater for flashes to high buildings than for flashes to open country; in the U.S.A., the ratios are 14 : 1 and 4 : 1 respectively.

The difference between the two types of discharge can be explained on the idea that the discharge commences where the field strength is greatest. In storms in open country the concentration of charge in the cloud, both the main negative and the lower positive charges, and the space charge below make the field strength greatest in or just below the cloud so that this is where the discharge begins, proceeding downwards. But, with a tall building, the concentration of the lines of force on the building gives rise to a very intense field there, and so the discharge is initiated just above the building. The continuing current, instead of a rapid return stroke, is accounted for by the fact that the earth and skyscraper form a good conductor, while the cloud does not. The skyscraper will cause a lightning flash to occur sooner than it would in the absence of the building and sometimes a flash occurs that otherwise would not have done so.

SCHONLAND and MALAN (1954) pointed out that it is rather surprising to find that the lengths of stepped leaders, the delay times and the pilot streamer velocities are almost exactly the same for the upward-moving discharges from the Empire State Building as for the downward-moving discharges in South African storms, especially as it would appear that in one case it is an advance of positive charge and in the other of negative charge. To avoid this unexpected result, SCHONLAND and MALAN suggested that the initial discharge might be, in fact, an upward movement of negative charge to the lower positive charge of the cloud, and this would be followed later by a movement of negative charge downwards from the main negative charge, along the previously ionized path. They pointed out that this would require that the electrical measurements would have missed the first few milliseconds of the discharge, and provide some evidence in favour of their suggestion, which would agree with the recent ideas of the preliminary process (see 14.10.) involving the lower positive charge in the initial stage. MÜLLER-HILLEBRAND (1961), discussing strokes to towers on Monte San Salvatore, Switzerland, showed that the explanation of SCHONLAND and MALAN could not hold in these cases.

PIERCE, ARNOLD and DENNIS (1962) consider that there is first a positive upward-moving leader from the building to the cloud; then there is a negative dart leader, down from the cloud, followed by a return stroke. Since this return stroke is preceded by a dart leader, not a stepped leader, it is similar to the second return stroke in a multiple flash, rather than to the first.

14.40. Discharges to Balloons

DAVIS and STANDRING (1947) made measurements on currents flowing through captive balloons, flying at 600 m in thundery weather. Milliammeter records of point-discharge currents showed continuous currents with kicks due to lightning flashes, some of which were not visible or audible; quantities of charge up to a few coulombs passed through the cable without the balloon being actually struck.

During actual lightning strokes, a drum oscillograph was used, measuring currents of a few thousand amperes over periods of the order of milliseconds; the quantities of charge were of the order of coulombs.

A single-sweep oscillograph gave finer details for 7 strokes, in which the currents were of kiloamp size lasting for 1 msec or less. The quantities of charge ranged from 0·5 to 30 C.

Magnetic indicators (see § 14.43.) were also used to give the currents concerned, and showed two cases of positive currents of over 60,000 A. A majority (about two-thirds) of the strokes brought negative charge down, but there were equal numbers of both signs with currents over 20,000 A and 5 out of 6 over 40,000 A were of positive polarity.

For balloons at 600 m it appeared that the lower limit of currents giving visible and audible lightning strokes occurred at about 500 A. Even more than for the skyscraper, a balloon at 600 m will initiate discharges which otherwise would not have occurred. Since the alti-electrograph has shown the presence of positive charge low down in the cloud, it can be seen that this is most likely to be the charge which discharges to the balloon and hence why positive discharges are prominent.

A continuing current was often observed as in the skyscraper measurements. In a number of cases the oscillograms suggested an upward leader from the balloon similar to that from a skyscraper, but lightning photographs would be needed for definite determination of the leader direction, and these were not taken.

In one case, the cable was vaporized and the necessary current could be calculated; if the average current had been 50,000 A it would have had to last 3·2 μsec, giving a charge of 160 C; or if the average current had been 200,000 A, the time would have been 200 μsec and the charge 40 C.

14.41. The Branching of Lightning Flashes

It is found by photographic and visual observations that most lightning flashes which show branching at all are branched downwards. SIMPSON (1926) found that 242 out of 442 photographs showed branching downwards and only 3 upwards. But MCEACH-

RON (1939) found that strokes to the Empire State Building were branched upwards.

Laboratory experiments on the sparking between two spheres showed the branching to be from the positive sphere and, on the basis of this result, SIMPSON (1926) concluded that the discharge from cloud to earth must involve a positive charge on the cloud base.

When measurements of potential-gradient changes had led to the definite conclusion that lightning brings down negative charges from the cloud, SCHONLAND and ALLIBONE (1931) made investigations of laboratory sparking under conditions closer to the natural conditions of lightning than the sparking between two spheres; they found that a discharge from a negative point to a positive plate could take place with branching towards the plate, if this was provided with small projections to form a source of ions by point discharge.

The branching takes place in the direction of the propagation of the discharge, independently of the polarity, but JENSEN (1933) found that discharges from a positive charge showed finer branching than those from a negative charge.

MALAN (1961) pointed out that apparent upward-branching, or a Y-shaped flash, would occur if a second, independent, stroke were to strike the main channel of an earlier stroke while it was still conducting, and then follow this to the ground. He gave a photograph of an N-shaped flash in which a second stroke struck a branch of the first.

14.42. The Klydonograph

This instrument uses the "Lichtenberg figures" produced by a rounded electrode bearing on a photographic film or plate, backed by a flat electrode; if corona discharge occurs, this affects the photographic plate. It can measure peak voltages between 2 kV and 18 kV, with an extension by the use of a potential divider, and also shows characteristic differences between positive and negative discharges. The klydonograph is easy to install and has been used, in particular, to give results for discharges to overhead power lines (LEWIS and FOUST, 1936).

14.43. The Magnetic Link

FOUST and KUEHNI (1932) introduced a "magnetic link" for the measurement of peak currents from lightning. This consists of a bundle of short steel strips of very high retentivity, placed, un-magnetized, close to a conductor in which it is expected that the current from a lightning discharge will flow. The remanent magnetism gives a good measure of the peak current, both in magnitude and direction. In order to measure currents of widely different magnitudes, two or more magnetic links are placed at different distances from the conductor.

WAGNER and MCCANN (1940) developed the magnetic link to give the rate of rise of current, with an instrument called the Fulchronograph. In this, a slotted aluminium rotating wheel carries a number of magnetic links which pass successively through a gap between narrow coils and become magnetized when a current is passing. Thus the rate of rise of current can be obtained from the remanent magnetism of the different links. The time scale available depends on the speed of rotation of the wheel, and the usual installation consists of two instruments with different speeds.

WAGNER and MCCANN also used a "surge front recorder" in which three resistance and inductance circuits with different time constants are connected across an inductance coupled with the main current. Magnetic links near each circuit give the currents induced, and the average rate of rise of the current can thus be obtained.

In a "magnetic surge integrator", WAGNER and MCCANN (1940) connected an inductance across a resistance in the main circuit, and used a magnetic link to measure the total surge.

WAGNER, MCCANN and BECK (1941) have collected the results from a considerable number of magnetic link observations.

More recently, the bundle of wires has been replaced by sintered links made from magnetic powder as used in coating magnetic recording tapes.

14.44. Use of Cathode-ray Oscillograph

For the measurement of the features of surges such as those in lightning flashes, the most suitable instrument to give full information about the surges is the cathode-ray oscillograph, but the high cost of its installation and operation allows of its use only where there is a good chance of a lightning discharge.

MCEACHRON (1939) has used the cathode-ray oscillograph on the Empire State Building and STEKOLNIKOV and VALEEV (1937) and DAVIS and STANDRING (1947) have used it with balloons.

14.45. General Results for Lightning Currents

The largest peak currents recorded are 220,000 A and 160,000 A reported by WAGNER, McCANN and BECK (1941); average values are around 20,000 A. The time taken to reach the peak value varies from 1 to 19 μsec with an average value of about 6 μsec. The fall to half value after the peak takes from 7 to 115 μsec, with an average value of about 24 μsec.

The total charge neutralized by a lightning flash reaches up to 164 C, measured by McEACHRON (1941) for a discharge from the Empire State Building, in which the greater part of the charge flowed in the continuing current after the first stroke. BRUCE and GOLDE (1942) gave an average value of 30 C, of which one-third neutralizes the space charge so that only 20 C comes to earth.

Further statistical results have been given by HAGENGUTH (1951).

14.46. Lightning Currents in the Ground

When lightning strikes the earth, the current will be distributed in the ground in a way which must depend on the impedances of the alternative paths, and if the ground is inhomogeneous the lightning current may travel more in one direction than another.

The lightning current produces a magnetic field, which may magnetize suitable rocks. GRAHAM, quoted by MALAN (1963), used measurements on the magnetization of rocks to deduce the currents in the earth, finding two components with currents 5 and 50 kA, following different paths.

14.47. Initiation of Lightning Discharge

The measurements of potential gradient and the results relating to the propagation of the discharge show that the lightning discharge must start at some point and then progress; at the starting point, the field strength must exceed the breakdown value.

It has been suggested that the intense field strength needed to start the discharge occurs between the main negative charge and the lower positive charge and then progresses down to the ground, so that the lower positive charge is essential to the initiation of a lightning flash to ground. This is in accord with the results on the preliminary processes (see § 14.10.). WICHMANN (1952b) (see § 12.29.) considered the production of this lower positive charge to be the fundamental difference between a thunderstorm and a shower, but KUETTNER (1950) has found the lower positive charge in showers and MALAN (1952) ascribes it to a secondary process (see § 16.37.).

The lower positive charge cannot be essential for the production of discharges within the cloud, excepting in so far as its occurrence must bring about a larger negative charge than would occur otherwise. Also discharges to a skyscraper (see § 14.39.) do not appear to involve the lower positive charge since the discharge brings down negative charge and starts at the building, not in the cloud; however the suggestion of SCHONLAND and MALAN (1954) (see § 14.39.) does involve the lower positive charge.

The results quoted in § 12.18. show that it is probable that a smaller field strength would be required to start a discharge inside a cloud than outside.

14.48. Artificial Initiation of Lightning Discharges

If a lightning discharge could be caused to flow into a conductor suitably equipped with measuring devices, a great deal more information could be obtained than is available from the usual indirect measurements on lightning discharges to unequipped points. Discharges to high buildings differ from normal discharges (see § 14.39.). An obvious technique would seem to be to use an earthed captive balloon, but this is not very successful because point discharge at points on and connected to the balloon releases so much charge that local conditions are altered and the balloon is protected by the space charge (BROOK *et al.* 1961).

The ideal would be suddenly to produce a conducting path and it was suggested by NEWMAN (1958) and BROOK *et al.* (1961) to fire as a projectile an earthed wire into a storm. BROOK *et al.* made small-scale experiments with a Van de Graaff generator and NEWMAN (1965) reported preliminary results for two successful firings from a sea-going laboratory on a yacht; further results are to be expected.

Accidental initiations of lightning discharges by plumes of water from explosions have been reported by YOUNG (1961) and others.

14.49. Radar Echoes from Lightning Channels

When a radar beam is directed towards a thunderstorm, there are echoes which are caused by the precipitation particles and which persist with slight alteration for some time. In addition, many observers have found echoes of short duration to be attributed to lightning or, more exactly, to the column of air ionized by a lightning flash. LIGDA (1950) was probably the first to make such observations, followed by BROWNE (1951), MARSHALL (1953), MILES (1953), JONES (1954) and others. LIGDA (1956) has given a thorough description of the techniques employed.

This method is much more useful for the detection of long discharges in and between clouds than for the detection of cloud-to-ground discharges which may be obscured by precipitation as far as radar is concerned, but which are detected visually or by electric field-change methods.

Radar echoes from lightning channels are seldom found with 3 cm waves but have been found from 10 cm to 50 cm.

14.50. Point of Fall

As discussed in § 14.15., the place where a lightning flash strikes the ground is determined by which upward streamer joins up with the downward leader, and therefore by the initiation and the rate of development of the upward streamers. It is clear that, other things being equal, the upward streamer is likely to start and develop most rapidly where the field strength is greatest and so a tree or other conductor projecting upwards is likely to give rise to the most developed upward streamer, just because of the concentration of lines of force on the conductor.

Another factor which must be important is the speed with which charge can be supplied to the conductor from the earth and this must depend on the resistance between the conductor and earth; since it is a question of a current in a very short time, it is the high-frequency resistance, involving skin effects, which comes into the matter.

There have been several observations which have appeared to show some connection between the point of fall of lightning and the geological structure of the ground, but other workers have denied this. If the phenomenon exists, it might be explained in terms of the conductivity of the ground. Observations, such as those of DAUZÈRE and BOUGET (1928), have suggested that the point of fall shows a preference for places where the natural ionization of the atmosphere is greater, presumably through abnormally large local radioactivity; these observations, too, have been questioned, but if true might be explained not in terms of the greater existing ionization, but rather, perhaps, that the greater ionizing power would produce more electrons and so give a greater chance of the upward streamer development, or, perhaps, that the conductivity in the ground is increased by the radioactivity.

Another observational result upon which doubt has been cast is that lightning is stated to strike more readily the one of a series of chimneys in which a fire is burning rather than others which are cold. It does not seem possible that the electric conductivity of the emerging plume could produce any effect, but it might be that the fire gives increased conductivity to the chimney as a whole.

There are some statistics showing that certain kinds of tree, e. g. oak, are more readily struck than others, e. g. beech. There may be various causes for this, such as projecting branches in an oak tree, or different resistances in the tree and root system; or it may be that, for example, the smooth bark of the beech allows the lightning flash to go to earth without much visible sign, and so it is not recognized that the tree has been struck.

14.51. Protection from Lightning

A lightning flash must hit the ground somewhere; in order to protect a given object it is necessary to provide that a flash which would have struck the object shall instead strike a lightning conductor which takes the flash harmlessly away to ground.

There have been two views as to the possible mode of action of a lightning conductor, both put forward by FRANKLIN himself. According to one idea, the points produce enough electric charge themselves to neutralize the charge in the cloud base, thus harmlessly discharging the cloud, and avoiding the occurrence of any lightning; the alternative viewpoint is that the lightning conductor attracts towards itself the inevitable flash and provides a safe path to earth.

FRANKLIN himself first favoured the neutralization theory, arguing from laboratory experiments in which a charged conductor can be discharged by the action of points on an earthed conductor nearby. However, this cannot be effective for a thunder cloud, as can be shown from a simple calculation; the quantity of charge involved in a lightning flash is of the order of 20 C, and the point-discharge current from a point beneath a cloud is never more than a few μA. Thus a single point would take some millions of seconds to discharge a thunder cloud, and even if there might be 1000 points below a cloud it would still take a good fraction of an hour and a lightning flash does not wait that long! In spite of such arguments, the neutralization theory of the action of a lightning conductor still persists in some textbooks of elementary physics.

It can be definitely stated that a good lightning conductor will convey harmlessly to ground flashes which would have hit within its "protective radius", that is to say within a cone from the top of the conductor reaching the ground a certain distance away. The protective radius is not a precise term since a conductor never protects absolutely completely all objects within a certain range, and it may give some measure of protection to objects at greater distances. The protective radius can be converted into a protective ratio, the ratio of the radius to the height of the conductor, i. e. the tangent of the angle of the cone from the top of the conductor. GOLDE (1945b) discussed the matter theoretically and from observational statistics and showed that the protective ratio should vary with the quantity of charge concerned in the leader stroke; his conclusion was that the average value is about 2. SCHONLAND (1964) gave an account of the precautions necessary to ensure that a conductor acts efficiently

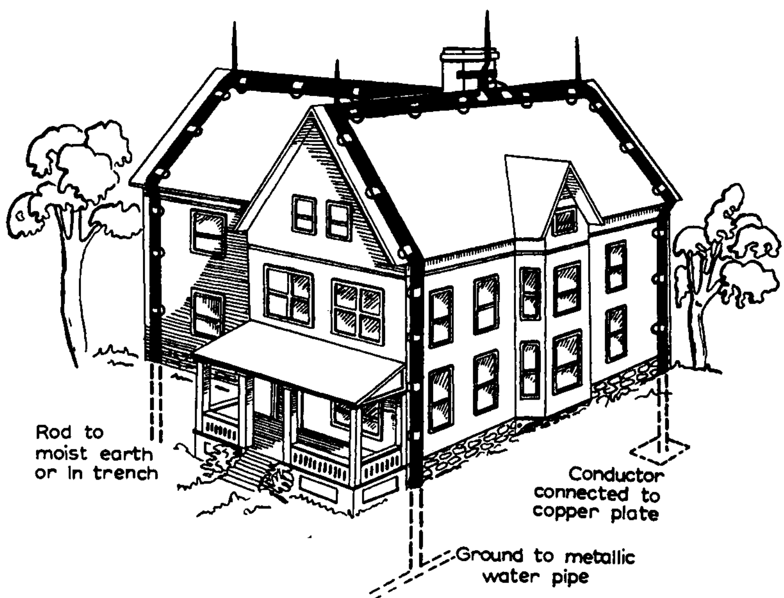


FIG. 70. Protection from lightning. (From SCHONLAND, 1964, Fig. 2, p. 125.)

and considered that a 45° cone, i.e. a protective ratio of 1, gives 99 per cent protection. For greater protection, e. g. for a powder magazine, it is necessary to have a "Faraday cage", i.e. a complete or nearly complete envelope of well-bonded and well-earthed metal.

Ever since the time of FRANKLIN, there have been some who have doubted the value of a lightning conductor, even holding that a pointed conductor might initiate a discharge that would otherwise not occur, and might then not succeed in making the discharge completely harmless, so that the erection of a conductor might actually be dangerous. However, it is now known that it is only a building of considerable height (probably over 100 m) which can provoke a discharge that would not otherwise occur, while for conductors on buildings of ordinary height there is no such effect. If a lightning flash hits a lightning conductor, one can be certain that, in the absence of the conductor, it would have struck close by. The only danger associated with lightning conductors is when they are defective in their connection to the ground; in such a case, a flash, which might have hit the ground harmlessly, may now hit the conductor and cause damage to the building holding it, by reason of a poor earth connection. These and many other points are dealt with in detail by SCHONLAND (1964).

MÜLLER-HILLEBRAND (1962a) has given an historical review of lightning conductors and has also given a detailed discussion of the economics of lightning protection, balancing the cost of installation against the chance of a strike and the value of the contents, and also considering the risks in using wire of smaller cross-section than has been the usual practice. A further discussion has been given by MÜLLER-HILLEBRAND (1965).

14.52. Aircraft and Lightning

The effects of lightning discharges upon aircraft were discussed in detail by NEWMAN (1953) and showed that metal aircraft are well protected from effects of lightning in general, the discharge passing through with little effect. The most vulnerable part of the aircraft is its radio equipment, of which the aerial has to be insulated from the main body of the aircraft; discharges through the aerial are particularly likely when a trailing wire is used, since this often gives a long path for the discharge. As a consequence of a strike to an aerial, there is often damage to the radio equipment itself and there have been cases where there have been secondary effects and it is possible that internal sparking so produced might cause explosions.

14.53. Discharges to Power Lines

Work on this subject has been extensively studied in America and reported on by HARDER and CLAYTON (1953). A large measure of protection is afforded by a shielding earthed wire above the power lines; a shielding of all but 0·1 per cent of the strokes is effected if the power lines are within a wedge of vertical angle 60° having the shielding wire as the apex. For effective shielding there must be a good earth connection to the shielding wire and in practice this means a resistance of less than a few hundred ohms.

A lightning discharge can be considered to supply a definite current, rather than to give the struck wire a definite voltage, and, because of the time taken for the production of ions to carry the current, the lightning discharge behaves as if it possessed a surge impedance of the order of 5000 Ω, so that resistances small compared with this in the earthing have little effect.

The magnetic link type of measuring apparatus (§ 14.43.) is of particular use in power-line work.

It is estimated that where the level of thunderstorm activity ("isoceraunic level") is 30 thunderstorm days per year, there are 10 ground strokes per square mile per year and an unshielded power line would receive one stroke per mile per year.

SCHONLAND (1962) quotes recent results with higher pylons, showing that the frequency of striking is over ten times as great as would be expected; although automatic switching has reduced the actual interruption of supplies, the problem is still of some importance.

14.54. The Spectrum of Lightning

In a recent survey of the subject, SALANAVE (1965) expressed surprise at the comparative paucity of work on lightning spectra. Spectra can be obtained either by the use of an ordinary spectrometre with a slit (WALLACE, 1960, who gives references to earlier work) or by using the flash itself as a slit, which SALANAVE claims to have several advantages.

SALANAVE (1965) described measurements with a moving film attached to the camera of the slitless spectrometer, so as to obtain time resolution.

The spectral lines and bands can be identified in most cases, and prominent features are lines due to N^+ at 5045 Å and N_2 at 4935 Å, the ratio of the intensities of which depends on temperature and electron density. Another very prominent line is the $H-\alpha$ at 6563 Å, which can be used with a filter for lightning photography in daylight and for counting cloud-to-ground flashes (see § 14.37.).

From relative intensities of lines it is possible to obtain an estimate of the temperature in the lightning channel. Results show values around 25,000°K. The electron density is then found to be about 10^{16} or 10^{17} per cm^3.

14.55. Bead Lightning

It is occasionally observed that a lightning flash appears to break up into luminous fragments of length some tens of metres. MALAN (1961) suggested that the phenomenon was to be accounted for in terms of a continuing discharge which would appear much brighter where the channel bends in the direction of the observer who would view it end-on, but UMAN (1962) suggested that it could be better explained as an example of the "pinch effect" observed in laboratory investigations of spark discharge as a periodic change in the radius of the discharging column, with a period of the order of microseconds; he has shown that there is a possibility of the same effect occurring for a lightning discharge.

14.56. Ball Lightning

Although electrical effects from ball lightning do not appear to have been observed, the phenomenon, though rare, seems to be well authenticated and a short account is not out of place. Accounts have been given by, among others, BRAND (1923), GOODLET (1937), ALIVERTI and LOVERA (1950) and HILL (1960). It may be remarked that SCHONLAND (1950) was sceptical about the phenomenon and believed that all the reports could be accounted for in terms of more ordinary processes; on the other hand, the present author has read and heard accounts, of remarkable consistency, from people who are unlikely to have been influenced by previous descriptions.

Ball lightning consists of ball-shaped material, usually 10–20 cm in diameter, but sometimes as small as 1–2 cm and sometimes as large as 150 cm. The ball is usually described as red and luminous, but occasionally as white and sharp; other colours have also been reported. The ball appears only in thunderstorms and is often, but not always, preceded by a lightning flash, the ball appearing sometimes at or near the point of striking; sometimes, however, the ball must be produced higher in the atmosphere and fall towards the earth.

A ball remains luminous for a period from a few seconds to a minute or more, and when it disappears there is a sharp-smelling mist which is blue by reflected light and brown by transmitted light, but may be white in moist air.

Some balls float in the air and avoid conductors but are attracted to closed spaces and sometimes enter houses by chimneys, windows or even cracks. They avoid human beings as they do other conductors and are much more frightening than dangerous.

Other balls remain attached to conductors and may roll along gutters, etc.; if one touches a human being, burns may be caused. Floating and attached balls sometimes change from one type into the other.

The balls sometimes disappear with a crack or even with an explosion, giving out streamers, and sometimes silently. Occasionally they are extinguished by another lightning flash and sometimes burst into two.

If a ball falls into a water tank, the heating is considerable and on one occasion it was possible to estimate the energy as from 4×10^6 to 10^7 J (GOODLET, 1937).

Further properties and examples of ball lightning have been given by SILBERG (1965).

14.57. Theories of Ball Lightning

Since ball lightning is started by a lightning flash, it is almost inevitable to look for the energy in that of lightning. But, since the ball lasts for periods of at least seconds, there must be some source of energy to maintain it for this period of time.

Some theories have been put forward which involve considering the ball as a "plasma" of ions and electrons, but any such plasma unless maintained in some way would disappear in a time of the order of milliseconds rather than seconds; if the ball has the very high temperature which could maintain a plasma condition, it would surely expand rapidly and not maintain its size and shape.

Other theories have suggested that the lightning flash starts off some chemical reaction which provides enough energy to keep itself going; suggestions have been made involving ozone, oxides of nitrogen or some kind of organic or metallic dust. NAUER (1953) has made a detailed study of this type of theory.

KAPITSA (1955) suggested the existence of intense radio-frequency electromagnetic waves which would give a system in resonance equilibrium if the wavelength is 3·65 times the diameter of the ball; this would require a frequency of about 300 Mc/s. PIERCE, NADILE and MCKINNON (1960) pointed out two difficulties in the theory; first, any source of the energy would have to be in the cloud, which is a long way from the situation of the ball near the ground; and, second, it would be necessary to have a definite frequency for the waves. They put forward the suggestion that a source near the ball and a definite frequency could be found in the pulsed nature of point discharge from points in connection with the ground in the positive potential gradient immediately after a flash.

SILBERG (1965) has given a more detailed discussion of the standing-wave theory. ANDERSEN (1965) suggested a source of energy in the collision of charged drops.

ARABADGY (1956) made the suggestion that ball lightning consists of atoms of xenon which have been disintegrated by high-energy particles in a lightning flash and which give out nuclear energy sufficient for a self-sustained nuclear reaction. There have been other, less specific, suggestions of nuclear energy as the source of the energy of ball lightning, but none can overcome several difficulties such as that of the constant size.

HILL (1960) considered that a fairly satisfactory theory can be built up in terms of the production of molecular ions, not electrons, by the lightning flash. These ions are distributed inhomogeneously and may become attached to dust particles. The luminosity would then be due to internal corona discharge between regions of different space charges.

SCHONLAND (1962) has been converted from his earlier scepticism and thinks the phenomena can be explained in terms of a slowly burning bubble of combustible material either arising from or catalysed by rock dust which has been affected by lightning currents flowing along wet fissures in rocks in preference to flowing in the more resistive rocks. This would explain why ball lightning is quite frequently reported from the Alps and never from South Africa, where the condition of the ground is very different.

MALAN (1962) thinks that some of the observations may be accounted for by persistence of vision after a very bright lightning flash.

CHAPTER 15

Atmospherics

15.1. Atmospherics

The term "atmospherics", sometimes shortened to "sferics", originally comprised all forms of radio interference in the form of the picking-up of signals of accidental origin. The term, as used here, is confined to signals which originate from lightning flashes, thus excluding such effects as radio waves from the sun, stars, etc.

Atmospherics have two features of importance, firstly in regard to the information they can give about the discharge at which they originate, and secondly in regard to information about propagation of waves in the atmosphere, including reflections at the ionosphere. In as far as these two can be considered separately, we shall be concerned here primarily with the first feature and shall mention the second only where it interacts with the first.

The first reception of atmospherics and the recognition that they are caused by lightning was due to POPOV (1896). Other early work was that of TOMMASINA (1900) and TURPAIN (1905); the first simultaneous measurement of atmospherics at two stations was probably that of ECCLES and AIREY (1911), and ECCLES (1912) first recognized the need for a reflecting region in the upper atmosphere to account for results.

A more detailed study of atmospherics was started by APPLETON *et al.* (1926), and the reflection of these from the ionosphere has been studied by SCHONLAND *et al.* (1940) and others.

Recent work on the relation of atmospherics to the lightning discharges causing them includes a thorough study by ISHIKAWA (1961).

15.2. Nature of Atmospherics

Since atmospherics are detected by radio receivers which are capable of receiving electromagnetic waves of some definite frequency, it is clear that they must comprise the radiation, as distinct from electrostatic and induction, components of the potential-gradient change (see § 14.7.).

Electromagnetic radiation is produced by the acceleration of electric particles and so the radiation from a lightning flash must give information about the accelerations of the particles concerned. As these accelerations show no definite periods, the radiation is in the form of a pulse, which, when split up into Fourier components, shows a very wide range of frequencies. Thus, when a tuned receiver is used, atmospherics will be observed no matter what frequency is used, although the relative intensities of the effects at different frequencies depends on the nature of the pulse and on the propagation of the different frequencies from the source to the receiver.

15.3. Classification of Atmospherics

CATON and PIERCE (1952) have divided atmospherics into 4 main classes: (a) an irregular high-frequency type, (b) a regular peaked type, (c) a regular smooth type, and (d) a long oscillatory train type. Type (a) is considered to originate in the stepped leader process while the other types come from the return stroke, the differences being caused by different reflection processes at the ionosphere.

LABY *et al.* (1940) accounted for some features of atmospherics in terms of reflections at the ionosphere, and HALES (1948) and BUDDEN (1951) considered the ground and ionosphere to form a wave guide. BUDDEN (1951) pointed out that the reflection and wave guide methods of approach are essentially the same, but one or other may be more suitable in a particular case.

15.4. Direction of Atmospherics

The first investigation of the direction of the origin of atmospherics appears to have been that of AUSTIN (1921), using a frame type of aerial. This type of measurement was continued with the development of "goniographs" with two vertical frame aerials at right-angles, for example in the instruments of BUREAU (1931) and LUGEON (1928).

The earlier work was concerned with the determination of the direction of origin of the most prominent groups of atmospherics and confirmed that they originated in thundery regions. With the development of the cathode-ray oscillograph, it was possible to determine the direction of each individual disturbance (WATSON WATT and HERD, 1926). With two or three such receivers, the source of the disturbance can be located accurately and this has been developed into a complex system, the "Sferics" organization, for the warning of aircraft (OCKENDEN, 1947).

The apparatus used and details of the work have been summarized by NORINDER (1953).

15.5. Location by Single-station Observations

WORMELL (1953b) discussed the possibility of locating a discharge from the observation of the wave-form of atmospherics as well as the use of loop aerials; his conclusion is that, for storms to the S.E. of the British Isles, they can be located at night to a greater degree of accuracy and at a greater range than is possible with the normal Sferics organization. However, for reasons which are not yet understood, atmospherics from storms to the S.W. are not of the same form and the distance cannot be determined reliably when it is greater than 1500 km. During daylight, because of the different effect of the ionosphere on reflection, the limit of distance in any direction is about 1000 km.

For storms within 100 miles, STERGIS and DOYLE (1958) found that great accuracy is attainable using frequencies from 0·5 to 1·5 Mc/s, the direction being accurate to about 2°.

15.6. Sudden Enhancement of Atmospherics

There occur occasional abrupt increases in the intensity of atmospherics, the intensity remaining high for a period which is sometimes only a few minutes and sometimes up to an hour or more, after which the intensity reverts to normal. These increases occur only during day-time and are coincident with solar flares.

The absence of much delay between the appearance of the flare and the enhancement of atmospherics shows that the cause must lie in radiation, rather than particle, emission from the flare. It is believed that the cause of the enhancement is the increased ionization, and consequent increased reflecting power for radio waves, of the D region of the atmosphere (60–90 km up) by X-rays of wavelengths less than 10 Å emitted in the flare.

The effect is most noticeable for frequencies around 27 kc/s which are normally not reflected at the D region but are attenuated by passage through it.

15.7. Relation between Radiation Frequency and Discharge Processes

The relative intensities of the radiation received on different frequencies from the same discharge depend partly on the nature of the discharge concerned and partly on the propagation of the disturbance from the source to the receiver. In order to eliminate the latter factor, MALAN (1958) made observations at distances of less than 50 km, so that there are no effects of ionospheric reflection; this also allowed simultaneous measurement of the electrostatic field change, which helped to identify the particular discharge process concerned. MALAN made measurements at various frequencies from 3 kc/s to 12 Mc/s. At higher frequencies, over 1 Mc/s, the radiation is fairly continuous during the discharge process, but shows a maximum for the preliminary stages of cloud–earth discharges; at lower frequencies, the radiation becomes confined to the return strokes of cloud–earth discharges with a smaller effect of the rapid components of cloud–cloud discharges, though this becomes insignificant when the freqency is as low as 3 kc/s. MALAN showed that these results are in agreement with what is expected from theoretical considerations of the discharge process.

JONES (1958) has identified different types of discharge from their atmospherics on 150 kc/s and 10 kc/s; he agreed with MALAN that it is only cloud–earth discharges which give much effect on 10 kc/s. JONES identified a high rate of reception of "directional pips" on 150 kc/s and no radiation on 10 kc/s as characteristic of a tornado and quoted observations as showing that there are inner cloud discharges in tornadoes, and few strokes to ground.

PIERCE, ARNOLD and DENNIS (1962) discussed in detail the "spectrum" emitted by the various components of a lightning flash.

15.8. Very Low Frequency Atmospherics

HOLZER and DEAL (1956) found that the mean signal amplitude of atmospherics at a low frequency, from 30 to 120 c/s, showed, when there were no local thunderstorms, a simple diurnal oscillation very similar in relative amplitude and phase to that of the atmospheric air–earth conduction current over the oceans and therefore, presumably, to that of the potential of the electrosphere. This correlation disappeared when higher frequencies were used.

HOLZER (1958) showed that the result could be accounted for by assuming that these low-frequency atmospherics are radiation from thunderstorms which can reach the observer from a large fraction of the earth, the ionosphere and the ground forming a "wave-guide" which gives only small attenuation.

LARGE and WORMELL (1958) measured disturbances at a range of frequencies between 5 and 320 c/s and found that down to 10–20 c/s the disturbances could be correlated with thunderstorms, but that at lower frequencies these effects are masked by those of local space charges such as were found by WHITLOCK and CHALMERS (1956) (see § 6.13.).

A thorough study of very low frequency atmospherics was made by PIERCE, ARNOLD and DENNIS (1962), dealing with both source and propagation problems. They concluded that there are three principal kinds of source, these being, in descending order of magnitude, (1) return strokes of discharges to ground, (2) leader strokes of flashes to ground and perhaps intracloud flashes, and (3) K changes (see § 14.26.) during both cloud and ground flashes. The wave-guide theory must be modified to include two modes; the effects of changes of ground conductivity and of the earth's magnetic field can be included.

15.9. Very High Frequency Atmospherics

BROOK and KITAGAWA (1964) measured atmospherics at frequencies of 420 Mc/s and 825 Mc/s and found that these are associated with the development of streamers in breakdown processes both in leader strokes and inside the cloud. No effect is found ascribable to the main return stroke itself and a dart leader gives no effect during the last 100 μsec of its movement.

15.10. Radio Emission from Clouds

There have been several reports of radio emission from clouds at times when there are no lightning flashes. For example, GIBSON (1957) found radiation on a wavelength of 0·86 cm (3·5 $\times$ 10^4 Mc/s) from clouds without lightning. SARTOR (1963) considered the phenomenon and suggested that the origin of the radiation might be the collision of charged water drops; he carried out laboratory experiments to confirm this.

15.11. "Whistlers"

In addition to the normal atmospherics which give sharp clicks in a receiver, there have been observed "whistlers" consisting of a signal of decreasing musical frequency and sometimes repeated a number of times at intervals of the order of seconds. These signals can sometimes be received on a long wire even without amplification.

STOREY (1953) interpreted them as signals which leave the earth along magnetic lines of force, return to the opposite hemisphere along the same line, then are reflected and retrace their paths a number of times, giving the successive repetitions. In order for this to be possible there would need to be an appreciable density of electrons at distances of several times the earth's radius, and this would give a dispersion of velocity for these signals, so as to produce the decrease in frequency. At the time of STOREY's theory it was not believed that the electron density would have the requisite value, but modern ideas render this more probable. CURTIS (1958) has attempted to distinguish between those lightning discharges which produce whistlers and those which do not.

HOFFMAN (1960) has suggested that whistlers, particularly those with "long trains", are generated by synchronous interaction between an energetic upward discharge from the top of a thunder cloud to the ionosphere (see § 14.38.) and the extraordinary component of an atmospheric that is delayed by a time of the order of 1 sec. This theory, he claims, can explain many features of whistler phenomena, e.g. the correlation between whistlers and multiple discharges.

CLARENCE (1965), in a survey of whistlers, gave the following conclusions: in the majority of cases, the source of a whistler is a lightning discharge between cloud and ground; multiple whistlers correspond to separate strokes in the flash; successive components of multiple whistlers show increasing dispersion. He considered that the last fact might be accounted for in terms of "runaway" electrons (see § 14.33.) moving upwards from the cloud.

CHAPTER 16

The Separation of Charge

16.1. Cloud Charges

The results discussed in earlier chapters have established that there are various concentrations of charge in different parts of different types of cloud; to establish and maintain these charges there must be processes of charge separation and it is the purpose of this chapter to discuss the theories that have been put forward to account for the phenomena.

While it is clear that one single process is unlikely to be found suitable to account for all the facts, it is equally unlikely that a number of different processes should be of approximately equal importance; thus we would hope to be able to discover a few distinct processes, among all those suggested, and to account for the facts in terms of these; from the point of view of scientific economy, the fewer such processes the better. It will be convenient to consider the separate results roughly in order of magnitude.

16.2. The Main Thunder-cloud Charges

Investigations of thunder clouds, as described earlier, have established that the main distribution of charge in a thunder cloud consists of an upper positive charge centred round a region where the temperature is about $-20°C$ and a lower negative charge centred at a temperature somewhat below zero. The quantities of free charge involved range between 10 and 100 C.

The external dissipating current from a thunder cloud is of the order of 1 A (0·5 A, GISH and WAIT 1950; 1·3 A, STERGIS, REIN and KANGAS, 1957b), but the actual charging current is several times this value, a large proportion of this being dissipated internally mainly by conduction. The values quoted are for a whole storm which normally consists of several cells in various stages of development.

16.3. Subsidiary Thunder-cloud Charges

In addition to the main charges in thunder clouds, there are certainly other charges present, of which the most important is the positive charge found in the base of some, perhaps all, thunder clouds at temperatures above the freezing point of water.

Measurements of potential gradients at the earth's surface during storms, and the alti-electrograph measurements themselves show that a simple picture of 2 or 3 charges is not sufficient to fit the detailed results. It is, however, difficult to discuss these measurements since there is, at the same time, relative motion of the cloud and the point of observation, and also motion of charges in the clouds.

16.4. Charges in Non-stormy Clouds

The potential gradients in non-stormy clouds are seldom sufficient to show effects with the alti-electrograph or similar methods, and so, in spite of the fact that conditions are much more steady, less is known about the charges in non-stormy clouds than about those in thunder clouds.

The general negative potential gradient below a cloud giving continuous rain shows that there is a negative charge in the base of the cloud, but it is only by indirect methods (e.g. that of § 10.14.) that it is yet possible to reach any conclusion about charges in the upper part of the cloud. If it should turn out that the electrical structure of the continuous rain cloud (nimbo-stratus) is similar to that of the main charges of the thunder cloud (cumulo-nimbus), then the same process of charge separation might be operating in both cases, though the rate of separation is much less in the nimbostratus. But if the electrical structure is different, then the process at work in the thunder cloud must be reduced to very small proportions in the continuous rain cloud; and there must exist in the latter some other process which remains insignificant in the thunder cloud. Thus the problem of the electrical nature of the continuous rain cloud is not only of interest in itself, but may throw light on the problem of the separation process in thunder clouds.

In non-raining clouds there is generally found to be little appreciable charge; but LECOLAZET (1946), on a few occasions, found effects to be attributed to small charges in some fine-weather cumulus clouds and GUNN (1952) has found charges on the droplets in non-raining cumulus (see § 13.7.). WHITLOCK and CHALMERS (1956) found effects of small charges in stratus and strato-cumulus clouds (see §§ 5.62., 13.7.).

16.5. Mechanisms of Separation

The separation of charge in clouds must take place in two stages. In the first place, charges of opposite signs must become attached to particles which will move relatively to one another; these charges may be derived from natural ions, perhaps drawn into the cloud from outside, or there may be some process by which there is a creation of equal and opposite charges on different particles.

Then there must be a separation of these charges in space, charges of one sign moving in one direction and those of opposite sign relatively in the other. It has usually been thought that this occurs by the agency of gravitation, larger particles moving more rapidly downwards against a rising air stream than smaller particles; if particles of different sign are produced at the same place, it is difficult, if not impossible, to suggest any alternative mechanism by which they can be separated in space. But if it is the ions which provide the charges, then these may become attached to larger particles at different places according to sign and so the separation in space might be accomplished by relative vertical motion of different parts of the cloud.

16.6. Relation between Precipitation and Electrification

There is a very strong connection between precipitation and electrical activity in clouds, for non-raining clouds show small, if any, electrical effects, and when a cloud is in the process of development, electrification and precipitation appear at about the same time (see § 12.24.).

This has suggested that electrification is caused by precipitation, but that is not the only logical conclusion; it may be that, as MOORE, VONNEGUT and BOTKA (1958) suggested, the precipitation is assisted by the electrification (see § 12.24.); or it may be that both precipitation and electrification are independent consequences of some other physical process in the cloud.

If electrification is caused by precipitation, that cannot be the only factor involved, since the extent of charge separation by no means corresponds to the amount of precipitation; a comparison of electrical effects in cumulo-nimbus and nimbo-stratus clouds shows that the relative effects are much greater than the ratio of the rates of rainfall.

Most of the workers in recent years have taken it for granted that precipitation is the main agent by which charges of different sign are separated in space by gravitational action. On this basis, most of the theories have attempted to make use of the precipitation in the initial process of charge separation.

An important argument in connection with the association of precipitation and electrification is that emphasized by MOORE, VONNEGUT and EMSLIE (1959) that the electric charge actually brought down to the ground by rain is only a very small fraction of the total charge within the cloud and this is still the case when the observing station is actually within the thunder cloud.

16.7. Connection with Ice

If the conclusion is accepted that the main separation of charge is connected with the fall of precipitation through the cloud, then there is evidence that, in the great majority of cases at least, the separation occurs where solid precipitation falls through a cloud of supercooled water drops.

The measurements of the alti-electrograph (§ 12.15.) and the Zugspitze measurements (§ 12.16.) show that the main separation of charge occurs at a level which is determined, not by the actual height, nor by the height above the cloud base, but by the temperature, the maximum separation occurring at about $-12°C$; at these levels, the precipitation occurs in the solid form and the cloud consists mainly, but not entirely, of supercooled water drops.

