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1. INTRODUCTION

by V. V. Beloussov

Seismology has been included in the list of main IGY disciplines according to the Resolution of the III. CSAGI Assembly in Brussels (September 1955). At that Assembly the main types of observations were also determined: SS — standard seismic observations (chiefly in the Arctic, Antarctic and other inadequately studied regions); mS — observations of microseisms on ordinary stations; ZmS — observations on the tripartite microseismic stations; detailed studies of the crustal structure with seismic methods. Besides these researches it was recommended to conduct special seismic studies: LP — observations of the long-period waves; Lg — measurements of phases of the Lg type; SA — measurements of strains in the Earth's crust. These observations were made as part of the IGY program.

The resolution to include seismology in the IGY program has considerably stimulated seismic researches and accelerated the development of nets of seismic stations in many countries. On the Antarctic continent, for instance, where there was not a single station in operation before the IGY, eleven stations carried out observations with modern equipment during the IGY.

The V. CSAGI Assembly, Moscow, August 1958, resolved to publish a special volume of the IGY Annals which would deal with seismic research in the IGY Program. Resolution VI of the Working Group on Seismology determined the list of chapters of the Volume and the persons responsible for this preparation. The general editorship of the volume has been entrusted to the CSAGI Reporter on Seismology.

The work for the preparation of the Volume was carried out, in general, according to the schedule, although it took longer than originally expected.

Though the Reporter received National Reports from many countries which participated in the IGY seismological program, it would, probably, be inappropriate to draw conclusions of the relative importance of contributions made by separate countries to the general program on the evidence of statistical data. In certain cases the observations of one properly located and equipped station can be of more value than observations from several stations with old equipment and insufficient reduction. Instead it might be useful to attempt an estimation of the general success of the IGY seismological program.

Though the main attention of the researches of the Earth's seismicity has been concentrated on such unstudied regions as the Arctic and especially the Antarctic, the efforts undertaken in other countries should also be noted. It is greatly regretted that no general recommendations were suggetsed prior to the IGY which would have provided for the regular study of the Earth's seismicity, starting with a certain level of magnitude. Thus the selection of stations, which announced their participation in the IGY and which regularly sent their data to the WDCs, was to some extent casual. Altogether 278 of the several hundred seismic stations that operate on the Earth announced their participation in the IGY. Not less than 50 stations were specially established in 17 countries for the IGY. The WDCs regularly received materials from 263 stations (including 21 stations not announced previously). The distribution of these stations according to the continents is as follows: Europe — 88, Asia with neighbouring islands — 51, Africa — 16, North and Central America — 49, South America — 15, Australia and Oceania — 33, Antarctica — 11 (the number of stations may be inaccurate in some cases; any errors will be noticeable when the volume of the IGY Annals with the final catalogue of data collected at the WDCs is published). The data, arranged according to countries, are given in Table 1.

It is noted with regret that not all the stations which made observations during the IGY submitted their data to the WDCs. It would have been useful to collect at the WDCs all the national seismic bulletins for the IGY period.

TABLE 1

1 — The number of announced stations; 2 — The number of stations that sent data more or less regularly to one or both WDCs (the brackets show irregular and infrequent dispatch)

Country	SS		mS	
	1	2	1	2
Argentina	11	2	0	3
Australia	12	11	8	5
Austria	4	2		
Belgium	7	1	2	1
Bolivia	1	1	0	1
Brazil	1	(1)		
Bulgaria	1	1	1	1
Canada	10	5	4	4
Ceylon	1	1		
Chile	10	5	3	0
Colombia	4	4	4	4
Congo	0	4		
Czechoslovakia	6	6	4	3
Denmark	5	3	4	3
East Africa	2	0		
Ecuador	1	0		
Ethiopia	1	1		
Finland	2	2		
France	19	15	11	1
German Dem. Rep.	1	2	1	2