However, as discussed in § 12.26., there are observations which appear to show lightning from clouds which cannot contain ice. If it is substantiated that these clouds are electrically similar to thunder clouds of the usual type, then ice cannot be a necessity for all thunder clouds.

KUETTNER (1950) stated that it appeared to be necessary for a thunderstorm that the cloud should extend to below the freezing-point level, so that precipitation could melt. However, there appear to be observations where this is not so. A thunderstorm does not develop when all the cloud particles are frozen and none remain supercooled.

It may well be that precipitation in thunder clouds is sometimes, or even often, initiated by the coalescence mechanism, but the growing precipitation particles are carried up to such a height that they freeze, so that when they fall they are in the solid rather than in the liquid state. Thus any conclusion that electrification may be related to the presence of solid precipitation does not imply that the initial mechanism of the formation of precipitation need involve ice.

The results for continuous rain and snow clouds (§ 10.25.) appear to require two distinct processes of charge separation. SMITH (1951 b) suggested that a few examples found by SIMPSON (1949) of fairly heavy rain with small electrical activity might be cases of precipitation which had originated by coalescence and had never been in the solid state; meteorological observations appeared to support SMITH's suggestion and SIVARAMAKRISHNAN (1960) has found similar cases. If this suggestion is confirmed, it would seem that both separation processes in the usual continuous rain involve the presence of ice, as if no ice is present there is little charging even by a lower process.

16.8. Requirements of Theory Involving Precipitation

If it is accepted that precipitation plays a necessary part in the process of charge separation for the main charges in the thunder cloud, any theory to account for this must satisfy certain conditions.

1. The process must give a positive upper charge and negative lower charge.
2. The process must give a rate of separation of charge of up to several amperes.
3. The process must operate at temperatures below the freezing point.
4. The process must be connected with precipitation in the solid form.
5. If the process operates in nimbo-stratus clouds, it must do so much less effectively than in cumulo-nimbus clouds.

From these requirements and from the need for any process suggested to fit in with existing knowledge of cloud phenomena, including the initiation of precipitation, certain general results can be deduced.

In the gravitational separation in space, for the heavier particles, which must carry down a negative charge, requirement (4) suggests that these must be solid precipitation particles. The lighter particles, carrying a positive charge, must move upwards in the rising air of the storm and might be ions in the air, or cloud droplets, or fragments of water or ice smaller than precipitation particles.

In order to account for (5) above it is necessary to postulate some property of the cloud, important in charge separation, that differs between cumulo-nimbus and nimbo-stratus clouds. The difference in vertical air currents cannot affect the extent of charge separation by gravitation, since that merely depends on a difference in velocity relative to the air. An important difference between cumulo-nimbus and nimbo-stratus clouds is the amount of turbulence in the clouds, and if it is this which determines the difference in the electrical behaviour of the two types of cloud, then the initial process of charge separation would need to be one in which turbulence plays a part.

16.9. Masking of Charges

In most discussions of the separation of charge, it has been considered that the main process of separation involves negatively charged precipitation, with positive charge remaining in the air. KUETTNER (1956) suggested that this is not, in fact, the case. KUETTNER's argument is that, if charge separation is going on continuously in air with no vertical motion, then the precipitation charge is continually being removed and the charge remaining is of opposite sign to the precipitation, and this will soon become larger than that on the precipitation. On the other hand, where there is an up-draught sufficient to carry up the precipitation particles. the charges on these will continue to grow and will not be masked by charges of opposite sign. On this basis, KUETTNER suggested that the main separation of charge in a thunder cloud must be of positive precipitation and negative smaller particles.

KUETTNER himself drew attention to a major difficulty in this theory, namely, that it would involve the thunder cloud giving a separation of charge in the wrong direction to account for the maintenance of the earth's charge or for the results of GISH and WAIT (1950) on the direction of the current above the cloud.

Another difficulty in this picture is the question of where would be found the negative charges that are moving up more rapidly in the up-draught than the precipitation particles.

On the other hand, this theory does give a neat explanation of the existence of the lower positive charge; if the separation of charge takes place only above a certain level, say the freezing-point level, then below this level the only charge present will be that on the precipitation.

16.10. Quantities of Charge

Assuming, as in the preceding section, that the separation in space is by the agency of gravitation, it can be seen that, depending on the actual particles concerned, the relative velocity of separation can be estimated.

After a lightning flash, the recovery of potential gradient must be initially at the rate at which the gravitational separation is operating. Thus, if the recovery rate gives the current of separation, presumably under conditions when there is no internal dissipating current, and the velocity of separation is known, it is possible to calculate the charges concerned. These charges, of opposite signs, exist in the same region of the cloud and become apparent as free charges only when they have been separated gravitationally.

GISH (1951) pointed out that the rate of recovery after lightning flashes amounted to currents of several amperes, and that the gravitational velocity of separation, even of precipitation particles and atmospheric ions, could not be more than a few m/s, so that the charge densities must be quite large. GISH made calculations to show that these charges might be close to the limit that could be carried by water and ice particles without breakdown.

GUNN (1954b) found that his own measurements of drop charges in clouds (GUNN, 1950) could give just sufficient charge density to meet the requirements.

WORMELL (1953a) made more precise calculations on these lines; if the electric moment of the lightning discharge, on the average 110 C.km, is to be regenerated at a rate $1/7\ \text{sec}^{-1}$ by means of a relative velocity of 11 m/sec, then the total amount of charge of either sign in the cloud must be about 1400 C, though only about 20 or 30 C is free charge, neutralized by a lightning flash, or available to give an external potential gradient. If the relative velocity is less than the rather high value of 11 m/s, then the amount of charge required is correspondingly greater. Similar calculations have been made by MASON (1953a).

These results have some important consequences in connection with theories of initial charge separation. Assuming that the separation in space is by gravitation, if the relative velocity is some m/s, the charges are approaching the limit which could be carried by the particles concerned. Any theory which involves a velocity much less than this would have to be ruled out completely, so it would be quite impossible that the charges of opposite sign could be carried on, for example, cloud droplets and atmospheric ions. Also the figures show that, to provide the total charge required, this must exist and therefore must be produced, over a quite wide range of heights, certainly over one or more km, thus ruling out any separation process confined to one temperature and so to one level.

If, on the other hand, the separation in space is other than gravitational, there might be larger relative velocities and so there might be smaller quantities of charge required. However, as will be discussed in § 16.31., the only type of theory which does not require gravitation as the agency of separation in space has some grave defects as an acceptable theory.

In an average thunderstorm, LATHAM and MASON (1961b) suggested that the generation of charge must take place at a rate of about 1 C per km^3 per min; this, it should be emphasized, is the rate of actual production of charges, or of use of charges already in the atmosphere.

16.11. Charges in Violent Storms

In normal thunderstorms, conditions are such that during the interval between one lightning flash and the next there is time for the potential gradient to be approaching to a steady state. However, there are some very violent storms, in which lightning flashes are so frequent that there is no question of a steady state. In a normal thunderstorm, if there were no lightning, the equilibrium condition would be that in which the potential gradients would be not very much greater than those which give breakdown, but in violent storms the equilibrium condition would involve much higher potential gradients.

From this argument and from the actual amounts of electricity involved in the discharges, it follows that in violent storms there must be a rate of charge separation that is much greater than in the normal storm. This must apply to both parts of the separation process, that of the original production or attachment of charges of different signs on particles of different motion, and also that of the separation in space of the differently charged particles.

Considering first the latter process only, then if the agency is gravitational, the relative velocity of motion of the particles still cannot be more than a few m/sec and so the much greater current of separation must involve much greater charge densities. If it were possible to calculate exactly what must be the current of separation in such cases and to estimate the number and sizes of drops and/or ice particles in the cloud, it would be possible to see whether these particles could, in fact, carry the charges concerned. If it should turn out that the potential gradients at the surface of the particles are larger than breakdown values, it must be concluded that, in these violent storms, separation in space cannot be by the agency of gravitation but must involve some process that can give larger relative velocities. There are too few large hailstones, which do fall faster, for these to provide the current.

Not only must the process of separation in space proceed sufficiently fast to provide the currents observed in violent storms, but it is also necessary for the primary process of charge production to work very rapidly to provide the necessary charges; if this process is one which depends on production of equal and opposite charges from neutral bodies, there may be no difficulty, but if it is one which depends upon ions in the atmosphere, it would be necessary to postulate some production of ions.

16.12. Classification of Theories Involving Gravitation

Assuming that separation in space is brought about by gravitation, the different theories of the initial separation of charge on to particles of different sizes can be classified simply according to the origin of the charges.

In one class of the theories, the process involved is concerned with the attachment, to particles of different sizes, of the natural ions that exist in the region, these having been originally produced by cosmic rays, radioactivity, etc. The theory concerned must give an account of how ions of one sign become attached preferentially to the heavier particles, while ions of opposite sign become attached to lighter particles or remain as ions in the air.

In the other class are theories according to which some process produces positive and negative charges from bodies previously neutral, the different signs being associated with particles of different size.

The rate at which charge can be separated according to a theory of the first class is limited by the supply of ions, while there is no such limit in the case of theories of the second class.

Of the theories to be considered here, those of GERDIEN (1905a), ELSTER and GEITEL (1885, 1913), WILSON (1929 and 1956), WALL (1948), GUNN (1935, 1954c) and FRENKEL (1944, 1946, 1947) fall into the first class and all the others into the second.

16.13. Gerdien's Theory

THOMSON (1898) suggested that if water vapour were found to condense more readily on ions of one sign than on ions of the other, this might form a means of obtaining a separation of charge in clouds; WILSON (1899) found that this difference did, in fact, occur for condensation on ions in super-saturated air from which nuclei had been removed, condensation occurring more readily on the negative ions. GERDIEN (1905a) based a theory of cloud charges on this, but it has since been realized that the conditions necessary for such a process, namely, fourfold super-saturation, are never to be found in the atmosphere because nuclei are always present, and GERDIEN's theory, which was never supported by WILSON himself, is no longer seriously considered.

16.14. Influence Theories

ELSTER and GEITEL (1885) put forward the idea that raindrops polarized in an existing electric field would charge by contact the cloud particles with which they came into contact. In the vertical air currents that occur during rain there would then arise a separation of charge between the larger raindrops and the smaller cloud particles. At first, ELSTER and GEITEL suggested that the cloud particles would take the charge of the top of the drop and, if this charge is that produced by polarization of the drop in the existing field, then the effect would be to reduce, rather than to increase, the existing potential gradient. Later, however, ELSTER and GEITEL (1913) considered the contacts to occur on the lower half of the drop, which would provide a building up of the potential gradient.

The difficulty with this theory is the doubt whether there can be electrical contact without actual coalescence of the drops. GSCHWEND

(1922) found evidence in favour of this but SCHUMANN's (1925) results were unfavourable and GOTT (1935) found no charging under conditions similar to those envisaged in the theory. More recently, SARTOR (1954) investigated coalescence phenomena for water drops in mineral oil with the same REYNOLDS number as for cloud droplets in air, thereby being able to use much larger drops; he found effects, on applying a vertical field, which suggested electrical contact without coalescence at the bottom of a larger drop in some cases and suggested a theory of charge separation similar to that of ELSTER and GEITEL. Although the same REYNOLDS number ensures similar hydrostatic effects, the surface phenomena may not be the same and the occurrence of the effect for large water drops in mineral oil does not necessarily mean that it will occur for small water drops in air.

SARTOR (1961) has been able to show from hydrodynamic and electrostatic calculations that a small drop, moving relatively to a large drop, comes closest to contact near the bottom of the large drop. He then assumed that there can be charge transfer without coalescence and went on to show that the theory would give a quantitative increase of charge separation in approximate agreement with thunderstorm phenomena. SARTOR's calculations were concerned with water drops not ice particles, but he considered that the process might work more efficiently with ice than with water; his theory would thus be able to account for warm thunderstorms (see § 12.26.), but would predict larger effects when ice is present.

SARTOR (1965) calculated that induction charging of colliding particles in a polarizing electric field should give rates of charge separation at least comparable with those given by other mechanisms.

16.15. Wilson's Theory

WILSON (1929) suggested that the separation of charge in thunder clouds might be due to ion capture by water drops, as discussed above. In the normal positive potential gradient, falling water drops can acquire a negative charge, if the positive ions move more slowly then the drops. Under suitable conditions, this process can give rise to an increase in the potential gradient and so to the observed polarity of clouds.

It has been objected that WILSON's process gives the observed polarity only as a consequence of the normal positive potential gradient, but that this potential gradient is maintained only by reason of the effects of thunderstorms. The argument has been likened to lifting oneself by the bootstraps! WILSON has answered this by pointing out that if clouds of either polarity were initially equally probable, the greater mobility of electrons over positive ions at electrospheric levels would result in the transfer of electrons more readily to positive cloud tops than of positive ions to negative cloud tops, thus initiating a positive potential for the electrosphere and so a positive fine-weather potential gradient, from which would follow the positive polarity of clouds.

Perhaps a more important objection to WILSON's theory lies in the very strict limitations to the conditions under which the process could produce a separation of charge. In the first place, the positive ions must be "slow", not "fast", according to the definition of § 3.11. But when the separation of charge has proceeded to the extent to which it does in thunder clouds, the potential gradient inside the cloud would be sufficient to make small ions fast and so the positive ions would destroy any negative charges on the drops and hinder the increased separation of charge; thus the process could continue to operate, as the separation builds up, only if small ions are absent. There is, at present, no evidence as to the sizes of ions within thunder clouds, though it is known that small ions do exist in cumulus clouds (see § 7.20.).

WILSON's process consists essentially in water drops bringing down again some of the negative charge which is moving upwards by conduction; it does not involve to the same extent the positive charge moving downwards by conduction, and in steady conditions in still air it follows that the resultant current is still bringing some positive charge downwards and some negative upwards and so is reducing the existing separation of charge.

However, if there is a vertical air current which is smaller than the vertical fall velocity of the drops but greater than the vertical conduction velocity of the positive ions, a separation of charge can take place; this rather strict limitation on the conditions to those with a vertical air velocity between certain values might be stricter than the meteorological conditions of thunderstorms could allow, but there is, as yet, insufficient knowledge to decide the matter.

WILSON's process makes use of the ions present in the atmosphere, and the question as to whether these are sufficient will be discussed later (§ 16.30.).

Wilson's theory, in its original form, was concerned with water drops and, when it was found that the main separation of charge occurs at temperatures below the freezing point, so that the precipitation particles are probably solid, it appeared that Wilson's theory could not be applicable. However, Chalmers (1947) showed that ice crystals can participate in Wilson's process in very much the same way as water drops, so this objection no longer holds.

In spite of the criticisms that can be levelled against Wilson's theory, it cannot be rejected outright as the explanation of the main separation of charge in thunder clouds; the experiments of Gott (1933, 1935) and of Mühleisen and Holl (1953) have shown that the charging process does occur, at any rate under some conditions. But it seems, from present knowledge, unlikely that the process is, in fact, the main one at work in the charging of thunder clouds.

16.16. Wall's Theory

Wall (1948) produced a variation of Wilson's theory, in which there is a reason other than the fine-weather field for the polarization of the precipitation particles concerned.

Wall assumed, without experimental basis, that ice crystals fall in such a way that there is a positive charge at the lower side. Then, as in Wilson's theory, negative charges reach the lower part of the crystal but, in suitable conditions, no positive charges reach the upper part. If the crystal grows at its lower side, fresh positive charges appear and again attract negative charges.

It appears that Wall's theory suffers under many of the disadvantages expressed above in regard to Wilson's theory; while it overcomes the difficulties of the definiteness of the polarity of the cloud, it does so only by the assumption of a phenomenon for which there is no experimental evidence. Soon after the theory was propounded, Israël (1950a) gave it serious consideration, but it is not now much regarded.

16.17. Electrical Cell Theories

Gunn (1935) considered a water droplet and the surrounding air containing water vapour and ions to act as an electrical concentration cell. He deduced that water drops on which vapour is condensing should become negatively charged while those which are evaporating should become positive. From the initial separation of charge as described, gravitation could produce a large-scale separation. GUNN's ideas were later modified, as discussed in § 16.18.

FRENKEL (1944, 1946, 1947) put forward a theory with a somewhat similar basis, and this has been quite widely discussed. A water droplet is supposed to have a potential difference across its surface layers, as a double layer with negative outwards; if an ion gets between the charges of the double layer it moves inwards if positive or outwards if negative. Alternatively, the question can be discussed in terms of BOLTZMANN's law; a negative ion has a less energy inside the droplet than outside, a positive ion a greater energy, and so the drop becomes charged negatively. The drops then fall under gravitation relatively to the surrounding air and this leads to a vertical separation of charge of the right polarity for a thunder cloud.

It seems that these theories do not satisfy a number of the requirements we have set out above for the process of the main separation of charge in the thunder cloud. In the first place, the process envisaged deals with liquid water, not ice, and so operates, if at all, at the wrong temperature; it does not seem possible to transfer FRENKEL's ideas to ice crystals, since it is unlikely that all surfaces of an ice crystal would show the same potential difference and a similar argument is probably also applicable to GUNN's theory. Also the process would operate whether the cloud is precipitating or not and so we might expect effects of similar magnitude to occur, not only with rain clouds, such as cumulo-nimbus and nimbo-stratus, but also with non-raining clouds such as stratus or strato-cumulus, whereas, in fact, these clouds give very small effects. The gravitational separation is that between cloud droplets and air, and so is much less than that which occurs between precipitation particles and air; in this respect, therefore, the processes under discussion are less effective in producing large-scale separation than are other suggested processes, and are almost certainly insufficient (see § 16.10.). As in WILSON's theory, the processes rely on natural ions and so are subject to the same doubt as to the adequacy of the supply of ions to provide sufficient current. SHEPPARD (1950) has discussed FRENKEL's theory in some detail and has drawn attention to other features on which it is open to criticism.

It would appear that these two theories have so many unsatisfactory features that they cannot merit serious consideration for the main process of separation of charge in thunder clouds. The favourable discussion given to FRENKEL's theory in some quarters (e.g. ISRAËL, 1950a) seems undeserved.

But, while these theories are inadequate to explain the main separation of charge in thunder clouds, they may well be worthy of consideration in connection with the very much smaller separation of charge that appears to exist in stratus and strato-cumulus clouds (see § 13.7.).

16.18. Diffusion Theory

GUNN (1954c) considered the capture of ions by drops due to the process of diffusion. He has shown that, in an atmosphere with equal numbers of ions of both signs, the greater diffusion coefficient of the negative ions will cause the drops to acquire negative charges. If the drops remain stationary in the atmosphere, the ions are gradually removed from the neighbourhood of the drops and the charges on the drops may fall to zero. But if the drops are falling relatively to the atmosphere, then there is no such depletion of ions from the neighbourhood of the drop and the charge acquired by a drop can become quite considerable. Experiments with metal spheres by PHILLIPS and GUNN (1954) have shown that the charging does take place in accordance with the theory.

It does not appear likely and indeed GUNN did not claim, that this theory could account for the separation of charge in thunder clouds, but it may be important in the charges of precipitation and of cloud droplets. However, more recently, GUNN (1956) has put forward a similar idea as the initial electrification process in thunderstorms.

16.19. Coagulation Effects

SHISHKIN (1965) has brought up to date the theories discussed in §§ 16.17. and 16.18., by removing some of the difficulties.

He considered first the charging of the cloud droplets, both by diffusion and by the effect of the potential difference at the air–water boundary. He then considered the coagulation of these cloud droplets on to falling particles, irrespective of whether the latter are liquid or solid, and found that large charges appear on the precipitation particles, without any need for change-of-state effects, and large potential gradients can arise.

This theory dispenses with the need for the presence of ice, but does not require the precipitation to be necessarily in liquid form.

16.20. Wilson's Convection Theory

WILSON (1956) brought forward a revised theory, in which his original theory (§ 16.15.) is invoked only as the first stage. He then imagined that the positive ions, moving upwards from below the cloud and mainly originating as point-discharge ions, are captured on cloud droplets and carried upwards in the cloud to give the positive upper charge of the cloud. Meanwhile, the negative ions, attracted to this positive charge from above the cloud, become attached in some way to the precipitation particles and fall to provide the negative charge in the lower part of the cloud.

This theory has an advantage over WILSON's earlier theory in that the ions required are provided from outside and there is not the limited supply that exists within the cloud.

The outstanding difficulty in accepting a theory such as this is to understand how it is that it is the negative ions, rather than the positive droplets present, which become attached to the precipitation particles.

16.21. The Breaking-drop Theory

LENARD (1892) had shown that splashing of water drops gives rise to a separation of charge, the water becoming positive and the air negative. SIMPSON (1909), NOLAN and ENRIGHT (1922) and others found that the separation also occurs when water drops are broken up by air currents. The very large vertical air currents in thunderstorms would be sufficiently large to cause the breaking of water drops which get into the part of the cloud concerned, so that there would be positively charged drops separated from negatively charged air. SIMPSON (1927) made calculations to show that the effect could produce a separation of charge of the magnitude found in thunder clouds.

With the alti-electrograph showing, first, that the main separation of charge gives upper positive and lower negative charges, and, second, that the separation occurs at temperatures below the freezing point, the breaking-drop theory had to be abandoned for the main separation. In spite of this, CHAPMAN (1952) further investigated the breaking of drops and confirmed the conclusion of SIMP-

SON that the quantities of charge available are sufficient for the requirements.

It is possible that the process may be concerned with the lower positive charge (see § 16.36.).

16.22. Early Theories Involving Ice

LUVINI (1884) made use of an observation of FARADAY (1843) that the impact of water against ice electrifies the ice positively and the water negatively. He considered that this might occur at the level of cirrus clouds and give rise to electrification in the air. SOHNCKE (1888) carried this idea further and applied it to clouds lower than cirrus; he suggested that the negative charge on the water would fall out in the rain and thus maintain the negative charge on the earth; however, results soon after this on the actual rain charges showed the wrong sign. SOHNCKE also suggested that the variation, with time of year, of the height of the freezing-point level would account for the annual variation of the potential gradient but he could not similarly explain the daily variation. Also he considered that turbulence within storm clouds could produce the observed electrification by the same process.

BRILLOUIN (1897) invoked a photo-electric effect on ice crystals in cirrus clouds to give the crystals a positive charge and the air a negative charge.

It may be remarked that these early theories concerned ice particles, but later, until 1937, theories all considered the water phase, probably because the presence of ice particles in cumulo-nimbus clouds was not realized.

16.23. Ice Impact Theory

As discussed in § 16.17., charge separation occurs where the precipitation is almost certainly in the solid form. When they had established this, SIMPSON and SCRASE (1937) suggested that the process of charge separation might be connected with the impact of ice crystals on one another, giving a negative charge to ice fragments, and a positive charge, in the form of ions, to the air. Such a process had already been suggested by SIMPSON (1919) to account for the large positive potential gradients found in blizzards (see § 5.66.).

As described in § 3.13., the observational evidence about charge separation on ice impact is extremely confusing. It is clearly very difficult to reproduce in laboratory experiments the conditions that exist within a cloud sufficiently closely to be certain of even the sign, much less the magnitude, of the charge separation. It therefore seems difficult to decide, from the experimental evidence available, whether a process of ice impact can account qualitatively and quantitatively for the charges of thunder clouds.

On the basis of their experiments, REYNOLDS (1954) and REYNOLDS, BROOK and GOURLEY (1957) have put forward a plausible theory of thunder-cloud electrification. The results quoted in § 3.17. lead to the following picture; a pellet of graupel which has grown in falling through the cloud by picking up supercooled water droplets will cause them to freeze and release latent heat, so that the pellet is warmer than its surroundings; also the pellet will have the same degree of contamination as the cloud droplets. But small ice crystals will be at the same temperature as the surroundings and, since they have grown by sublimation from the vapour, they will also be less contaminated. Therefore both on the score of temperature and that of contamination, the graupel pellet will be more conductive than the small ice crystals with which it impacts, and hence, from the laboratory results, the pellet becomes negatively charged and the crystals positively. Attempts to estimate the quantitative significance of this process require more accurate knowledge of conditions within the cloud than are available, but it seemed that the quantities of charge available might be about correct.

The more recent work of HUTCHINSON (1960), LATHAM and MASON (1961b) and EVANS and HUTCHINSON (1963) have led to the conclusion that ice-impact effects are much smaller than were found by REYNOLDS (1954) and so, if the earlier quantitative estimates were valid, the recent work would suggest that ice impact would not be able to account for the charge separation.

Calculations based on reasonable assumptions of the numbers of collisions to be expected between ice crystals and soft hail particles, together with their own observational results for the charge transfer on collisions, led LATHAM and MASON (1961b) to the conclusion that the rate of production of charge in a thunder cloud by this process would be of the order of 10^{-4} C per km^3 per min, while the quantity required for a thunderstorm would be 10^4 times this. REYNOLDS and BROOK (1962) have pointed out that the measurements of REYNOLDS, BROOK and GOURLEY (1957) did, in fact, give a much greater rate of separation of charge, by a factor of 10^5, on the impact of an ice crystal and a hailstone, and that this would, if correct, give an ample rate of charging for a thunderstorm. LATHAM (1965a) has shown that it may be possible to reconcile these differences.

Qualitatively, the ice-impact theory satisfies to a large extent the requirements listed earlier (§ 16.8.), and in one respect it appears to be superior to most others, in that it would lead to the expectation of much larger effects in the very turbulent cumulo-nimbus cloud than in the much quieter nimbo-stratus cloud. Quantitatively, the evidence is contradictory and until this matter is satisfactorily settled it will not be possible to decide on the merits of the theory.

LATHAM and MASON (1962) found the effect was not much increased by the presence of electric fields of the order of magnitude found in thunder clouds.

TAKAHASHI (1962) discussed the generation of charge in a thunderstorm in terms of the breaking of the splinters formed on freezing, not at the time of freezing as in the theory of LATHAM and MASON (1961b) (see § 16.27.), but later by impact after the splinters had settled to a temperature gradient, with the outer end colder and hence positively charged.

Recently, REITER and CARNUTH (1965) have suggested that electrification by ice impact or ice breaking can occur through a concentration difference of NO_3^- ions; they suggest that the NO_3^- ions are formed by point discharge from the points of ice crystals, so that there is feedback and the electrification is magnified; this process would require enough initial electrification by some other method to produce the first point discharge; they described laboratory measurements which agreed with their theory.

16.24. Ice Splinter Theory

From their results on the charges on ice splinters formed during the growth of ice crystals from vapour and during the sublimation of ice crystals (see § 3.20.), FINDEISEN and FINDEISEN (1943) built up a theory of charge separation in clouds, with two primary regions of separation, an upper region where the ice particles grow and obtain a positive charge, the negative charge on the splinters being, somehow, dissipated, and a lower region where the evaporating ice particles obtain a negative charge. The effect of the potential gradient on the splinter charges was considered to give a building up of the charge separation, somewhat similarly to the mechanism of WILSON's process.

ISRAËL (1950a) objected to the FINDEISENS' theory because they had found, in addition to the splinter charges, charges of greater magnitude due to riming and they took no account of these in their theory. A further difficulty lies in the dissipation of the negative charge on the splinters in the upper part of the cloud; this might be expected to rise in the up-draught and give a negative charge at the top of the cloud, but this has not been observed.

As in all ice crystal measurements, it is difficult to be sure that there is no effect of the support carrying the ice, and there is the extra complication that the FINDEISENS cooled their support to temperatures much lower than would occur in clouds. In addition, somewhat similar experiments by KRAMER (1949) and KUMM (1951) have given results which contradict those of the FINDEISENS.

The results of LATHAM and MASON (1961b) involving splinters formed during riming will be discussed in § 16.27.

16.25. Theories Involving Glazing

Based in their experimental work on the production of charges in the glazing process (see § 3.18.), WORKMAN and REYNOLDS (1950, 1953) suggested that, in a suitable temperature range, supercooled cloud droplets meeting falling ice particles will be frozen in part (not wholly because of the latent heat required) and the remaining water will splash off, carrying away positive charge and leaving the ice particles negatively charged. They have gone on to give a distinctive feature to their theory in an explanation of the transfer of negative charge from the precipitation to the actual cloud droplets in the lower part of the cloud, so that the charge is in the cloud itself, rather than in the rain. When the falling ice particle reaches below the 0°C level, the surrounding cloud droplets are not supercooled and now the water splashing off at impact carries a negative charge, so that the cloud droplets in the region acquire a negative charge. These droplets, rising in the up-draught of the storm, will help to give more negative charge to the falling ice particles. Since this augmentation of charge separation can occur only if the cloud extends below the 0°C level, WORKMAN and REYNOLDS claimed that their theory could explain why they found that clouds could develop into thunderstorms only if they extend below the 0°C level, but other observers have found cases of thunderstorms completely above the 0°C level. WORMELL (1955) suggested that the charge carried off by the water might be the effect of the potential gradient, as in the theory of ELSTER and GEITEL (see § 16.14.); in this case, there would be no change of sign at the 0°C level and no "cycling".

LUEDER (1951 b) put forward a similar theory and carried out calculations to show that the amount of charging to be expected from this process is several times greater than that from WILSON's ion-capture process, and is of the right order for thunderstorm charges. MASON (1953 a) and REYNOLDS (1953) also made calculations showing the experimental results on the rate of charging are sufficient to account for the charges of thunder clouds. KANO (1954) has made more detailed calculations and obtained results for the distribution of charge with height which are in reasonable agreement with GUNN's (1947, 1950) observed results.

REYNOLDS (1954), though part author of the original theory, has raised two objections to it. In the first place, it is probable that glazing, as defined, does not extend to temperatures below about -15°C (see § 3.3.), whereas charge separation certainly occurs at lower temperatures than this; so it would be necessary that riming should also produce charge separation, but he had no experimental evidence for this. Also, REYNOLDS found that charging occurs only if there are present both water droplets and ice crystals; this shows that glazing does not, in itself, produce charge separation, but is only an incidental accompaniment, providing, according to REYNOLDS's picture, a temperature difference (see § 16.23.). WORKMAN (1965) has argued that, in spite of arguments against it, the theory is still worthy of consideration.

While this discussion shows that the separation of charge may not be caused primarily by the process of glazing, the calculations mentioned are based on the experimental observations of charge separation under conditions which may approach those in clouds, and may remain valid, whatever is the primary cause of charge separation.

16.26. Calculations on Glazing Process

In addition to the calculations made by LUEDER (1951 b), MASON (1953 a), REYNOLDS (1953) and KANO (1954), there have been somewhat different calculations, in greater detail, by TWOMEY (1957). He started with an assumption, based on his own observations (TWOMEY, 1956), that an ice particle cannot acquire a charge greater than that which would give it a surface potential gradient of

70,000 V/m if it were spherical; he pointed out that an ice particle of size 300 μ could be charged to more than this by a single impact with a 1 μ supercooled water droplet if LUEDER's figure of $-0{\cdot}2$ C per cc of supercooled water is correct. Thus the basic assumption of TWOMEY's calculations as regards electrical effects is the figure of 70,000 V/m. He then considered the growth of ice particles and the charge acquired, finally reaching values of the rate of charge separation which agree quite well with the observed phenomena; other consequences of the theory are also satisfactory, among them the conclusion that the negative charge is to be found in the cloud droplets rather than in the precipitation.

16.27. Theories Involving Riming

MASON (1953 b) put forward a theory similar to that of WORKMAN and REYNOLDS (1950, 1953), but the positive charge was considered to go into the air as ions, rather than to be carried off by the fragments of water splashing off. The essential process is then riming, in which the whole of the water droplet is frozen, rather than glazing, in which some remains unfrozen and splashes off.

From the measurements of LATHAM and MASON (1961 b), showing that there is separation of charge on riming, they had support for this theory, modified in that the positive charge would be carried off, not as ions, but in the small splinters produced in the freezing of the droplets (see § 3.11.). The quantities of charge appear to be about of the right magnitude to account for the charges in thunder clouds; LATHAM and MASON (1962) found that the effects are not much altered by the presence of electric fields of the magnitudes which exist in thunder clouds.

This appears one of the most satisfactory of the theories yet put forward, but there remains the problem whether it can explain effects in very violent storms (§ 6.11.) and whether it can account for the differences between the electrification in cumulo-nimbus and nimbo-stratus clouds.

16.28. Theory Involving Freezing

CHALMERS (1943) suggested that the sudden freezing of water droplets might, by destroying a free water surface, give just the opposite electrical effect to the breaking of water drops, which produces new free water surfaces; then it would be expected that the ice would acquire a negative charge and the air a positive. At the time there was no experimental evidence for the effect, but it has since been discovered by MASON and MAYBANK (1960), though the explanation is different.

CHALMERS (1943) originally put forward the suggestion in connection with the freezing of a small number of water droplets that is often the first stage in the formation of precipitation (see § 3.5.) and which must occur at somewhere about the same level as the main charge separation. But it is clear that electrical effects in riming and glazing would be connected with the same process of charge separation.

The results of MASON and MAYBANK (1960) show that the charge separation on freezing is too small for there to be any significance in the thunder cloud of the freezing of the few droplets which initiate precipitation, but the effects in the riming of the many droplets encountered during precipitation are important (see § 16.27.).

16.29. Other Theories

Some other theories may be mentioned, but, as they depend on hypothetical processes of charge separation which have not yet been experimentally verified, they will not be discussed in detail.

ROSSMANN (1948) suggested electrification of ice particles by a pyroelectric effect and built up a thunderstorm theory on this basis. But there was no experimental basis for such an effect and MASON and OWSTON (1952) showed that ice has no pyroelectric effect.

KUMM (1951) suggested that the polar nature of ice crystals might cause the crystal points to present a positively charged surface to the atmosphere; they would thus attract negative ions in the same way as would water droplets on the theory of FRENKEL (§ 16.17.); they might also give charge separation on fragmentation. But there is no evidence for such charge orientation.

PÜHRINGER (1961) suggested that electromagnetic induction in conducting water drops in clouds moving in the earth's magnetic field would give rise to an induced e.m.f. which would give the electrostatic field of the thunder cloud; the magnitude of the effect would depend on the capacitances of the cloud drops, but it would seem unlikely that the effect would be great enough.

16.30. Discussion of Theories Involving Gravitation and Natural Ions

In the preceding sections, a number of theories have been considered, many of which have been put forward to explain the main separation of charge in thunder clouds; in a number of cases, evidence has been produced to show that the process can operate under the conditions that exist in thunder clouds, and so can play some part in the charge separation. Some of those who have discussed the problem have thought that perhaps two or more processes might be in operation to a more or less equal extent, possibly in different regions of the cloud. However, it would be a rather unlikely coincidence if two or more processes happened to be of approximately equal importance, and so it seems logical to look for one process to which can be ascribed the main separation of charge in the thunder cloud, other processes perhaps playing minor parts.

Considering first those theories which depend on the natural ions existing in the atmosphere, WORMELL (1953a) made calculations which show that the normal rate of production of ions within the volume of the thunder cloud is insufficient to maintain the observed rate of regeneration of separated charge. Thus, if natural ions are the source of the charges separated, there must be some mechanism by which ions initially produced outside the volume of the cloud are brought into the cloud to provide some of the charge. Alternatively, this type of process might act with ions produced within the cloud by some mechanism which does not operate in the absence of the cloud; for example, it has been suggested that one lightning flash within the cloud might produce enough ions for the building-up of charges to cause the next flash, the first flash arising from the separation of the charges on ions which have accumulated over a longer period; it is, however, doubtful if enough ions exist under equilibrium conditions in a developing thunder cloud to give the first flash. Otherwise, it would be necessary to postulate some process within the cloud which would provide an adequate supply of ions; for example, REYNOLDS (1954) suggested that CHAPMAN's (1952) results show that the breaking of water drops might provide sufficient ions, which would rise in the up-draught and then be separated by WILSON's process. This does not appear to be a very attractive theory, but probably cannot be immediately discarded.

If one accepts WICHMANN's views (see § 12.29.), supported by more recent evidence from the preliminary processes in lightning (see § 14.10.), that the lower positive charge is the essential feature of the thunder cloud, then it might be possible to consider an ion-capture process to be responsible for the "main", upper separation of charge, and some other process to trigger off the lightning flash. However, the supply of ions might still be inadequate.

ISRAËL (1950 a), in discussing theories of the separation of charge in clouds, chose as the most promising of those then available two theories involving natural ions, those of FRENKEL (1944) and WALL (1948). In view of what has been written above, such a choice would not now appear justified; in addition to the difficulties discussed in regard to the supply of ions, WALL's theory depends on a property of ice crystals for which there is as yet no experimental evidence, and FRENKEL's process would operate at the wrong temperature, in all clouds, whether precipitating or not, and with too small a velocity of separation.

16.31. Discussion of Theories Involving Charge-production Processes

A charge-production process produces equal amounts of charge of the two signs at the same place, and the opposite signs are, or become, attached to particles which are of different nature (material, size, etc.). It is very difficult to imagine any process other than gravitation which can then segregate the particles carrying the different signs.

The theories of this kind which require any discussion can be divided into those which involve impacts of ice particles and those which involve a change of state, in particular that from liquid to solid or water to ice.

The original ice-impact theory of SIMPSON and SCRASE (1937) has been developed in terms of differences of temperature and of contamination by REYNOLDS (1954) and would be a very serious contender for the place of the most acceptable theory but for the results of LATHAM and MASON (1961 a, b), which showed that the separation of charge was much less than REYNOLDS found, and not sufficient to give the observed charge separation in a thunder cloud; these results agreed with HUTCHINSON's (1960) and EVANS and HUTCHINSON's (1963) failure to find the quantities of charge reported by REYNOLDS. However, until further experiments, preferably on similar lines to those of REYNOLDS, confirm or otherwise the conclusions of LATHAM and MASON, it would not be possible completely to eliminate the ice-impact theory.

WORKMAN and REYNOLDS (1950, 1953), followed by LUEDER (1951a, b) with their theory of charging due to glazing, have been able to calculate that the process would give sufficient charging. But REYNOLDS (1954) pointed out that the process of glazing probably cannot occur at temperatures below about −15°C, while there is no doubt that charge separation does still go on at these temperatures.

MASON (1953a) suggested riming, rather than glazing, as the important process, and the discovery of charged splinters by MASON and MAYBANK (1960) and the measurement of the charge production by them by LATHAM and MASON (1961b), showing sufficient charge for the thunder cloud, has led to this process being now that most favoured.

One point in favour of ice impact rather than riming is that this would more easily account for the much less charging in nimbostratus than in cumulo-nimbus clouds, since the turbulence is so much less and hence there would be fewer impacts.

In favour of riming or glazing theories is the fact that they give a mechanism by which the charge on the precipitation particles can be transferred to cloud particles, since, at temperatures above the freezing point, the impact of cloud drops on a hailstone splashes off negative charges thus removing the charge on the hailstone.

However, there still remains the awkward fact that observations exist (see § 12.26.) that thunderstorms sometimes occur when there can be no ice present in the cloud.

16.32. Theories not Involving Gravitation

Since the theories that involve gravitation as the agency for the separation in space of charges of different signs on particles of different properties may come into difficulties in regard to the amounts of charge separated on to different particles but not yet separated in space, particularly in the cases of very violent storms, it is necessary to consider theories in which some agency other than gravitation separates the particles in space. In the past, such theories have been neglected and it has rather been taken for granted that gravitation is the operating force; the connection between charge separation and precipitation has supported this view.

If the separation in space is not by gravitation, then it is necessary to find some other form of relative motion of different particles that can effect the separation in space. In the thunder cloud there is certainly relative motion of different parts of the cloud, there being an up-draught in some parts and a down-draught in others; if, therefore, it is possible to find some way in which charges of one sign become attached to particles moving in the up-draught and charges of opposite sign to those moving in the down-draught, then it would be possible to get relative velocities of separation many times those possible by gravitational action.

The up-draught is the main feature of the initial stages of the thunderstorm, but there is no definite down-draught within the storm until the precipitation has become well developed. Observations of thunderstorms (see § 12.23.) show that electrical activity occurs as soon as precipitation is observed and probably well before the down-draught has become established, so it is not likely that this down-draught can be directly concerned in the separation of charge.

The idea was first put forward by GRENET (1947), amplified in 1959, and independently and in more detail by VONNEGUT (1955) that the agency of separation in space is relative motion between the main up-draught inside the cloud and a down-draught round the edges of the cloud; the existence of such a down-draught is by no means universally admitted by meteorologists. VONNEGUT suggested that the up-draught would bring in, from below, the positive charges which arise (see § 2.25.) from variations in conductivity; at a later stage, the negative potential gradient beneath the cloud becomes sufficiently large to give point discharge at points close to the ground and therefore produces positive ions, which are also drawn up into the cloud; thus the up-draught contains a positive charge. Above the cloud, negative ions are drawn down from the electrosphere by the positive charge that has accumulated at the top of the cloud, and these negative ions are then considered to become attached to the cloud droplets which move in the down-draught. When it gets to the bottom of the cloud this negative charge gives a potential gradient which causes point discharge and gives the positive ions.

VONNEGUT and MOORE (1958c) attempted to find corroboration for this theory by producing a negative space charge below a developing thunder cloud; according to the theory, the resulting cloud should be of opposite polarity to the normal storm cloud. Results of a preliminary experiment were inconclusive.

It should be pointed out that WILSON's (1956) theory (see § 16.20.), though based on convection like VONNEGUT's, is still a theory involving gravitation, since the negative charges are attached to precipitation particles.

Although VONNEGUT's theory depends upon ions in the atmosphere, it does not rely upon ions produced by cosmic rays, etc., within th volume of the cloud; WORMELL's (1953a) criticism (see § 16.30.) does not therefore apply.

The outstanding difficulty in VONNEGUT's theory appears to be that of explaining how it is that positive charges, which have risen in the up-draught, do not go down again in the down-draught, but remain at the top of the cloud, while the down-draught contains negative charges attracted from above the cloud; and, in the same way, it is not explained how it is that the up-draught at the bottom of the cloud should contain the positive ions coming up from below but not the negative charges that have come down in the down-draught.