Country	SS		mS	
	1	2	1	2
German Fed. Rep.	9	6	2	1
Ghana	1	0		
Greece	1	1	1	1
Guatemala	1	0		
Hungary	4	1	0	3
Iceland	3	4	3	2
India	15	16	4	6
Indonesia	3	1	1	1
Iran	1	0	1	0
Ireland	1	1	1	0
Israel	2	1		
Italy	0	2	11	4
Japan	6	6	9	14
Lebanon	0	1		
Malgasy Rep.	0	1		
Mexico	10	10	3	3
Mongolia	1	1	1	0
Morocco	1	0	2	0
Netherlands	2	3	1	1
New Zealand	17	16	1	(1)
Norway	2	1	1	0
Pakistan	6	5	2	1
Peru	4	1	1	1
Philippines	8	2	2	0
Poland	12	6	1	3
Portugal	7	6	3	2
Rhodesia and Nyasaland	3	2		
Rumania	6	6	1	2
El Salvador	0	3		
South African Rep.	6	6	3	3
Spain	6	5	9	4
Sweden	4	4	2	2
Switzerland	0	4		
Turkey	0	3		
United Arab Rep.	1	1		
United Kingdom	16	13	3	4
U.S.A.	7	27		
U.S.S.R.	6	21	21	19
Venezuela	0	1		
Dem. Rep. of Viet-Nam	1	1	1	1
Rep. Viet-Nam	1	(1)	1	0
Yugoslavia	5	2	4	2

It should be noted that the equipment of stations was extremely diverse, in addition to modern seismographs old installations with mechanical registration continued to operate, and not all the stations followed the recommendations of the CSAGI Guide in the presentation of their data. Here the influence of definite traditions in the preparation of seismic bulletins can be noticed. However, according to preliminary data, the WDCs in general collected rather copious and very valuable material.

With regard to the major regions in which seismic researches took place, the work in the Arctic has in general confirmed the earlier data about the existence of a considerably active seismic belt in the Polar Basin. The establishment of many new stations in the Arctic has improved the accuracy of the determination of the position of the epicenters and has helped us to follow in detail the transition of this belt to the seismic zones of the North Atlantic on the one side and to those of Eastern Siberia on the other. The interesting coincidence of the extension of this belt with the sub-marine ridges of the Polar Basin allows further conclusions about the tectonic characteristics of this region.

In spite of the high sensitivity of the stations operating in the Antarctic they have registered only a few weak shocks which might be considered of tectonic origin. The combination of the observed considerable recent geological movements with the almost complete aseismicity of Antarctica presents an interesting tectonic puzzle. Only recently have ideas been put forward which provide an explanation of this situation based on the peculiar characteristics of the deep thermal regime of Antarctica. Further study of the earthquake belt surrounding Antarctica will undoubtedly be very useful. We should pay high tribute to the efforts of the seismologists of the U. S. A., Australia, New Zealand, U.K, France and the U.S.S.R. who have undertaken the difficult task of making stationary seismic observations in the extremely difficult condition of the ice continent.

One of the sections of the IGY program which promised results with direct practical significance is the study of microseisms. The connection of microseismic oscillations with the passing of intense cyclones and sea waves has been determined with considerable reliability, and it seemed that the precise tracing of disastrous hurricanes and typhoons from the registration of microseisms was near solution.

In many countries interesting observations of microseisms, originating in the Atlantic, Pacific and Indian oceans and in some of the inner basins have been conducted during the IGY. These observations, however, did not lead to comparable results, and a simple method of tracing sea storms has not yet been found. Theoretical difficulties and doubts have also arisen. As a result it appeared that it was impossible to give a single picture of the formation and propagation of microseisms, though a series of important researches were carried out in Japan, the U.S.S.R., the U.K., Czechoslovakia, the U.S.A. Norway, Denmark, Australia, India, and in other countries.

The WDCs have collected a considerable amount of material of microseismic observations obtained at more than a hundred standard seismic stations and also at a few special tripartite stations in the U.S.S.R. and Japan.

Out of 39 countries, which announced that they would make observations of microseisms, 31 have sent data to the WDCs and, in addition, four countries that did not previously announce microseismic observations in their programs (Argentina, Bolivia, Hungary, El Salvador). The WDCs A and/or B have more or less continuous series of microseismic observations conducted at 110 stations, which compares favourably with the 138 stations announced (see Table 1). The recommendations of the WDC Guide in respect of the materials were generally taken into account.

The list of all the stations with an indication of the amount of materials forwarded can be found in the well-known catalogues of the WDCs A and B. The final Catalogue of data held in all IGY World Data Centres will make one of the volumes of the *IGY Annals*.

It should be noted that no attempt was made to make a fairly full analysis of all the data from the main oceanic basins.

Although the material collected during the IGY has proved to be insufficient for the solution of the problem, it will undoubtedly be of considerable help in future microseismic researches with the new methods that have developed since the IGY.

To a certain extent this situation is reflected in the compilation of the parts of Chapter 4, "Microseisms", as it proved rather difficult to give corresponding summaries of the results of microseismic studies for the main oceanic basins. As a result the composition of Chapter 4 differs from the scheme proposed at the CSAGI Moscow Assembly. In particular summarizing material is not given for the Pacific and Indian oceans. Instead an account is made of the results of the national programs of the U.S.S.R. and India in these regions.

However, in order to give the fullest possible summary to the interested researchers of the studies carried out in these regions, short reports about the study of microseisms in Japan and Australia, taken from the National Reports, have been added to the material submitted by the compilers of this section.

In agreement with the Resolution of the CSAGI Fifth Assembly, Chapter 5 "Structure of the Earth's Crust" includes National Reports of countries that conducted studies of the Earth's crust with seismic methods during the IGY. These methods consist mainly of deep seismic sounding and its different modifications, application of records of industrial explosions at permanent and temporary stations, and the detailed study of records of earthquake. The National Reports on these researches were presented by Australia, Canada, Finland, France, Western Germany, Hungary, Rumania, the U. S. A. and the U.S.S.R.

In order to give a better impression of the scope of the IGY researches of the Earth's crust using seismic methods the National Reports in Chapter 5 are preceded by a map of the regions of these researches. The map has been compiled by P. S. VEITSMAN and I. P. KOSMINSKAYA. An explanatory note to the map is supplied with a bibliography containing the major references.

The acceleration of studies of the structure of the Earth's crust on the oceans and on continents indicates the increased seismic interest in researches of the Earth's interior. It can be noted with satisfaction that the fulfilment of the IGY program in the study of the Earth's crust yielded abundant observational material which made it possible to establish the presence of new types of crustal structure that differ from the classical two-layers scheme, gave new data for the solution of the problem of the origin of continents and oceans, and provided material for further studies planned by the international "Upper Mantle Project".

The study of the Antarctic's structure is, no doubt, one of the most significant achievements of the IGY. These difficult researches were conducted with the participation of only a few countries: Australia, France, Japan, the U.S.S.R., the United Kingdom and the U.S.A. But the results of the accomplished traverses have greatly altered our knowledge about the height of the Antarctic continent and about the thickness of ice which makes its upper layers. Not everything is quite clear, though. The results of the Fuchs—Hillary Expedition and the data of the succeeding American and Soviet expeditions, and the measurements of the ice thickness in the region of the South Pole are not all in agreement. This can lead to partial revision of some observations, but it is unlikely that it will significantly change the general understanding we now have of the relief of Antarctica under the ice. No doubt these studies will provide a useful basis for future seismological and geological researches in the Antarctic.

The last Chapters of the Volume are dedicated to the comparatively new methods of research of the Earth's interior. The IGY program was important not only in the acquirement of new data, but also because it showed the wide range of possibilities in such studies. Research in these fields, until recently carried out almost entirely by seismologists in the U.S.A. is now conducted in many countries.

Recent years have been characterized by the widening of the range of recorded seismic waves. Additional information has been derived from seismograms using waves which, like the Lg and Rg waves, are always present in the records but were not identified until recently and were not used in processing. On the other hand the study of rapid and slow deformations brings seismology closer to geodetic disciplines than a study of the recent crustal movements.

The program of seismic researches during the IGY was far from complete and it did not include many interesting developments. For instance, the numerous determinations of the thickness of the Earth's crust using observations of ordinary surface waves have not been gathered into a single summary. No planning was made for the comparative study of the seismic régime of weak earthquakes in different regions; this might be a work for the future. Nevertheless, the researches that were carried out, make an important link in the inseparable chain of all the sciences about the Earth and the materials collected should for many years still be the source of new researches and discoveries.