The pros and cons of the convection theory have been thoroughly argued by VONNEGUT (1965) and LATHAM (1965b).

16.33. Modification of Cloud Charges by Artificially Charged Air

In order to provide evidence in regard to VONNEGUT's theory of thunderstorm electrification, VONNEGUT *et al.* (1962) introduced artificial charges into the air below small cumulus clouds and investigated the charges found in the clouds. Not only did they find, as expected, charges in the clouds of the same sign as those released below, but they also found regions in which there existed charges of opposite sign, in the same way as would be expected on VONNEGUT's theory. If thunderstorms involve similar types of circulation of air to those in the cumulus clouds, then the mechanism proposed by VONNEGUT might well operate similarly, but to a greater degree, in thunderstorms.

VONNEGUT *et al.* calculated that the energy of charges stored in a cloud would be proportional to the fifth power of the linear dimensions.

16.34. Internal and External Currents

During a thunderstorm there are currents both within the storm cloud and outside it. It is probably allowable to consider the storm as a quasi-static state when averages are taken over the active period of the storm. Then the average currents above, within and below the storm are equal, including in these currents the generating current, in whatever way it is acting, and dissipating currents both in the form of conduction currents and in the form of lightning.

If the charges concerned are first produced, by whatever means, within the cloud, then the current of charge separation in the cloud, i.e. the generating current, must be balanced, on the average, by the various dissipating currents inside and outside the cloud. Therefore it follows that the generating current must be greater than the total current through the cloud, measured above, within or below the cloud.

On the other hand, if the charges are first produced outside the cloud and are drawn into it, as suggested by the theories of GRENET and VONNEGUT (§ 16.32.) and of WILSON (§ 16.20.), the whole of the generating current must be supplied by currents external to the cloud; since there must be some dissipating currents outside the cloud, it follows (see CHALMERS, 1961) that the generating current is less than the external currents and probably less than the total current through the cloud.

The evidence available is in favour of the generating current being greater than the total current. The recovery of potential gradient after lightning flashes (see § 12.11.) suggests a generating current of about 3 A for storms in England. On the other hand, for storms in America the measurements of the current above the storm (see § 11.17.) have given values of 0·5 A and 1·3 A.

It would be most desirable to have results for generating currents and external currents for the same storms, when the above argument would be able to discriminate between theories involving charge production inside and outside the cloud.

16.35. The Lower Positive Charge

The lower positive charge has been found definitely in only a proportion of thunder clouds, but it appears to be confined to a fairly small volume and so its existence may well have been missed in other cases. WICHMANN (1952b) has suggested that the lower positive charge is the distinguishing feature of the thunder cloud, and CLARENCE and MALAN (1957) (see § 14.10.) have concluded that lightning discharges are initiated between the negative and the lower positive charge; this suggests that the lower positive charge is present in all thunder clouds. The charge has been found where the temperature is at or above 0°C, so that liquid water may play a part in its production.

Two general types of theory have been put forward for the production of the lower positive charge. The first type is based on the assumption that the lower positive charge is produced by a process quite distinct from that responsible for the main thunder-cloud charges; as most theories of this type involve liquid water, it is convenient to group these as "water theories". In the other type of theory it is assumed that the positive charge arises, not by some distinct process of charge separation, but by secondary effects which cause the positive charge to become concentrated in the base of the cloud; we can call these "secondary theories".

16.36. Water Theories

Two of the processes already discussed might be suitable to explain the production of the lower positive charge, since they would operate at the correct temperature.

SIMPSON's (1909, 1927) process of the breaking of drops (see § 16.21.) might operate in the violent up-draught, and if ice particles grow to large sizes in the cloud, they would form water drops on melting which would be too large to be stable and would break up; thus the charging proceess would occur at the melting level.

DINGER and GUNN (1946) found a production of charge during the process of melting (see § 3.18.), again a process which would operate at the correct level in the cloud.

Since FINDEISEN and FINDEISEN (1943) had found a positive charge on a deposit by riming, WICHMANN (1952b) suggested that the lower positive charge might be produced by this process. Results, quoted in § 3.13., that the impact of water, which is not supercooled, on ice gives the ice a positive charge, have led to the suggestion that this is the origin of the lower positive charge where the falling ice crystals or pellets are in a region of water drops not supercooled, but where the ice has not yet metlted.

If, as has been suggested, the lower positive charge is a feature which distinguishes the thunderstorm from the shower, then the process which produces this positive charge must be one which occurs only in the very violent conditions of the thunderstorm; this would, perhaps, suggest that it is a violent process, such as the breaking of drops, rather than a less violent process, such as melting. However, the effect of the thunderstorm conditions may be such as to concentrate the positive charge in a limited volume, whereas in other conditions it is more dispersed.

The results for the continuous-rain cloud (§ 13.4.), suggest that in this case also, there are two processes of separation, the lower one being concerned with melting. It would then appear that, if it is the same process in the two types of cloud, a similar lower positive charge should appear also in the shower cloud.

It is tempting to suggest that, if the lower positive charge is to be ascribed to a distinct process of charge separation, this should be a physical process of a similar type to that invoked for the upper process. Thus if the upper process involves ice impact, with fracture of ice crystals, the lower process might be the breaking of water drops; but if the upper process is a change of state process, involving glazing or riming, then the lower process might be melting. It is, however, to be realized that, while in the continuous-rain cloud the two processes are of approximately equal importance, in the thunderstorm, the upper process is considerably more important; this might be taken to mean that the two processes are of different natures.

KUETTNER (1950) found at the Zugspitze that the positive charge is centred at the 0°C level; but ice particles would melt only at levels below this, because of latent heat requirements, so that it may be incorrect to assume that the precipitation must be liquid for the lower separation of charge.

16.37. Secondary Theories

SMITH (1951a), MALAN and SCHONLAND (1951b) and MALAN (1952) have suggested that the lower positive charge might be due to ions from point discharge at the earth's surface in the strong negative potential gradient, which could get caught in the strong up-draught of the thunder cloud and then become immobilized by attachment to cloud particles in the base of the cloud. Before such a theory could gain acceptance, it would be necessary to calculate potential gradients and ion paths to find out whether a consistent scheme can be worked out; this has not yet been attempted. If this idea can be substantiated, it may be that the removal of the point-discharge ions by the wind will be able to account for the anomaly of the absence of increase of field as measured by the alti-electrograph (see § 9.23.).

WALL (1948) suggested that the process of melting could give rise to a concentration of positive charge as a secondary effect by the following action. In the base of the cloud, there are negative ice particles falling and positive ions rising, with negative charge predominating, at temperatures just below 0°C; a little lower, where the ice particles have melted, the velocity of fall of the particles will be considerably greater as the water drops suffer much less air resistance than the ice particles, and if they carry the same charges as higher up, the charge on the precipitation in unit volume of space would be much less, and so the total volume charge in this region may be positive instead of negative. As in the case of the last suggestion, detailed calculations would be needed to show if this idea if feasible. If it should be, it would give a positive charge at the correct temperature. This suggestion and the last depend upon the ions from point discharge and so would not give the lower positive charge in a cloud which does not produce point discharge.

WALL (1948) also suggested that, as it is known that precipitation in a thunder cloud produces a down-draught over a limited region, this might bring down to the base of the cloud the upper positive charge. KUETTNER's (1950) results (see § 12.16.) would appear to be much more in favour of this form of a "secondary" theory than the first, since a down-draught could not collect ions as suggested.

16.38. Subsidiary Charge Centers

In a number of cases, such as some of the alti-electrograph records, the potential gradients due to thunder clouds can be explained only if, in addition to the two main charges and the lower positive charge, there are other charge concentrations in the cloud. SIMPSON's (1949) "patterns" (see § 5.69.) may also require for their explanation distributions of charge different from those of the simple thunder cloud, though WHITLOCK and CHALMERS (1946 found that some of the "pattern" effects can be ascribed to space charges from point discharge and not to effects inside the cloud.

There is probably not yet enough knowledge to make it profitable to discuss such subsidiary charge centres in the same way as the lower positive charge, namely whether they are to be considered as arising from a distinct process of separation or from secondary movements of the charges produced by the primary process. If the latter, the down-draughts in thunderstorm cells or wind shear in the cloud might cause sufficient relative motion to give rise to subsidiary charge centres.

16.39. Charge Separation in Clouds Giving Continuous Rain and Snow

If, as discussed in § 16.15., the origin of the charges in continuous rain and snow may lie in the clouds, then we can discuss the possible processes at work. In the continuous-snow cloud, the snow must receive a negative charge, the positive charge presumably remaining behind in the cloud, and it may be that it is the same process in operation, to a smaller extent, that occurs for the main separation in a thunder cloud; since there is not sufficient charge to produce a high enough potential gradient to give point discharge, there is not the change of sign of the precipitation caused by WILSON's process as discussed in § 10.20. In order to account for the negative potential gradient when the precipitation current is small or zero, it would be necessary to assume that some, or even all, of the negative charge can remain behind in the base of the cloud and does not fall.

For continuous rain, since this usually starts as snow, we must have the same process as with snow, together with another separation of charge operating in the opposite direction. This latter process may be the same as that which operates on the base of the thunder cloud; but as this process, which is very evident in the storm cloud, is much less so in the less violent shower cloud, it seems unlikely that it should reappear in the still less violent continuous rain cloud. If, however, it is the same process, the possibilities discussed in § 16.36. remain. If it is not the same process, then one may suggest that it is the melting effect discovered by DINGER and GUNN (1946) which gives the rain the positive charge; it is unlikely that the breaking-drop process of SIMPSON (1909, 1927) could act in the quiet conditions of the continuous-rain cloud.

16.40. Charge Separation near the Ground in Continuous Rain and Snow

As discussed in § 10.25., the separation of charge in continuous rain and snow may occur at or close to the ground. ISRAËL and

LAHMEYER (1948) and SMITH (1955) have suggested splashing at the ground as the origin of the charges associated with continuous rain, but there are objections to this as discussed in § 10.16. SIMPSON (1915) suggested that rain might receive its positive charge by the breaking of drops in the turbulence close to the ground, due to eddies caused by buildings, etc., but later (1949) expressed strong doubts as to whether the air velocities would be large enough. If a process such as this accounts for the rain charges, one would require a similar process of opposite sign to occur with snow and this would be provided by the process of ice impact as suggested by SIMPSON and SCRASE (1937), but it does not seem very clear how it would fit in with the results of REYNOLDS (1954) that the electrical effects of ice impact depend upon temperature and contamination differences.

16.41. Charge Separation in Non-raining Cumulus Clouds

In non-raining cumulus clouds, GUNN (1952) found that the larger droplets carry a small excess of positive charge, while the corresponding negative charges reside mainly in large ions (see § 13.7.). It is not easy to suggest a process for this separation of charge; if it is the same as occurs in the continuous-rain cloud, the process could only be due to some property of water, since no part of the cloud is ice and no precipitation falls out of the cloud. If the droplets are evaporating, GUNN's own early ideas (§ 16.17.) might apply. Or it may be a manifestation of the diffusion process of charging (§ 16.8.), with charges of opposite sign to those suggested for thunder clouds.

16.42. Charge Separation in Stratus Clouds

WHITLOCK and CHALMERS's (1956) results indicating that there is sometimes a concentration of negative charge in the base of non-raining stratus and strato-cumulus clouds (see § 5.61.), present another problem. The droplets might acquire charge by FRENKEL's process, or by GUNN's if they are growing; or WILSON's process might be operating. We have already discussed evidence (§ 7.20.) that there are small ions in such clouds, but small ions in a potential gradient of 1000 V/m would move only at 0·15 m/sec, so that they would remain "slow" compared with the droplets, and if there is a vertical up-draught of a speed between this and the fall velocity of the drops, WILSON's process could occur.

The distinction of sign between the charges on the drops in cumulus and stratus clouds is difficult to understand.

16.43. Control of Thunderstorm Electrification

If and when the mechanism of thunderstorm electrification has been elucidated, it may become possible to devise methods of controlling the phenomenon, at least for limited periods and over a limited area; but until the mechanism is understood, there is little hope for practical success. REYNOLDS (1954) discussed some suggestions, including one of introducing into the cloud some contaminant that would neutralize the effect of natural contaminants, assuming that these are in fact responsible for the separation of charge. GUNN (1951) suggested the dispersion of radioactive material into a cloud in sufficient quantity to increase the conductivity so that the charges separated can be dissipated by internal conduction currents without reaching the magnitudes necessary for a discharge. REYNOLDS (1954) also discussed suggestions to prevent the BERGERON process (§ 3.5.) operating either by inhibiting the production of ice particles or by providing so many that none can grow large.

If it should turn out to be correct that the lower positive charge is essential to the lightning flash to ground, then the method of control would have to be one which would prevent the development of this lower positive charge and would depend on what is found to be the mechanism of its production.

CHAPTER 17

Conclusion

17.1. Outstanding Problems

From time to time it has been suggested, even by workers in the subject, that the time is approaching when there will no longer remain important unsolved problems in atmospheric electricity. In this chaper, some of the remaining problems will be mentioned, and it will be obvious that there remains a great deal yet to be done.

One of the major problems that has occupied many workers in the subject has been that of the maintenance of the negative charge on the earth; in general, this appears to have been successfully solved although there are several minor details yet to be discussed.

The other major problem, that of the origin of the charges in thunderstorms, can by no means be considered near to solution, and it might yet turn out that the main process at work is one which has not been considered up to the present.

Various other outstanding problems will also be discussed.

17.2. The Maintenance of the Earth's Charge

Although all the evidence available appears to agree with the hypothesis that it is thunderstorms which maintain the negative charge on the earth, it would be desirable to accumulate evidence which could make more precise the estimates of the different factors bringing charge to the whole surface of the earth, such as those of WAIT (1950) and ISRAËL (1953a) given in § 11.16. It would be necessary to make estimates, of greater accuracy than are now possible, of the various factors in the less accessible parts of the world, e. g. the Arctic and Antarctic, deserts, mountain regions and tropical forests.

It would be very desirable to have more results for the total current over thunderstorms, similar to those of GISH and WAIT (1950) and of STERGIS, REIN and KANGAS (1957b); the averages of these two sets of results differ by a factor of 2·6 and if, as appears quite likely, this depends on a difference in the peculiarities of the storms investigated, it would be very useful for the same type of measurement to be made with many different kinds of thunderstorms in various locations. The question of ocean thunderstorms is important and measurements of currents above these would confirm, or the reverse, that WHIPPLE and SCRASE (1936) were justified in the inclusion of these storms in their discussion, and that WICHMANN's (1951) objection is not valid.

In discussing the maintenance of the earth's charge, no account has been taken of continuous-precipitation clouds; these are much more frequent than thunderstorms, but it has been assumed that the current through each is so small that the total is insignificant. This should be verified, or the reverse, by measurement of the currents above such clouds in different situations.

17.3. Thunderstorm Problems

There are many problems yet to be solved regarding the electrical phenomena of the thunderstorm, and the answers to these will inevitably suggest more questions.

One question of fundamental importance is that of the warm thunderstorm. Are the observations quoted in § 12.26. "freaks" or are there many such clouds which are completely at temperatures above 0°C and which give lightning? If so, is the electrical structure of such clouds similar to that of the normal thunder cloud which can contain ice? Can one find a place in the world where such storms are frequent enough for it to be feasible to set up apparatus to ascertain all the features concerned, including the verification of temperatures by means of balloon measurements? If it can be firmly established that there is lightning from such clouds, then it must follow that ice is not essential in this type of storm, and if it is established that such clouds have similar electrical properties to normal thunder clouds, then ice can no longer be regarded as necessary for the separation of charge in all thunder clouds. It therefore seems that a thorough investigation of warm thunderstorms should be very high on the priority list for those concerned with thunderstorm problems.

Another problem is that of the lower positive charge. Is it an essential feature of all thunderstorms? Does it appear in showers which do not develop into thunderstorms? Whereabouts in the lower part of the cloud is the positive charge situated? Does it occur in the down-draught associated with the precipitation? If so, it might have a direct connection with the precipitation and, from the temperature, might be connected in some way with the process of melting. On the other hand, is the positive charge to be found in the up-draught? In this case, it might just represent a concentration of positive ions from point discharge below the cloud.

Perhaps related to this problems is that of whether or not a cloud has to extend below the level of the freezing point before it can become a thunder cloud. In other words, must there be a possibility for the melting of precipitation in the base of the cloud?

Yet another problem is that of extremely violent thunderstorms. Is it possible to get any estimate of the rate of separation of charge in such storms? It would then be possible to test suggested theories not only as to their adequacy for normal storms, but also as to whether they could possibly provide the rate of charge separation at its maximum.

Problems of conductivity and ionic content inside thunder clouds do not appear simple to solve, but if they could be solved they would give most useful information about various processes within the cloud.

Individual thunderstorms differ widely from one another, and much of our present knowledge is obtained from a variety of different storms, so that different "average" values may not be comparable. It would therefore be most desirable to set up some sort of "combined operation" in which a number of different techniques are used to obtain information about the same individual storm and the same flashes of lightning within the storm. It would be necessary to choose some place where, at the appropriate season, thunderstorms occur regularly and reliably, and to use all the various methods, electrical and optical, that can be of value. Even then, it might well be found that different storms have different features, but one could hope to know a good deal about a limited number of storms. Following on this, it might be desirable to repeat the exercise at some other place, where the storms originate differently and might have different characteristics.

17.4. Possible Processes of Charge Separation

It is very desirable that there should be more laboratory experiments on electrical phenomena connected with ice, such as effects of ice impact, of freezing, riming and glazing, and any connection between these, under conditions approaching as closely as possible the conditions in a cloud. The final tests of a theory of charge separation must be, first, whether the process can be demonstrated to occur under the conditions that actually exist in nature; second, whether the amount of charge separation agrees with that found in clouds; third, it must be able to give the rate of separation found in very violent storms; and, fourth, it must not be such that one would expect greater charges than are actually found in nimbostratus and non-raining clouds.

The suspended hailstone, as used by WORKMAN and REYNOLDS (1953), and the natural supercooled fog, as used by LUEDER (1951 a), appear to be methods of approaching natural conditions, and it would be desirable to use as large a volume of cloud as possible, to avoid any wall effects. Any effects of polarization should also be investigated by the application of an electric field.

Whereas some writers have expressed the view that there may be a number of processes each taking a part in the electrification of a thunderstorm, the author's opinion is contrary to this. With all the powers of 10 from zero to infinity at her disposal for different processes, it is unlikely that Nature will have chosen several processes to have approximately equal effects; it would seem much more probable that one or two, or possibly even three, processes would predominate in thunderstorms; if the lower positive charge is the result of a separate process, then there must be two processes with not very different effects in opposite directions. The aim of the worker in this field must be to identify the important processes from among those which have been suggested already and those which may be suggested in the future, and the author is not tempted to indulge in any crystal-gazing (not even ice crystals!).

17.5. Instrumented Bomb Method

In order to obtain more precise information regarding the charges in a thunder cloud, it is necessary to take instruments into the cloud. The use of aircraft is limited by the danger involved and it is difficult to be certain that a balloon will go into the portion of the cloud that is of interest. An alternative would seem to be to drop, through the cloud from above, a small "bomb", carrying something similar to radio-sonde instruments, adapted to give the potential gradient; this would be able to move at a sufficient speed to overcome the currents in the cloud that affect aircraft and balloons, and, by choice of the dropping-point, could be made to penetrate the centre of the storm. A suitable arrangement would be necessary to prevent damage being done by the bomb, and also to salvage the instruments.

17.6. Intermediate Ions

There are some problems yet to be solved in relation to intermediate ions. One is the question of the conditions under which the intermediate ions form a separate group; then there is the question as to whether some special substance such as sulphuric acid, is required for their appearance. Further work is also needed in respect of HOGG's (1939a) discovery of the unit size of ion; does this result hold wherever intermediate ions are found and what is the significance of this particular unit size?

There is also scope for investigation of the production and destruction of intermediate ions; are there combination coefficients and equilibrium equations as for the small and large ions? Are there uncharged particles of the same size as the intermediate ions and, if so, how can they be detected? Do intermediate ions act as centres for condensation of water vapour and if so do they require any degree of super-saturation before they become effective?

17.7. Equilibrium of Ionization

The problem of the ionization state of the lower atmosphere has been clarified by the discovery that equilibrium is not reached for the large ions under normal conditions in the atmosphere, so that the equations referring to equilibrium cannot hold. Thus there appears to be little hope of relating the ion and nucleus content of the atmosphere to the combination coefficients and other properties of the ions and nuclei. It is, at best, far from the source of nuclei that there can be any equilibrium of ionization and it seems very likely that nuclei will be much modified at such distances and that any properties that may be found for them will be different from the properties of newly formed nuclei.

An approach to the problem of the properties of nuclei by measurements under controlled conditions in laboratory experiments seems to be getting somewhat outside the scope of atmospheric electricity and of this book. But if definite results can be obtained from such experiments and then applied to the consideration of the non-equilibrium state, it may be possible to explain measurements of numbers of charged and uncharged nuclei.

17.8. Potential-gradient Measurements

It seems that future advances from potential-gradient measurements are most likely to come from their use to show the presence of space charges, both in fine weather, as in the work of MÜHLEISEN (1953), and in disturbed weather.

If a number of potential-gradient recorders could be installed at different places, it would be possible to follow the motion of space charges, extending the work of WHITLOCK and CHALMERS (1956). A further development would be the use of measurements of potential gradient above the ground, on masts or balloons.

There appears to be a possibility of using electrical measurements to give indications of small-scale air motion, since ionization produced in a volume of air appears to persist after several km of travel. Air could easily be artificially ionized, e.g. by the method used by LARGE and PIERCE (1957).

17.9. Conductivity of the Air

A problem that has arisen in regard to conductivity is the investigation of abnormal conductivities below clouds, presumably caused by extra ions from point discharge. The evidence on this is conflicting and measurements in aircraft and at various places on the ground, particularly on high ground, are desirable. If the positive-ion conductivity is found to be increased in strong negative potential gradients, this would confirm the space charge expected from point discharge and might even give a method of measuring this, but would make it still more difficult to explain the failure to find the expected increase of potential gradient with height below clouds, with the alti-electrograph (see § 9.23.). On the other hand, if there is no increase in conductivity, the problem arises as to what has become of the ions released in the point-discharge process.

More measurements of the conductivity above thunder clouds would be desirable to confirm the conclusion of GISH and WAIT (1950) and of STERGIS, REIN and KANGAS (1957b) that there are no abnormalities, so that any currents between the tops of clouds and the electrosphere are simple conduction currents and not discharges.

If the difficulties of conductivity measurement inside a cloud could be overcome, this would lead to useful information (see § 17.4.).

17.10. Air-Earth Current

After the measurements of the NOLANS (1937) and HOGG (1939b), it seemed that the problems of the two methods of measuring the air–earth current had been satisfactorily solved. But the more recent measurements of LAW (1963) and of HIGAZI and CHALMERS (1966) (see §§ 7.6., 8.15.), near the ground, and of KRAAKEVIK and CLARK (1958), within and above the *austausch* region (see § 8.17.), have shown that the convection current in the lower atmosphere is not to be neglected.

A major difficulty in dealing with this problem in the lowest regions of the atmosphere is that there has hitherto been no means of measuring directly the air–earth current at any level other than that of the ground. If, however, the suggestion in § 8.16. is successful, this should give useful information.

A comprehensive survey of the air–earth current near the ground by direct and indirect methods would determine the importance of convection currents.

17.11. Point Discharge

It is important, both from considerations of the transfer of charge between clouds and ground, and in connection with relations between precipitation currents and point discharge, to be able to determine the true point-discharge current density corresponding to the measured point-discharge current through a particular point. To obtain accurate information on this matter it will be necessary to determine the point-discharge current through a tree, and this has been discussed in § 9.14., where some methods of measurement are described. Another possibility would be to use the tree as the primary of a transformer wound on a magnetic core round it; the secondary current would then depend on changes in the tree current and integration should give the current; or, if the point-discharge current to the tree is in the form of pulses, these might be recognized by the transformer.

If there is a total point-discharge current below a storm amounting to even 1/10 A, and even if spread over a wide area, modern magnetic techniques should be able to detect this.

If, during a period of fairly steady point discharge, measurements could be made, by means of balloons or aircraft, of the potential gradient at various levels, then the point-discharge current density should be obtainable from the result of § 9.18.; this is what it might have been hoped would result from measurements of the widths of alti-electrograph traces (see § 9.21.).

More work, both experimental and theoretical, is required in connection with the relationship between point-discharge current and potential gradient; one consequence of such knowledge might be a resolution of the anomaly of the widths of the alti-electrograph traces. If a balloon with a radio-sonde or with a recorder could give, at the same time, measurements of point discharge, as in the alti-electrograph, and of potential gradient, as in the methods of, for example, KOENIGSFELD and PIRAUX (1950) or CURRIE and KREIELSHEIMER (1960), the necessary information might be obtained.

17.12. Precipitation Currents

In stormy weather it is quite clear that the precipitation current to the ground is insignificant in comparison with other currents. The question then arises as to how the precipitation loses any charge it may have carried in the cloud, and how it has acquired the charge it shows on reaching the ground. The general inverse relation between precipitation current and potential gradient, and more particularly the mirror-image effect, show that the precipitation receives its charge from the point-discharge ions, but, while the mirror-image effect suggests that it is quite close to the ground that the charge is acquired, the actual magnitude of the charge per drop is too large for this, if charge is obtained by the ion-capture process of WILSON (1929) (see § 3.11.). The question of time delays in the mirror-image effect (see § 10.27.) must be related to this problem, but is complicated. If it is possible to measure precipitation charges at different heights below the cloud, this would give valuable information.

There is an apparent discrepancy between the results of SMITH (1955) and of HUTCHINSON and CHALMERS (1951) as to the signs of charges on single raindrops. If, as suggested in § 10.17., the discrepancy is just one of sampling, then further measurements, perhaps with even more drops in a short interval of time, and with a lower limit of charge measurement than SMITH's, might solve the problem and verify the surmise that it is the smaller drops which contribute most to the inverse relation.

For conditions of low potential gradient the problems are different, since here it is probable that the precipitation current is the largest current; thus the mirror-image effect in this case has a different origin from that in storms. Of considerable importance to the theory is the difference between the effects for snow as found by CHALMERS (1956) and by REITER (1965), in particular the potential gradient for zero current. It would be of considerable interest if the conclusion could be verified that the direction of the current, or the sign of the potential gradient, above a cloud depends on whether the precipitation is in the form of rain or snow.

17.13. The Nimbo-stratus Cloud

It is a surprising fact that there is little definite knowledge of the electrical structure of the nimbo-stratus cloud which gives continuous precipitation. Because of its much more prominent electrical phenomena, the cumulo-nimbus storm cloud has been much more extensively studied.

It is to be hoped that measurements will be made, for the nimbo-stratus cloud, of the same kind as have been made for the cumulo-nimbus cloud, but with more sensitive apparatus. For example, the alti-electrograph, either in its original form or in the radio-sonde variation, could be used but would require a much longer trailing wire to give sufficient potential difference to produce point discharge; instead of the 20 m wire used in the alti-electrograph, wires of 100 m or more would be needed (but CHAPMAN (1952) mentions the use of 300 m). This would limit the use of the method to places where such a length of wire would not cause damage to electric overhead cables, etc., unless a device could be provided to uncoil the wire when a sufficient height had been reached and to coil it up again on descent.

Measurements, such as those of GISH and WAIT (1950) or STERGIS, REIN and KANGAS (1957b), of potential gradient above the clouds would give an unambiguous answer as to the direction of the current through the cloud, and hence of the direction of the process of separation. From the total current measurements at the ground (§§ 10.22. and 10.24.) it is to be expected that the sign of the potential gradient above the cloud will be different according to whether the precipitation is snow or rain.

One factor that may need consideration, but which has not been discussed in Chapter 13, is that horizontal motion is much greater than vertical motion in connection with nimbo-stratus clouds. Therefore it may be necessary to consider horizontal motion of charge and it would then be no longer possible to use the principle of the quasi-static state in its simple form as applied to horizontally stratified conditions. If there should turn out to be discrepancies between the directions of currents above and below the clouds, it would be worth while to consider the effects of horizontal motion.

Aircraft have flown through thunder clouds and have made measurements in them (GUNN, 1950), so it should not be too difficult to arrange for measurements of potential gradient to be made at places within nimbo-stratus clouds.

Another important feature that requires investigation is whether there is any difference in electrical structure between nimbo-stratus clouds in which the precipitation has been initiated by the ice-crystal mechanism and those in which it has been initiated by coalescence, in particular those in which there can be no ice; the results of SMITH (1951 b), discussed in § 16.7., suggest that there may be a significant difference, and this would have far-reaching consequences in the discussion of the origin of charges in nimbo-stratus clouds, and possibly also in cumulo-nimbus clouds. Whereas warm thunderstorms are rare, and possibly non-existent, warm clouds giving non-thundery rain are well authenticated and sufficiently common for an investigation into their electrical structure to be a feasible proposition in a suitable locality.

17.14. The Lightning Discharge

With the discovery, by CLARENCE and MALAN (1957), of the preliminary processes in the discharge, one of the main problems outstanding, that of the discrepancies in the time scale between visual and electrical observations, has now been solved, and the remaining problems are fewer.

The precise mechanism of the stepped leader is not yet certainly known and there appears to be a discrepancy between the results in South Africa and in England as to the relative positions of the charges taking part in successive strokes of a multiple flash.

Confirmation is desirable of the suggestion of CLARENCE and MALAN (1957) that the first stage of a lightning flash to ground is breakdown between the negative and lower positive charge. If the location of the lower positive charge could be made more precise, that might be useful.

In view of the suggestion of SCHONLAND and MALAN (1954) that the first process of the upward stepped leader from the Empire State Building may take negative charge up to the lower positive charge (see § 14.39.), further investigations of such discharges are desirable. Although only 4 per cent of lightning flashes in England bring down to the ground a positive charge (see § 14.30.) and perhaps even fewer in South Africa, it would be very interesting to have Boys camera photographs of these, identified as such by simultaneous electrical measurements, so as to confirm, or otherwise, the expectation of SCHONLAND and MALAN that the characteristics of positive streamer advance would differ from those of negative streamer advance.

NEWMAN (1958) has made novel suggestions for the investigation of lightning. He proposes to use a schooner travelling at sea under thunder clouds and to fire a wire conductor some 300 m by means of a rocket so that the wire will initiate a discharge from the cloud; the current in the wire can then be investigated in detail in the same way as discharges to high buildings have been investigated, but with the advantage that the conductor has not previously been present to affect the cloud. NEWMAN also proposes to apply a high potential to the lower end of the wire, after the initial discharge, so as to investigate the characteristics of the channel. He also hopes that the radiation from the discharge will act as an artificial atmospherics generator. Preliminary results (NEWMAN, 1965) have shown that many of these hopes are justified, and further results are eagerly awaited.

17.15. Instrumentation

It would appear that there should be two quite separate aims for the recording of results in the future. On the one hand, there are measurements of parameters which do not vary rapidly, and in this case the aim should be to take fairly frequent sample values, convert them into digits, print out these digits on paper or magnetic tape and then analyse them with a computer in whatever way appears desirable. If it were possible to use, not sample values, but average values over continuous periods, this would be an improvement on what has been done hitherto. It would also be an improvement if the different parameters could be sampled at the same time, or, better, averaged over the same period, before being digitalized, instead of the present system of taking samples of the different parameters in turn at different times.

On the other hand, there are cases where there are rapid changes and continuous recording becomes necessary. In certain cases, e.g. lightning phenomena, the changes are exceedingly rapid and a very open time-scale is needed; if recording is carried out continuously, there must be considerable waste of recording material. The alternatives are to use some form of triggering, so that recording starts only when certain conditions, e.g. a high potential gradient, are satisfied, or else to use some form of recording which can be re-used, for example the magnetic tape method of CLARENCE and MALAN (1951), from which the interesting portions can be extracted.

17.16. The Outer Atmosphere

As already pointed out, the electrosphere acts as a complete electrostatic screen and the phenomena of atmospheric electricity are confined to the region between this screen and the earth. Whatever electrical effects may exist further out in the upper atmosphere, they cannot be of any importance to the phenomena discussed in this book.

However, the advent of rockets and other vehicles capable of taking measuring instruments to levels beyond the conducting layers of the electrosphere and ionosphere has now made it meaningful to ask what are the electrical conditions beyond these layers and whether the earth and its atmosphere as a whole carries any charge on the outside of the conducting layers.

At any rate close above what can be termed the top of the ionosphere, the atmosphere still remains a good conductor in which it is probable that any electrical potential differences are small.

17.17. Theoretical Considerations

An important feature of many of the theoretical arguments presented here is the use of what we have termed the principle of the quasi-static state. This serves well in giving a general picture of average conditions if it is permissible to assume the continuous action of the processes concerned; but in most phenomena of atmospheric electricity, the processes do not go on at the same rate for very long and it is often likely that a steady, quasi-static state will not be reached; an example of this is seen in the case of ionization equilibrium (§ 4.36.) and WICHMANN (1952b) has drawn attention to the same difficulty in regard to the thunderstorm. It seems fairly certain that similar considerations will apply in discussing such phenomena as the mirror-image effect (§ 10.25.), and the travel of point-discharge ions below thunder clouds (§ 9.23.). It is clear that the mathematical methods that will be needed will be of considerably greater complexity than those used up to the present, and it may well be that solutions will be obtainable only by numerical computation, probably with electronic computers.

One further point that must not be left out of consideration is that the application of the laws of electricity, in particular those of electrostatics, to phenomena in the atmosphere involves a very wide extrapolation from the small-scale phenomena for which the laws have been experimentally verified, and it is possible that such an extrapolation may not be justified, or, to put it otherwise, that there may exist terms which are quite insignificant in laboratory phenomena but which cannot be neglected on the atmospheric scale. And there might also be the possibility that it is incorrect to treat a quasi-static condition by the laws of static electricity—a similar difference has arisen in the theory of metallic conduction. While there is at present no evidence for any breakdown of the physical laws assumed to govern the phenomena, it is clearly only an assumption, subject to verification, that these laws still hold accurately at the large distances involved; but one cannot, with any profit, predict what deviations from the laws might be needed.

17.18. The Future

The solution of many of the problems set forth in the preceding pages must come from the use of some form of aircraft. The high speed of travel of aeroplanes has advantages from the point of view of rapid sampling of different regions, but at the same time makes it impossible to study any one region in detail. In some cases, the best aircraft to use might be a captive balloon, in others perhaps a helicopter (could this be a gigantic field mill?), and in others again a free balloon; ROSSMANN (1950) has put forward the advantages of a towed glider.

Although aircraft measurements must play a large part in the future work on atmospheric electricity, there is still room for much work by the earth-bound, particularly, perhaps, in the "extraordinary" parts of the world, e.g. mountain tops, arctic regions, deserts, etc.

If and when the questions posed in this chapter may have been answered, it is not to be anticipated that atmospheric electricity will be a completed subject. Every question answered will pose fresh problems, as yet unformulated, but it is safe to predict that some of the problems will be concerned with more precise connections between electrical and meteorological phenomena.

Appendix 1

SINCE many results have been given in terms of electrostatic units, it is convenient to have available a list of conversion factors to the M.K.S. system of units used throughout the present work:

Potential	1 E.S.U. = 330 V (volts)
Charge	1 E.S.U. = $3{\cdot}33 \times 10^{-10}$ C (coulombs)
Capacitance	1 E.S.U. (1 cm) = $1{\cdot}11 \times 10^{-12}$ F (farads)
Resistance	1 E.S.U. = 9×10^{-11} Ω (ohms)
Conductivity	1 E.S.U. = $1{\cdot}11 \times 10^{10}$ Ω^{-1} (mhos)

Appendix 2
Average Values of Results

Quantity	Units	Kew	Average land station	Oceans
Potential gradient	V/m	365	130	126
Air–earth current	A/m^2	$1{\cdot}12 \times 10^{-12}$	$2{\cdot}4 \times 10^{-12}$	$3{\cdot}7 \times 10^{-12}$
Conductivity	$\Omega^{-1}\ m^{-1}$	$3{\cdot}0 \times 10^{-15}$	$1{\cdot}8 \times 10^{-14}$	$2{\cdot}8 \times 10^{-14}$
Columnar resistance	Ω/m^2	$4{\cdot}0 \times 10^{17}$	$1{\cdot}9 \times 10^{17}$	$1{\cdot}2 \times 10^{17}$
Number of nuclei	cm^{-3}	40,000	4,000	4,500 (?)
Positive small ions	cm^{-3}	205	750	640
Negative small ions	cm^{-3}	155	680	575
Rate of production of ions	$cm^{-3}\ sec^{-1}$	10 (?)	9·5	1·5
Space charge	C/m^3	10^{-11}	10^{-11}	?

More detailed results of these and other quantities have been given by ISRAËL (1952).

References

IN THE references, certain collections of papers and articles are referred to several times; the complete references are not given, but an abbreviated form is used. These are:

Thunderstorm Electricity = *Thunderstorm Electricity*, 1953, H. R. BYERS (ed.), Univ. of Chicago Press, Chicago.

Compendium of Meteorology = *Compendium of Meteorology*, 1951, T. F. MALONE (ed.), Amer. Met. Soc., Boston.

Wentworth Conf. = *Proceedings of the Conference on Atmospheric Electricity, held at Wentworth-by-the-Sea, Portsmouth, N.H., May 1954*, R. E. HOLZER and W. A. SMITH (eds.), 1955, Geophys. Res. Dir., Air Force Cambridge Research Center, Bedford, Mass.

Art. Stim. Rain = *Artificial Stimulation of Rain*, 1957, H. WEICKMANN and W. A. SMITH (eds.) (Proceedings of the 1st Conference on Physics of Clouds and Precipitation), Pergamon Press, New York.

Rec. Adv. = *Recent Advances in Atmospheric Electricity*, 1958, L. G. SMITH (ed.) (Proceedings of the 2nd Wentworth Conference, May 1958), Pergamon Press, New York, pp. 1–631.

Problems = *Problems of Atmospheric and Space Electricity*, 1965, S. C. CORONITI (ed.) (Proceedings of 3rd International Conference on Atmospheric and Space Electricity, Montreux, May 1963), Elsevier, Amsterdam, pp. 1-616.

Tokyo Conf. = *International Conference on Cloud Physics, Tokyo and Sapporo, May–June, 1965.*

The numbers in brackets after each reference give the section in which the work is referred to.

ACCARDI, L. (1952). Nuovo apparacchio per la registrazione del campo elettrico atmosferico, *Goefis. Pur. Appl.* **22**, 84–8. [5.25., 5.27.]

ADAMSON, J. (1960). The compensation of the effects of potential-gradient variations in the measurement of the atmospheric air–earth current, *Quart. J.R. Met. Soc.* **86**, 252–8. [5.25., 5.27., 8.8., 10.4.]

ADKINS, C. J. (1959a). Measurements of the atmospheric potential gradient on a Canadian glacier, *Quart. J.R. Met. Soc.* **85**, 60–4. [5.65., 5.66.]

ADKINS, C. J. (1959b). The small ion concentration and space charge near the ground, *Quart. J.R. Met. Soc.* **85**, 237–52. [3.24., 4.12., 6.11.]

References

AITKEN, J. (1880). On dust, fog and clouds, *Trans. R. Soc. Edinb.* **30**, 337–68. *Collected Scientific Papers*, Cambridge Univ. Press, 1923, pp. 34–64. [4.3., 4.6.]

AITKEN, J. (1888–9). On improvements in the apparatus for counting the dust particles in the atmosphere, *Proc. R. Soc. Edinb.* **16**; *Collected Scientific Papers*, pp. 207–35. [4.7.]

D'ALIBARD, T. F. (1752). Letter to Acad. de. Sci. [1.3.]

ALIVERTI, G. and LOVERA, G. (1950). Über einige in Oberitalien beobachtete Kugelblitze, *Arch. Met., Wien* A, 3, 77–83. [14.56.]

ALLEE, P. A. and PHILLIPS, B. B. (1959). Measurements of cloud droplet charge, electric field and polar conductivities in supercooled clouds, *J. Met.* **16**, 405–510. [13.7.]

ALLIBONE, T. E. (1948). Development of the spark discharge, *Nature, Lond.* **161**, 970–1. [14.11.]

ALLSOPP, H. L. (1952). An investigation of point-discharge currents, *S. Afr. J. Sci.* **48**, 244–6. [9.9.]

ANDERSEN, W. H. (1965). Energy source for ball lightning, *J. Geophys. Res.* **70**, 1291–3. [14.57.]

ANDERSON, R. V. (1965). In Discussion, *Problems*, p. 143. [13.3.]

ANDREE, S. A. (1890). Observations faites au Cap Thordsen, *Met. Z.* **7**, 29–32. [5.56.]

ANGENHEISTER, G. (1914). Die luftelektrischen Beobachtungen am Samoa-Observatorium, 1912–13, *Nachr. Ges. Wiss., Göttingen*, 191–206. [5.40.]

APPLEMAN, R. (1957). *J. Met.* **14**, 89–90. [12.26.]

APPLETON, E. V. (1925). Discussion on ionisation in the atmosphere, *Proc. Phys. Soc., Lond.* **37**, 48D–50D. (11.7.]

APPLETON, E. V. and BOWEN, E. G. (1933). Source of atmospherics and penetrating radiation, *Nature, Lond.* **132**, 965. (12.31.]

APPLETON, E. V. and CHAPMAN, F. W. (1937). On the nature of atmospherics. IV, *Proc. Roy. Soc.* A, **158**, 1–22. [14.9.]

APPLETON, E. V., WATSON WATT, R. A. and HERD, J. F. (1926). On the nature of atmospherics. III, *Proc. Roy. Soc.* A, **111**, 654–77. [12.7., 15.1.]

ARABADJI, V. I. (1956). Ball lightning (in Russian), *Sci. Notes Gorky Minsk Pedagog. Inst.*, No. 5. [14.57.]

ARABADJI, V. I. (1959) On the electrical attributes of storm precipitation(in Russian), *Akad. Nauk SSSR* **127**, 298–301. [10.5., 10.8., 10.10.]

ARNOLD, H. R., PIERCE, E. T. and WHITSON, A. L. (1965). The effect of a living tree upon the fair weather potential gradient, *J. Atmosph. Terr. Phys.* **27**, 429–30. [5.18.]

ATLAS, D. (1958). Radar lightning echoes and atmospherics in vertical cross-section, *Rec. Adv.*, pp. 441–59. [14.38.]

VAN ATTA, L. C., NORTHRUP, D. L., VAN ATTA, C. M. and VAN DE GRAAFF, F.J. (1936). The design, operation and performance of the Round Hill electrostatic generator, *Phys. Rev.* **49**, 761–76. [5.25., 5.27.]

AULT, J. O. and MAUCHLY, S. J. (1926). Ocean magnetic and electric observations obtained aboard the *Carnegie*, 1915–21, *Pub. Carnegie Inst.* **175**, 197–286. [11.19.]

AUSTIN, L. W. (1921). Determination of the direction of atmospheric disturbances or static in radio telegraphy, *J. Franklin Inst.* **5** (191) 619–20. [15.4.]

AWE, O. (1964). Errors in correlation between time series, *J. Atmosph. Terr. Phys.* **26**, 1239–55. [2.41.]

BALDIT, A. (1911). Sur les charges électriques de la pluie au Puy-en-Veley, *C. R. Acad. Sci., Paris* **152**, 807–9; **154**, 731-3. [10.2., 10.8.]

BALDIT, A. (1912). Nouvelles observations sur les charges électriques de la pluie en 1911 au Puy-en-Velay, *Radium, Paris* **9**, 92–101. [10.2., 10.8.]

BANERJI, S. K. (1932). The electric field of overhead thunderstorms, *Phil. Trans.* A, **231**, 1–27. [10.2., 10.8.]

BANERJI, S. K. (1938). Does thunderstorm rain play any part in the replenishment of the earth's negative charge? *Quart. J. R. Met. Soc.* **64**, 293–9. [10.2., 10.8.]

BANERJI, S. K. and LELE, S. R. (1952). Electric charges of rain-drops, *Proc. Nat. Inst. Sci., India* **18**, 93–124. [10.5., 10.10., 10.12., 10.14.]

BARKLIE, R. H. D. (1956). Values of combination coefficients. Unpublished. [4.34.]

BARKLIE, R. H. D., WHITLOCK, W. and HABERFIELD, G. (1958). Observations on the reactions between small ions and (a) cloud droplets, (b) Aitken nuclei, *Rec. Adv.*, pp. 223–9. [3.23., 4.34., 13.7.]

BARLOW, J. S., FREY, G. W. and NEWMAN, J. B. (1954). Very low frequency noise power from the lightning discharge, *J. Franklin Inst.* **258**, 187–203. [14.17.]

BARNARD, V. (1951). The approximate mean height of the thundercloud charges taking part in a flash to ground, *J. Geophys. Res.* **56**, 33–5. [12.8.]

BARRÉ, M. (1953). Propriétés électriques du blizzard, *Ann. Geophys.* **9**, 164–83. [10.29.]

BATTAN, L. J. (1953). Formation and spread of precipitation in convective clouds, *J. Met.* **10**, 311. [3.6.]

BAUER, L. A. (1920). Vertical electric currents and the relation between terresttrial magnetism and atmospheric electricity, *Terr. Magn. Atmos. Elect.* **25**, 145–62. [8.29.]

BECCARIA, G. B. (1775) *Dell' elettricita terrestre atmosferica a cielo sereno*, Turin. [1.5., 5.3.]

BEHACKER, M. (1910). Zur Berechnung des Erdfeldes unter der Voraussetzung homogener Ionisierung der Atmosphäre, *S.B. Akad. Wiss., Wien* **119**, 675–84. [8.19.]

BELIN, R. E. (1948). A radio-sonde method for atmospheric potential gradient measurements, *Proc. Phys. Soc., Lond.* **60**, 381–7. [9.13.]

BENNDORF, H. (1900). Über die Störungen des normale atmosphärischen Potentialgefälles durch Bodenerhebungen, *S.B. Akad. Wiss., Wien* **110**, 923–40. [5.18.]

BENNDORF, H. (1902). Über ein mechanisch registrierendes Elektrometer für luftelektrische Messung, *S.B. Akad. Wiss., Wien* **111**, 457–512. [2.51.]

BENNDORF, H. (1906). Über gewisse Störungen des Erdfeldes mit Rücksicht auf die Praxis luftelektrischer Messungen, *S.B. Akad. Wiss., Wien* **115**, 425–56. [5.18.]

BENNDORF, H. (1910). Luftelektrische Beobachtungen an Graz, *S.B. Akad. Wiss., Wien* **119**, 89–110. [10.2., 10.13.]

BENNDORF, H. (1912). Über die Beobachtungen der Niederschlagselektrizität, *S.B.Bayer. Akad. Wiss.*, 402–29. [10.3.]

BENT, A. E. (1955). Variations in atmospheric electricity at Mount Washington, *Mt. Wash. Obs. News. Bull.* **27**, 1–8. [5.14., 5.71.]

BENT, R. B. (1964). The testing of apparatus for ground fair-weather space charge measurements, *J. Atmosph. Terr. Phys.* **26**, 313–8. [6.5.]

BENT, R. B., COLLIN, H. L., HUTCHINSON, W. C. A. and CHALMERS, J. A.(1965). Space charges produced by point discharge from trees during a thunderstorm, *J. Atmosph. Terr. Phys.* **27**, 67–72. [6.14., 9.14.]

BENT, R. B. and HUTCHINSON, W. C. A. (1965). Electric space charges over melting snow on the ground, *J. Atmosph. Terr. Phys.* **27**, 91–9. [3.19., 6.14.]

BENT, R. B. and HUTCHINSON, W. C. A. (1966). Electric space charge measurements and the electrode effect within the height of a 21 m mast. *J. Atmosph. Terr. Phys.* **28**, 53–73 [2.31., 6.8., 6.15., 6.16.]

BERGER, K. (1961). Gewitterforschung auf dem Monte San Salvatore, *Elektrotech. Z.* **82**, 249-60. [14.23.]

BERGERON, T. (1933). On the physics of cloud and precipitation, *Gen. Ass. Int. Un. Geod. (Lisbon)*, pp. 156-70. [3.5.]

BERNDT, G. (1913). Luftelektrische Beobachtungen in Argentinien, *Met. Z.* **30**, 363–6. [10.2.]

BEST, A. C. (1943). *The Size Distribution of Rain Drops*, Met. Res. Ctee. M.R.P., 352. [10.16.]

BEST, A. C. (1950a). The size distribution of rain drops, *Quart. J. R. Met. Soc.* **76**, 16–36, [10.16., 10.20.]

BEST, A. C. (1950b). Empirical formula for the terminal velocity of water drops falling through the atmosphere, *Quart. J. R. Met. Soc.* **76**, 302–11. [10.20.]

BEST, A. C. (1953). The relation between rain current and rate of rainfall, *J. Atmosph. Terr. Phys.* **3**, 285–8. [10.20.]

BLAKE, L. J. (1883). Über Elektrizitätsentwicklung bei der Verdampfung und über die elektrische Neutralität des von ruhigen elektrisierten Flüssigkeitsoberflächen aufsteigenden Dampfes, *Ann. Phys. Chem.* **19**, 518 [3.21.]

BLANCHARD, D. C. (1961). The electrification of the atmosphere by particles from bubbles in the sea, *Woods Hole Oceanographic Institute*, Reference 61–9. [4.25., 8.17.]

BLANCHARD, D. C. (1964). Charge separation from saline drops on hot surfaces, *Nature, Lond.* **201**, 1164–6. [3.21.]

BLANCHARD, D. C. (1965). Charge separation at the surface of the sea, *Tokyo Conf.*, pp. 324-8. [5.55.]

BOURDEAU, R. E. WHIPPLE, E. C. and CLARK, J. F. (1959). Analytic and experimental electrical conductivity between the stratosphere and the ionosphere, *J. Geophys. Res.* **64**, 1363–70. [7.15.]

BOWEN, E. G. (1950). The formation of rain by coalescence, *Austral. J. Sci. Res.* A, **3**, 193. [3.6.]

BOYS, C. V. (1926). Progressive lightning, *Nature, Lond.* **118**, 749–50.[14.2., 14.38.]

BOYS, C. V. (1929). Progressive lightning, *Nature, Lond.* **124**, 44–5. [14.2.]

BRADBURY, N. E. and MEURON, H. J. (1938). The condensation nuclei, *Terr. Magn. Atmos. Elect.* **43**, 231–40. [4.8.]

BRAND, W. (1923). *Der Kugelblitz*, Henri Grand, Hamburg. [14.57.]

BRASEFIELD, C. J. (1959). Some observations of low-level ion clouds, *J. Geophys. Res.* **64**, 141–8. [6.7., 6.13., 6.15.]

BREWER, A. W. (1953). An electrometer valve voltmeter of wide range, *J. Sci. Instrum.* **30**, 91–2. [5.12., 5.13., 5.17., 6.7.]

BRICARD, J. (1949). L'èquilibre ionique de la basse atmosphère, *J.Geophys. Res.* **54**, 39–52. [4.29., 4.31., 7.18.]

BRILLOUIN, M. (1897). L'èlectricité atmosphèrique, *Ecl. Élect.* **12**, 577–99. [16.22.]

BROOK, M., ARMSTRONG, G., WINDER, R. P. H., VONNEGUT, B. and MOORE C. B. (1961). Artificial initiation of lightning discharges, *J. Geophys. Res.* **66**, 3967–9. [14.48.]

BROOK, M. and KITAGAWA, N. (1960a) Some aspects of lightning activity and related meteorological conditions, *J. Geophys. Res.* **65**, 1203–10. [14.22., 14.30.]

BROOK, M. and KITAGAWA, N. (1960b). Electric-field changes and the design of lightning-flash counters, *J. Geophys. Res.* **65**, 1927–32. [14.37.]

BROOK, M. and KITAGAWA N. (1964). Radiation from lightning discharges in the frequency range 400 to 1000 Mc/s, *J. Geophys. Res.* **69**, 2431–4. [15.9.]

BROOK, M., KITAGAWA, N. and WORKMAN, E. J. (1962). Quantitative study of strokes and continuing currents in lightning discharges to ground, *J. Geophys. Res.* **67**, 649–59. [14.23., 14.24.]

BROOK, M. and VONNEGUT, B. (1960). Visual confirmation of the junction process in lightning discharges, *J*; *Geophys. Res.* **65**, 1302–3. [14.25.]

BROOKS, C. E. P. (1925). The distribution of thunderstorms over the globe, *Geophys. Mem., Lond.* **24**. [11.7., 11.14., 11.19.]

BROWN, J. G. (1930). The relation of space charge and potential gradient to the diurnal system of convection in the lower atmosphere, *Terr. Magn. Atmos. Elect.* **35**, 1–15. [6.5., 6.8.]

BROWN, J. G. (1935). The local variation of the earth's electric field, *Terr. Magn. Atmos. Elect.* **40**, 413–24. [5.45.]

BROWN, J. G. (1936). The effect of wind upon the earth's electric field at the surface, *Terr. Magn. Atmos. Elect.* **41**, 279–85. [5.45.]

BROWN, J. G. (1937). The effect of overcast sky upon the local diurnal variation of the earth's electric field, *Terr. Magn. Atmos. Elect.* **42**, 163–4. [5.45.]

BROWNE, I. C. (1951). A radar echo from lightning, *Nature, Lond.* **167**, 438. [14.49.]

BROWNE, I. C., PALMER, H. P. and WORMELL, T. W. (1954). The physics of rainclouds, *Quart. J. R. Met. Soc.* **80**, 291–327. [3.3., 3.6.]

BRUCE, C. E. R. (1944). The initiation of long electrical discharges, *Proc. Roy. Soc.* A, **183**, 228–42. [14.13.]

BRUCE, C. E. R. (1959). Cosmic thunderstorms, *J. Franklin Inst.* **268**, 425–45. [12.35.]

BRUCE, C. E. R. and GOLDE, R. H. (1942). The lightning discharge, *J. Instn. Elect. Engrs.* **88**, Pt. II, 487–505. [14.17., 14.19., 14.20., 14.21.,14.28., 14.39., 14.45.]

BRYANT, G. W. and FLETCHER, N. H. (1965). Thermoelectric effects in ice, *Tokyo Conf.*, 357–60. [3.12.]

BUDDEN, K. G. (1951). The propagation of a radioatmospheric, *Phil. Mag.* **42**, 1–19. [15.3.]

BUILDER, G. (1929). The effect of condensation nuclei in atmospheric-electric observations, *Terr. Magn. Atmos. Elec.* **34**, 281–6. [4.35.]

BUREAU, R. (1931). Électricité atmosphèrique - un radiogoniomètre enregistreur; son application aux parasites atmosphèriques, *C. R. Acad. Sci., Paris* **192**, 170–2. [15.4.]

BYERS, H. R. (ed.) (1953). *Thunderstorm Electricity*, Univ. of Chicago Press, Chicago. [Pref.]

BYERS, H. R. (1965). The relation of lightning and thunderstorms to meteorological conditions, *Problems*, pp. 491–6. [3.8.]

BYERS, H. R. and BRAHAM, R. R. (1953). Thunderstorm structure and dynamics, *Thunderstorm Electricity*, pp. 46–65. [3.6., 3.8.]

LE CADET, G. (1898). Étude de champ électrique de l'atmosphère, *Ann. Univ. Lyons* **35**. [1.12.]

CAIRNS, J. E. I. (1933). Penetrating radiation from thunderclouds, *Nature, Lond.* **132**, 174. [12.31.]

CALLAHAN, R. C., CORONITI, S. C., PARZIALE, A. J. and PATTEN, R. (1951). Electrical conductivity of air in the troposphere, *J. Geophys. Res.* **56**, 545–51. [7.14., 7.18.]

CATON, P. G. F. and PIERCE, E. T. (1952). The waveforms of atmospherics, *Phil. Mag.* **43**, 393–409. [15.3.]

CHALMERS, J. A. (1939). A note on theories of the electric field below clouds, *Quart. J. R. Met. Soc.* **65**, 237–43. [9.23.]

CHALMERS, J. A. (1943). The separation of electricity in clouds, *Phil. Mag.* **34**, 63–7. [16.28.]

CHALMERS, J. A. (1944). Electric fields below clouds, *Quart. J. R. Met. Soc.* **70**, 121–7. [9.18., 9.23.]

CHALMERS, J. A. (1946). The ionisation in the lowest regions of the atmosphere, *Quart. J. R. Met. Soc.* **72**, 199–205. [7.9., 8.20.]

CHALMERS, J. A. (1947). The capture of ions by ice particles, *Quart. J. R. Met. Soc.* **73**, 324–34. [3.11., 16.15.]

CHALMERS, J. A. (1949). *Atmospheric Electricity*, Clarendon Press, Oxford. [4.35., 11.16.]

CHALMERS, J. A. (1951). The origin of the electric charge on rain, *Quart. J. R. Met. Soc.* **77**, 249–59. [9.24., 10.14., 10.20.]

CHALMERS, J. A. (1952a) Negative electric fields in mist and fog, *J. Atmosph. Terr. Phys.* **2**, 155–9. [4.17., 5.23., 5.59., 6.16.]

CHALMERS, J. A. (1952b). The relation between point discharge current and field, *J. Atmosph. Terr. Phys.* **2**, 292–300 [9.17.]

CHALMERS, J. A. (1952c). Point discharge currents, *J. Atmosph. Terr. Phys.* **2**, 301–5. [9.25., 11.16.]

CHALMERS, J. A. (1952d) Electric charges from ice friction, *J. Atmosph. Terr. Phys.* **2**, 337–9. [3.13.]

CHALMERS, J. A. (1953a) Atmospheric electric conductivity near the earth's surface, *J. Atmosph. Terr. Phys.* **3**, 223–4, [7.8.]

CHALMERS, J. A. (1953b). The charge on the ionosphere, *J. Atmosph. Terr. Phys.* **3**, 345–6. [11.5.]

CHALMERS, J. A. (1953c). The effective separation of discharging points, *J. Atmosph. Terr. Phys.* **3**, 346–7. [9.24.]

CHALMERS, J. A. (1953d). Induced charges and the fine-weather field, *J. Atmosph. Terr. Phys.* **4**, 84–5. [11.6.]

CHALMERS, J. A. (1953e). The agrimeter for continuous recording of the atmospheric electric field, *J. Atmosph. Terr. Phys.* **4**, 124–8. [5.29., 8.8.]

CHALMERS, J. A. (1954). Atmospheric electricity, *Rep. Progr. Phys.* **17**, 101–34. [4.6.]

CHALMERS, J. A. (1956). The vertical electric current during continuous rain and snow, *J. Atmosph. Terr. Phys.* **9**, 311–21. [10.4., 10.17., 10.22., 10.24., 11.20., 13.2., 13.3., 17.12.]

CHALMERS, J. A. (1957a). Effects of condensation nuclei in atmospheric electricity, *Geofis. Pur. Appl.* **36**, 211–17. (5.46., 8.23.]

CHALMERS, J. A. (1957b). Point-discharge current, potential gradient and wind speed, *J. Atmosph. Terr. Phys.* **11**, 301–2. [9.16.]

CHALMERS, J. A. (1961). A criterion for thunderstorm theories, *J. Atmosph. Terr. Phys.* **21**, 174–6. [16.34.]

CHALMERS, J. A. (1962a). The measurement of vertical electric current in the atmosphere, *J. Atmosph. Terr. Phys.* **24**, 297–302. (8.12.]

CHALMERS, J. A. (1962b). The relation of point-discharge current to potential gradient and wind-speed, *J. Atmosph. Terr. Phys.* **24**, 339–44. [9.19.]

CHALMERS, J. A. (1962c). Point-discharge currents through a living tree during a thunderstorm, *J. Atmosph. Terr. Phys.* **24**, 1059–63. [9.14.]

CHALMERS, J. A. (1963). Taking the mystery out of Gauss's theorem, *Contemporary Phys.* **4**, 456-8. [2.4.]

CHALMERS, J. A. (1964). On the conductivity of the air in thunderstorms, *J. Geophys. Res.* **69**, 357–9 and 362. [7.11., 7.20., 12.14.]

CHALMERS, J. A. (1965a) The relation between precipitation current and potential gradient, *J. Atmosph. Terr. Phys.* **27**, 899–905 [10.27.]

CHALMERS, J. A. (1965b) The effect of wind on point-discharge pulses, *J. Atmosph. Terr. Phys.* **27**, 1037–38 [9.4.]

CHALMERS, J. A. and LITTLE, E. W. R. (1940). The electricity of continuous rain, *Terr. Magn. Atmos. Elect.* **45**, 451–62. [9.23., 10.4., 10.8., 10.13., 10.14.]

CHALMERS, J. A. and LITTLE, E. W. R. (1947). Currents of atmospheric electricity, *Terr. Magn. Atmos. Elect.* **52**, 239–60. (8.6., 9.9., 10.4., 10.6., 10.11., 10.17., 10.22., 10.23., 11.10., 11.12., 11.13., 11.16.]

CHALMERS, J. A. and MAPLESON, W. W. (1955). Point discharge currents from a captive balloon, *J. Atmosph. Terr. Phys.* **6**, 149–59. [9.11.,9.13.,9.16., 9.17.]

CHALMERS, J. A. and PASQUILL, F. (1938). The electric charges on single raindrops and snowflakes, *Proc. Phys. Soc., Lond.* **50**, 1–16. [10.5., 10.8., 10.10., 10.23.]

CHAPMAN, F. W. (1939). Atmospheric disturbances due to thundercloud discharge. I, *Proc. Phys. Soc., Lond.* **51**, 876-94. [14.11., 14.27.]

CHAPMAN, S. (1952). *Thundercloud Electrification Studies.* II, Report No. V.C.-603-P-1 to U.S. Office of Naval Research, Cornell Aeronautical Lab., Buffalo, N.Y. [9.15., 9.22., 16.21., 16.30., 17.13.]

CHAPMAN, S. (1953). Thundercloud electrification in relation to rain and snow particles, *Thunderstorm Electricity*, pp. 207–30. [3.13., 9.21., 9.22., 12.15., 12.18.]

CHAPMAN, S. (1956). *Electrostatic Field Measurements, Corona Discharge and Thunderclouds*, CAL Report **68**, Cornell Aeronautical Lab., Buffalo, N.Y. [5.54., 9.15., 9.16.]

CHAUVEAU, B. (1900). Études de la variation de l'électricité atmosphérique, *Annales de B.C.M.* **5**, 1. [1.10., 5.65.]

CHAUVEAU, B. (1902). Recherches sur l'électricité atmosphérique, *Mem.* **2**, 103. [6.4.]

CHAUVEAU, B. (1922, 1924, 1945). *Électricité Atmosphérique*, Gaston Doin et Cie, Paris. [1.17., 5.37., 5.42., 5.54.]

CHIPLONKAR, M. W. (1940). Measurement of point discharge current during disturbed weather at Colaba, *Proc. Indian. Acad. Sci.* A, **12**, 50. [9.9., 9.13., 9.15.]

CLARENCE, N. D. (1965). Whistlers as a phenomenon to study space electricity, *Problems*, pp. 514–27. [14.33., 15.11.]

CLARENCE, N. D. and MALAN, D. J. (1951). Magnetic tape recording of electrostatic field changes, *Proc. Phys. Soc., Lond.* B, **64**, 529. [2.51., 14.4., 17.15.]

CLARENCE, N. D. and MALAN, D. J. (1957). Preliminary discharge processes in lightning flashes to ground, *Quart. J. R. Met. Soc.* **83**, 161–72. [12.27., 14.9., 14.10., 16.35., 17.14.]

CLARK, J. F. (1949). Rocket applications of electrostatic generating voltmeters, *Instruments* **22**, 1007. [5.25., 5.27.]

CLARK, J. F. (1957). Airborne measurement of atmospheric potential gradient, *J. Geophys. Res.* **62**, 617–28. [5.38., 11.3., 11.4.]

CLARK, J. F. (1958). The fair-weather atmospheric electric potential and its gradient, *Rec. Adv.*, pp. 61–73. [11.4.]

COLE, R. K. and PIERCE, E. T. (1965). Electrification in the earth's atmosphere for altitudes between 0 and 100 kilometers, *J. Geophys. Res.* **70**, 2735–49. [4.43.]

COLLADON, D. (1826). Ablenkung der Magnetnadel durch den Strom einer gewöhnlichen Elektrisier-Maschine und der atmosphärischen Elektrizität, *Ann. Phys. Chem.* **8**, 336–52. [9.6.]

COLLIN, H. L. (1962). Sign discrimination in field mills, *J. Atmosph. Terr. Phys.* **24**, 743–5. [5.27.]

COLLIN, H. L. and RAISBECK, H. A. (1964). Errors in recorded precipitation current caused by splashing, *J. Atmosph. Terr. Phys.* **26**, 1107–13. [10.26.]

CORONITI, S. C. (1960). Airborne conductivity meter, *Geofis. Pur. Appl.* **47**, 79–83. [7.3.]

CORONITI, S. C. (1961). The effect of aerosols on the electrical conductivity of the atmosphere, *Geofis. Pur. Appl.* **48**, 124–8. [7.14.]

CORONITI, S. C. (ed.) (1965). *Problems of Atmospheric and Space Electricity*, pp. 1–616, Proceedings of 3rd International Conference on Atmospheric and Space Electricity, Montreux, May 1963, Elsevier, Amsterdam[1.21., 1.23.]

CORONITI, S. C. and HEATON, E. (1953). The electrical conductivity of the atmosphere over the Pacific Ocean, *Trans. Amer. Geophys. Un.* **34**, 833–5. [7.14.]

CORONITI, S. C., NAZAREK, A., STERGIS, C. G., KOTES, D. E., SEYMOUR, D. W. and WERME, J. V. (1954). Balloon borne conductivity meter, *Instrumentation for Geophys. Res.* **3**, 1–9, Geophys. Res. Dir., AFCRC 1, Cambridge, Mass. [7.3.]

CORONITI, S. C., PARZIALE, A. J., CALLAHAN, R. C. and PATTEN, R. (1952). Effect of aircraft charge on airborne conductivity measurements, *J. Geophys. Res.* **57**, 197–205. [7.14.]

COULIER, M. (1875). Note sur une nouvelle propriété de l'air, *J. Pharm. Chim., Paris* **22**, 165. [4.6.]

COULOMB, C. A. (1795). *Mem. Acad. Sci., Paris*, p. 616. [1.14., 4.1.]

CROSS, A. S. (1953). Two electrostatic field meters, *Brit. J. Appl. Phys.*, Supp. **2**, S. 47–S. 50. [5.25., 5.27.]

CROZIER, W. D. (1963a). Electrode effect during night-time low-wind conditions, *J. Geophys. Res.* **68**, 3451–8. [6.8., 8.19.]

CROZIER, W. D. (1963b). Measuring atmospheric potential with passive antennas, *J. Geophys. Res.* **68**, 5427–9. [5.17.]

CROZIER, W. D. (1964). The electric field of a New Mexico dust devil, *J. Geophys. Res.* **69**, 5427–9. [5.62.]

CROZIER, W. D. (1965). Atmospheric electrical profiles below three metres, *J. Geophys. Res.* **70**, 2785–92. [6.14.]

CURRIE, D. R. and KREIELSHEIMER, K. S. (1960). A double field mill for the measurement of potential gradients in the atmosphere, *J. Atmosph. Terr. Phys.* **19**, 126–35. [5.32., 17.11.]

CURTIS, H. O. and HYLAND, M. C. (1958). Aircraft measurements of the ratio of negative to positive conductivity, *Rec. Adv.*, pp. 111–17. [7.18.]

CURTIS, H. W. (1958). The nature of lightning discharges which accompany whistlers, *Rec. Adv.*, pp. 619–23. [15.11.]

DAHL, H. (1951). On the sensitivity of generating electrostatic D.C. voltmeters, *Beretn. Michelsens Inst., Vidensk* **14**, [5.25.]

DAUNDERER, A. (1907). Luftelektrische Messungen, *Phys. Z.* **8**, 281–6. [6.7 6.8.]

DAUNDERER, A. (1909). Über die in den unteren Schichten der Atmosphäre vorhandene freie elektrische Raumladung, *Phys. Z.* **10**, 113–8. [6.7., 6.8.]

DAUZÉRE, C. and BOUGET, J. (1928). Sur l'ionisation intense de l'air dans les lieux fréquement foudroyés, *C. R. Acad. Sci., Paris* **186**, 1744–5.[14.50.]

DAVIS, M. H. (1965). The effect of electric charges and fields on the collisions of very small cloud drops, *Tokyo Conf.*, pp. 118–20. [3.6.]

DAVIS, R. and STANDRING, W. G. (1947). Discharge currents associated with kite balloons, *Proc. Roy. Soc.* A, **191**, 304–22. [5.68., 9.11., 9.16., 14.40., 14.44.]

DEMON, L. (1953). Premiers resultats obtenus au cours de printemps, 1953, *Journ. Res. C.N.R.S.* **24**, 126–7. [5.32.]

DIECKMANN, M. (1912). Experimentelle Untersuchungen aus dem Grenzgebiet zwischen drahtloser Telegraphie und Luftelektrizität, *Luftfahrt u. Wiss.* **1**, 59. [9.6.]

DINGER, J. E. (1964). Electrification accompanying melting of ice and snow, *Quart. J. R. Met. Soc.* **90**, 208. [3.19.]

DINGER, J. E. and GUNN, R. (1956). Electrical effects associated with a change of state of water, *Terr. Magn. Atmos. Elect.* **51**, 477–94. [3.19., 16.36., 16.39.]

DODSON, W. B. (1950). A discussion of methods for the measurement of space charge density. Unpublished report from Res. Lab. American Air Filter Co. Inc., Louisville, Kentucky. [6.4.]

DOLEZALEK, F. (1901). Über ein einfaches und empfindliches Quadrantelektrometer, *Z. InstrumKde*, **21**, 345–50. [2.45.]

DOLEZALEK, H. (1956). *On Antennas for Atmospheric Electrical Measurements*, Report on AF (61) 514–640. Tech. Note 4. [5.15.]

DOLEZALEK, H. (1958). Problems in atmospheric electric synoptic investigations, *Rec. Adv.*, pp. 195–212. [2.35.]

DOLEZALEK, H. (1960a). Zur Berechnung des luftelektrischen Stromkreises I. Ist das Schönwetterfeld positiv oder negativ zu nennen? *Geofis. Pur. Appl.* **45**, 267–72. [2.7., 2.9.]

DOLEZALEK, H. (1960b). Zur Berechnung des luftelektrischen Stromkreises II. Über die Gültigkeit des Ohmschen Gesetzes in der Atmosphäre, *Geofis. Pur. Appl.* **45**, 273–97. [2.30.]

DOLEZALEK, H. (1960c). Zur Methodik luftelektrischer Messungen. VI. Das Ersatzschaltbild für den radioaktiven Kollektor zur Messung der luftelektrischen Feldstärke, *Beitr. Geophys.* **69**, 87–94. [5.4.]

DOLEZALEK, H. (1960d). Zur Berechnung des luftelektrischen Stromkreises III. Kontrolle des Ohmschen Gesetzes durch Messung, *Geofis. Pur. Appl.* **46**, 125–44. [8.15.]

DOLEZALEK, H. (1962). *Atmospheric Electric Parameter Study Survey on an Effect Relating Atmospheric Electric Variations with Formation and Dissipation of Fog, Tech. Rep. Contr.* NONR 3388(00), (AVCO Corp. Res. and Adv. Dev. Dis). [7.12.]

DOLEZALEK, H. (1963). Passive antenna and collector antenna for the measurement of the atmospheric electric potential, *J. Geophys. Res.* **68**, 5181. [5.17.]

DOLEZALEK, H. (1964). *Study of Atmospheric Electric Tracing of High-altitude Irregularities*, pp. 1–61, Final Report Contract AF 19 (628)–3289, AVCO Corp., Wilmington Mass. (5.56.]

DORNO, C. (1911). *Studie über Licht und Luft des Hochgebirges*, J. Vieweg und Sohn, Braunschweig, Verlag. [8.21., 8.24.]

DOYLE, A., MOFFETT, D. R. and VONNEGUT, B. (1964). Behaviour of evaporating electrically charged droplets, *J. Coll. Sci.* **19**, 136–43. [3.22.]

EBERT, H. (1901). Aspirations apparat zur Bestimmung des Ionengehaltes der Atmosphäre, *Phys. Z.* **2**, 662–6. [2.54., 4.11.]

EBERT, H. (1902). Galvanometrische Messung des elektrischen Ausgleichs zwischen den Ionenladungen der Atmosphäre und der Ladung der Erdoberfläche, *Phys. Z.* **3**, 338–9. [8.4.]

EBERT, H. (1904). Über die Ursache des normalen atmosphärischen Potentialgefälles und der negativen Erdladung, *Phys. Z.* **5**, 135–40. [4.17., 4.24.]

EBERT, H. and LUTZ, C. W. (1908.) Die Freiballon im elektrischen Felde der Erde, *Beitr. Phys. Frei. Atmos.* **2**, 183–204. [5.37.]

ECCLES, W. H. (1912). On the diurnal variations of the electric waves occurring in nature, and on the propagation of electric waves round the bend of the earth, *Proc. Roy. Soc.* A, **87**, 79–99. [15.1.]

ECCLES, W. H. and AIREY, H. M. (1911). Note on the electric waves occurring in nature, *Proc. Roy. Soc.* A, **85**, 145–50. [15.1.]

EIGEN, M. and DE MAEYER, L. (1958). Self-dissociation and protonic charge transport in water and ice, *Proc. Roy. Soc.* A, **247**, 505–33. [3.12.]

EINSTEIN, A. (1905). Über die von der molekularkinetischen Theorie der Wärme geforderte Bewegung von in ruhenden Flüssigkeiten suspendierten Teilchen, *Ann. Phys., Lpz.* **17**, 549–60. [2.14., 7.18.]

ELSTER, J. and GEITEL, H. (1885). Über die Elektrizitätsentwicklung bei der Regenbildung, *Ann. Phys. Chem.* **25**, 121–31. [16.12., 16.14.]

ELSTER, J. and GEITEL, H. (1888). Über eine Methode, die elektrische Natur der atmosphärischen Niederschläge zu bestimmen, *Met. Z.* **5**, 95–100. [10.2., 10.8., 10.13., 10.23.]

ELSTER, J. and GEITEL, H. (1899a). Über die Existentz elektrischer Ionen in der Atmosphäre, *Terr. Magn. Atmos. Elect.* **4**, 213–34. [1.14., 4.1.]

ELSTER, J. and GEITEL, H. (1899b). Beiträge zur Kenntnis der atmosphärischen Elektrizität, *Phys. Z.* **1**, 245–9. [1.14., 4.1., 7.2.]

ELSTER, J. and GEITEL, H. (1901). Über eine fernere Analogie in dem elektrischen Verhalten der natürlichen und der durch Becquerel-Strahlen abnorm leitend gemachten Luft, *Phys. Z.* **2**, 590–3. [4.19.]

ELSTER, J. and GEITEL, H. (1913). Zur Influenztheorie der Niederschlagselektrizität, *Phys. Z.* **14**, 1287–92. [16.12., 16.14.]

ERMAN, P. (1804). Observations et doutes sur l'électricité atmosphèrique, *J. Phys.* **59**, 95–105. [1.9.]

ETTE, A. I. I. (1966a) Anomalous electrical behaviour of a tree "looking in" between electrodes, *J. Atmosph. Terr. Phys.* **28**, 285–94 [9.14.]

ETTE, A. I. I. (1966b). Measurement of electrode by-passing efficiency in living trees, *J. Atmosph. Terr. Phys.* **28**, 295–302 [9.14.]

EVANS, D. G. and HUTCHINSON, W. C. A. (1963). The electrification of freezing water droplets and of colliding ice particles, *Quart. J. R. Met. Soc.* **89**, 370–5. [3.15., 16.23., 16.31.]

EVERLING, E. and WIGAND, A. (1921). Spannungsgefälle und vertikaler Leitungsström in der freien Atmosphäre nach Messungen bei Hochfahrten im Freiballon, *Ann. Phys.* **66**, 261–82. [8.17.]

EXNER, F. (1900). *Sur les recherches récentes relative à l'électricité atmosphèrique*, Rapport au Congrès Intern, Phys. (Paris). Also various papers in *S. B. Akad. Wiss. Wien*, 1886–90. [1.11., 3.21.]

FALCONER, R. E. (1953). *A Correlation between Atmospheric Electricity and the Jet Stream*, Rep. Gen. Elect. Res. Lab., Schenectady, No. R.L. 900. [5.71.]

FARADAY, M. (1843). 18 Reihe von Experimentaluntersuchungen über Elektrizität, *Ann. Phys., Lpz.* **60**, 321–47. [3.18., 16.22.]

FEDEROV, E. K. (1951). Electric charges of the particles of precipitation (in Russian), *C.R. Acad. Sci. URSS* **78**, 1131–4. [10.5., 10.10., 10.12.]

FELDENKRAIS, M. (1937). Voltmètre rotatif pour la mesure de très hautes tensions, *J. Phys. Radium* **8**, 383–4. [5.26., 5.27.]

FETERIS, P. J. (1952). Detailed observations of thunderstorms, *Weather* **7**, 7, 35–9. [12.27.]

FINDEISEN, W. (1940). Über die Entstehung der Gewitterelektrizität, *Met. Z.* **57**, 201–15. [3.18.]

FINDEISEN, W. and FINDEISEN, E. (1943). Untersuchungen über die Eissplitterbildung an Rief-Schichten, *Met. Z.* **60**, 145–54. [3.17., 3.20., 16.24., 16.36.]

FISCHER, H. J. (1962). Die elektrische Spannung zwischen Ionosphäre und Erde, Thesis. Tech. Hochsch, Stuttgart. (5.47., 11.4., 11.7.]

FITZGERALD, D. R. (1956). *Theoretical and Experimental Studies of Convective Cloud Electrification*, Report Contract AF 19(604), 1552, Univ. of Chicago. [12.8., 13.5.]

FITZGERALD, D. R. and BYERS, H. R. (1958). Aircraft observations of convective cloud electrification, *Rec. Adv.*, pp. 245–68, *J. Atmosph. Terr. Phys.* **15**, 254–61. [13.5.]

FITZGERALD, D. R. and BYERS, H. R. (1962). *Aircraft Electrostatic Measurement Instrumentation and Observations of Cloud Electrification*, Report Contract AF 19(604) 2189, Univ. of Chicago. [5.38., 12.19.]

FITZGERALD, D. R. and CUNNINGHAM, R. M. (1965). Multiple aircraft studies of the electrical properties of thunderstorms, *Tokyo Conf.* [12.21.]

FLEISCHER, R. and FALCONER, R. E. (1960). *Cosmic Noise and the Atmosphere*, Obs. of Rensselaer Polytech. Inst., Troy, N.Y., pp. 191–207. [12.32.]

FLEMING, J. A. (ed). (1939). *Terrestrial Magnetism and Electricity*, McGraw-Hill, N.Y. [4.7., 4.11., 7.3., 7.13.]

FLUEGGE, R. A. and PILIÉ, R. J. (1965). *Studies and Experimental Investigation of Space Charge Measuring Equipment*, Final Report. Contract AF 19(628)-1680. Cornell Aeronautical Lab. Buffalo, N.Y. [10.5.]

FORD, J. W. (1958). Paper at 2nd Wentworth Conf. Unpublished [5.32.]

FORREST, J. S. (1943). The determination of the location and frequency of thunderstorms by a radio method, *Quart. J. R. Met. Soc.* **69**, 33–44. [14.37.]

FOSTER, H. (1950). An unusual observation of lightning, *Bull. Amer. Met. Soc.* **31**, 140. [12.26.]

FOUST, C. M. and KUEHNI, H. P. (1922). The surge-crest ammeter, *Gen. Elect. Rev.* **35**, 644–8. [14.43.]

FRANKLIN, B. (1750). Letter to Roy. Soc. [1.3.]

FRANKLIN, B. (1752). *Phil. Trans. Roy. Soc.* **47**, 289. [1.3., 1.5.]

FREIER, G. D. (1960a). The electric field of a large dust devil, *J. Geophys. Res.* **65**, 3504. [5.62.]

FREIER, G. D. (1960b). The coalescence of water drops in an electric field, *J. Geophys. Res.* **65**, 3979–85. [12.24.]

FREIER, G. D. (1961). Auroral effects on the earth's electric field, *J. Geophys. Res.* **66**, 2695–702. [5.56]

FREIER, G. D. (1962). Conductivity of the air in thunderstorms, *J. Geophys. Res.* **67**, 4683–92. [7.11., 7.20., 12.14.]

FRENKEL, J. I. (1944). A theory of the fundamental phenomena of atmospheric electricity, (in Russian), *J. Phys., Moscow* **8**, 285–304. [16.12., 16.17., 16.30.]

FRENKEL, J. I. (1946). Influence of water drops on the ionization and electrification of air (in Russian), *J. Phys. Moscow* **10**, 151–8. [16.12., 16.17.]

FRENKEL, J. I. (1947). Atmospheric electricity and lightning, *J. Franklin Inst.* **243**, 287–307. [16.12., 16.17.]

FRENKEL, J. I. (1949). *Theoriya yavelenii atmosfernogo elektrischestva* (*Theory of phenomena of atmospheric electricity*), Gostelkhizdat. [2.21.]

FUCHS, J. (1955). Report on some observations on atmospheric electricity, *Wentworth Conf.*, pp. 135–9. [5.64.]

FÜRTH, R. (1955). On the theory of diffusion and sedimentation on condensation nuclei in cylindrical containers, *Geofis. Pur. Appl.* **31**, 80–99. [4.39.]

GANE, P. G. and SCHONLAND, B. F. J. (1948). The ceraunometer. A device to record the number of lightning discharges within a given radius, *Weather* **3**, 174–8. [14.37.]

GDALEVICH, G. L. (1965). Rocket experiments aimed at detecting an electric field in the ionosphere, *Problems*, pp. 566–72. [5.72.]

GERASHIMOVA, T. W. (1939). Atmospheric-electric measurements on Elbrus, *Isv. Akad. Nauk*, *SSSR*, Ser. Geogr. Geofiz. 4–5. [9.9.]

GERDIEN, H. (1903). Registrierung der Niederschlagselektrizität im Göttinger geophysikalischen Institut, *Phys. Z.* **4**, 837–42. [10.3., 10.8.]

GERDIEN, H. (1905a). Der Elektrizitätshaushalt der Erde und der unteren Schichten der Atmosphäre, *Phys. Z.* **6**, 647–66. [16.12., 16.13.]

GERDIEN, H. (1905b). Demonstration eines Apparates zur absoluten Messung der elektrischen Leitfähigkeit der Luft, *Phys. Z.* **6**, 800–1. [2.54., 7.2., 7.3.]

GHERZI, E. (1961). *The Action of Different Air Masses on the Potential of the Atmosphere at the Geophysical Observatory of Jean de Brebeuf College, Montreal, Canada*, pp. 3–49. [5.52.]

GIBSON, J. E. (1957). Some observations of microwave radiation from clouds, *Mem. Rep.*, **693**, 1–4, U.S. Naval Res. Lab. Washington, D.C. [15.10]

GILBERT, H. W. and SHAW, T. E. (1925). Electrical charges arising at a liquid–gas interface, *Proc. Phys. Soc., Lond.* **37**, 195–213. [3.21., 4.17.]

GILL, E. W. B. and ALFREY, G. F. (1952). Production of electric charge on water drops, *Nature. Lond.* **169**, 203–4. [3.15.]

GISH, O. H. (1931). Significance of geoelectric data from the polar regions, *Trans. Amer. Geophys. Un.* **12**, 140–2. [2.36.]

GISH, O. H. (1939). Atmospheric electricity, in *Terrestrial Magnetism and Electricity* (J. A. FLEMING, Ed.) pp. 149–230. McGraw-Hill, New York: [4.35.]

GISH, G. H. (1942). Further evidence of a latitude effect in potential gradient, *Terr. Magn. Atmos. Elect.* **47**, 323–4. [5.43.]

GISH, O. H. (1944a). Discussion of atmospheric-electric observations at Huancayo, Peru, during the solar eclipse, January 25, 1944, *Terr. Magn. Atmos. Elect.* **49**, 123–4. [5.54.]

GISH, O. H. (1944b). Evaluation and interpretation of the columnar resistance of the atmosphere, *Terr. Magn. Atmos. Elect.* **49**, 159–68. [2.36., 7.13., 7.17.]

GISH, O. H. (1951). Universal aspects of atmospheric electricity, *Compendium of Meteorology*, pp. 101–19. [11.3., 16.10.]

GISH, O. H. and SHERMAN, K. L. (1929). A leak-free method of measuring air potentials, *Terr. Magn. Atmos. Elect.* **34**, 231–7. [5.9., 5.12., 5.13.]

GISH, O. H. and SHERMAN, K. L. (1936). Electrical conductivity of air to an altitude of 23 km, *Nat. Geog. Soc. Stratosphere*, Ser. 2, 94–116. [7.13., 7.14., 7.17., 7.18., 11.3., 12.22.]

GISH, O. H. and SHERMAN, K. L. (1940). Ionic equilibrium in the troposphere and lower stratosphere, *Gen. Ass. Int. Un. Geod. (Washington). Terr. Magn. Atmos. Elect.*, Bull. 474–91, [4.34.]

GISH, O. H. and WAIT, G. R. (1950). Thunderstorms and the earth's general electricification, *J. Geophys. Res.* **55**, 473–84, [5.38., 7.14., 8.25., 9.8., 9.24., 11.5., 11.6., 11.17., 11.19., 12.21., 12.22., 12.28., 14.24., 14.38., 16.2., 16.9., 17.2., 17.9., 17.13.]

GOCKEL, A. (1917). Luftelektrische Beobachtungen im schweizerischen Mittelland, im Jura und in den Alpen, *N. Denkschr. Schweiz. Naturf. Ges.* **54**, 1–75. [7.10.]

GOHLKE, W. and NEUBERT, U. (1940). Bemerkung zur Hoch- und Höchstspannungsmessung, *Z. Tech. Phys.* **21**, 217–22. [5.26.]

GOHLKE, W. and NEUBERT, U. (1942). Messungen an einem Schwingsvoltmeter, *Z. Tech. Phys.* **23**, 70. [5.26.]

GOLDE, R. H. (1945 a). Frequency of occurrence of lightning flashes to earth, *Quart. J. R. Met. Soc.* **71**, 89–109. [11.14., 11.16., 14.34.]

GOLDE, R. H. (1945 b). On the frequency of occurrence and the distribution of lightning flashes to transmission lines, *Trans. Amer. Inst. Elect. Engrs.* **64**, 1-9. [14.15., 14.51.]

GOLDE, R. H. (1947). Occurrence of upward streamers in lightning discharges, *Nature, Lond.* **160**, 383–7. [14.15.]

GOODLET, B. L. (1937). Lightning, *J. Instn. Elect. Engrs.* **81**, 1–26. [14.56.]

GOTO, M. (1952). A newly designed differential electrometer and its application to the simultaneous measurement of air–earth current and potential gradient, *J. Geomagn. Geoelect., Kyoto* **3**, 22–3. [5.29., 5.30.]

GOTO, M. (1957). Simultaneous measurements of the air–earth current and the atmospheric electric potential gradient, *J. Geomagn. Geoelect., Kyoto* **9**, 133–9. [8.5., 8.7., 8.11.]

GOTT, J. P. (1933). On the electric charge collected by water drops falling through ionized air in a vertical electric field, *Proc. Roy. Soc.* A, **142**, 248–68. [3.11., 16.15.]

GOTT, J. P. (1935). On the electric charge collected by drops falling through a cloud of electrically charged particles in a vertical electric field, *Proc. Roy. Soc.* A, **151**, 665–84. [3.11., 16.14., 16.15.]

GRAY, S. (1935). *Phil. Trans.* **39**, 400. [1.2.]

GRAYSTONE, P. (1954). Atmospheric electricity and jet streams, *Met. Mag., Lond.* **83**, 347. [5.71.]

GRENET, G. (1947). Essai d'explication de la charge électrique des nuages d'orages, *Ann. Geophys.* **3**, 306–7. [16.32.]

GRENET, G. (1959). Le nuage d'orage: machine électrostatique, *La Météorologie*, **53**, 45–53. [16.32]

GROOM, K. N. and CHALMERS, J. A. (1964). The current through a radioactive point in the atmosphere, *J. Atmosph. Terr. Phys.* **26**, 603–6. [5.14.]

GROSS, G. W. (1965). The Workman–Reynolds effect and ionic transfer processes at the ice-solution interface, *J. Geophys. Res.* **70**, 2291–300. [3.15.]

GSCHWEND, P. O. (1922). Beobachtungen über die elektrische Ladung einzelner Regentropfen und Schneeflocken, *Jb. Radioakt.* **17**, 62–79. [10.5., 10.8., 10.10., 10.23., 16.14.]

GUNN, R. (1932). Principles of a new portable electrometer, *Phys. Rev.* **40**, 307. [5.26.]

GUNN, R. (1935). The electricity of rain and thunderstorms, *Terr. Magn. Atmos. Elect.* **40**, 79–106. [16.12., 16.17.]

GUNN, R. (1947). The electrical charge on precipitation at various altitudes and its relation to thunderstorms, *Phys. Rev.* **71**, 181–6. [10.5., 10.12., 10.18., 12.17., 16.25.]

GUNN, R. (1948). Electric field intensity inside of natural clouds, *J. Appl. Phys.* **19**, 481–4. [5.38., 9.21., 12.18.]

GUNN, R. (1949). The free electrical charge on thunderstorm rain and its relation to droplet size, *J. Geophys.Res.* **54**, 57–63. [10.5., 10.8., 10.10.]

GUNN, R. (1950). The free electrical charge on precipitation inside an active thunderstorm, *J. Geophys. Res.* **55**, 171–8. [10.5., 10.12., 10.18., 16.10., 16.25., 17.13.]

GUNN, R. (1951). Precipitation electricity, *Compendium of Meteorology*, pp. 128–35. [16.43.]

GUNN, R. (1952). The electrification of cloud droplets in non-precipitating cumulus, *J. Met.* **9**, 397–402. [10.18., 13.7., 16.4., 16.41.]

GUNN, R. (1954a). Electric field meters, *Rev. Sci. Instrum.* **25**, 432–7. [5.26., 5.27., 5.29.]

GUNN, R. (1954b). Electric field regeneration in thunderstorms, *J. Met.* **11**, 130–8. [2.7., 12.13., 14.32., 16.10.]

GUNN, R. (1954c). Diffusion charging of atmospheric droplets by ions, and the resulting combination coefficients, *J. Met.* **11**, 339–47. [4.31., 13.8., 16.12., 16.18.]

GUNN, R. (1955a). Droplet-electrification processes and coagulation in stable and unstable clouds, *J. Met.* **12**, 511–8. [13.7., 13.8.]

GUNN, R. (1955b). The statistical electrification of aerosols by ionic diffusion, *J. Coll. Sci.* **10**, 107–19. [4.30.]

GUNN, R. (1956). Initial electrification processes in thunderstorms, *J. Met.* **13**, 21–9. [16.18.]

GUNN, R. (1957). The electrification of precipitation and thunderstorms, *Proc. I.R.E.* **45**, 1331–58. [12.18.]

GUNN, R. and ALLEE, P. A. (1954). A three thousand cubic meter cloud chamber, *Bull. Amer. Met. Soc.* **35**, 180–81. [4.15.]

GUNN, R. and DEVIN, C. (1953). Raindrop charge and electric field in active thunderstorms, *J. Met.* **10**, 279–84. [10.5., 10.8., 10.10., 10.12., 10.13., 10.27.]

GUNN, R., HALL, W. C., KINZER, G. D. *et al.* (1946). Technical reports on precipitation static, *Proc. Inst. Radio Engrs., N.Y.* **34**, 156–77; 234–54. [5.38., 10.30.]

HACKING, C. A. (1954). Observations on the negatively charged column in thunderclouds, *J. Geophys. Res.* **59**, 449–53. [12.8., 14.24.]

HAGENGUTH, J. H. (1951). The lightning discharge, *Compendium of Meteorology*, pp. 136–43. [14.45.]

HALES, A. L. (1948). A possible mode of propagation of "slow" or tail component in atmospherics, *Proc. Roy. Soc.* A, **193**, 60–71. [15.3.]

HALLIDAY, E. C. (1932). The polarity of thunderclouds, *Proc. Roy. Soc.* A, **138**, 205–29. [12.7.]

HALLIDAY, E. C. (1934). Thunderstorms and penetrating radiation, *Proc. Camb. Phil. Soc.* **30**, 206–15. [12.31.]

HARDER, D. L. and CLAYTON. T. M. (1953). Power line protection, *Thunderstorm Electricity*, pp. 335–44. [14.53.]

HARNWELL, G. P. and VAN VOORHIS, S. N. (1933). Electrostatic generating voltmeter, *Rev. Sci. Instrum.* **4**, 540–2. [5.25.]

HARPER, W. R. (1940). A rigorous theory of the recombination of both large and small ions in gases at high pressures, *Phil. Mag.* **29**, 434–48.[4.31.]

HARRIS, D. L. (1955). Effects of radioactive debris from nuclear explosions on the electrical conductivity of the lower atmosphere, *J. Geophys. Res.* **60**, 45–62; *Wentworth Conf.*, pp. 140-6. [7.22.]

HATAKEYAMA, H. (1949). On the disturbance of the atmospheric potential gradient caused by the smoke-cloud of the volcano Yakeyama, *J. Geomagn. Geoelect., Kyoto* **1**, 48–51. [5.55.]

HATAKEYAMA, H., KOBAYASHI, J., KITAOKA, T. and UCHIKAWA, K. (1958). A radiosonde instrument for the measurement of atmospheric electricity and its flight results. *Rec. Adv.*, pp. 119–35. [5.37., 7.3.]

HATAKEYAMA, H. and UCHIKAWA, K. (1951). On the disturbance of the atmospheric potential gradient caused by the eruption smoke of the volcano Aso, *Gen. Ass. Int. Un. Geod.* (*Brussels*); *Ass. Terr. Magn. Elect.* [5.5.]

HEAVISIDE, O. (1902). Telegraphy I. Theory, *Encycl. Brit.* 10th Edn., **33**, 213–8. [2.21.]

HENDERSON, J. E., GOSS, W. H. and ROSE, J. E. (1935). The use of the rotary voltmeter for measurements up to 830 kilovolts, *Rev. Sci. Instrum.* **6**, 63–5. [5.26., 5.27.]

HEPBURN, F. (1957). Atmospheric waveforms with very low-frequency components below 1 kc/s known as slow tails, *J. Atmosph. Terr. Phys.* **10**, 266–87. [14.17.]

HERATH, F. (1914). Die Messung der Niederschlagselektrizität durch das Galvanometer, *Phys. Z.* **15**, 155–9. [10.4., 10.8.]

HESS, V. F. (1911). Messungen der durchdringenden Strahlen bei zwei Freiballonfahrten, *S.B. Akad. Wiss., Wien* **120**, 1575–84. [1.15., 4.19.]

HESS, V. F. and O'DONNELL, G. A. (1951) On the rate of ion formation at ground level and at one meter above ground, *J. Geophys. Res.* **56**, 537–62. [7.9.]

HESS, V. F., PARKINSON, W. D. and MIRANDA, H. A. (1953). *Beta Ray Ionization from the Ground*, Report No. 4 to AFCRC, Fordham Univ. [4.21.]

HICKS, W. W. (1956). Air ion generation, metering and physiological effects, *J. Franklin Inst.* **281**, 209. [2.37.]

HIGAZI, K. A. and CHALMERS, J. A. (1966). Measurements of atmospheric electrical conductivity near the ground, *J. Atmosph. Terr. Phys.* **28**, 327–30 [7.6., 8.15., 17.10.]

HILL, E. L. (1960). Ball lightning as a physical phenomenon, *J. Geophys. Res.* **65**, 1947–52. [14.56., 14.57.]

HILL, R. D. (1963a). Determination of charges conducted in lightning strokes, *J. Geophys. Res.* **68**, 1365–75. [14.18., 14.35.]

HILL, R. D. (1963b). Investigation of electron runaway in lightning, *J. Geophys. Res.* **68**, 6261–6. [14.33.]

HOBBS, P. V. (1964). The effect of air bubbles in ice on charge transfer produced by asymmetrical rubbing, *J. Atmos. Sci.* **21**, 706–7. [3.13.]

HOCKING, L. M. (1959). The collision efficiency of small drops, *Quart. J. R. Met. Soc.* **85**, 44–50. [3.6.]

HODGES, D. B. (1954). A comparison of the ratio of change of current in the step and return processes of lightning flashes, *Proc. Phys. Soc., Lond.* B, **67**, 582–4. [14.12.]

HOFFMAN, W. C. (1960). The current-jet hypothesis of whistler generation, *J. Geophys. Res.* **65**, 2047–54. [14.38., 15.11.]

HOFFMANN, K. (1923). Bericht über die in Ebetofthafen auf Spitzbergen in den Jahren 1913–14 durchgeführten luftelektrischen Messungen, *Beitr. Phys. Frei. Atmos.* **11**, 1–19. [5.43.]

HOGG, A. R. (1934a). Atmospheric electric observations, *Beitr. Geophys.* **41**, 1–31. [4.16., 4.20.]

HOGG, A. R. (1934b) Some observations on the average life of small ions and atmospheric ionization equilibria, *Beitr. Geophys.* **41**, 32–5. [4.20., 4.34.]

HOGG, A. R. (1935). Continuous observations of the rate of production of small ions in the atmosphere, *Beitr. Geophys.* **43**, 359-78. [4.20.]

HOGG, A. R. (1939a). The intermediate ions of the atmosphere, *Proc. Phys. Soc., Lond.* **51**, 1014–27. [4.4., 17.6.]

HOGG, A. R. (1939b). The conduction of electricity in the lowest levels of the atmosphere, *Mem. Commonw. Solar Obs., Australia* **7**. (7.6., 7.7., 7.8., 7.9., 8.13., 8.15., 8.20., 17.10.]

HOGG, A. R. (1950) Air–earth current observations in various localities, *Arch. Met., Wien* A, **3**, 40–55. [8.24., 8.25., 11.8.]

HOGG, A. R. (1955). A survey of air-earth current observations, *Wentworth Conf.*, pp. 86-90. [2.36., 8.25.]

HOLL, W. and MÜHLEISEN, R. (1954). Ein Kondensationskernzähler mit kontinuierlicher Übersättigung, *Naturwiss.* **13**, 300–2. [4.9.]

HOLL, W. and MÜHLEISEN, R. (1955). On the equilibrium of ionisation in air containing nuclei, *Geofis. Pur. Appl.* **31**, 115–8. [4.38.]

HOLZER, R. E. (1955). A program of simultaneous measurement of air–earth current density, *Wenthworth Conf.*, pp. 96–100. [5.46., 8.27., 15.8.]

HOLZER, R. E. and DEAL, O. E. (1956). Low audiofrequency electromagnetic signals of natural origin, *Nature, Lond.* **177**, 536–7. [14.37., 15.8.]

HOLZER, R. E. and SAXON, D. S. (1952). Distribution of electrical conduction currents in the vicinity of thunderstorms, *J. Geophys. Res.* **57**, 207–16. [12.11., 12.21., 12.22., 14.30.]

HOLZER, R. E. and SCHILLING, G. F. (1954). Electrical energy in the earth's atmosphere, *World Power Conf. (Rio de Janeiro)* **5**, 1–7. [14.34.]

HOLZER, R. E. and SMITH, W. A. (eds.). (1955). *Proceedings on the Conference on Atmospheric Electricity held at Wenthworth-by-the-Sea, Portsmouth. N.H., May 1954*, Geophys. Res. Dir. AFCRC, Bedford, Mass. [1.19.]

HORNER, F. (1960). The design and use of instruments for counting local lightning flashes, *Proc. Inst. Elec. Engrs.* B, **107**, 321–30. [14.37.]

HUTCHINSON, W. C. A. (1951). Point-discharge currents and the earth's electric field, *Quart. J. R. Met. Soc.* **77**, 627–32. [9.10., 9.15.]

HUTCHINSON, W. C. A. (1960). Ice-crystal contact electrification, *Quart. J. R. Met. Soc.* **86**, 406-7. [3.17., 16.23., 16.31.]

HUTCHINSON, W. C. A. and CHALMERS, J. A. (1951). The electric charges and masses of single raindrops, *Quart. J. R. Met. Soc.* **77**, 85–95. [9.24., 10.5., 10.10., 10.12., 10.13., 10.16., 10.17., 10.20., 10.26., 17.12.]

IMIANITOV, I. M. (1949a). Measuring the horizontal component of the atmospheric-electric field (in Russian), *Izv. Akad. Nauk SSSR. Ser. Geog. Geofiz.* **13**, 320–40. [5.58.]

IMIANITOV, I. M. (1949b). Avtorskoe svidstelstvo, No. 86672. [5.33.]

IMIANITOV, I. M. (1958). *Instruments and Methods for Measuring Atmospheric Electricity*, Translated by Amer. Meteor, Soc. [2.45.]

IMIANITOV, I. M., MIKHAILOVSKAIA, V. V., ZIGANOV, M.P. and STRELTSOVA, M.B. (1956). An instrument for prolonged measurements of the strength of the electric field of the atmosphere under complex meteorological conditions (in Russian), *Izv. Akad. Nauk SSSR. Ser. Geofiz.* **9**, 1121–7. [5.27.]

IMMELMANN, M. N. S. (1938). Point discharge currents during thunderstorms, *Phil. Mag.* **25**, 159–63. [9.9.]

ISHIKAWA, H. (1961). Nature of lightning discharges as origins of atmospherics, *Proc. Res. Inst. Atmosph. (Nagoya Univ.)* **8**A, 1–274. [5.27., 14.24., 14.28., 15.1.]

ISRAËL, H. (1931). Zur Theorie und Methodik der Größenbestimmung von Luftionen, *Beitr. Geophys.* **31**, 171–216. [4.11.]

ISRAËL, H. (1941). Ionen und Kerne: eine kritische Studie, *Beitr. Geophys.* **57**, 261–282. [4.16., 4.37.]

ISRAËL, H. (1943). Die Unruhe des elektrischen Feldes, *Met. Z.* **60**, 56–62. [5.34.]

ISRAËL, H. (1947). Extraterrestrische Einflüsse auf das luftelektrische Feld, *Wiss. Arb. Dtsch. Met. Dienstes in fr. Zone.* **1**, 62-7. [2.56.]

ISRAËL, H. (1948a). Atmosphärische Elektrizität, *F.I.A.T. Review of German Science*, 1939–46; *Geophysics*, Pt. I, Sect. 5, pp. 165–83. [Pref.]

ISRAËL, H. (1948b). Zum Ionisationsgleichgewicht in atmosphärischer Luft, *Met. Rdsh.* **1**, 344. [4.36.]

ISRAËL, H. (1949). Die Tagesvariation des elektrischen Widerstandes der Atmosphäre, *Ann. Géophys.* **5**, 196–210. [5.45., 7.17.]

ISRAËL, H. (1950a). Das Gewitter, *Probleme der kosmischen Physik*, **25**. [Pref., 12.28., 16.16., 16.17., 16.24., 16.30.]

ISRAËL, H. (1950b). Zur Entwicklung der luftelektrischen Grundanschauungen, *Arch. Met. Wien* A, **3**, 1–16. [4.17., 6.13.]

ISRAËL, H. (1951). Radioactivity of the atmosphere, *Compendium of Meteorology*, pp. 155–61. [4.44.]

ISRAËL, H. (1952). *Zahlenwerte und Funktionen*, Landolt-Bornstein, 6th Edn, **3**; *Astronomie und Geophysik*, pp. 707–17. [App. II.]

ISRAËL, H. (1953a). Bemerkung zum Energieumsatz im Gewitter, *Geofis. Pur. Appl.* **24**, 3–11. [7.17., 11.3., 11.14., 11.16., 17.2.]

ISRAËL, H. (1953b). The atmospheric electric field and its meteorological causes, *Thunderstorm Electricity*, pp. 4–23. [5.45.]

ISRAËL, H. (1954). Der luftelektrische Vertikalstrom, *Ann. Geophys.* **10**, 93–116. [8.15., 8.22.]

ISRAËL, H. (1955a). An experimental problem in the atmospheric-electrical synopsis, *J. Atmosph. Terr. Phys.* **6**, 322–7. [8.9.]

ISRAËL, H. (1955b). Synoptical researches on atmospheric electricity, *Wentworth Conf.*, pp. 11–19. [2.34., 8.26.]

ISRAËL, H. (1957). *Atmospheric Electric and Meteorological Investigations in High Mountain Ranges*, Contract AF 61(514)–640. Final Report. [2.35., 5.46., 5.50., 8.27.]

ISRAËL, H. (1959a). Atmospheric electrical agitation, *Quart. J. R. Met. Soc.* **85**, 91–104. [5.57., 6.13., 6.15., 13.7.]

ISRAËL, H. (1959b). Luftelektrische Wirkungen des Jet-stream? *Atti. Congr. Sci. Internat. sulle Correnti a Getto (Jet-Stream) e Ondulatorie, Torino.* [5.71.]

ISRAËL, H. (1962). What is the "sign" of the atmospheric-electric field? *Arch. Met. Wien* A, **13**, 281–2. [2.7.]

ISRAËL, H. and FRIES, G. (1956). Luftelektrische Untersuchungen während der Sonnenfinsternis am 20. Juni 1954, *Z. Geophys.* **20**, 137–49. [5.54.]

ISRAËL, H. and KASEMIR, H. W. (1949). In welcher Höhe geht der luftelektrische Ausgleich vor sich? *Ann. Géophys.* **5**, 313–24. [11.18.]

ISRAËL, H. and KASEMIR, H. W. (1952). Beispiele für das Verhalten luftelektrischer Elemente bei Nebel, *Arch. Met., Wien* A, **5**, 71–85. [7.11., 7.20., 13.9.]

ISRAËL, H., KASEMIR, H. W. and WIENART, K. (1955). Luftelektrische Tagesgänge und Massaustausch im Hochgebirge der Alpen, *Arch. Met., Wien* A, **3**, 357–81. [5.45., 5.50.]

ISRAËL, H., KASEMIR, H. W. and WIENERT, K. (1955). Luftelektrische Tagesgänge und Massaustausch im Hochgebirge der Alpen, *Arch. Met., Wien* A, **8**, 72–94. [5.50.]

ISRAËL, H. and KNOPP, R. (1962). Zum Problem der Ladungsbildung beim Verdampfen, *Arch. Met., Wien* A, **13**, 199–206. [3.21.]

ISRAËL, H. and LAHMEYER, G. (1948). Das Auswahlprinzip der luftelektrisch "ungestörten Tage", *Terr. Magn. Atmos. Elect.* **53**, 373–86. [5.41., 16.40.]

ISTED, G. A. (1954). Atmospheric electricity and long distance very high frequency scatter transmissions, *Marconi Rev.* **17**, 37–60. [14.38.]

ITO, K., KATO, T. and IWAI, A. (1955). Local lightning flash counter within 20 km, *Proc. Res. Inst. Atmosph. (Nagoya Univ.)* **3**, 69–74. [14.37.]

JACCARD, C. (1959). Étude thérorique et expérimentale des propriétés électriques de la glace, *Hev. Phys. Acta.* **32**, 89–128. [3.12.]

JACCARD, C. (1963). Thermoelectric effects in ice crystals, *Phys. Kondens. Materia* **2**, 143–51. [3.12.]

JENSEN, J. C. (1933.) The branching of lightning and the polarity of thunderclouds, *J. Franklin Inst.* **216**, 707–48. [14.41.]

JHAWAR, D. S. and CHALMERS, J. A. (1965). Point discharge from multiple points, *J. Atmosph. Terr. Phys.* **27**, 367–71. [9.13.]

JOLIVET, J. (1959). La charge électrique des gouttes de pluie dans les Antilles francaises, *Ann. Géophys.* **15**, 153–74. [10.3., 10.5., 10.8., 10.10.]

JONES, H. L. (1955). *Research on Tornado Identification*, U.S. Signal Corps. Contract DA-36-039. [12.33.]

JONES, H. L. (1958). The identification of lightning discharges by sferic characteristics, *Rec. Adv.*, pp. 543–56. [15.7.]

JONES, R. F. (1954). Radar echoes from lightning, *Quart. J. R. Met. Soc.* **80**, 579–82, (14.49.]

JONES, M. W. and GIESECKE, A. A. (1944). Atmospheric electric observations at Huancayo, Peru, during the solar eclipse, January 25th, 1944, *Terr. Magn. Atmos. Elect.* **49**, 119–22. [5.54.]

JONES, O. C., MADDEVER, R. S. and SANDERS, J. H. (1959a). Radiosonde measurement of the vertical electric field and polar conductivity, *J. Sci. Instrum.* **36**, 24–28. [5.32., 5.37., 7.3.]

JONES, O. C., MADDEVER, R. S. and SANDERS, J. H. (1959b). Sealed-room experiments on the equilibrium of ionization in air, *J. Atmosph. Terr. Phys.* **17**, 134–44. [4.38., 4.39.]

KACHURIN, L. G. and BEKRYAEV, V. I. (1960). Investigation of the electrification of crystallizing water (in Russian), *Dokl. Akad. Nauk SSSR* **130**, 57–60. [3.15.]

KÄHLER, K. (1908). Registrierung der Niederschlagselektrizität mit dem Benndorf-Elektrometer, *Phys. Z.* **9**, 258–60. [10.2., 10.23.]

KÄHLER, K. (1912). Die Schwankungen des elektrischen Leitvermögens der Atmosphäre und des vertikalen Leitungsstroms in Potsdam, *Phys. Z.* **13**, 1216–21. [8.21., 8.24.]

KÄHLER, K. (1927). Die elektrische Raumladung der Atmosphäre in Potsdam, *Met. Z.* **44**, 1–5. [6.4., 6.8.]

KÄHLER, K. (1934). Das Klima im Zimmer, *Naturwiss.* **23**, 461–3. [7.23.]

KANO, M. (1954). A note on the Workman–Reynolds theory of thunderstorm charge generation, *Pap. Met. Geophys.* **5**, 47–53. [16.25., 16.26.]

KAPITSA, P. L. (1955). The nature of ball lightning (in Russian), *Doklady Akad. Nauk SSSR* **101**, 245–8. [14.57.]

KASEMIR, H. W. (1944). *Die Zylinderfeldstärkenmühle mit kleinen Segmenton*, Flugfunk-Forschungsinst. Oberpfallenhohen E.V. Aussenstelle Grifelfirs, B.H.F. Ref. III A/1306. [5.26.]

KASEMIR, H. W. (1950). Zur Strömungstheorie des luftelektrischen Feldes I, *Arch. Met., Wien* A, **3**, 84–97. [2.29., 5.53.]

KASEMIR, H. W. (1951 a). An apparatus for simultaneous registrations of potential gradient and air–earth current, *J. Atmosph. Terr. Phys.* **2**, 32–7. [5.14., 8.8.]

KASEMIR, H. W. (1951 b). Die Feldkomponentenmühle, *Tellus* **3**, 240–57. [5.38.]

KASEMIR, H. W. (1955). Measurement of the air–earth current density, *Wentworth Conf.*, pp. 91–5. [8.9.]

KASEMIR, H. W. (1956). Zur Strömungstheorie des luftelektrischen Feldes III. Der Austauschgenerator, *Arch. Met., Wien* A, **9** 357–70. (5.46., 8.23.]

KASEMIR, H. W. (1960). *A Radiosonde for Measuring the Air–Earth Current Density*, USASRDL Tech. Rep. 2125. [8.18.]

KASEMIR, H. W. (1962). *Analysis of the Electrostatic Field of a Lightning Stroke*, U.S. Army Signal Res. and Dev. Lab., Fort Monmouth, N.J. [14.28.]

KASEMIR, H. W. (1963). *On the Theory of the Atmospheric Electric Current Flow. IV*, U.S. Army Electronics Res. and Dev. Lab., Fort Monmouth, N.J. [2.29.]

KASEMIR, H. W. (1964). *The Cylindrical Field Mill*, U.S. Army Electronics Lab., Fort Monmouth, N.J. (5.26., 5.38.]

KASEMIR, H. W. (1965a). *Does the Atmospheric Electric Current Flow Penetrate the Ionosphere and Enter Outer Space*? U.S. Army Electronics Lab., Fort Monmouth, N.J. [2.21.]

KASEMIR, H. W. (1965b). The thundercloud, *Problems*, pp. 214-31. [2.29., 12.21.]

KASEMIR, H. W. and RUNHKE, L. H. (1958). Antenna problems of measurement of the air–earth current, *Rec. Adv.*, pp. 137–347; USASRDL Tech. Rep. 2022. [7.2., 8.14.]

KAWANO, M. (1958a). The influence of the vertical distribution of the nuclei content on the vertical distribution of the air resistivity in the exchange layer, *J. Met. Soc., Japan* **36**, 67–72. [5.45.]

KAWANO, M. (1958b). The result of observation of the rate of ion pair production during the solar eclipse, Apr. 19, 1958. *J. Geomagn. Geoelect.* **9**, 210–11. [4.20.]

KAY, A. E. (1950). The application of electronic methods to measurements in atmospheric electricity, Ph. D. Thesis (Durham University). [8.8., 10.3.]

KEEFE, D. and NOLAN, P. J. (1961). Influence of image forces on combination in aerosols, *Geofis. Pur. Appl.* **50**, 155–60. [4.31.]

KEEFE, D. and NOLAN, P. J. (1962). Combination coefficients of ions and nuclei, *Proc. R. Irish Acad.* **62**, 43–53. [4.31., 4.40.]

KEEFE, D., NOLAN, P. J. and RICH, T. A. (1959). Charge equilibrium in aerosols according to the Boltzmann law, *Proc. R. Irish Acad.* **60**, 27–45. [4.30., 4.31., 4.36.]

KELVIN, LORD. (1859a). *On the Necessity for Incessant Recording, and for Simultaneous Observations in Different Localities to Investigate Atmospheric Electricity*, Report of BA (Aberdeen meeting); *Papers on Electricity and Magnetism*, Macmillan, London. pp. 227–9, [1.10.]

KELVIN, LORD (1859b). *Proc. Lit. Phil. Soc., Mnchr*; *Pap. on Elec. and Mag.*, pp. 200–3. [1.10., 5.6.]

KELVIN, LORD (1860a). Electricity, atmospheric, *Nichol's Cyclopaedia*, 2nd Ed; *Pap. on Elec. and Mag.*, pp. 192–208. [1.10., 6.13.]

KELVIN, LORD. (1860b). Atmospheric electricity, *Roy. Instn. Lect.*; *Pap. on Elec. and Mag.*, pp. 208–26. [1.10., 1.17., 2.21., 2.51., 5.65., 10.1.]

KELVIN, LORD. (1860c). Notes on atmospheric electricity, *Phil. Mag., B.A.*; *Pap. on Elec. and Mag.*, pp. 316–20. [1.10., 5.65.]

KELVIN, LORD. (1862). Observations on atmospheric electricity, *Proc. Lit. Phil. Soc., Mnchr.*; *Pap. on Elec. and Mag.*, pp. 230–5. [6.4.]

KELVIN, LORD. (1881). *Congres. Int. des Électriciens*, Paris. [1.12.]

KENNEDY, H. (1913). The large ions in the atmosphere, *Proc. R. Irish Acad.* A, **32**, 1–6. [4.34.]

KENNEDY, H. (1916). The large ions and condensation nuclei from flames, *Proc. R. Irish Acad.* A, **33**, 58–74. [4.34.]

KENNELLY, A. E. (1902). On the elevation of the electrically conducting strata of the earth's atmosphere, *Elect. World. N.Y.* **39**, 473. [2.21.]

KIKUCHI, K. (1965). On the positive electrification of snow crystals in the process of their melting, *Tokyo Conf.*, pp. 338–42. [3.19.]

VON KILINSKI, E. (1950). Die Registrierung der luftelektrischen Feldstärke, *Z. Met.* **4**, 677–81. [5.25., 5.27.]

VON KILINSKI, E. (1952). Ein Gerät zur Dauerregistrierung des luftelektrischen Vertikalstromes, *Z. Met.* **6**, 271–7. [8.10.]

VON KILINSKI, E. (1958). Verbesserung der Auffangvorrichtung des Potsdamer Vertikalstromgerätes, *Z. Met.* **12**, 352–4. [2.42., 8.10.]

KINMAN, T. D. (1954). The measurement of atmospheric space charge, ARL (Teddington). Unpub. Rep. [6.5.]

KIRKMAN, J. R. and CHALMERS, J. A. (1957). Point discharge from an isolated point. *J. Atmosph. Terr. Phys.* **10**, 258–65. [9.12., 9.15., 9.16., 9.17.]

KIRKPATRICK, P. (1932). Further developments of the rotary voltmeter, *Rev. Sci. Instrum.* **3**, 340. [5.26., 5.27.]

KIRKPATRICK, P. and MIYAKE, F. (1932). A generating voltmeter for the measurement of high potentials, *Rev. Sci. Instrum.* **3**, 1–8. [5.26., 5.27.]

KITAGAWA, N. and BROOK, M. (1960). A comparison of intracloud and cloud-to-ground lightning discharges, *J. Geophys. Res.* **65**, 1189–202. [14.30.]

KITAGAWA, N., BROOK, M. and WORKMAN, E. J. (1962). Continuing currents in cloud-to-ground lightning discharges, *J. Geophys. Res.* **67**, 637–47. [14.19., 14.20., 14.21., 14.23.]

KITAGAWA, N., IIZUKA, T., MURAI, K. and KOYABASHI, M. (1953). On some aspects of lightning discharges observed by electrostatic fluxmeter, *Geophys. Mag., Tokyo* **25**, 237–42. [14.9., 14.28.]

KITAGAWA, N. and KOYABASHI, M. (1958). Distribution of negative charge in the cloud taking part in a flash to ground, *Pap. Met. Geophys.* **9**, 99–105. [14.20., 14.21., 14.24.]

KOBAYASHI, M., KITAGAWA, N., IKEDA, T. and SATO, Y. (1958). Preliminary studies of variation of luminosity and field-change due to lightning flashes, *Pap. Met. Geophys.* **9**, 29–34. [14.26.]

KOBAYASHI, J. and KYOZUKA, M. (1962). On the specially designed radiosondes used in Japan during the World Meteorological Intervals as a programme of the International Geophysical Year. Pt. 2. Balloon-borne electrometer and its applications. *Geophys. Mag.* **31**, 195–236. [5.32.]

KOENIGSFELD, L. (1953). Investigations of the potential gradient at the earth's ground surface and within the free atmosphere, *Thunderstorm Electricity*, pp. 24–45. [5.37., 5.54., 6.12.]

KOENIGSFELD, L. (1955a). La mesure de la conductibilité par radio-sonde, *Inst. R. Met., Belg.* B, **18**, 1–8. [7.3., 7.13.]

KOENIGSFELD, L. (1955b). Study of the variation of potential gradient with altitude and correlated meteorological conditions, *Wentworth Conf.*, pp. 21–6. [5.37.]

KOENIGSFELD, L. and PIRAUX, P. (1950). Un nouvel électromètre portatif pour la mesure des charge électrostatiques per système électronique. Son application à la radiosonde, *Mem. Inst. Met., Belg.* **45**, [5.37., 17.11.]

KOENIGSFELD, L. and VAN DER SCHUEREN, A. (1965). Atmospheric electricity observations at King Baudouin's base, *Problems*, pp. 74–6. [5.56.]

KOHLRAUSCH, K. W. F. (1909). Luftelektrische Beobachtungen über dem Atlantischen Ozean und in Porto Rico, *S.B. Akad. Wiss., Wien* **118**, 25–67. [10.4.]

KOLHÖRSTER, W. (1913). Messungen der durchdringenden Strahlen im Freiballon in größeren Höhen, *Phys. Z.* **14**, 1153–5. [1.15., 4.19.]

KORNBLUCH, I. H. and GRIFFIN, J. E. (1955). Artificial air ionization in physical medicine, *Am. J. Phys. Med.* **34**, 618. [2.37.]

KRAAKEVIK, J. H. (1958a). The airborne measurement of atmospheric conductivity, *J. Geophys. Res.* **63**, 161–9. [7.14., 7.17., 7.18., 8.17., 11.3.]

KRAAKEVIK, J. H. (1958b). Electrical conduction and convection currents in the troposphere, *Rec. Adv.*, pp. 75–88. [8.17.]

KRAAKEVIK, J. H. (1961). Measurements of current density in the fair weather atmosphere, *J. Geophys. Res.*, **66**, 3735–48. [8.17., 11.10]

KRAAKEVIK, H. J. and CLARK, J. F. (1958). Airborne measurements of atmospheric electricity, *Trans. Amer. Geophys. Un.* **39**, 827–34. [8.1., 8.17., 11.2., 11.4., 17.10.]

KRAMER, C. (1949). Elektrische Ladingen aan berijpte opperflakken, *Meded. Ned. Met. Inst.* **54**, A, No. 102. [3.20., 16.24.]

KRASNOGORSKAYA, N. V. (1960). Investigation of electrification processes of cloud particles and precipitation (in Russian), *Bull.* (*Izv*) *Acad. Sci. USSR Geophys.*, Ser. 1, **89**. [3.11.]

KRASNOGORSKAYA, N. V. (1961). An investigation of atmospheric electricity at Mt. Elbrus, Trans. from *Isv. Geophys.*, Ser. 1961, pp. 1862–74. [7.11., 10.10., 10.13., 13.7.]

KREIELSHEIMER, K. (1947). A radiosonde method for potential-gradient measurements in the atmosphere, *N.Z.Science Congress*, pp. 91–8. [9.22.]

KREIELSHEIMER, K. and BELIN, R. (1946). Radio-sonde recording of potential gradient, *Nature, Lond.* **157**, 227–8. [9.22.]

KREIG, J. J. (1960). *An Improved Version of a Simple Thunderstorm Warning Device*, USASRDL Tech. Rep. No. 2105, pp. 1–11. [14.36.]

KROENING, J. L. (1960). Ion-density measurements in the stratosphere, *J. Geophys. Res.* **65**, 145–57. [4.14.]

KRUEGER, A. P. (1962). Air ions and physiological function, *J. Gen. Physiol.* **45** (Suppl.), 233 [2.37.]

KRUEGER, A. P., KOTAKA, S. and ANDRIESE, P. C. (1962). Studies on the effect of gaseous ions on plant growth, *J. Gen. Physiol.* **45**, (Suppl.), 879 [2.37.]

KUETTNER, J. (1950). The electrical and meteorological conditions inside thunderclouds, *J. Met.* **7**, 322–32. [10.19., 12.8., 12.16., 12.17., 14.47., 16.7., 16,36., 16.37.]

KUETTNER, J. (1956). The development and masking of charges in thunderstorms, *J. Met.* **13**, 456–70. [16.9.]

KUMM, A. (1951). Über die Entstehung von elektrischen Ladungen bei Vorgängen in der kristallinien Eisphase, *Arch. Met., Wien* A, **3**, 382–407. [3.20., 16.24., 16.29.]

LABY, T. H., MCNEILL, J. J., NICHOLLS, F. G. and NICKSON, A. F. B. (1940). Wave form, energy and reflexion by the ionosphere of atmospherics, *Proc. Roy. Soc.* A, **174**, 145–63. [15.3.]

LAMONT, J. (1852). *Jahresber. Münchn. Sternw.*, p. 82. [9.6.]

LAMONT, J. (1862). *Der Erdstrom und der Zusammenhang desselben mit dem Magnetismus der Erde*, Leipzig, p. 15. [9.6.]

LANDSBERG, H. (1938). Atmospheric condensation nuclei, *Beitr. Geophys.*, Suppl. **3**, 155–252. [4.6.]

LANGEVIN, P. (1903). Recombinaison et mobilités des ions dans les gaz, *Ann. Chim. Phys.* **28**, 433–530. [4.31., 4.33.]

LANGEVIN, P. (1905). Sur les ions de l'atmosphère, *C. R. Acad. Sci., Paris* **140**, 232–4. [1.14., 4.3.]

LARGE, M. I. and PIERCE, E. T. (1955). The fine structure of natural point discharge currents, *Quart. J. R. Met. Soc.* **81**, 92–5. [9.4.]

LARGE, M. I. and PIERCE, E. T. (1957). The dependence of point-discharge currents on wind, as examined by a new experimental approach, *J. Atmosph. Terr. Phys.* **10**, 251–7. [9.16., 17.8.]

LARGE, M. I. and WORMELL, T. W. (1958). Fluctuations in the vertical electric field in the frequency range from 1 cycle per second to 500 cycles per second *Rec. Adv.*, pp. 603–7. (15.8.]

LARSEN, A. (1906). *Photographing Lightning with a Moving Camera*, Smithsonian Inst. Ann. Rep. for 1905, pp. 119–27. [14.20., 14.21.]

LATHAM, J. (1963a). The electrification of frost deposits, *Quart. J. R. Met. Soc.* **89**, 265–70. [3.15.]

LATHAM, J. (1963b). Electrification produced by the asymmetric rubbing of ice on ice, *Brit. J. Appl. Phys.* **14**, 488–90. [3.14.]

LATHAM, J. (1963c). Charge transfer associated with temperature gradients in ice, *Nature, Lond.* **200**, 1087. [3.14.]

LATHAM, J. (1964a). The electrification of snowstorms and sandstorms, *Quart. J. R. Met. Soc.* **90**, 91–5. [3.14., 5.62.]

LATHAM, J. (1964b). Charge transfer associated with temperature gradients in ice crystals grown in a diffusion chamber, *Quart. J. R. Met. Soc.* **90**, 266–74. [3.12., 3.14.]

LATHAM, J. (1965a). Electrification produced by the growth of soft hail in thunderstorms, *Tellus* **17**. 204–12. [3.17., 16.23.]

LATHAM, J. (1965b). Electrification of frost deposits, *Quart. J. R. Met. Soc.* **91**, 369–72. [16.32.]

LATHAM, J. (1965c). The effect of air bubbles in ice on charge transfer produced by asymmetric rubbing, *J. Atm. Sci.* **22**, 325–8. [3.13.]

LATHAM, J. and MASON, B. J. (1961a). Electric charge transfer associated with temperature gradients in ice, *Proc. Roy. Soc.* A, **260**, 523–36. [3.12., 3.14., 3.15., 3.17., 16.31.]

LATHAM, J. and MASON, B. J. (1961b). Generation of electric charge associated with the formation of soft hail in thunderclouds, *Proc. Roy. Soc.* A, **260**, 537–49. [3.14., 3.15., 3.17., 3.18., 16.10., 16.23., 16.24., 16.27., 16.31.]

LATHAM, J. and MASON, B. J. (1962). Electrical charging of hail pellets in a polarizing electric field, *Proc. Roy. Soc.* A, **266**, 387–401. [16.23., 16.27.]

LATHAM, J. and STOW, C. D. (1965a). Electrification produced by collision between ice crystals and soft hail pellets in a thunderstorm, *Quart. J. R. Met. Soc.* **91**, 99–100. [3.14., 3.17.]

LATAHM, J., and STOW, C. D. (1965b). Electrification of evaporating ice crystals, *Tokyo Conf.*, pp. 352–6. [3.20.]

LAUTNER, R. (1929). Die luftelektrischen Verhältnisse am Zugspitzgipfel in 2960 m, *D. Met. Jahrb. für Bayern*, 1928, 1–19. [8.24.]

LAW, J. (1963). The ionisation of the atmosphere near the ground in fair weather, *Quart. J. R. Met. Soc.* **89**, 107–21. [5.46., 6.8., 8.15., 8.20., 8.23., 17.10.]

LECOLAZET, R. (1945). Sur un montage électrométrique pour la mesure du champ électrique pour une méthode de zéro, *C. R. Acad. Sci., Paris* **220**, 531–3. [5.9.]

LECOLAZET, R. (1946). Resultats d'une étude préliminaire de l'effect des cumulus sur le champ électrique terrestre, *Météorologie*, 429–33. [5.9., 16.4.]

LECOLAZET, R. (1948a). Étude expérimentale de l'état électrique des cumulus de beau temps; interprétation des résultats, *Ann. Géophys.* **4**, 81–95. [5.39., 7.14., 7.19.]

LECOLAZET, R. (1948b). Les phénomènes électriques à la frontière de deux atmosphéres differentes, *Ann. Géophys.* **4**, 181–92. [7.19.]

LECOLAZET, R. and PLUVINAGE, P. (1948). États de regimes permanents électrodynamiques dans l'atmosphère, *Ann. Géophys.* **4**, 96–108. [7.19., 7.20.]

LEMONNIER, L. G. (1752). Observations sur l'électricité de l'air, *Mem. Acad. Sci* .**2**, 233. [1.5., 5.3.]

LEMSTRÖM, S. (1882–4). *Exploration Internationale des Régions Polaires.* [9.6.]

LEMSTRÖM, S. [1898). *Expédition Polaire Finlandaise*, (Helsingfors) **3**. [9.6.]

LEMSTRÖM, S. (1900). *Acta Soc. Sci. Fenn.* **29**, No. 8. [9.6.]

LENARD, P. (1892). Über die Elektrizität der Wasserfälle, *Ann. Phys., Lpz.* **46**, 584–636. [3.24., 4.17., 5.65., 10.2., 10.26., 16.21.]

LENARD, P. (1914). Probleme komplexer Moleküle, *Sitz.Ber. Heidelberger Akad. Wiss.* **5**A. [3.21.]

LETTAU, H. (1941). Anwendung neuer Ergebnisse der Austauschlehre auf zwei luftelektrische Fragen, *Beitr. Geophys.* **57**, 365–83. [8.20.]

LEWIS, W. W. and FOUST, C. M. (1936). Lightning surges in transmission lines–natural lightning, *Gen. Elect. Rev.* **39**, 543–55. [14.42.]

LIGDA, M. G. H. (1950). Lightning detection by radar, *Bull. Amer. Franklin Met. Soc.* **31**, 279–83. [14.49.]

LIGDA, H. G. H. (1956). The radar observation of lightning, *J. Atmosph. Terr. Phys.* **9**, 329–46. [14.49.]

LINKE, F. (1904). Luftelektrische Messungen bei 12 Ballonfahrten, *Abh. Ges. Wiss., Göttingen* **3**, 5. [5.37.]

LINSS, F. (1887). Über einige die Wolken- und Luftelektrizität betreffende Probleme, *Met. Z.* **4**, 345–62. [11.1.]

LOCK, B. H. (1956). Experimental observations of the number of thunderstorm discharges (in Russian), *Trudy Leningrad. Glav. Geofiz. Obs.* **58**, 47–52. [14.37.]

LOEB, L. B. (1953). Experimental contributions to the knowledge of charge generation, *Thunderstorm Electrification*, pp. 150–92. [12.20.]

LUDLAM, F. H. (1951). The production of showers by the coalescence of cloud droplets, *Quart. J. R. Met. Soc.* **77**, 402–17. [3.6.]

LUDLAM, F. H. (1957). Noctilucent clouds, *Tellus* **9**, 341–64. [7.15.]

LUEDER, H. (1943). Elektrische Registrierung von heranziehenden Gewittern und die Feinstruktur des luftelektrischen Gewitterfeldes, *Met. Z.* **60**, 340–51. [5.25., 5.27.]

LUEDER, H. (1951 a). Ein neuer elektrischer Effekt bei der Eisbildung durch Vergraupelung in natürlichen unterkühlten Nebeln, *Z. Angew. Phys.* **3**, 247–87. [3.18., 16.31., 17.4.]

LUEDER, H. (1951 b). Vergraupelungselektrisierung als eine Ursache der Gewitterelektrizität, *Z. Angew. Phys.* **3**, 288. [16.25., 16.26., 16.31.]

LUGEON, J. (1928). Nouvelle méthode de sondage électromagnétique vertical et quasi-horizontale de l'atmosphère, *Arch. Sci. Phys. Nat.* **5**, 239–59. [15.4.]

LUGEON, J. and BOHNENBLUST, M. (1956). Radiosondages du gradient potential et de la conductibilité électrique de l'air, *Ann. Stat. Cent. Suisse de Met.* [5.37., 7.3.]

LUHR, O. and BRADBURY, N. E. (1931). Corrected values for the coefficient of recombination of gaseous ions, *Phys. Rev.* **37**, 998–1000. [4.34.]

LUTZ, C. W. (1911). Luftelektrische Messungen am erdmagnetischen Observatorium (Sternwarte) München, *S. B. Bayer. Akad. Wiss.*, 305–60. [8.24.]

LUTZ, C. W. (1939). Die wichtigsten luftelektrischen Größen für München, *Beitr. Geophys.* **54**, 337–47. [8.15.]

LUTZ, C. W. (1941). Über die Spitzentladung bei Gewittern und Schauern, *Beitr. Geophys.* **57**, 317–33. [9.10., 12.5.]

LUTZ, C. W. (1944). Über den Beitrag der Spitzentladung zur Aufrechterhaltung der negativen Erdladung, *Beitr. Geophys.* **60**, 9. [9.9.]

LUVINI, (1884). *Sept Études* (Turin). [16.22.]

MCCANN, G. D. (1944). The measurement of lightning currents in direct strokes, *Trans. Amer. Inst. Elect. Engrs.* **63**, 1157–64. [14.19., 14.39.]

MCCLELLAND, J. A. and GILMOUR, A. (1920). Further observations of the electric charge on rain, *Proc. R. Irish Acad.* A, **35**, 13–39. [10.2., 10.23.)

MCCLELLAND, J. A. and KENNEDY, H. (1912). The large ions in the atmosphere, *Proc. R. Irish Acad.* A, **30**, 72–91. [4.12.]

MCCLELLAND, J. A. and NOLAN, J. J. (1912). The electric charge on rain, *Proc. R. Irish Acad.* A, **29** 81–91; **30**, 61–71. [10.2., 10.23.]

MACCREADY, P. B. and PROUDFIT, A. (1965 a). Observations of hydrometeor charge evolution in thunderstorms, *Quart. J. R. Met. Soc.* **91**, 44–53. [3.19.]

MACCREADY, P. B. and PROUDFIT, A. (1965 b). Self-charging of melting ice, *Quart. J. R. Met. Soc.* **91**, 54–9. [3.19.]

MCEACHRON, K. B. (1939). Lightning to the Empire State Building. I, *J. Franklin Inst.* **227**, 149–217. [14.39., 14.41., 14.44.]

MCEACHRON, K. B. (1941). Lightning to the Empire State Building. II, *Trans. Amer. Inst. Elect. Engrs.* **60**, 885–90. [14.39., 14.45.]

MCGREEVY, G. and O'CONNOR, T. C. (1962). On the production of Aitken nuclei by sunlight at St. Moritz, *Geofis. Meteor.* **10**, 3–7. [4.3.]

MACKERRAS, D. (1963). *Thunderstorm Observations Related to Lightning-flash Counter Performance*, Progress Report to the Electrical Research Board (Dept. of Electrical Engineering, University of Queensland). [14.37.]

MACKY, W. A. (1931). Some investigations on the deformation and breaking of water drops in strong electric fields, *Proc. Roy. Soc.* A ,**133**, 565–87. [12.20.]

MACKY, W. A. (1937). Measurement of normal atmospheric electric potential gradients using a valve voltmeter, *Terr. Magn. Atmos. Elect.* **42**, 77–86. [5.25., 5.27.]

MAGONO, C. and KIKUCHI, K. (1961). On the electric charge of relatively large natural cloud particles, *J. Met. Soc., Japan* **39**, 258–68. [13.7.]

MAGONO, C. and KIKUCHI, K. (1963). On the positive electrification of snow crystals in the process of their melting, *J. Met. Soc., Japan* **41**, 270–7. [3.19]

MAGONO, C. and KOENUMA, S. (1958). On the electrification of water drops by breaking due to the electrostatic induction under a moderate electric field, *J. Met. Soc., Japan* **36**, 108–11. [3.24.]

MAGONO, C. and ORIKASA, K. (1960). On the surface electric field during rainfall, *J. Met. Soc., Japan* **38**, 182–94. [10.27.]

MAGONO, C. and ORIKASA, K. (1961). On the surface electric field caused by the space charge of charged raindrops, *J. Met. Soc., Japan* **39**, 1–11. [10.15.]

MAGONO, C., ORIKASA, K. and OKABE, H. (1957). The charge on precipitation elements and surface electric potential gradient, *J. Fac. Sci., Hokkaido Univ.* **1** (7) 7–20. [10.5., 10.13., 10.23.]

MAGONO, C. and SAKURAI, K. (1963). On the electric charge on drifting snow pellets, *J. Met. Soc., Japan* **41** 211–17. [10.23.]

MAGONO, C. and SHIOTSUKI, Y. (1964). On the effect of air bubbles in ice on frictional charge separation, *J. Atmos. Sci.* **21**, 666–70, [3.13.]

MAGONO, C. and TAKAHASHI, T. (1959). The electrification of antennae by snow and rain falls, *J. Met.* **16**, 388–92. [10.29.]

MAGONO, C. and TAKAHASHI, T. (1963). Experimental studies on the mechanism of electrification of graupel pellets, *J. Met. Soc., Japan* **41**, 197–210. [3.17., 3.18.]

MALAN, D. J. (1937). Sur les décharges orageuses dans la haute atmosphère, *C. R. Acad. Sci., Paris* **205**, 812–13. [14.38.]

MALAN, D. J. (1952). Les décharges dans l'air et la charge inférieur positive d'un nuage orageux, *Ann. Géophys.* **8**, 385–401. [12.1., 12.21., 14.24., 14.47., 16.37.]

MALAN, D. J. (1954). Les décharges orageuses intermittantes et continues de la colonne de charge négative, *Ann. Géophys.* **10**, 271–81. [14.23. 14.38.]

MALAN, D. J. (1956). The relation between the number of strokes, stroke interval and the total durations of lightning discharges, *Geofis. Pur. Appl.* **34**, 224–30. [14.20., 14.21.]

MALAN, D. J. (1957). The theory of lightning photography and a camera of new design, *Geofis. Pur. Appl.* **38**, 250–60. [14.2.]

MALAN, D. J. (1958). Radiation from lightning discharges and its relation to the discharge process, *Rec. Adv.*, pp. 557–63. [15.7.]

MALAN, D. J. (1961). Unusual types of lightning discharge, *Ann. Géophys.* **17**, 388–94, (14.41., 14.55., 14.57.]

MALAN, D. J. (1962). Lightning counter for flashes to ground, *Proc. Int. Conf. at C.E.A.Res. Labs., 1962*, Butterworth, London. [14.37.]

MALAN, D. J. (1963). *Physics of Lightning*, pp. 1–176, English Univ. Press, London. [14.1., 14.17., 14.46.]

MALAN, D. J. and COLLENS, H. (1937). Progressive lightning. III. The fine structure of return lightning strokes, *Proc. Roy. Soc.* A, **162**, 174–203. [14.3.]

MALAN, D. J. and SCHONLAND, B. F. J. (1947). Progressive lightning VII. Directly correlated photographs and electrical studies of lightning from near thunderclouds, *Proc. Roy. Soc.* A, **191**, 485–503. [14.25., 14.27., 14.38.]

MALAN, D. J. and SCHONLAND, B. F. J. (1950). An electrostatic fluxmeter of short response time for use in studies of transient field changes, *Proc. Phys. Soc., Lond.* B, **63**, 402–8. (5.25., 5.27., 5.34., 12.4., 12.7.]

MALAN, D. J. and SCHONLAND, B. F. J. (1951a). The electrical processes in the intervals between the strokes of a lightning discharge, *Proc. Roy. Soc.* A, **206**, 145–63. [14.12., 14.25.]

MALAN, D. J. and SCHONLAND, B. F. J. (1951b). The distribution of electricity in thunderclouds, *Proc. Roy. Soc.* A, **209**, 158–77. [14.24., 16.37.]

MALONE, T. F. (ed.). (1951). *Compendium of Meteorology*, Amer. Meteor. Soc. Boston. [Pref., 3.3.]

MAPLESON, W. W. and WHITLOCK, W. S. (1955). Apparatus for accurate and continuous measurement of the earth's electric field, *J. Atmosph. Terr. Phys.* **7**, 61–72. [5.25., 5.27.]

MARKGRAF, H. (1924). Luftelektrische Ergebnisse, *Met. Z.* **41**, 165–70. [7.10.]

MARSHALL, J. S. (1953). Frontal precipitation and lightning observed by radar, *Canad. J. Phys.* **31**, 194–203. [14.49.]

MARSHALL, J. S. and PALMER, W. M. (1948). *J. Met.* **5**, 165. [10.20.]

MARWICK, T. C. (1930). The electric charge on rain, *Quart. J. R. Met. Soc.* **56**, 39–44. [10.2., 10.23.]

MASON, B. J. (1953a). On the generation of charge associated with graupel formation in thunderclouds, *Quart. J. R. Met. Soc.* **79**, 501–9. [16.10., 16.25., 16.26., 16.31.]

MASON, B. J. (1953b). A critical examination of theories of charge generation in thunderstorms, *Tellus* **5**, 446–60. [16.27.]

MASON, B. J. (1957). *The Physics of Clouds*, pp. 481, Univ. Press, Oxford. [3.3.]

MASON, B. J. and LUDLAM, F. H. (1950). The microphysics of clouds, *Rep. Progr. Phys.* **14**, 147–95. [3.3.]

MASON, B. J. and MAYBANK, J. (1960). The fragmentation and electrification of freezing water drops, *Quart. J. R. Met. Soc.* **86**, 176–85. [3.15., 3.17., 16.28., 16.31.]

MASON, B. J. and OWSTON, P. G. (1952). Ice crystals of spiral form grown from the vapour, *Phil. Mag.* **43**, 911–2. [16.29.]

MATTEUCCI, C. (1850). *Ann. Chem. Phys.* **28**, 285. [1.14.]

MATTHEWS, J. B. and MASON, B. J. (1963). Electrification accompanying melting of ice and snow, *Quart. J. R. Met. Soc.* **89**, 376–80. [3.19.]

MATTHEWS, J. B. and MASON, B. J. (1964). Electrification produced by the rupture of large water drops in an electric field, *Quart. J. R. Met. Soc.* **90**, 275–86. [3.24.]

MATTHIAS, A. (1926). Fortschritt in der Aufklärung der Gewittereinflüsse auf Leitungsanlagen, *Elektrizitätswirtschaft* **25**, 297–308. [5.26., 5.27.]

MATTHIAS, A. (1929). Der gegenwärtige Stand der Blitzschutzfrage, *Elektrotech. Z.* **50**, 1469–74. [14.20.]

MAUCHLY, S. J. (1923). Diurnal variation of the potential gradient of atmospheric electricity, *Terr. Magn. Atmos. Elect.* **28**, 61–81. [5.43.]

MAUCHLY, S. J. (1926). Studies in atmospheric electricity based on observations made on the *Carnegie*, 1915–21, *Carnegie Inst. Wash. Yr. Bk.* **25**, 227. [8.24.]

MAUND, J. E. and CHALMERS, J. A. (1960). Point discharge from natural and artificial points, *Quart. J. R. Met. Soc.* **86**, 85–90. [5.68., 9.14., 11.12.]

MAURAIN, C., HOMERY, MLLE, G. and GIBAULT, T. (1930). Sur le courant électrique vertical atmosphèrique, *C. R. Acad. Sci., Paris.* **191**, 87–9. [8.24.]

MEEK, J. R. (1939). The mechanism of the lightning discharge, *Phys. Rev.* **55**, 972–7. [14.13.]

MEEK, J. M. and PERRY, F. R. (1944–45). The lightning discharge, *Rep. Progr. Phys.* **10**, 314–57. [14.1., 14.13.]

MEINHOLD, H. (1951). Die elektrische Ladung eines Flugzeuges bei Vereisung in Quellwolken, *Geofis. Pur. Appl.* **19**, 176–8. [3.18.]

MICHNOWSKI, S. (1957). Point discharge in the interchange of electric charge between the earth and the atmosphere, *Acta Geophys. Polon.* **5** 123–4. [9.9.]

MICHNOWSKI, S. (1963). On the observation of lightning in warm clouds, *Ind. J. Met. Geophys.* **14**, 320–2. [12.26.]

MILES, V. G. (1953). Radar echoes associates with lightning, *J. Atmosph. Terr. Phys.* **3**, 258–62. [14.49.]

MILNER, J. W. and CHALMERS, J. A. (1961). Point discharge from natural and artificial points (II), *Quart. J. R. Met. Soc.* **87**, 592–6. [2.51., 9.14., 9.16.]

MIRANDA, H. A. (1958). The ionization due to beta radiation from the atmosphere, *J. Geophys. Res.* **63**, 147–55. [4.21.]

MISAKI, M. (1950). A method of measuring the ion spectrum, *Pap. Met. Geophys.* **1**, 313–8. [4.11.]

MOORE, C. B., LEAVITT, P. R., VONNEGUT, B. and VRABLIK, E. A. (1962). *Some Atmospheric Electric Instruments for Use in Air Force Operations*, Report on Contract AF 19(604)–7413 (Arthur D. Little, Inc.). [14.36.]

MOORE, C. B., VONNEGUT, B. and BOTKA, A. T. (1958). Results of an experiment to determine initial precedence of organized electrification and precipitation in thunderstorms, *Rec. Adv.*, pp. 331–60. [12.18., 12.24., 16.6.]

MOORE, C. B., VONNEGUT, B. and EMSLIE, A. G. (1959). *Observations of Thunderstorms in New Mexico*, Report on Contract NONR-1684. (Arthur D. LITTLE, Inc.). [16.6.]

MOORE, C. B., VONNEGUT, B., MACHADO, J. A. and SURVILAS, H. J. (1962). Radar observations of rain gushes following overhead lightning strokes, *J. Geophys. Res.* **67**, 207–20. [12.25.]

MOORE, C. B., VONNEGUT, B., SEMONIN, R. G., BULLOCK, J. W. and BRADLEY, W. (1962) Fair-weather atmospheric electric potential gradient and space charge over Central Illinois, Summer, 1960, *J. Geophys. Res.* **67**, 1061–71. [6.8., 6.13.]

MOORE, C. B., VONNEGUT, B., STEIN, B. A. and SURVILAS, H. T. (1960). Observations of electrification and lightning in warm clouds, *J. Geophys. Res.* **65**, 1907–10. [12.26.]

MOORE, C. B., VONNEGUT, V., VRABLIC, E. A. and MCCAIG, D. A. (1964). Gushes of rain and hail after lightning, *J. Atmosph. Sci.* **21**, 646–65. [12.25.]

MORGAN, W. A. (1960). Determination of the straight line of best fit to observational data of two related variates when both sets of values are subject to error, *Quart. J. R. Met. Soc.* **86**, 107–13. [2.40.]

MÜHLEISEN, R. (1941). Zur Methodik der luftelektrischen Potentialmessung: Einfluß des Windes bei radioaktiven Kollaktoren, *Z. Naturf.* **6**, 667–71. (5.13., 5.15.]

MÜHLEISEN, R. (1953). Die luftelektrischen Elemente im Großstadtbereich, *Z. Geophys.* **29**, 142–60. [4.17., 5.44., 5.59., 6.13., 8.7., 17.8.]

MÜHLEISEN, R. (1955). Luftelektrische Beobachtungen während der Sonnenfinsternis am 30. 6. 54, *Geofis. Pur. Appl.* **30**, 106–12. [5.54.]

MÜHLEISEN, R. (1956). On the deviations of the course of elements of atmospheric electricity on continents from the world-wide course, *J. Atmosph. Terr. Phys.* **8**, 146–57. [5.46., 6.13., 8.17., 8.23.]

MÜHLEISEN, R. (1958). Elektrische Ladungen an Aerosol- Dunst- und Nebelteilchen, *Ber. der. deut. Wetterd.* **51**, 62–6. [3.21., 3.23., 5.46., 6.10., 8.23.]

MÜHLEISEN, R. (1959). Die luftelektrischen Verhältnisse im Küstenaerosol I, *Arch. Met. Wien* A, **11**, 93–108. [5.63.]

MÜHLEISEN, R. (1961a). Electrode effect measurements above the sea, *J. Atmosph. Terr. Phys.* **20**, 79–81. [8.15.]

MÜHLEISEN, R. (1961b). Über die elektrische Aufladung von Aerosol- und Dunstteilchen, *Geofis. Pur. Appl.* **50**, 140–47. [3.23.]

MÜHLEISEN, R. and FISCHER, H. J. (1958). Radiosonden für luftelektrische Messungen, *Arch. Tech. Mess.* **274**, 229–32. [2.50.]

MÜHLEISEN, R. and HOLL, W. (1952). Eine neue Methode zur Messung der elektrischen Raumladungsdichte der Luft, *Geofis. Pur. Appl.* **22**, 3–8. [6.4.]

MÜHLEISEN, R. and HOLL, W. (1953). Elektrische Aufladung fallender Wassertropfen, *Geofis. Pur. Appl.* **25**, 61–70. [3.11., 16.15.]

MÜLLER-HILLEBRAND, D. (1954). The ion capture on polarized drops, *Ark. Geofis.* **2**, 197–225. [3.11.]

MÜLLER-HILLEBRAND, D. (1959). Lightning counter and the results obtained in Sweden during the thunderstorm period 1958, *T.V.F.* **30**, 1–16. [14.37.]

MÜLLER-HILLEBRAND, D. (1961). The physics of the lightning discharge, *Elektrotech. Z. A*, **82**, 232–49. [14.1., 14.39.]

MÜLLER-HILLEBRAND, D. (1962). The protection of houses by lightning conductors. An historical review, *J. Franklin Inst.* **274**, 34–54. [14.51.]

MÜLLER-HILLEBRAND, D. (1965). Lightning protection, *Problems*, pp. 407–29. [14.51.]

MURPHY, T. (1958). The precision nucleus counter (Pollak type) with automatic recording, *Geofis. Pur. Appl.* **41**, 194–200. [4.8.]

MURRAY, W. H. (1951). *Undiscovered Scotland*, pp. 198–200, J. M. Dent, London. [12.30.]

NAKAYA, U. and TERADA, T. (1934). On the electrical nature of snow particles, *J. Fac. Sci., Hokkaido Univ.* **1**, 181. [10.23.]

NAUER, H. (1953). Model test on ball lightning, *Z. Angew. Phys.* **5**, 441–52. [14.57.]

NEUBERT, U. (1938). Selbsterregender elektrostatischer Generator mit in Pressgas laufenden Ladungsbändern, *Z. Phys.* **110**, 334–51. [5.26.]

NEWMAN, M. M. (1953). Aircraft and lightning, *Thunderstorm Electricity*, pp. 328–34. [14.52.]

NEWMAN, M. M. (1958). Lightning discharge channel characteristics and related atmospherics, *Rec. Adv.*, pp. 475–84. [14.48., 17.14.]

NEWMAN, M. M. (1965). Use of triggered lightning to study the discharge process in the channel and application to V.L.F. propagation studies, *Problems*, pp. 482–90. [14.48., 17.14.]

NICHOLS, E. H. (1916). Investigation of atmospheric electrical variations at sunrise and sunset, *Proc. Roy. Soc.* A, **92**, 401–8. [5.46.]

NICHOLSON, G. F. (1951). An introduction to the rationalized MKS system of units, *Brit. J. Appl. Phys.* **2**, 177–82. [2.3.]

NOLAN, J. J. (1950). The control of the electrical conductivity of the lower atmosphere, *Arch. Met., Wien.* A, **3**, 17–28. [4.36., 4.37.]

NOLAN, J. J., BOYLAN, R. K. and DE SACHY, G. P. (1925). The equilibrium of ions in the atmosphere, *Proc. R. Irish Acad.* A, **37**, 1–12. (4.11., 4.16., 4.28.]

NOLAN, J. J. and ENRIGHT, J. (1922). Experiments on the electrification produced by breaking-up water with special reference to Simpson's theory of the electrification of thunderclouds, *Proc. R. Dublin Soc.* **17**, 1–11. [3.24., 16.21.]

NOLAN, J. J. and GUERINI, V. H. (1933). The diffusion coefficients and velocities of fall in air of atmospheric condensation nuclei, *Proc. R. Irish Acad.* A, **43**, 5-24. [4.38., 4.40.]

NOLAN, J. J. and NOLAN, P. J. (1931). Observations on atmospheric ionisation at Glencree, Co. Wicklow, *Proc. R. Irish Acad.* A, **40**, 11–59. [4.12.]

NOLAN, J. J. and NOLAN, P. J. (1937). Atmospheric electrical conductivity and the current from air to earth, *Proc. R. Irish Acad.* A, **43**, 79–93. [5.21., 7.5., 7.7., 8.15., 17.10.]

NOLAN, J. J. and O'KEEFFE, J. G. (1933). Multiply charged large ions, *Proc. R. Irish Acad.* A, **41**, 26–40. [4.16.]

NOLAN, J. J. and DE SACHY, G. P. (1927). Atmospheric ionisation, *Proc. R. Irish Acad.* A, **37**, 71–94. [4.12., 4.28., 4.32., 4.34., 4.35.]

NOLAN, P. J. (1929). The law of recombination of ions and nuclei, *Proc. R. Irish Acad.* A, **38**, 49–59. [4.38.]

NOLAN, P. J. (1940). Estimation of the air-earth current, *Proc. R. Irish Acad.* A, **46**, 65–75. [7.3., 7.5., 7.7., 8.15., 13.9.]

NOLAN, P. J. (1941). Experiments on condensation nuclei, *Proc. R. Irish Acad.* A, **47**, 25–38. [4.34.]

NOLAN, P. J. (1943). The recombination law for weak ionisation, *Proc. R. Irish Acad.* A, **49**, 67–90. [4.34.]

NOLAN, P. J. (1955). The photo-electric counter and its application for measuring diffusion coefficients, *Wentworth Conf.*, pp. 113–9. [4.30.]

NOLAN, P. J. (1956). The equilibrium of ionization in the atmosphere and nuclear combination coefficients, *J. Atmosph. Terr. Phys.* **9**, 295–303. [4.38.]

NOLAN, P. J. and KUFFEL, E. (1955). Coagulation and diffusion of enclosed nuclei, *Geofis. Pur. Appl.* **31**, 97–106. [4.39.]

NOLAN, P. J. and O'CONNOR, F. C. (1955a). The photoelectric nucleus counter as hygrometer, *Geofis. Pur. Appl.* **31**, 32–6. [4.8.]

NOLAN, P. J. and O'CONNOR, T. C. (1955b). Size, mobility and charge of multiply charged ions, *Proc. R. Irish Acad.* A, **57**, 161–71. [4.16.]

NOLAN, P. J. and POLLAK, L. W. (1947). Calibration of a photo-electric nucleus counter, *Proc. R. Irish Acad.* A, **51**, 9–31. [4.8., 4.9.]

NORINDER, H. (1921). Researches on the height variation of the atmospheric electric potential gradient in the lowest layers of the air, *Geogr. Ann., Stockholm* **1**, 1–96, [6.7., 6.8.]

NORINDER, H. (1947). Some aspects and recent results of electro-magnetic effects of thunderstorms, *J. Franklin Inst.* **244**, 109–30. and 167–207. [14.8.]

NORINDER, H. (1951). *Joint Commission on Radio Meteorology*, Int. Conf. Sci. Un. [14.17.]

NORINDER, H. (1953). Long-distance location of thunderstorms, *Thunderstorm Electricity*, pp. 267–327. [15.4.]

NORINDER, H., ISRAËL, H. and SIKSNA, R. (1954). Das luftelektrische Potentialgefälle und der Vertikalstrom in Uppsala, *Ark. Geofys.* **2**, 109–38. [5.49., 5.59., 5.69.]

NORINDER, H. and KNUDSEN, E. (1961). Lightning discharge paths photographed in daylight and analysed by simultaneous variations of magnetic field components, *Ark. Geofys.* **3**, 375–90. [14.20., 14.31.]

NORINDER, H. and SALKA, O. (1951). Mechanism of positive spark discharges with long gaps in air at atmospheric pressure, *Ark. Fys.* **3**, 347–86. [12.20.]

NORINDER, H. and SIKSNA, R. (1949). The measurements of the density of small ions, *Ark. Geofys.* **1**, 1–32. [4.11.]

NORINDER, H. and SIKSNA, R. (1951). Ionic density of the atmospheric air near the ground during thunderstorm conditions, *Ark. Geofys.* **1**, 453–72. [4.12.]

NORINDER, H. and SIKSNA, R. (1954). Experiments concerning electrification of snow, *Ark. Geofys.* **2**, 59–89. [3.13.]

NORINDER, H. and SIKSNA, R. (1955). Electric charges measured in the air when blowing snow, *Ark. Geofys.* **2**, 343–69. [3.13.]

NOTO, H. (1939). Some studies of thunderstorms. Part IV. The relation between the sign of the potential gradient and that of the charge on rain and snowflakes, *Jap. J. Astro. Geophys.* **17**, 101–18. [10.13.]

NUKIYAMA, D. and NAKATA, K. (1926). Kite-balloon currents, *Proc. Phys. Math. Soc., Japan* **8**, 5. [9.11.]

OBOLENSKY, W. N. (1925). Über elektrische Ladungen in der Atmosphäre, *Ann. Phys., Lpz.* **77**, 644–66. [6.5., 6.8.]

OBOLENSKY, W. N. (1926). Der elektrische Zustand der unteren Atmosphärenschichten an klaren Tagen zu Pavlovsk (1916 bis 1920), *Met. Z.* **43**, 173–7. [8.21.]

O'BROLCHAIN, C. (1930). The comparison of Schweidler's two methods of determining the average life of small ions in air, *Beitr. Geophys.* **29**, 1–21. [4.41.]

OCKENDEN, C. V. (1947). Sferics, *Met. Mag., Lond.* **76**, 78–86. [15.4.]

O'CONNOR, T. C. and FLANAGAN, V. P. V. (1961). The measurement of size distributions in Aitken nuclei, *Geofis. Pur. Appl.* **50**, 60–66. [4.40.]

O'CONNOR, T. C. and SHARKEY, W. P. (1960). Ionization equilibrium in maritime air, *Proc. R. Irish Acad.* A, **61**, 15–27. [4.36.]

O'DONNELL, G. A. (1952). Electric conductivity and small ion concentration of the atmosphere at one metre above ground and conductivity at ground level, *J. Atmosph. Terr. Phys.* **2**, 201–15. [7.8.]

OGAWA, T. (1960a). Types of diurnal variation of the air–earth current, *J. Geomagn.* **9**, 165–73. [8.22.]

OGAWA, T. (1960b). Electricity in rain, *J. Geomagn. Geoelect.* **12**, 21–31. [10.20., 10.27.]

Ogawa, T. and Brook, M. (1964). The mechanism of the intracloud lightning discharge, *J. Geophys. Res.* **69**, 5141–50. [14.30.]

Owolabi, I. E. and Chalmers, J. A. (1965). Correlations of precipitation currents, *J. Atmosph. Terr. Phys.* **27**, 303–8. [10.27.]

Page, H. (1961). The suppression of corona- and precipitation-interference in VHF reception, *Proc. I.E.E.* **108**, B, 469–70. [10.30.]

Paltridge, G. W. (1964). Measurement of the electrostatic field in the stratosphere, *J. Geophys. Res.* **69**, 1947–54. [5.37.]

Paltridge, G. W. (1965). Experimental measurements of the small-ion density and electrical conductivity of the stratosphere, *J. Geophys. Res.* **70**, 2751–61. [4.14., 7.13., 8.18.]

Paramanoff, N. A. (1950). The unitary variation of the potential gradient of atmospheric electricity (in Russian), *C. R. Acad. Sci. USSR* **70**, 37–8. [5.43.]

Parkinson, W. D. (1958). Factors controlling the atmospheric conductivity at the Huancayo Magnetic Observatory, *Terr. Magn. Atmos. Elect.* **53**, 305–17. [4.22.]

Parkinson, W. D. (1959). Review of the literature on the ion balance of the lower atmosphere since 1939, *Gen. Ass. Int. Un. Geod. (Oslo) Ass. Terr. Magn. Atmos. Elect. Bull.* **13**, 307–13. [4.34.]

Pauthenier, M. and Cochet, R. (1953). Évolution d'une gouttellette d'eau chargée dans un nuage à temperature positive, *Rev. Gen. de l'Élect.* **62**, 255–62. [3.11.]

Pearce, D. C. and Currie, B. W. (1949). Some qualitative results on the electrification of snow, *Canad. J. Res.* A, **27**, 1–8. [3.13.]

Peltier, A. (1840). Observations et recherches expérimentales sur les causes qui concourent à la formation des trombes, *C. R. Acad. Sci., Paris* **10**, 712. (9.6., 9.11.]

Peltier, A. (1942). Recherches sur la cause des phénomènes électriques de l'atmosphère et sur les moyen d'en recueiller la manifestation, *Ann. Chim. Phys.* **4**, 385. [1.9.]

Pelton, F. *et al.* (1953). *Investigation of Means to Maintain Zero Electrostatic Charge on Aircraft*, Final Report RA-766-P-10. Cornell Aeronautical Lab. Contract AF-19(122)-475. [5.38.]

Perry, F. R. (1944). Discussion on Multiple lightning strokes, *Quart. J. R. Met. Soc.* **70**, 169. [14.19.]

Perry, F. R., Webster, G. H. and Baguley, P. W. (1942). The measurement of lightning voltages and currents in Nigeria, *J. Instn. Elect. Engrs.* **89**, 185–209. [9.9.]

Peters, W. J. (1923). Results from magnetic east-west paths round the earth, *Terr. Magn. Atmos. Elect.* **28**, 83–8. [8.29.]

Phillips, B. B., Allee, P. A., Pales, J. C. and Woessner, R. H. (1955). An experimental analysis of the effect of air pollution on the conductivity and ion balance of the atmosphere, *J. Geophys. Res.* **60**, 289–96. [4.15., 7.18.]

Phillips, B. B. and Gunn, R. (1954). Measurements of the electrification of spheres by moving ionized air, *J. Met.* **11**, 348–51. [16.18.]

Phillips, B. B. and Kinzer, G. D. (1958). Measurements of the size and electrification of droplets in cumuliform clouds, *J. Met.* **15**, 369–74. [10.18., 13.7.]

PIDDUCK, F. B. and SAS, R. K. (1947). *The Metre-kilogram-second System of Electrical Units*, Methuen, London. [2.3.]

PIERCE, E. T. (1952). Discussion on Thunderbolts, the electric phenomena of thunderstorms, *Nature, Lond.* **169**, 562. [14.9.]

PIERCE, E. T. (1955a). Electrostatic field-changes due to lightning discharges, *Quart. J. R. Met. Soc.* **81**, 211–28. [12.7., 12.9., 14.9., 14.20., 14.21., 14.24., 14.27., 14.28., 14.29., 14.30., 14.31.]

PIERCE, E. T. (1955b). The development of lightning discharges, *Quart. J. R. Met. Soc.* **81**, 229–39. [14.9., 14.11., 14.13., 14.24., 14.27., 14.28.]

PIERCE, E. T. (1956). The influence of individual variations in the field changes due to lightning discharges upon the design and performance of lightning flash counters, *Arch. Met., Wien* A, **9**, 78–86. [14.36., 14.37.]

PIERCE, E. T. (1957). Nuclear explosions and a possible secular variation of the potential gradient in the atmosphere, *J. Atmosph. Terr. Phys.* **11**, 71–2. [5.48.]

PIERCE, E. T. (1958). Some topics in atmospheric electricity, *Rec. Adv.*, pp. 5–16. [4.23., 7.9., 8.20., 11.6., 14.17.]

PIERCE, E. T. (1959). Some calculations on radioactive fallout with especial references to the secular variations in potential gradient at Eskdalemuir, Scotland, *Geofis. Pur. Appl.* **42**, 145–51. [5.48.]

PIERCE, E. T. (1960). Atmospherics from lightning flashes with multiple strokes, *J. Geophys. Res.* **65**, 1867–71. [14.17.]

PIERCE, E. T. (1964). Perturbations produced by jet aircraft in the earth's electric field, *J. Appl. Meteor.* **3**, 805–6. [6.13.]

PIERCE, E. T., ARNOLD, H. R. and DENNIS, A. S. (1962). *Very Low Frequency Atmospherics due to Lightning Flashes*, Report to USAF on Contract AF 33(657)-7009, Stanford Res. Inst. Menlo Park, Cal., U.S.A. [11.19., 14.31., 14.39., 15.7., 15.8.]

PIERCE, E. T., NADILE, R. M. and MCKINNON, P. J. (1960). *An Experimental Investigation of Negative Point-plane Corona and its Relation to Ball Lightning*, Tech. Rep. Contract AF 19(604)-7342, Avco Cpn., Wilmington, Mass. [9.4., 14.57.]

PIERCE, E. T. and WHITSON, A. L. (1964). The variation of potential gradient with altitude above ground of high radioactivity, *J. Geophys. Res.* **69**, 2895–8. [5.37.]

PIERCE, E. T. and WORMELL, T. W. (1953). Field changes due to lightning discharges, *Thunderstorm Electricity*, pp. 251–66. [14.9.]

PLUVINAGE, P. (1946). Étude théorique et expérimentale de la conductibilité électrique dans les nuages non orageux, *Ann. Géophys.* **2**, 31–54 and 160–78. [7.20., 13.8., 13.9.]

PLUVINAGE, P. and ROCHÉ, P. (1947). Le nombre de noyau de condensation à l'interieur d'un nuage, *Ann. Géophys.* **3**, 141–52. [7.20.]

PLUVINAGE, P. and STAHL, P. (1953). La conductibilité électrique dans l'air de l'inlandsis groenlandais, *Ann. Géophys.* **9**, 76–85. [7.21., 8.15.]

POLLAK, L. W. (1952). A condensation nuclei counter with photographic recording, *Geofis. Pur. Appl.* **22**, 1–9. [4.7.]

POLLAK, L. W. (1959). Counting of Aitken nuclei and applications of counting results, *Int. J. Air. Poll.* **1**, 293–306. [4.8.]

POLLAK, L. W. and METNIEKS, A. L. (1957). Photoelectric condensation nucleus counters of high precision for measuring low and very low concentrations of nuclei, *Geofis. Pur. Appl.* **37**, 174–82. [4.8.]

POLLAK, L. W. and METNIEKS, A. L. (1959). New calibration of photo-electric nucleus counters, *Geofis. Pur. Appl.* **43**, 285–301. [4.8.]

POLLAK, L. W. and METNIEKS, A. L. (1962). On the validity of Boltzmann's distribution law for the charges of aerosol particles in electrical equilibrium, *Geofis. Pur. Appl.* **53**, 111–33. [4.37.]

POLLAK, L. W. and MORGAN, W. A. (1940). On various photo-electric nucleus counters, *Irish Met. Serv.* (unpub.) [4.8.]

POLLAK, L. W. and MURPHY, T. (1952). Sampling of condensation nuclei by means of a mobile photo-electric counter, *Arch. Met., Wien* A, **5**, 100–9. [4.8.]

POLLAK, L. W. and O'CONNOR, T. C. (1955a). Decay of aerosols in small and very small vessels and a static method for determining the size of nuclei, *Geofis. Pur. Appl.* **31**, 66–79. [4.39., 4.40.]

POLLAK, L. W. and O'CONNOR, T. C. (1955b). A photoelectric nucleus counter of high precision. *Geofis. Pur. Appl.* **32**, 139–46. [4.8.]

POLLOCK, J. A. (1915a). The nature of large ions in the air, *Phil. Mag.* **29**, 514–26. [4.12.]

POLLOCK, J. A. (1915b). A new type of ion in the air, *Phil. Mag.* **29**, 636–46. [1.14., 4.4.]

POPOV, A. S. (1896). Apparatus for reception and recording of electrical oscillations (in Russian), *J. Russ. Phys. Chem. Soc.* **28**, 7–9. [15.1.]

PRUPPACHER, H. E. (1963a). The effect of an external electric field on the supercooling of water drops, *J. Geophys. Res.* **68**, 4463–74. [3.16.]

PRUPPACHER, H. E. (1963b). The effects of electric fields on cloud physical processes, *Z.A.M.P.* **14**, 590–9. [3.16.]

PÜHRINGER, A. (1961). Die elektromagnetische Induktion als Grundlage einer Gewittertheorie, *Arch. Met., Wien* A, **12**, 262–70. [16.29.]

QUETELET, A. (1849). *Sur le climat de la Belgique*, **3**, 26, Brussels. [9.6.]

RAMSAY, M. W. and CHALMERS, J. A. (1960). Measurements on the electricity of precipitation, *Quart. J. R. Met. Soc.* **86**, 530–9. [10.3., 10.4., 10.13., 10.17., 10.21., 10.22., 10.27.]

RANGS, — (1942). Gerät für Messung der luftelektrischen Feldstärke, *Z.W.B.*, No. 699, BHF Ref. IIB/6022. [5.25., 5.27., 9.11.]

RASTOGI, R. G. (1962). Thunderstorms and sporadic-E layer ionization over Ottawa, Canada, *J. Atmosph. Terr. Phys.* **24**, 533–40. [14.38.]

RAYLEIGH, LORD. (1879). The influence of electricity on colliding water drops, *Proc. Roy. Soc.* **28**, 406–9. [12.24.]

REITER, R. (1955a). Ergebnisse luftelektrischer Messungen auf dem Zugspitzplatt im 2500 m Seehöhe, *Geofis. Pur. Appl.* **30**, 155–9. [5.61., 5.65., 5.66., 13.3.]

REITER, R. (1955b). Einflüsse meteorologischer Feinvorgänge auf das Verhalten luftelektrischer Größen in verschiedenen Höhenstufen, *Met. Rdsch.* **8**, 127–33. [2.26.]

REITER, R. (1957). Das Verhalten luftelektrischer Grössen während Niederschlag in verschiedenen Höhenstufen, synoptisch betrachtet, *Met. Rdsch.* **10**, 63–73. [10.27.]

REITER, R. (1959). Zum Einfluss der natürlichen und künstlichen Radioaktivität des Aerosols auf luftelektrische Elemente unter Berücksichtigung des Grades der Luftverunreinigung, *Z. Aerosol-Forsch u. Ther.* **8**, 3–23. [5.48.]

REITER, R. (1960). Das luftelektrische Erscheinungsbild des Südföhns in den Nordalpen nach synoptisch-klimatologischen Untersuchung im Wettersteingebirge 1955 bis 1959, *Arch. Met., Wien* A, **12**, 72–124. [5.60.]

REITER, R. (1965). Precipitation and cloud electricity, *Quart. J. R. Met. Soc.* **91**, 60–72. [10.21., 10.23., 10.25., 10.28., 13.3., 13.7., 17.12.]

REITER, R. and CARNUTH, W. (1965). An atmospheric-electric feed-back process as a possible contribution to thunderstorm electrification, *J. Atmosph. Terr. Phys.* **27**, 673–705. [16.23.]

REMILLARD, W. J. (1960). *The Acoustics of Thunder*, Acoustics Research Lab. Harvard (Contract Nonr-186624), Tech. Memo. 40, pp. 145. [12.34.]

REYNOLDS, G. T. (1956). Cosmic rays and elementary particles, Notes on lectures at Imperial College, London. [11.15.]

REYNOLDS, S. E. (1953). Thunderstorm precipitation growth and electrical charge generation, *Bull. Amer. Met. Soc.* **34**, 117–23. [16.25., 16.26.]

REYNOLDS, S. E. (1954). *Compendium of Thunderstorm Electricity*, U.S. Signal Corp. Research Report (Socorro, N.M.). [3.3., 3.6., 3.17., 3.20., 10.19., 12.20., 12.23., 14.38., 16.23., 16.25., 16.30., 16.31., 16.40., 16.43.]

REYNOLDS, S. E. and BROOK, M. (1956). Correlation of the initial electric field and the radar echo in thunderstorms, *J. Met.* **13**, 376–80. [12.24.]

REYNOLDS, S. E. and BROOK, M. (1962). Thunderstorm electrification, *Quart. J. R. Met. Soc.* **88**, 550–1. [16.23.]

REYNOLDS, S. E., BROOK, M. and GOURLEY, M. F. (1957). Thunderstorm charge separation, *J. Met.* **14**, 426–36. [3.14., 3.17., 16.23.]

REYNOLDS, S. E. and NEILL, H. W. (1955). The distribution and discharge of thunderstorm charge centers, *J. Met.* **12**, 1–12. [5.28., 12.8., 14.24.]

RICH, T. A. (1955). A photo-electric nucleus counter with size discrimination, *Geofis. Pur. Appl.* **31**, 60–65. [4.8.]

RICH, T. A., POLLAK, L. W. and METNIEKS, A. L. (1962). On the time required for aerosols to reach electrical equilibrium, *Geofis. Pur. Appl.* **51**, 217–24. [4.36.]

RIECKE, E. (1903). Beiträge zur Lehre der Luftelektrizität, *Ann. Phys.* **12**, 52–84. [7.2.]

ROBINSON, G. D. (1951). Thunderstorm electrification, *Met. Mag., Lond.* **80**, 15–18. [12.18.]

DE ROMAS, J. (1753). Mémoire ou, après avoir donné un moyen aisé pour élever fort haut, et à peu de frais un corps électrisable isolé, on rapport les observations frappantes qui prouvent que plus le corps isolé est élevé au-dessus de la Terre, plus la feu d'électricité est abondant, *Mem. Acad., Bordeaux*. [1.3., 9.11.]

ROSSMANN, F. (1948). Vom Ursprung der Gewitterelektrizität, *Met. Rdsch.* **1**, 193–5. [16.29.]

ROSSMANN, F. (1950). Luftelektrische Messung mittels Segelflugzeugen, *Ber. dtsch. Wetterd., U.S.Zone*, No. 15. [5.39., 7.14., 12.18., 17.18.]

RUDGE, W. A. D. (1914). On some sources of disturbance of the normal atmospheric potential gradient, *Proc. Roy. Soc.* A, **90**, 571–82. [5.62.]

RUHNKE, L. H. (1961 a). *The Effect of Mismatching in the Measurement of Air–Earth Current Density*, U.S.A.S.R.D.L. Tech. Rep. 2232. [8.9.]

RUHNKE, L. H. (1961 b). The change of the small ion density due to condensation nuclei and the relation to the extinction coefficient of light, *Proc. Int. Conf. on Ionization of the Air*, pp. 2–13. [4.38., 4.40.]

RUHNKE, L. H. (1962). Electrical conductivity of air on the Greenland ice-cap, *J. Geophys. Res.* **67**, 2767–72. [7.21., 8.15.]

RUMI, G. G. (1957). VHF radar echoes associated with atmospheric phenomena, *J. Geophys. Res.* **62**, 547–64. [14.38.]

RUSSELTVEDT, N. (1925). Instrumente und Apparate für die luftelektrischen Untersuchungen an dem meteorologischen Observatorium in As, *Beih. Jb. Norweg. Met. Inst.*, 11–15. [5.29.]

RUTTENBERG, S. and HOLZER, R. E. (1955). Atmospheric electrical measurements in the Pacific Ocean, *Wentworth Conf.*, pp. 101–8. [5.40., 8.28.]

SAGALYN, R. G. (1958 a). The production and removal of small ions and charged nuclei over the Atlantic Ocean, *Rec. Adv.* pp. 21–41. [7.16., 7.18.]

SAGALYN, R. C. (1958 b). Significance of the ratio of the polar conductivities in regions of variable pollution content, *Rec. Adv.*, pp. 235–7. [7.18.]

SAGALYN, R. C. (1965). Space electricity: physical problems and experimental techniques, *Problems*, pp. 548–65. [5.72.]

SAGALYN, R. C. and FAUCHER, G. F. (1954). Aircraft investigation of the large ion content and conductivity of the atmosphere and their relation to meteorological factors, *J. Atmosph. Terr. Phys.* **5**, 253–72. [4.13., 5.39., 7.14., 7.17., 7.21., 8.25.]

SAGALYN, R. C. and FAUCHER, G. F. (1956). Space and time variations of charged nuclei and electrical conductivity of the atmosphere, *Quart. J. R. Met. Soc.* **82**, 428–43. [4.39., 8.22.]

SALANAVE, L. E. (1965). The photographic spectrum of lightning: determinations of channel temperature from slitless spectra, *Problems*, pp. 371–83. [14.54.]

SALANAVE, L. E. and BROOK, M. (1965). Lightning photography and counting in daylight, *J. Geophys. Res.* **70**, 1285–9. [14.37.]

SAPSFORD, H. B. (1937). Influence of pollution on potential gradient at Apia, *Terr. Magn. Atmos. Elect.* **42**, 153–8. (5.44.]

SARTOR, D. (1954). A laboratory investigation of collision efficiencies, coalescence and electrical charging of simulated cloud droplets, *J. Met.* **11**, 91–103. [16.14.]

SARTOR, J. D. (1960). Some electrostatic cloud-droplet collision efficiencies, *J. Geophys. Res.* **65**, 1953–58. [12.24.]

SARTOR, J. D. (1961). Calculations of cloud electrification based on a general charge-separation mechanism, *J. Geophys. Res.* **66**, 831–8. [16.14.]

SARTOR, J. D. (1963). Radio emission from clouds, *J. Geophys. Res.* **68**, 5169–72. [15.10.]

SARTOR, J. D. (1965). Induction charging thunderstorm mechanism, *Problems*, pp. 307–10. [16.14.]

DE SAUSSURE, H. B. (1779). *Voyages dans les Alpes* (Geneva). [1.6.]

SAYERS, J. (1938). Ionic recombination in air, *Proc. Roy. Soc.* A, **169**, 83–101. [4.34.]

SCHAEFER, V. J. (1955). Atmospheric electricity associated with jet streams, *Wentworth Conf.*, pp. 59–66. [5.14., 5.71.]

SCHERING, H. (1906). Der Elster-Geitelsche Zerstreuungsapparat und ein Versuch quantitativer absoluter Zerstreuungsmessungen, *Ann. Phys., Lpz.* **20**, 174–95. [7.2., 7.4.]

SCHILLING, G. F. (1955). On the variation of electrical conductivity of air with elevation, *Wentworth Conf.*, pp. 53–7. [7.21.]

SCHILLING, G. F. and CARSON, J. (1953). On the electrical conductivity of air inside buildings, *Arch. Met., Wien* B, **5**, 52–7. [2.37., 7.23., 7.24.]

SCHILLING, G. F. and HOLZER, R. E. (1954). Über mögliche Zusammenhänge zwischen Bioklimatologie und Luftelektrizität, *Wett. u. Leben.* **6**, 13–16. [2.37., 7.23., 7.24.]

SCHINDELHAUER, F. (1913). Über die Elektrizität der Niederschläge, *Phys. Z.* **14**, 1292. [10.2., 10.23., 11.13.]

SCHMIDT, A. (1924). Zur Frage der elektrischen Vertikalströme, *Z. Geophys.* **1**, 281–4. [8.29.]

SCHMIDT, A. (1939). Zur Frage der hypothetischen, die Erdoberfläche durchdringenden elektrischen Ströme, *Beitr. Geophys.* **55**, 292–302. [8.29.]

SCHOLZ, J. (1931 a). Ein neuer Apparat zur Bestimmung der Zahl der geladenen und ungeladenen Kerne, *Z. InstrumKde.* **51**, 505–22. [4.7.]

SCHOLZ, J. (1931 b). Theoretische Untersuchungen über die Feld- und Ionenverteilung in einem stromdurchflossenen Gas, das auch schwer bewegliche Elektrizitätsträger enthält, *S.B. Akad. Wiss., Wien* **140**, 49–66. [8.19.]

SCHOLZ, J. (1935). Polarlichtuntersuchungen auf Franz-Josefs-Land, *Beitr. Geophys.* **44**, 145–56. [5.56.]

SCHONLAND, B. F. J. (1928a). The polarity of thunderclouds, *Proc. Roy. Soc.* A, **118**, 233–51. [12.1., 12.4., 12.7., 14.31.]

SCHONLAND, B. F. J. (1928b). The interchange of electricity between thunderclouds and the earth, *Proc. Roy. Soc.* A, **118**, 252–62. [9.6., 9.8., 9.14., 9.15., 9.24., 10.4., 10.8., 10.11.]

SCHONLAND, B. F. J. (1930). Thunderstorms and the penetrating radiation, *Proc. Roy. Soc.* A, **130**, 37–63. [12.31.]

SCHONLAND, B. F. J. (1932). *Atmospheric Electricity*, pp. 100, Methuen, London [1.17.]

SCHONLAND, B. F. J. (1937). The diameter of the lightning channel, *Phil. Mag.* **23**, 503–8. [14.18.]

SCHONLAND, B. F. J. (1938). Progressive lightning. IV. *Proc. Roy. Soc.* A, **164**, 132–50. [14.3., 14.11., 14.13., 14.19.]

SCHONLAND, B. F. J. (1950). *The Flight of Thunderbolts*, Clarendon Press, Oxford. [14.56.]

SCHONLAND, B. F. J. (1953a). *Atmospheric Electricity*, 2nd Ed., Methuen, London. pp. 95. [Pref., 5.44.]

SCHONLAND, B. F. J. (1953b). The pilot streamer in lightning and the long spark. *Proc. Roy. Soc.* A, **220**, 25–38. [14.11., 14.13., 14.34.]

SCHONLAND, B. F. J. (1956). The lightning discharge, *Handbuch der Physik* (ed. S. FLÜGGE) **22**, 576–628. [14.1., 14.20., 14.21.]

SCHONLAND, B. F. J. (1962). Lightning and the long electric spark, *Advancement of Science*, **19**, 1–8. [14.13., 14.28., 14.53., 14.57.]

SCHONLAND, B. F. J. (1964). *The Flight of Thunderbolts*, 2nd Edn., Clarendon Press, Oxford. [1.1., 1.3., 1.4., 14.1., 14.34., 14.51.]

SCHONLAND, B. F. J. and ALLIBONE, T. E. (1931). Branching of lightning, *Nature, Lond.* **128**, 794–5. [1.18., 14.41.]

SCHONLAND, B. F. J. and COLLENS, H. (1934). Progressive lightning, *Proc. Roy. Soc.* A, **143**, 654–74. [14.2., 14.3.]

SCHONLAND, B. F. J. and CRAIB, J. (1927). The electric field of South African thunderstorms, *Proc. Roy. Soc.* A, **114**, 229–43. [12.7.]

SCHONLAND, B. F. J. and ELDER, J. S. (1941). Anticipatory triggering devices for lightning and static investigations, *J. Franklin Inst.* **231**, 39–47. [14.27.]

SCHONLAND, B. F. J., ELDER, J. S., HODGES, A. B., PHILLIPS, W. E. and VAN WYK, J. W. (1940). The wave-form of atmospherics at night, *Proc. Roy. Soc.* A, **176**, 180–202. [15.1.]

SCHONLAND, B. F. J., HODGES, D. B. and COLLENS, H. (1938). Progressive lightning. V. A comparison of photographic and electrical studies, *Proc. Roy. Soc.* A, **166**, 56–75. [14.3., 14.9.]

SCHONLAND, B. F. J. and MALAN, D. J. (1954). Upward stepped leaders from the Empire State Building, *J. Franklin Inst.* **258**, 271–5. [14.39., 14.47., 17.14.]

SCHONLAND, B. F. J., MALAN, D. J. and COLLENS, H. (1935). Progressive lightning. II. *Proc. Roy. Soc.* A, **152**, 595–625. [14.3.]

SCHONLAND, B. F. J., MALAN, D. J. and COLLENS, H. (1938). Progressive lightning. VI. *Proc. Roy. Soc.* A, **168**, 455–69 [14.3.]

SCHONLAND, B. F. J. and VILJOEN, J. P. T. (1933). On a penetrating radiation from thunderclouds, *Proc. Roy. Soc.* A, **140**, 314–33. [12.31.]

SCHUCHARD, E. A. (1935). *Univ. of Washington Engrng. Expt. Sta. Bull.*, No. 80 (Seattle). [5.26., 5.27.]

SCHUMANN, T. E. W., (1925). On Elster and Geitel's theory of the electrification of rain drops, *Phys. Rev.* **26**, 105–10. [16.14.]

VON SCHWEIDLER, E. (1908). Über die Ionenverteilung in den untersten Schichten der Atmosphäre, *S.B. Akad. Wiss., Wien* **117**, 653–64. [8.19.]

VON SCHWEIDLER, E. (1918, 1919). Über das Gleichgewicht zwischen ionenerzeugenden und ionenvernichtenden Vorgängen in der Atmosphäre, *S.B. Akad. Wiss., Wien* **127**, 953–67; **128**, 947–955. [4.41.]

VON SCHWEIDLER, E. (1924). Über die Charakteristik des Stromes in schwach ionisierten Gasen, *S.B. Akad. Wiss., Wien* **133**, 23–7. [4.41.],

VON SCHWEIDLER, E. (1929). *Einführung in die Geophysik*, **2**, 291–375, Springer Berlin. [5.37., 8.19.]

SCORER, R. S. (1963). Cloud nomenclature, *Quart. J. R. Met. Soc.* **89**, 248–53. [3.2.]

SCRASE, F. J. (1933). The air–earth current at Kew Observatory, *Geophys. Mem., Lond.* **58** [7.10., 8.7., 8.21., 8.24., 11.1., 11.10., 13.9.]

SCRASE, F. J. (1934). Observations of atmospheric electricity at Kew. A survey of results obtained from 1843 to 1931, *Geophys. Mem., Lond.* **60.** [5.10., 5.18., 5.42., 5.47.]

SCRASE, F. J. (1935a). The charged and uncharged nuclei in the atmosphere and their part in atmospheric ionization, *Geophys. Mem., Lond.* **64.** [4.34., 4.35., 7.10., 8.24.]

SCRASE, F. J. (1935b). Some measurements of the variation of potential gradient with height near the ground at Kew Observatory, *Geophys. Mem., Lond.* **67**. [6.7., 6.8., 8.15.]

SCRASE, F. J. (1938). Electricity on rain, *Geophys. Mem., Lond.* **75**. [10.2., 10.4., 10.6., 10.8., 10.9., 10.13., 10.14., 11.13., 12.29.]

SERBU, G. P. and TRENT, E. M. (1958). A study of the use of atmospheric electric measurements in fog forecasting, *Trans. Amer. Geophys. Un.* **39**, 1034–2. [7.12.]

SHEPPARD, P. A. (1933). Discussion of results, atmospheric electricity, *Brit. Polar Yr. Exped. Fort Rae*, Roy. Soc., London, pp. 311–8. [5.56., 5.66., 8.24.]

SHEPPARD, P. A. (1950). The cumulo-nimbus and thunderstorm. II. Electrical phenomena, *Sci. Progr., Lond.* **88**, 488–96. [16.17.]

SHERMAN, K. L. (1934). *Trans. Amer. Geophys. Un.* **15**, 141 [5.56.]

SHERMAN, K. L. (1937). Atmospheric electricity at College Fairbanks Polar Year Station, *Ferr. Magn. Atmos. Elect.* **42**, 371–90. [5.43.]

SHERMAN, K. L. and GISH, O. H. (1937). Electrical potential gradient and conductivity of air near Rapid City, South Dakota, *Terr. Magn. Atmos. Elect.* **42**, 289–99. [5.43.]

SHISHKIN, N. S. (1965). The role of the coagulation of charged particles in the development of thunderstorm phenomena, *Problems*, pp. 268–79. [16.19.]

SILBERG, P. A. (1965). A review of ball lightning, *Problems*, pp. 436–54. [14.56., 14.57.]

SIMPSON, G. C. (1905). Atmospheric electricity in high latitudes, *Phil. Trans.* A, **205**, 61–97. [5.43.]

SIMPSON, G. C. (1909). On the electricity of rain and its origin in thunderstorms, *Phil. Trans.* A, **209**, 379–413. [3.24., 10.2., 10.9., 10.13., 10.23., 11.13., 16.21., 16.36., 16.39.]

SIMPSON, G. C. (1910). Earth–air electric currents, *Phil. Mag.* **19**, 715–25. [8.6.]

SIMPSON, G. C. (1915). The electricity of atmospheric precipitation, *Phil. Mag.* **30**, 1–12. [16.40.]

SIMPSON, G. C. (1919). *Brit. Antarc. Exped.* 1910–13. *Meteorology* (*Calcutta*), **1**, 302–12. [3.13., 5.43., 5.66., 16.23.]

SIMPSON, G. C. (1926). On lightning, *Proc. Roy. Soc.* A, **111**, 56–67. [14.41.]

SIMPSON, G. C. (1927). The mechanism of a thunderstorm, *Proc. Roy. Soc.* A, **114**, 376–401. [1.18., 12.1., 12.6., 16.21., 16.36., 16.39.]

SIMPSON, G. C. (1942). The electricity of cloud and rain, *Quart. J. R. Met. Soc.* **68**, 1–34. [10.8., 10.23.]

SIMPSON, G. C. (1949). Atmospheric electricity during disturbed weather, *Geophys. Mem., Lond.* **84**, 1–51. [5.64., 5.66., 5.68., 5.69., 9.24., 10.2., 10.4., 10.13., 10.14., 10.16., 10.17., 10.20., 10.21., 10.22., 10.23., 10.27., 11.20., 16.7., 16.38., 16.40.]

SIMPSON, G. C. and ROBINSON, G. D. (1940). The distribution of electricity in thunderclouds II, *Proc. Roy. Soc.* A, **177**, 281–329. [9.20., 9.24., 12.4., 12.15., 12.21., 14.24.]

SIMPSON, G. C. and SCRASE, F. J. (1937). The distribution of electricity in thunderclouds, *Proc. Roy. Soc.* A, **161**, 309–52. [1.18., 3.13., 9.20., 9.21., 12.4., 12.15., 16.23., 16.31., 16.40.]

SIMPSON, G. C. and WRIGHT, C. S. (1911). Atmospheric electricity over the ocean, *Proc. Roy. Soc.* A, **85**, 175–99. [5.65.)

SIVARAMAKRISHNAN, M. V. (1952). Changes of atmospheric electric potential gradient at Poona during disturbed weather, *Indian J. Met. Geophys.* **4**, 62–75. [5.69.]

SIVARAMAKRISHNAN, M. V. (1957). Point discharge current, the earth's electric field and rain charges during disturbed weather at Poona, *Indian J. Met. Geophys.* **8**, 379–90. [9.9., 9.13., 9.15., 10.27.]

SIVARAMAKRISHNAN, M. V. (1960). The relation between raindrop size distribution, rate of rainfall and the electric charge carried down by rain in the tropics, *Indian J. Met. Geophys.* **11**, 258–68. [10.20., 10.21., 16.7.]

SIVARAMAKRISHNAN, M. V. (1961). The origin of electric charge carried by thunderstorm rain, *Indian J. Met. Geophys.* **12**, 447–64. [10.13., 10.14., 10.16., 10.20.]

SMIDDY, M. and CHALMERS, J. A. (1958). The double fieldmill, *J. Atmosph. Terr. Phys.* **12**, 206–10. [5.32., 6.7.]

SMIDDY, M. and CHALMERS, J. A. (1960). Measurement of space charge in the lower atmosphere, *Quart. J. R. Met. Soc.* **86**, 79–84. [3.24., 6.5., 6.8., 6.11., 8.15.]

SMITH, L. G. (1951a). Dissertation *Rain Electricity*, Cambridge Univ. [5.27., 9.24., 12.12., 16.37.]

SMITH, L. G. (1951b). In Discussion on Cloud Physics, *Quart. J. R. Met. Soc.* **77**, 683. [16.7., 17.13.]

SMITH, L. G. (1953). On the calibration of conductivity meters, *Rev. Sci. Instrum.* **24**, 998. [7.3.]

SMITH, L. G. (1954). Electric field meter with extended range, *Rev. Sci. Instrum.* **25**, 510–13. [5.25., 5.27., 5.34., 12.4., 12.7.]

SMITH, L. G. (1955). The electric charge of raindrops, *Quart. J. R. Met. Soc.* **81**, 23–47. [5.65., 10.5., 10.12., 10.14., 10.17., 10.20., 10.26., 16.40., 17.12.]

SMITH, L. G. (1957). Intracloud lightning discharges, *Quart. J. R. Met. Soc.* **83**, 103–11. [14.30.]

SMITH, L. G. (ed.) (1958). *Recent Advances in Atmospheric Electricity*, Proceedings of 2nd Wentworth Conference, Pergamon Press, New York. [1.19.]

SMITH, L. G. (1959). Air-earth current by Wilson's method, (Unpublished). [8.5.]

SMYTHE, F. S. (1950). *Climbs in the Canadian Rockies*, pp. 251–3, Hodder and Stoughton, London. [12.30.]

SOHNCKE, L. (1886). Der Ursprung der Gewitterelektrizität und der gewöhnlichen Elektrizität der Atmosphäre, *Ann. Phys., Lpz.* **28**, 551. [3.18.]

SOHNCKE, L. (1888). Gewitterelektrizität und gewöhnliche Luftelektrizität, *Met. Z.* **5**, 413–25. [16.22.]

STÄGER, A. (1925a). Experimentaluntersuchungen über Kontaktelektrisierung von staub- und wolkenförmig zerteilten Körpern, speziell von Schnee als gewitterbildenden Faktor, *Ann. Phys., Lpz.* **76**, 49–70. [3.13.]

STÄGER, A. (1925b). Weitere Untersuchungen über Kontaktelektrisierung bei fein zerteilten Körpern, besonders bei Schnee, *Ann. Phys., Lpz.* **77**, 225–40. [3.13.]

STEKOLNIKOV, L. J. and VALEEV, C. (1937). *Confl. Int. Gr. Res. Élect., Paper,* **330**. [14.44.]

STERGIS, C. G. (1954). Study of atmospheric ions in a non-equilibrium system. *J. Geophys. Res.* **59**, pp. 63–6. [4.36.]

STERGIS, C. G. and DOYLE, J. W. (1958). Location of near lightning discharges, *Rec. Adv.*, pp. 589–97. [15.5.]

STERGIS, C. G., CORONITI, S. C., NAZAREK, A., KOTES, D. E., SEYMOUR, D. W. and WERME, J. V. (1955). Conductivity measurements in the stratosphere, *J. Atmosph. Terr. Phys.* **6**, 233–42. [7.13., 12.22.]

STERGIS, C. G., REIN, G. C. and KANGAS, T. (1957a). Electric field measurements in the stratosphere, *J. Atmosph. Terr. Phys.* **11**, 77–82. [5.37., 7.13., 11.5.]

STERGIS, C. G., REIN, G. C. and KANGAS, T. (1957b). Electric field measurements above thunderstorms, *J. Atmosph. Terr. Phys.* **11**, 83–90. [9.8., 9.24., 11.6., 11.17., 11.19., 12.28., 14.38., 16.2., 17.2., 17.9., 17.13.]

STEWART, K. H. (1960). Some recent changes in atmospheric electricity and their cause, *Quart. J. R. Met. Soc.* **86**, 399–405. [5.48.]

STOCKILL, J. (1954). Electric charge or rain, (Unpublished) [10.3.]

STOREY, L. R. O. (1953). An investigation of whistling atmospherics, *Phil. Trans. Roy. Soc.* A, **246**, 113–41. [15.11.]

STOTT, D. and HUTCHINSON, W. C. A. (1965). The electrification of freezing water drops, *Quart. J. R. Met. Soc.* **91**, 80–6. [3.15.]

SULLIVAN, A. W. and WELLS, J. D. (1957). A lightning stroke counter, *Bull. Amer. Met. Soc.* **38**, 291–4. [14.37.]

SVERDRUP, H. U. (1927). *Magnetic, Atmospheric Electric and Auroral Results, Maud Expedition, 1918-25*, Publ. Carnegie Instn. Washington D.C., No. 175, **6**, 425–60. [5.66.]

SWANN, W. F. G. (1913). The atmospheric potential gradient and a theory as to the cause of its connection with other phenomena in atmospheric electricity, *Terr. Magn. Atmos. Elect.* **18**, 163–82. [8.19.]

SWANN, W. F. G. (1914a). The theory of electrical dispersion into the free atmosphere, with a discussion of the theory of the Gerdien conductivity apparatus, and of the theory of the collection of radioactive deposit by a charged conductor, *Terr. Magn. Atmos. Elect.* **19**, 81–92. [7.2.]

SWANN, W. F. G. (1914b). On certain new atmospheric-electric instruments and methods, *Terr. Magn. Atmos. Elect.* **19**, 171–85. [5.40.]

SWANN, W. F. G. (1915). The atmospheric-electric observations on the third cruise of the *Carnegie*, 1914, Report and discussion, *Terr. Magn. Atmos. Elect.* **20**, 13–48. [5.65.]

SWANN, W. F. G. (1955). The present state of atmospheric electricity, *Wentworth Conf.*, pp. 3–10. [1.14., 11.4.]

TAKAGI, M. (1961). The mechanism of discharges in a thundercloud, *Proc. Res. Inst. Atmospherics, Nagoya Univ.* **8** B, 1–106. [14.24., 14.25., 14.30.]

TAKAHASHI, T. (1962). Electric charge generation by breaking of ice piece, *J. Met. Soc., Japan* **40**, 277–86. [3.15., 16.23.]

TAKEUTI, T., ISHIKAWA, H. and TAKAGI, M. (1958). On the mill type electric field meter, *Proc. Res. Inst. Atmospheric, Nagoya Univ.* **5**, 64–70. [5.25.]

TAKEUTI, T., ISHIKAWA, H. and TAKAGI, M. (1960). On the cloud discharge preceding the first ground stroke, *Proc. Res. Inst. Atmospherics, Nagoya Univ.* **7**, 1–6. [14.10.]

TAMM, F. (1901). Über den Einfluß des Luftdruckes und der Luftfeuchtigkeit auf die Entladung statischer Elektrizität aus Spitzen, *Ann. Phys. Lpz.* **6**, 259–79. [9.21.]

TAMURA, Y. (1954). An analysis of electric field after lightning discharges, *J. Geomagn. Geoelect., Kyoto* **6**, 34–46. [12.11. 12.12.]

TAMURA, Y. (1956). *Electrification of Minor Shower Clouds*, Report on Contract AF 62(502)-1219. [13.6.]

TAMURA, Y. (1958). Investigations on the electrical structure of thunderstorms, *Rec. Adv.*, pp. 269–76. [12.8. 14.24.]

THOMAS, D. G. A. and FINCH, H. W. (1950). A simple vibrating condenser voltmeter, *Electronic Engrg.* **22**, 395–8. [2.47.]

THOMSON, J. J. (1898). On the charge of electricity carried by the ions produced by Röntgen rays, *Phil. Mag.* **46**, 528–45. [16.13.]

THOMSON, J. J. and THOMSON, G. P. (1928. 1933). *The Conduction of Electricity through Gases*, Univ. Press, Cambridge. [4.1.]

TOMMASINA, M. T. (1900). *Atti. Accad. Gionia di. sci. nat., Catonia* **1**. [15.1.]

TORRESON, O. W. (1939). Condensation nuclei in the atmosphere in Peru, and their relation to atmospheric-electrical and meteorological observations, *Terr. Magn. Atmos. Elect.* **44**, 59–74. [4.2., 8.24.]

TORRESON, O. W., GISH, O. H., PARKINSON, W. C. and WAIT, G. R. (1946). *Scientific Results of Cruise VII of the Carnegie during 1928-29*, Pub. Carnegie Instn., Washington, D.C., No. 568. [11.8., 11.19.]

TORRESON, O. W. and WAIT, G. R. (1934). Measurements of total nuclei, of uncharged nuclei and of large ions in the free atmosphere at Washington, D.C., *Terr. Magn. Atmos. Elect.* **39**, 47–64. [4.35.]

TORRESON, O. W. and WAIT, G. R. (1948). Atmospheric electric results from Huancayo Observatory, Peru. 1924–34, *Res. Dept. Terr. Magn. Pub.*, 175. [5.45.]

TRICHEL, G. W. (1938). The mechanism of the negative point to plane corona near onset, *Phys. Rev.* **54**, 1078–84. [9.4.]

TRUMP, J. G., SAFFORD, F. J. and VAN DE GRAAFF, R. J. (1940). Generating voltmeter for pressure-insulated H. V. sources, *Rev. Sci. Instrum.* **11**, 54–6. [5.25.]

TUMA, J. (1899). Luftelektrizitätsmessungen im Luftballon, *S.B. Akad. Wiss., Wien* **108**, 227–60. [1.12.]

TURPAIN, A. (1905). Appareil pour l'observation et l'enregistrement automatique des orages, *J. Phys.* **4**, 512–3. [15.1.]

TWOMEY, S. (1956). The electrification of individual cloud droplets, *Tellus* **8**, 445–51. [13.7., 16.26.]

TWOMEY, S. (1957). Electrical charge separation in subfreezing cumuli. *Tellus* **9**, 384–93. [16.26.]

UCHIKAWA, K. (1961). Atmospheric electric phenomena in the upper air over Japan (Part I and II), *Geophys. Mag. Japan.* **30**, 617–72. [8.18.]

UMAN, M. A. (1962). Bead lightning and the pinch effect, *J. Atmosph. Terr. Phys.* **24**, 43–5. [14.55.]

UMAN, M. A. (1964). The diameter of lightning, *J. Geophys. Res.* **69**, 583–95. [14.18.]

VENKITESHWARAN, S. P., DHAR, N. C. and HUDDAR, B. B. (1953). On the measurement of the electrical potential gradient in the upper air over Poona by radiosondes. *Proc. Indian Acad. Sci.* **37**, 260. [5.37.]

VENKITESHWARAN, S. P., GUPTA, B. K. and HUDDAR, B. B. (1954). Measurement of the electrical potential gradient in the atmosphere by radiosonde, *Indian J. Met. Geophys.* **5**, 253–6. [7.3.]

VENTKITESWARAN, S. P. and HUDDAR, B. B. (1956). Variation of electrical potential gradient at Poona from 31st October, 1953 to 2nd November, 1953, *Indian J. Met. Geophys.* **7**, 61–4. [5.37.]

VERZÁR, F. (1953). Kondensationskernzähler mit automatischer Registrierung, *Arch. Met., Wien* A, **5**, 372–6. [4.8.]

VOGLER, G. (1959). Über den Einfluß eines Emissionseffektes auf das Messergebnis der elektrischen Leitfähigkeit der Luft, *Geofis. Pur. Appl.* **43**, 250–4. [7.18.]

VOLTA, A. (1782). Del modo di render sensibilissima la più debole electricità sia naturale sia artificale, *Phil. Trans. Roy. Soc.* **72**, 237–80. (1.7., 1.8., 3.21., 5.16.]

VONNEGUT, B. (1955). Possible mechanism for the formation of thunderstorm electricity, *Wentworth Conf.*, pp. 169–81. [16.32.]

VONNEGUT, B. (1956). *The Electrical Theory of Tornadoes*, Report on Contract Nonr-1684(00). [12.33.]

VONNEGUT, B. (1960). Electrical theory of tornadoes, *J. Geophys. Res.* **65**, 203–12. [12.33.]

VONNEGUT, B. (1965). Electrification of frost deposits, *Quart. J. R. Met. Soc.* **91**, 372–4. [16.32.]

VONNEGUT, B. and MCCAIG, D. A. (1960). Airplane instrument for measurement and vectorial presentation of electrical potential gradient, *J. Geophys. Res.* **65**, 1959–64. [5.38.]

VONNEGUT, B., MAYNARD, K., SYKES, W. G. and MOORE, C. B. (1961). Technique for introducing low-density space charge into the atmosphere, *J. Geophys. Res.* **66**, 823–30. [6.16.]

VONNEGUT, B. and MOORE, C. B. (1958a). *Giant Electrical Storms*, Report on Contract Nonr-1684(00). [12.33.]

VONNEGUT, B. and MOORE, C. B. (1958b). *A Study of Techniques for Measuring the Concentration of Space Charge in the Lower Atmosphere*, Report on Contract AF 19(604)-1920, pp. 1–163. [5.28., 5.29., 6.3., 6.4., 6.5., 6.6.]

VONNEGUT, B. and MOORE, C. B. (1958c). Preliminary attempts to influence convective electrification in cumulus clouds by the introduction of space charge into the lower atmosphere, *Rec. Adv.*, pp. 317–31. [6.16., 16.32.]

VONNEGUT, B. and MOORE, C. B. (1961). Apparatus using radioactive probes for measuring the vertical component of atmospheric potential gradient from an airplane, *Bull. Amer. Met. Soc.* **42**, 773–7. [5.8.]

VONNEGUT, B., MOORE, C. B. and BLUME, M. (1957). Preliminary investigation of the distribution of space charge in the lower atmosphere, *Art. Stim. Rain*, pp. 131–41. [5.38., 6.12.]

VONNEGUT, B., MOORE, C. B. and HARRIS, C. K. (1956). *Stabilization of a High-voltage Discharge by a Vortex*, Report on Contract Nonr-1684(00). [12.33.]

VONNEGUT, B., MOORE, C. B. and HARRIS, C. K. (1961). Agrimeter for measurement of atmospheric electrical potential gradient, *J. Met.* **18**, 812–5. [5.29.]

VONNEGUT, B., MOORE, C. B., SEMONIN, R. G., BULLOCK, J. W., STAGGS, D. W. and BRADLEY, W. E. (1962). Effect of atmospheric space charge on initial electrification of cumulus clouds, *J. Geophys. Res.* **67**, 3909–27. [16.33.]

VONNEGUT, B., MOORE, C. B., STOUT, G. E., STAGGS, D. W., BULLOCK, J. W. and BRADLEY, W. E. (1962). Artificial modification of atmospheric space charge, *J. Geophys. Res.* **67**, 1073–83. [6.16.]

WADDEL, R. C. (1948). An electric field meter for use on aeroplanes, *Rev. Sci. Instrum.* **19**, 31–5. [5.25., 5.27.]

WAGNER, C. F. and MCCANN, G. D. (1940). Lightning phenomena, *Elect. Engng., N.Y.* **60**, 438–43. [14.43.]

WAGNER, C. F., MCCANN, C. D. and BECK, E. (1940). Field investigations on lightning, *Trans. Amer. Inst. Elect. Engrs.* **60**, 1222–9. [14.43., 14.45.]

WAIT, G. R. (1931). Diurnal variation of concentration of condensation nuclei and of certain atmospheric-electric elements at Washington, D.C., *Terr. Magn. Atmos. Elect.* **36**, 111–31. [4.22.]

WAIT, G. R. (1932). The Aitken pocket nuclei-counter, *Beitr. Geophys.* **37**, 429–39. [4.7.]

WAIT, G. R. (1935a). Computed and observed rates of small-ion production in the atmosphere, *Terr. Magn. Atmos. Elect.* **40**, 209–14. [4.35.]

WAIT, G. R. (1935b). The intermediate ion of the atmosphere, *Phys. Rev.* **48**, 383. [4.4.]

WAIT, G. R. (1946). Some experiments relating to the electric conductivity of the lower atmosphere, *J. Wash. Acad. Sci.* **36**, 321–43. [4.34.]

WAIT, G. R. (1950). Measurements by airplane of the electric charge passing vertically through thunderstorms to ground, *Arch. Met., Wien* A, **3**, 70–6. [11.16., 17.2.]

WAIT, G. R. and MCNISH, R. G. (1934). Atmospheric ionization near the ground during thunderstorms, *Month. Weath. Rev.* **62**, 1–4. [4.20.]

WAIT, G. R. and PARKINSON, W. D. (1951). Ions in the atmosphere, *Compendium of Meteorology*, pp. 120–7. [4.34.]

WAIT, G. R. and TORRESON, O. W. (1935). Large-ion content and the small-ion content of air in occupied rooms, *J. Amer. Soc. Heat. Ventil. Engrs.* **7**, 105. [7.24.]

WALL, E. (1948). Das Gewitter, *Wett. u. Klim.* **1**, 8–21, 64–74, 193–204, 321–6. [12.28., 16.12., 16.15., 16.30., 16.37.]

WALL, W. (1710). Experiments on the luminous qualities of amber, diamonds and gumi lac. *Phil. Trans.* **26**, p. 71. [1.2.]

WALLACE, L. (1960). Note on the spectrum of lightning in the region 3670 to 4280 Å, *J. Geophys. Res.* **65**, 1211–14. [14.54.]

WALTER, B. (1903). Über die Entstehungsweise des Blitzes, *Ann. Phys., Lpz.* **10**, 393–407. [14.2.]

WALTER, B. (1936). Über das Abspringen des Blitzes von einer fertigen Bahnlinie, *Ann. Phys., Lpz.* **25**, 124–42. [14.19.]

WANG, C. P. (1963). Lightning discharges in the tropics. 1. Whole discharges, *J. Geophys. Res.* **68**, 1943–49. [12.9., 14.31.]

WATSON, R. E. (1929). Measurements of the effective electrical conductivity of the air and the earth's electric field at and near ground level by means of the Wilson Universal electrometer, *Geophys. Mem., Lond.* **45**, [5.18., 5.21., 8.15.]

WATSON WATT, R. A. and HERD, J. F. (1926). An instantaneous direct-reading radiogoniometer, *J. Inst. Elect. Engrs.* **64**, 611–22. [15.4.]

WEBB, W. L. and GUNN, R. (1955). The net electrification of natural cloud droplets at the earth's surface, *J. Met.* **12**, 211–14. [10.18., 13.7., 13.9.]

WEBER, L. 1886–92. Mitteilungen, betreffend die im Auftrage des Elektrotechnischen Vereins aus geführten Untersuchungen über atmosphärische Elektrizität, *Elektrotech. Z.* **7**, 445; **9**, 189; **10**, 387; 521; 571, **13**, 239. [9.6., 9.11.]

WEICKMANN, H. K. (1933). Observational data on the formation of precipitation in cumulo-nimbus clouds, *Thunderstorm Electricity*, pp. 66–138. [3.3.]

WEISE, G. (1905). Atmosphärisch-elektrische Ströme in vertikalen Leitern unter Berücksichtigung meteorologischer Elemente, *Beibl. Ann. Phys.* **29**, 210–11. [9.6.]

WEISS, E. (1906). Beobachtungen über Niederschlagselektrizität, *S. B. Akad. Wiss., Wien* **115**, 1285–1320. [10.4., 10.23.]

WEISS, R. and STEINMAURER, R. (1937). Messungen der Luftionen in Innsbruck, *Beitr. Geophys.* **50**, 238–51. [4.4.]

WEISSNER, J. (1895). Beiträge zur Kenntnis des tropischen Regens, *S. B. Akad. Wiss. Wien* **104** B, 1397–1434. [10.5.]

WHIPPLE, E. C. (1964). Electricity in the terrestrial atmosphere above the exchange layer, *Problems*, pp. 123–39. [4.43.]

WHIPPLE, F. J. W. (1929a). On the association of the diurnal variation of electric potential gradient in fine weather with the distribution of thunderstorms over the globe, *Quart. J. R. Met. Soc.* **55**, 1–17. [1.17., 5.43., 5.47., 11.6., 11.7.]

WHIPPLE, F. J. W. (1929b). Potential gradient and atmospheric pollution; the influence of summer time, *Quart. J. R. Met. Soc.* **55**, 351–61. [5.44.]

WHIPPLE, F. J. W. (1932). On potential gradient and the air-earth current, *Terr. Magn. Atmos. Elect.* **37**, 355–9. [8.20.]

WHIPPLE, F. J. W. (1933). Relations between the combination coefficients of atmospheric ions, *Proc. Phys. Soc., Lond.* **45**, 367–80. [4.33., 4.34., 4.41.]

WHIPPLE, F. J. W. and CHALMERS, J. A. (1944). On Wilson's theory of the collection of charge by falling drops, *Quart. J. R. Met. Soc.* **70**, 103–20. [3.11., 10.20.]

WHIPPLE, F. J. W. and SCRASE, F. J. (1936). Point discharge in the electric field of the earth, *Geophys. Mem., Lond.* **68**, 1–20. [5.19., 5.34., 5.43., 7.10., 9.9., 9.10., 9.11., 9.15., 9.16., 9.17., 9.18., 9.21., 9.23., 9.24., 11.7., 11.12., 12.7., 12.10., 12.12., 17.2.]

WHITLOCK, W. S., and CHALMERS, J. A. (1956). Short period variations in the atmospheric electric potential gradient, *Quart. J. R. Met. Soc.* **82**, 325–36. [5.61., 5.68., 5.69., 6.13., 6.15., 8.26., 13.7., 14.32., 15.8., 16.4., 16.38., 16.42., 17.8.]

WICHMANN, H. (1951). Die Weltgewittertätigkeit und das luftelektrische Feld der Erde, *Arch. Met., Wien* A, **3**, 290–302. [9.11., 9.25., 11.7., 17.2.]

WICHMANN, H. (1952a). Die Messung des luftelektrischen Feldes mit radioaktiven Potentialsonden, *Arch. Met., Wien* A, **5**, 86–99. [5.15.]

WICHMANN, H. (1952b). Zur Theorie des Gewitters, *Arch. Met., Wien* A, **5**, 187–230. [12.29., 14.47., 16.35., 16.36., 17.17.]

WIGAND, A. (1914). Measurements of the electrical conductivity in the free atmosphere up to 9,000 m in height, *Terr. Magn. Atmos. Elect.* **19**, 93–101. [7.13.]

WIGAND, A. (1926). Ladungsmessungen an natürlichem Nebel, *Phys. Z.* **27**, 803–8. [13.9.]

WIGAND, A. and FRANKENBERGER, E. (1930). Über Beständigkeit und Koagulation von Nebel und Wolken, *Phys. Z.* **31**, 204–15. [13.9.]

WILDMAN, P. J. L. (1965). A device for measuring an electric field in the presence of ionisation, *J. Atmosph. Terr. Phys.* **27**, 417–23. [5.72.]

WILKINS, E. M. (1964). The role of electrical phenomena associated with tornadoes, *J. Geophys. Res.* **69**, 2435–47. [12.33.]

WILLIAMS, J. C. (1958). Some properties of the lower positive charge in thunderclouds, *Rec. Adv.*, pp. 425–9. [12.27.]

WILLIAMS, D. P. and BROOK, M. (1963). Magnetic measurements of thunderstorm currents. 1. Continuing currents in lightning, *J. Geophys. Res.* **69**, 3243–7. [14.8., 14.23.]

WILSON, C. T. R. (1899). On the comparative efficiency as condensation nuclei of positively and negatively charged ions, *Phil. Trans.* A, **193**, 289–308. [16.13.]

WILSON, C. T. R. (1900). On the leakage of electricity through dust-free air, *Proc. Camb. Phil. Soc.* **11**, 32. [1.14., 4.1.]

WILSON, C. T. R. (1901). On the ionisation of atmospheric air, *Proc. Roy. Soc.* A, **68**, 151–61. [4.19.]

WILSON, C. T. R. (1906). On the measurement of the air–earth current and on the origin of atmospheric electricity, *Proc. Camb. Phil. Soc.* **13**, 363–82. [5.21., 8.5.]

WILSON, C. T. R. (1916). On some determinations of the sign and magnitude of electric discharges in lightning flashes, *Proc. Roy. Soc.* A, **92**, 555–74. [1.18., 5.22., 5.34., 5.35., 8.5., 10.4., 12.7.]

WILSON, C. T. R. (1920). Investigations on lightning discharges and on the electric field of thunderstorms, *Phil. Trans.* A, **221**, 73–115. [1.17., 9.6., 11.1., 12.7., 12.11.]

WILSON, C. T. R. (1923). The acceleration of beta particles in strong electric fields, *Proc. Cambridge Phil. Soc.* **22**, 534–8. [14.33.]

WILSON, C. T. R. (1925). The electric field of a thundercloud and some of its effects, *Proc. Phys. Soc., Lond.* **37**, 32D–37D. [9.18., 12.1., 12.31.]

WILSON, C. T. R. (1929). Some thundercloud problems, *J. Franklin Inst.* **208**, 1–12. [3.11., 10.14., 10.22., 16.12., 16.15., 17.12.]

WILSON, C. T. R. (1956). A theory of thunder cloud electricity, *Proc. Roy. Soc.* A, **236**, 297–317. [16.12., 16.20., 16.32.]

WILSON, R. R. (1957). Atmospheric signals caused by cosmic-ray showers. *Phys. Rev.* **108**, 155–6. [5.70.]

WINKLER, J. H. (1746). Die Stärke der elektrischen Kraft des Wassers, Breitkopf, Leipzig [1.2.]

WITKOWSKI, A. (1902). Spostrzenia nad elektrycznosia atmosferyczna w Zakopanem, *Bull. Acad. Sci., Cracovai* **7**, 13. [5.10.]

WOESSNER, H. H., COBB, W. E. and GUNN, H. (1958). Simultaneous measurements of the positive and negative light-ion conductivities to 26 kilometers, *J. Geophys. Res.* **63**, 171–80. [7.13.]

WOOD, C. A. (1951). Unusual lightning, *Weather* **6**, 63. [14.38.]

WORKMAN, E. J. (1965). Thunderstorm electricity, *Problems*, pp. 296–303. [16.25.]

WORKMAN, E. J., BROOK, M. and KITAGAWA, N. (1960). Lightning and charge storage, *J. Geophys, Res.* **65**, 1513–18. [14.20., 14.21.]

WORKMAN, E. J. and HOLZER, R. E. (1939). Recording generating voltmeter for the study of atmospheric electricity, *Rev. Sci. Instrum.* **10**, 160–3. [5.29.]

WORKMAN, E. J., HOLZER, R. E. and PELSOR, G. T. (1942). *The Electrical Structure of Thunderstorms*, N.A.C.A. Tech. Notes, No. 864. [12.8., 12.17.]

WORKMAN, E. J. and REYNOLDS, S. E. (1949a). Electrical activity as related to thunderstorm cell growth, *Bull. Amer. Met. Soc.* **30**, 142–4. [12.23.]

WORKMAN, E. J. and REYNOLDS, S. E. (1949b). Time of rise and fall of cumulus cloud tops, *Bull. Amer. Met. Soc.* **30**, 359–61. [3.8.]

WORKMAN, E. J. and REYNOLDS, S. E. (1950). Electrical phenomena occurring during the freezing of dilute aqueous solutions and their possible relationship to thunderstorm electricity, *Phys. Rev.* **78**, 254–9. [3.15., 16.25., 16.27., 16.31.]

WORKMAN, E. J. and REYNOLDS, S. E. (1953). Structure and electrification, *Thunderstorm Electricity*, pp. 139–49. [3.18., 16.25., 16.27., 16.31., 17.4.]

WORMELL, T. W. (1927). Currents carried by point discharge beneath thunderclouds and showers, *Proc. Roy. Soc.* A, **115**, 443–55. [9.6., 9.7., 9.9., 11.19.]

WORMELL, T. W. (1930). Vertical electric currents below thunderstorms and showers, *Proc. Roy. Soc.* A, **127**, 567–90. [5.34., 9.6., 9.9., 11.9., 11.10., 11.12., 11.13., 11.14., 11.16., 11.19.]

WORMELL, T. W. (1939). The effects of thunderstorms and lightning discharges on the earth's electric field, *Phil. Trans.* A, **238**, 249–303. [5.34., 9.24., 11.14., 12.4., 12.7., 12.9., 12.12.]

WORMELL, T. W. (1953a). Atmospheric electricity; some recent trends and problems, *Quart. J. R. Met. Soc.* **79**, 3–50. [Pref. 3.11., 9.8., 11.12., 11.14., 11.16., 12.10., 12.11., 14.9., 16.10., 16.30., 16.32.]

WORMELL, T. W. (1953b). Lightning, *Quart. J. R. Met. Soc.* **79**, 474–89. [12.11., 14.1., 14.9., 14.27., 15.5.]

WORMELL, T. W. (1955). Theories of thunderstorm electrification—some general conclusions, *Wentworth Conf.*, pp. 157–61. [16.25.]

WORMELL, T. W. and PIERCE, E. T. (1948). Atmospherics, *J. Instn. Elect. Engrs.* **95**, (pt. III) 331. [14.9.]

WRIGHT, H. L. (1933). The influence of atmospheric suspensoids upon the earth's electric field as indicated by observations at Kew Observatory, *Proc. Phys. Soc., Lond.* **45**, 152–71. [7.10.]

WRIGHT, H. L. (1936). The size of atmospheric nuclei; some deductions from measurements of the number of charged and uncharged nuclei at Kew Observatory, *Proc. Phys. Soc., Lond.* **48**, 675–689. [4.2., 4.31.]

YAGLOU, C. P. and BENJAMIN, L. C. (1934). *J. Amer. Soc. Heat. Vent. Engrs.* **6**, 25. [7.24.]

YAGLOU, C. P., BENJAMIN, L. C. and CHOATE, S. P. (1931). Changes in ionic content of air in occupied rooms ventilated by natural and mechanical methods, *Trans. Amer. Soc. Heat. Vent. Engrs.* **38**, 191. [7.24.]

YOKOUTI, K. (1939). Point discharge in the atmospheric electricity, *J. Met. Soc., Japan* **17**, 73–5. [9.9.]

YOUNG, G. A. (1961). *A Lightning Stroke of an Underwater Explosion Plume*, U.S. Nav. Ord. Lab. Rept. 61–43. [14.48.]

YRIBERRY, A. J. (1954). Registro del campo electrostatico tormentoso en San Miguel (Argentina) y en Kew (Inglaterra), *Acta Scient. San Miguel Obs. de Fis. Cos.*, No. 2. [9.15.]

YUNKER, E. A. (1940). The mobility spectrum of atmospheric ions, *Terr. Magn. Atmos. Elect.* **45**, 127–32. [4.4.]

ZEILINGER, P. R. (1935). Über die Nachlieferung von Radium-Emanation aus dem Erdboden, *Terr. Magn. Atmos. Elect.* **40**, 281–94. [4.17.]

ZELENY, J. (1900). The velocity of ions produced in gases by Röntgen rays, *Phil. Trans.* A, **195**, 193–234. [2.54.]

Index

Abbreviations 20
Absolute values of potential gradient 9
Activity, thunderstorm 297
Aeroplanes, measurement of potential gradient in 159
Agitation 174
Agrimeter 148
  theory of 149
Air
  conductivity of 11, 439
  mass, correlation with potential gradient 171
  total ionization of 101
Aircraft
  ion counting in 95
  and lightning 388
  measurement of conduction current in 228
  measurement of conductivity in 207
  point discharge from 252
  thunderstorm measurements 331
Air–earth conduction current 216, 440
Air–earth current
  annual variation 234
  comparison of methods 227
  correlation with storms 299
  direct method 218
  diurnal variation 232
  electronic methods 222
  in fine weather 216
  GOTO's method 224
  indirect method 225
  matching circuit 222
  measurements
    in aircraft 228
    near ground 227
    on mountains 237
    over oceans 237
    with radio-sonde 229
  methods of measurement 217
  relation with conductivity 235
  SCRASE's method 220
  SIMPSON's measurements 219
  synoptic use of measurements of 236
  and terrestrial magnetism 237
  two components of 228
  VON KILINSKI's method 223
  WILSON's measurements 218
  to a wire 226
Aitken nuclei 88
Altielectrograph 256
  results 324
  width of traces 258
Amplification, direct current 55
Analysis, choice of periods for 161
Annual variation
  of air–earth current 234
  of conductivity 203
  of potential gradient 168
Antenna, passive 134
Application of BOLTZMANN's law 106
Approach to equilibrium, rate of 114
Artificial initiation of lightning 384
Artificial space charges 194
Atmosphere
  as condenser 13
  ions in 11
  outer 445
  radioactivity of 121
  upper, potential gradients in 183
Atmospheric electricity
  measurements of 47
  relation with meteorology 45

Atmospheric electricity (*cont.*)
relation with upper atmosphere 46
Atmospheric ionization 12
Atmospherics 393
classification of 394
direction of 394
high-frequency 397
location by single-station observations 395
low-frequency 396
nature of 393
sudden enhancement in 395
Aurora, effects of, on potential gradient 46, 173
Austausch 44
Auto-correlation 49
Auto-generator 233
Average values of results 450
Ayrton shunt 54

Balance sheet, electrical 300
Ball lightning 390
theories of 391
Balloons
discharges to 379
measurements
of conductivity with 205
of ions with 96
of potential gradients with 157
point discharge from 247
Bead lightning 390
β-ray ionization 101
Blizzards, charges from 179, 290
BOLTZMANN's law, application of 106
Bomb, instrumented 437
Bound charge
electronic method for 140
measurement of potential gradient by 137
BOYS' camera 350
Branching of lightning flashes 380
Breakdown potential gradients 331
Breaking-drop theory 414
Bright-band phenomenon 71
Buildings, conductivity inside 214

Cables, high-tension, potential-gradient effects from 175
Cage method for space charge 185
Calibration of field machine 152
Camera, moving 350
Capacitative effects 40
Capillary electrometer 140
Capricorn expedition 237
Capture of ions by water drops 71
Carnegie research ship 162, 237
Cathode follower 55
Cathode-ray oscillograph for lightning investigations 382
Cell, thunder-cloud 70
Centres of charge, heights of 367
Change of conductivity, relation with space charge 38
Change of electric moment 372
Change of potential gradient
measurement of 155
slow final 366
small rapid 370
Channel, lightning, diameter of 363
Charge
on earth 7
correlation with thunderstorms 297
maintenance of 292, 434
on electrosphere 296
lower positive, in clouds 337, 427
separation of 399
transfer of 292
on wires in snowfall 290
Charge centres, heights of 367
Charge separation
classification of theories of 407
in continuous rain and snow clouds 431
discussion of theories of 422
in non-raining cumulus clouds 432
possible processes of 437
requirements of theory of 403
in stratus clouds 432
Charge transfer in non-stormy rain and snow 307
Charged and uncharged nuclei 92
ratio of 115
Charged droplets, evaporation of 82
Charges
from blizzards 179, 290

Charges (*cont.*)
cloud 399
connection with precipitation 401
on droplets in non-raining clouds 346
location of 26
main thunder-cloud 399
masking of 404
in mist and fog 349
in non-stormy clouds 400
quantities in clouds 332, 405
on single drops 268
on snow, sleet and hail 284
subsidiary 400
in violent storms 406
Charging of water drops 71
Classification of theories of charge separation involving gravitation 401
Climatology, electric 45
Close discharges, potential-gradient effects of 370
Close flashes, recovery after 322
Cloud
charges 399
discharge to electrosphere from 377
droplet electrification, theory of 349
and ground flashes 374
levels, space charges at 191
physics 65
types 64
Clouds
conductivity of 204, 212
contamination in 69
currents above 305
electrification of 66
non-raining, charges on droplets in 346
non-stormy 343
charge in 400
electrification of 343
polarity of 309
potential-gradient variations under 176
potential gradients below 176, 259
potential gradients inside 330
radar study of 71
radio emission from 397
raindrop charges in 279
total current below 261
Coagulation
effects on charge separation 413
of nuclei 117
Coalescence, initiation of precipitation by 68
Coast, potential-gradient effects at 176
Collector of potential 124
Columnar resistance 210
for potential of electrosphere 294
Combination coefficients 104, 107, 111
Condensation 65
electrical effects of 82
nuclei 88
Condenser, atmosphere as 13
cylindrical 61, 196
Conduction current 216, 300
two components of 228
Conduction currents, distribution of, near storms 333
Conductivities, polar, ratio of 211
Conductivity
of the air 11, 195, 439
inside buildings 214
change of, related to space charge 38
close to ground 200
in clouds 212
effect of nuclear explosions on 214
of fogs and clouds 204
of ice 74
indirect determination of 201
ionic 27, 195
measurement
in aircraft 207
in balloons 205
by GERDIEN's method 196
by HOGG's method 200
in large enclosed sphere 211
on mountains 213
by the NOLANS' method 199
by rockets 209
over sea 290
under trees 201
by wire-cage method 199

Conductivity (*cont.*)
and physiological phenomena 214
and potential gradient
calculations 212
variation with height 37
relation of, with air–earth current 235
in thunderclouds 324
variations of
annual 203
cause of, close to earth 201
diurnal 202
Conductor
leakage from 195
lightning 3, 386
movable 4
in potential gradient 32
Contact of ice 76
Contact potentials 52
Contamination in clouds and precipitation 69
Continuing currents 366
Continuous nucleus counter 91
Continuous-rain and snow clouds, charge separation in 431
Continuous recording 59
Control of thunderstorm electrification 433
Convection currents 231, 301
Convection theories of separation of charge 414, 424
Cooperation, international 15
Correlation of meteorological and physiological phenomena 46
Cosmic noise and thunderstorms 340
Cosmic rays
currents from 303
ionization by 100
Cosmic-ray showers, effects of on potential gradient 182
Cosmic thunderstorms 341
Counters
ion 93
nucleus 90
Counting of lightning flashes 376
Critical damping 54
Cumulonimbus 64, 70
Cumulus
clouds, trade-wind 346
stage of thunder cloud 70
Curie 121
Current
air–earth 216, 440
conduction 216, 300
convection 231, 301
correlation with storms 299
displacement 218
in ground from lightning 383
ionic 29
lightning 383
from radioactive equalizer 132
in return stroke 362
saturation 30
total, below cloud 261
Currents
above clouds 305
continuing 366
convection 231
horizontal 306
precipitation 263, 302, 441
rain 263
sign convention for 25
Cylindrical condenser 61, 196

Dart leader 360
Deposition 65
electrical effect of 81
Destruction of ions 103
Development
of nimbo-stratus cloud 71
of thunder cloud 69
Diameter of lightning channel 363
Differentiating circuit 58
Diffusion coefficient 28
Diffusion current 231
Diffusion, eddy 29
Diffusion theory of charge separation 413
Digitization 60
Direct-current amplification 55
Direction of atmospherics 394
Discharges
to balloons 379
bringing down positive charge 371
from cloud to electrosphere 377
duration and frequency of 366

Discharges (*cont.*)
initiation of 383
lightning 302, 443
minor 374
to power lines 389
processes, related to radiation frequency 396
to skyscraper 378
Discharging points, effective separation of 260
Discussion of theories of charge separation 422
Displacement current 218
Dissipating stage of thunder cloud 70
Dissipation coefficient 118
Dissipation of fog, related to conductivity 204
Disturbed weather 44
potential gradients in 177
Diurnal variation
of air–earth current 232
of conductivity 202
of potential gradient 161
origin of 166
in undisturbed conditions 162
Double dipole theory 329
Double field machine 153
Droplets
in clouds, theory of electrification of 349
in non-raining clouds, charges on 346
Drops
charges on 268
in clouds, charges on 274
Duration of discharges 366
Dust storms, potential-gradient effects of 176
Dynamic equilibrium 35

E layer, spasmodic ionization in 378
Early theories 6
Earth
charge on 7
correlation with thunderstorms 298
connections 52
maintenance of charge on 292, 434
vertical potential gradient of 122
Eclipses, effect on potential gradient of 172
Eddy diffusion 29
Effective resistance of equalizer 124
Effective separation of points 260
Electric climatology 45
field effect on freezing 78
field strength 22
moment
changes of 372
destroyed by lightning 321
space constant 20
synopsis 45
Electrical balance sheet of earth **300**
Electrical cell theories of charge separation 411
Electrical effects
of condensation 82
of deposition 81
of evaporation 82
of freezing 77
of glazing 79
of ice impact 75
of melting 80
of riming 78
of splashing 83
of sublimation 81
with water drops 83
Electrical images 31
Electrical measurements of lightning 356
Electrical properties of ice 74
Electricity
of fine weather 3
identification of lightning with 1
space 16
from thunder clouds 2
Electrification
of clouds 66, 343
relation with ice 402
relation with precipitation 334, 401
thunderstorm, control of 433
Electrode effect 42, 230
Electro-geodesy 9
Electrolytic tank 60

Electrometer
capillary 140
measurement of rain charge 264
quadrant 54
universal portable 138
valve 56
Electronic method
for air–earth current 222
for bound charge 140
with radioactive equalizer 131
with water-dropper 128
Electronics, use of 15
Electrons, runaway 375
Electrosphere 13, 33
charge on 296
discharges from cloud to 377
potential of 35, 293
methods of determining 293
Electrostatic field of lightning 354
Electrostatic fluxmeter 145
formulae 20
precipitation method for space charge 188
Empire State Building, discharges to 378
Energy of lightning 375
Enhancement of atmospherics 395
Equalizer
effect of leaks on 125
effective resistance of 124
radioactive 129
effective resistance of 128
material of 129
Equalizers, potential 123
Equalizing layer 34
Equilibrium
of ionization 103, 104, 438
rate of approach to 114
Equipotentials near balloon 157
Evaporation 65
of charged droplets 82
electrical effects of 82
Explorer II balloon 205
Exposure factor 135

Fair-weather conduction current 216
Feldvariograph 156
Field 22
Field machine 142
calibration and reduction factor for 152
double 153
sign determination in 146
Field mill 143
Filtration method for space charge 186
Fine weather
current 216
electricity of 3
fundamental formulae for 36
potential gradient 10
variations in 155
values of space charge 189
Flame equalizer 7, 134
Flashes
branching of 380
cloud and ground 374
intracloud 372
lightning, counting of 376
Fluctuations of potential gradient
due to space charges 174
short-period 174
Fluxmeter, electrostatic 145
Fog
charges in 349
conductivity changes related to 204
conductivity of 204
potential gradients in 179
Föhn winds, effects on potential gradient of 175
Formulae, electrostatic 20
Free charge in thunder clouds 332
Free space, permittivity of 21
Free space charge 184
Freezing 65
effect of electric field on 78
electrical effects of 77
potential 77
theory involving 420
Frequency
of discharges 366
of positive potential gradients in storms 314, 326, 327
Fulchronograph 382
Fulgamites 376
Fundamental formulae for fine weather 36

Fuse as equalizer 134

Galvanometer, critically damped 54
Glazing 66
  electrical effects of 79
  process, calculations on 419
  theories of charge separation 418
Gliders, measurement of potential gradient in 160
Ground
  currents from lightning in 383
  ionization close to 202
  ions from 103
Gush of rain after lightning 335

Hail, charges on 284
Heaviside layer 13
Height, variation of potential gradient and conductivity with 37
Heights of charge centres 332, 367
High-frequency atmospherics 397
High-tension cables, effect of potential gradient of 175
Historical survey 1
Horizontal currents 306
Horizontal potential gradients 174
Humidity, relation of electrical effects with 10

Ice
  conductivity of 74
  connection with electrification 402
  contact 76
  early theories involving 415
  electrical properties of 74
  impact
    electrical effects of 75
    theory of charge separation 415
  initiation of precipitation from 68
  splinter theory of charge separation 417
  thermoelectric properties of 74
Identification of lightning with electricity 1
Images, electrical 31
Impact, ice 75
Induction voltmeter 147
Influence theory of thunderstorm electricity 408
Initiation
  of lightning discharge 383
  of precipitation
    by coalescence 68
    from ice 68
Instrumentation 444
Instrumented bomb 437
Instruments, measuring 53
Insulation 50
Integrating circuit 58
Integration method for potential of electrosphere 295
Intermediate ions 87, 438
International co-operation 15
Intervals between strokes 365
Intracloud flashes 372
Inverse-square term in lightning field 354
Inverted test plate 315
Ion-capture theory of charge separation 409
Ion counter 93
Ion counting
  in aircraft 95
  in large sphere 97
  results of 94
Ion densities and space charge 190
Ionic conductivity 27, 195
Ionic current 29
Ionization
  β-ray 101
  close to the ground 102
  by collision 239
  by cosmic rays 100
  equilibrium of 103, 438
  origin of atmospheric 12
  production of 98
  by radioactivity 99
  rate of 100, 113
  total, of air 101
  up to 100 km 120
  variation of, with level 102
Ionosphere 13
Ions
  in the atmosphere 11
  capture of, by water drops 71
  counting of, with aircraft 95

Ions (*cont.*)
destruction of 103
distinction between large and small 88
from ground 103
intermediate 87, 438
large 87
life of 119
and nuclei 85
point-discharge, effect on potential gradient 179
from sea 103
small 85
nature of 86
velocity of 30
Isoceraunic level 307
Isolated point, discharge from 248

Jet stream and potential gradient 182
Junction leaders 369

KELVIN's contributions 8
Kites, point discharge from 247
Klydonograph 381

Large ions 87
distinction from small 88
Leader
α- and β-types 358
dart 360
junction 369
stepped 358
theories of 359
Leakage from conductor 195
Leak-free method with radioactive equalizers 129
Leaks of equalizer, effects of 125
Life of ion 119
Lightning
and aircraft 388
artificial initiation 384
ball 390
theories of 391
bead 390
changes of electric moment due to 321
close discharges of 322
conductor 3, 386
currents 383
in ground 383
discharge, initiation of 383
discharges 302, 350, 443
to balloons 379
from cloud to electrosphere 377
energy in 375
to power lines 389
to skyscraper 378
electric moment destroyed by 321
electrical measurements 353, 356
electrostatic field of 354
flashes
branching of 380
counting of 376
general results 383
gush of rain after 335
identification with electricity 1
inverse-square term in field of 355
investigation of 350
investigation with cathode-ray oscillograph 382
magnetic measurements of 355
photographic measurements on 352
point of fall 385
potential-gradient changes due to 316
preliminary processes 357
protection from 386
radar echoes from 384
radiation field of 355
recovery of potential gradient after 321
spectrum of 389
warning of 376
Lines of force 24, 25
with non-raining cloud 67
Local effects on mean value of potential gradient 170
Local time effects on potential gradient 164
Location
of atmospherics by single-station observations 395
of charges 26
Low-frequency atmospherics 396

Lower positive charge 337, 427
  secondary theories of 429
  water theories of 428

M.K.S. units 19
Magnetic link 381
Magnetic measurements of lightning 355
Magnetic surge integrator 382
Magnetic tape recorder 353
Main stroke 361
Main thunder cloud charges 399
Maintenance of earth's charge 292, 434
Masking of charges 404
Matching circuit 222
Mature stage of thunder cloud 70
Mean value of potential gradient, local effects on 170
Measurements of atmospheric electricity 47
Measuring instruments 53
Mechanism of charge separation 401
Melting 65
  electrical effects of 80
Meteorological phenomena, correlation with physiological 46
Meteorology, relation with atmospheric electricity 45
Microvoltameter 242
Military applications 14
Minor discharges 374
Minor shower clouds 346
Mirror-image effect 288
Mist, charges on 349
Mobility 27
  of small ions 86
Modification of cloud charges 426
Moisture and space charge 190
Moment, electric, of lightning discharge 372
Mountain thunderstorms 339
Mountains, measurement on
  of air–earth current 237
  of conductivity 213
  of potential gradient 170
Movable conductor 4
Moving camera 350
Multiple points, discharge from 248
  station recording of potential-gradient changes 320
  strokes 363
Multiply charged nuclei 97, 106

Nature of small ions 86
Negative feedback 56
Negative potential gradient in mist and fog 175
Nimbostratus
  clouds 71, 343, 442
    development of 71
  rain cloud 345
  snow cloud 344
Non-raining clouds, charges on droplets in 346
Non-raining cumulus clouds, charge separation in 432
Non-stormy clouds 343
  charges in 400
  electrification 343
Non-stormy rain, transfer of charge in 307
Nuclear explosions, effect on conductivity of 214
Nuclei
  charged and uncharged 92
    ratio of 115
  coagulation of 117
  condensation 88
  decay of 117
  multiply charged 97, 106
  original method of counting 90
  removal of 116
  sizes of 118
Nucleus counter
  AITKEN's 90
  continuous 91
  photoelectric 91
Number of strokes in multiple flash 364
Numerical values 26

Oceans, measurement of air–earth current over 237
OHM's law 25
  validity of 42

Onset of fog, related to conductivity 204
Origin
of atmospheric ionization 12
of charges on rain and snow 286
of local diurnal variation of potential gradient 166
Outer atmosphere 445
Outstanding problems 434

Packets, space charge 191
Passive antenna 134
Patterns, potential-gradient 180
Penetrating radiation and thunderstorms 339
Permittivity of free space 20
Photoelectric effect 415
Photoelectric nucleus counter 91
Photographic recording 59
Physical storm 337
Physiological phenomena
and conductivity 214
correlation with meteorological 46
Piezoelectric measurement of potential gradient 155
Pilot streamer 357
Point discharge 239, 301, 446
from aircraft 252
CHAPMAN's formula 251
current 239
early measurement of 242
effects
of space charge 242
of wind speed 251
general space-charge theory of 253
ions, effect on potential gradient of 179
from isolated points 248
from kites and balloons 247
laboratory and atmospheric conditions 241
measurement of potential gradient 137
measurements concerning thunder clouds by 316
from multiple points 248
pulsed nature of 241
ratio of signs 245
relation
with potential gradient 250
with rain current 277, 281
SCHONLAND's measurements 244
simplified theory 241
theory
of isolated point 254
of widespread discharge 253
time lags 246
from trees 249
WORMELL's measurements 242
Point of fall 385
Polar conductivities, ratio of 211
Polarity of thunder cloud 309
Pole-finding paper 257
Positive charge, lower 337, 427
discharge to earth 371
potential gradients, frequency of 314, 324, 325
Potential 8, 20
of electrosphere 35
methods of determination 293
of thunder cloud 375
Potential equalizers 123
effective resistance of 124
Potential gradient 22, 122
absolute values of 9
for breakdown 331
change
measurements of 155, 318
related to space charge 188
changes
due to lightning 316
multiple-station recording of 320
slow final 366
choice of periods for analysis of 161
below clouds 176
and conductivity
calculations 212
variation of height of 37
conductor in 32
diurnal variation of 161
earth's vertical 122
effect of relaxation time 171
effects
of aurora 173
of close discharges 370

Potential gradient (*cont.*)
effects (*cont.*)
at coast 176
of cosmic-ray showers 182
during eclipses 172
of dust storms 176
of Föhn winds 175
from high-tension cables 175
of jet stream 182
of point-discharge ions 179
of volcanic eruptions 173
local-time effects 164
origin of 166
low
rain current with 283
total current with 283
measurement
in aeroplanes 159
in balloons 157
by bound charge 137
in gliders 160
on mountains 170
by point-discharge 137
at sea 160
measurements 122, 439
patterns 180
piezoelectric measurement of 155
point discharge, relation with 250
pre-discharge 321
recovery of
after close flashes 322
after lightning 321
related to space charge 24
relation to precipitation charges 274
short-period fluctuations of 172
sign determination of, in field machines 146
storms, correlation with 298
due to thunder cloud 311
types of measurement of 122
variations
annual 168
diurnal 161, 162, 166
in fine weather 193
secular 169
under cloud 176
Potential gradients
below clouds 176, 259
inside clouds 330
in disturbed weather 177
in fog 179
horizontal 172
multiple-station recording of 320
positive, frequency of 314, 324, 325
in quiet rain 177
in snow 178
in upper atmosphere 183
Potentials in thunder clouds 375
Power lines, discharges to 389
Precipitation
charge, relation to potential gradient 274
charges
in clouds 279
at Zugspitze 280
connection of charges with 401
contamination in 68
currents 263, 302, 441
initiation
by coalescence 68
of, from ice 68
relation with electrification 335, 401
static 290
Pre-discharge potential gradient 321
Preliminary processes in lightning 357
Primary space charge 184
Primitive notions 1
Principle of quasi-static state 35
Production of ionization 99
Protection from lightning 386
Pulses due to space charges 191
Pyroelectric effect 421

Quadrant electrometer 54
Quantities
of charge in clouds 405
of free charge in thunder clouds 332
Quasi-static state 35
Quiet rain, potential gradients in 177
Quiet snow, potential gradients in 178

Radar
echoes from lightning 384
observations of thunder clouds 333
study of clouds 71
Radiation
field of lightning 355
frequency, related to discharge processes 396
Radio emission from clouds 397
Radioactive equalizer 128
current from 132
electronic method with 131
leak-free method with 129
limitations of 132
material for 129
Radioactivity
of atmosphere 121
ionization by 99
Radio-sonde method
for conduction current 229
for potential gradient 258
Radiosonding 58
Rain
continuous observations on 270
gush of, after lightning 335
non-stormy, charge transfer in 307
origin of charge on 286
quiet, potential gradients in 177
space charge on 276
space charges during 190
Rain charges
in clouds 279
measurement of, by electrometer 264
numbers of single drops 273
quantities of water with 272
ratio of positive to negative 271
on single drops 268
times of rainfall with 272
per unit volume 272
Rain current
to completely exposed receiver 266
with low potential gradient 283
measurement of
direct 265
by electrometer 264
need for continuous measurement of 270
relation with point discharge 277, 281
Rain currents 263, 273
Rain drop charges in clouds 279
Rain electricity, importance of 263
Rain results, classification of 270
Raised sphere 156
Rate
of approach to equilibrium 114
of ionization 100, 113
Ratio
of charged and uncharged nuclei 114
of polar conductivities 211
of signs
in point discharge 245
in rain current 271
of small to large ions 116
Rationalized M.K.S. system of units 19
Receiver for rain current, completely exposed 266
Recombination coefficient 104
Recovery of potential gradient
after close flashes 322
after lightning 321
Reduction factor 9, 10, 135
for field machines 152
Regeneration time 323
Relaxation time 39, 323
effect on potential gradient of 171
Removal of nuclei 116
Requirements of theory of charge separation involving precipitation 403
Resistance
columnar 210
effective of equalizer 124
Return stroke 361
current in 362
Reversal distance 312
Riming 65
electrical effects of 79
theories involving 420
Rocket measurements of conductivity 209
Runaway electrons 375

St. Elmo's fire 239
Saturation current 30
Sea
conductivity over 209
ions from 103
measurement of potential gradient at 160
Secondary space charge 184
Secondary theories of lower space charge 429
Secular variation of potential gradient 169
Separation of charge 399
mechanism of 401
Sferics 393
Shower clouds 290
minor 346
Showers, distinction from storms 338
Sign convention
for currents 25
for field 23
Sign determination with field machines 146
Single drops
charges on 268, 277
numbers of 273
Single-station observations for location of atmospherics 395
Sizes of nuclei 118
Skyscraper, discharges to 378
Sleet, charges on 284
Slow final potential-gradient change 366
Small ions 85
distinction from large ions 88
nature of 86
ratio of, to large 116
Small rapid potential-gradient changes 370
Snow
charges on 284
cloud, nimbo stratus 344
non-stormy, charge transfer in 307
origin of charges in 286
potential gradients in 178
total current in 285
Snowfall, charge on wires during 290
Space charge 24, 171, 184
filtration method 186
free 184
and ion densities 190
measurement
by cage 185
by electrostatic precipitation 188
by filtration 186
methods of measurement 185
and moisture 190
packets 191
primary and secondary 184
of rain 276
related to change of conductivity 38
related to change of potential gradient 24, 188
theory of point discharge 252
values in fine weather 189
Space charges 24
artificial 194
at cloud levels 191
potential-gradient fluctuations due to 193
produced by natural phenomena 192
during rain 190
Space electricity 16, 17
Specific conductivity 29
Specific resistance 29
Spectrum of lightning 389
Speed of response of water-dropper 127
Sphere, large
conductivity measurements in 211
ion counting in 97
Splashing, electrical effects of 83, 287
Splinter theory 417
Static, precipitation 290
Statistical methods 49
Stepped leader 358
Storms
correlation
with current 299
with potential gradient 298
number of 307
and showers, distinction between 338
violent, charges in 406

Stormy-weather phenomena 13
Stratus clouds, charge separation in 432
Streamer
  pilot 357
  upward 360
Stretched wire 135
Stroke, main or return 361
Strokes
  multiple 363
  number of 364
Sublimation 65
  electrical effects of 81
Subsidiary thunder-cloud charges 400, 430
Sudden changes 41
Sudden enhancement of atmospherics 395
Summer time, effect on potential gradient of 165
Sunrise effect 168, 233
Sunspots 46
Supply current 11
Surface charges 24
Surge front recorder 382
Synopsis, electric 45
Synoptic use of measurements 236

Temperature effects on ice impact 76
Temperatures at charge centres 325
Terrestrial magnetism and air–earth current 237
Theoretical considerations 445
Theories, early 6
Thermoelectric properties of ice 74
Thunder 341
Thunderbolts 1
Thunder cloud 14, 309
  conductivity in 324
  development of 69
  electricity from 2
  lower positive charge in 337, 427
  main charges in 399
  measurement
    by point discharge 316
    of potential gradient due to 313
  methods of investigation of 310
  polarity of 309
  potential gradient due to 311
  quantities of free charge in 332
  subsidiary charges in 400
Thunderstorm
  activity
    correlation with earth's charge 297
    index 366
  distribution of conduction currents over 333
  electrification, control of 433
  main charges in 399
  problems 435
  radar observations of 333
  subsidiary charges in 400, 436
Thunderstorms
  cosmic 341
  and cosmic noise 340
  mountain 339
  number of 307
  and penetrating radiation 339
  warm cloud 336
Time constant of atmosphere 39
Time intervals 156
Time lags in point discharge 246
Times of rainfall and rain charges 272
Tornadoes 340
Total current
  below cloud 261
  with low potential gradient 283
  in snow 285
Total ionization of air 101
Trade-wind cumulus clouds 346
Traffic-jam effect 67
Transfer of charge 292
  in non-stormy rain and snow 307
  total 303
Tree, point discharge from 249
Trees, measurement of conductivity under 201
Turrets 70
Types of cloud 64

Undisturbed conditions, diurnal riation in 162
Undisturbed days 161
Unitary variation 167

Units
M.K.S. 19
systems of 18
Universal portable electrometer 138
Upper atmosphere
potential gradients in 183
relation with atmospheric electricity 46
Upward streamer 360

Validity of OHM's law 42
Variability of results 48
Variation of potential gradient and conductivity with height 37
Velocities of ions 30
Vertical potential gradient, earth's 122
Vertical rain current 273
Vibrating-reed electrometer 55
Violent storms, charges in 406
Volcanic eruptions, effects on potential gradient of 173
Voltmeter, induction 147

Warm cloud thunderstorms 336
Warning of lightning 376
Water droplets, electrical effects with 83
Water-dropper 125
effect of leaks with 127
effective resistance of 127
electronic method with 128
speed of response of 127
use of 127
Water drops, capture of ions by 71
Water theories of lower positive charge 428
Waterfall electricity 98
Whistlers 397
Widespread point discharge 253
Wind-speed effect on point discharge 251
Wire-cage method for conductivity 199
Wire, current to 226
Wires, charge on, in snowfall 290

Zugspitze, thunderstorm measurements at 326

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