The Editor of the Seismological Volume of the IGY Annals acknowledges with deep gratitude the efforts of all those who participated in the work for the compilation of this first summary of results of seismic observations of the IGY program, and takes this opportunity to add a word of praise to the whole army of observers, technicians, computers and researchers, whose labour made it possible to conduct and process these observations.

2. LA SEISMICITE DE L'ANTARCTIQUE

par J. P. Rothé

2.1. Les stations séismologiques

Au cours de l'Année Géophysique Internationale (AGI) quinze stations séismologiques ont fonctionné au Sud du 45e parallèle Sud. L'effort fait pendant l'AGI a donc été important, mais reste encore insuffisant si on désire mettre en évidence une activité séismique régionale. Les stations sont encore trop distantes les unes des autres et trop peu nombreuses: le Tableau 1 permet de comparer la densité des stations séismologiques dans l'hémisphère nord et dans l'hémisphère sud entre les pôles et le 45e parallèle.

Tableau 1
Nombre de stations séismologiques

Latitude	Hémisphère Nord	Hémisphère Sud
45°–50°	51	1
50°–60°	46	1
60°–70°	15	8
70°–80°	6	4
80°–90°	2	1
	120	15

On trouvera ci-dessous la description des stations antarctiques et subantarctiques classées par latitude décroissante.

South Pole: $\varphi = 90°$ S $\lambda = 0°$ (à environ 300 mètres du Pôle, le long du méridien 10°E), $h = 2760$ m (9200 feet)

soubassement : glace sur une épaisseur de 2500 m (8300 feet)

appareils : un Benioff vertical (enregistrement sur film)
$T_0 = 1s2$; $T_g = 0s5$; déroulement 15 mm/mn
un Benioff vertical (enregistrement sur papier)
$T_0 = 1s1$; $T_g = 0s5$; rapport d'amortissement 17 : 1
déroulement 60 mm/mn
un Wilson–Lamison horizontal
$T_0 = 6s0$

Pays ayant la station en charge: U.S.A. (Coast and Geodetic Survey)

Byrd : $\varphi = 79°59'0$ S $\lambda = 120°00'6$ W; $h = 1510$ m
soubassement: la station se trouve sur une couche de glace de 3500 m d'épaisseur
appareils: un Benioff vertical
$T_0 = 1s0$, $T_g = 0s5$; déroulement 60 mm/mn
un Wilson–Lamison NS
$T_0 = 6s2$, $T_g = 3s5$; déroulement 30 mm/mn
un Wilson–Lamison EW
$T_0 = 7s2$, $T_g = 3s5$; déroulement 30 mm/mn
Pays ayant la station en charge: U.S.A. (Coast and Geodetic Survey)

Scott Base : $\varphi = 77°51'$ S $\lambda = 166°48'$ E; $h = 33$ m
soubassement:
appareils: un Benioff vertical
$T_0 = 0s6$; $T_g = 0s2$ et $25s0$; enregistrement sur film de 35 mm avec un déroulement de 15 mm/mn, agrandi 3 fois
un Benioff horizontal NS
$T_0 = 0s5$; $T_g = 0s2$ et $10s0$; enregistrement comme plus haut
un Benioff horizontal EW
$T_0 = 0s6$; $T_g = 0s2$ et $25s0$; enregistrement comme plus haut
Voir: T. HATHERTON, *New Zealand I. G. Y. Antartic Expeditions, Scott Base and Hallett Station*, New Zealand Department of Scientific and Industrial Research, Bulletin 140, Wellington 1961, pp. 107—116.
Pays ayant la station en charge: Nouvelle Zélande

Halley Bay : $\varphi = 75°31'$ S, $\lambda = 26°36'$ W; $h = 220$ m
soubassement: station creusée à 10 mètres de profondeur dans la neige sur l'*ice-shelf*; l'ice-shelf mesure environ 140 m d'épaisseur et repose sur une couche d'eau d'environ 80 mètres; la station est à une cinquantaine de kilomètres au large de la côte.
appareils: Willmore à courte période à trois composantes
$T_0 = 1s$; $T_g = 0{,}25s$; amortissement électromagnétique.
Amplification: $1{,}2 \times 10^5$ pour une période de 1 s.
Vitesse de déroulement: 60 mm/mn.
Voir: J. MACDOWALL and E. M. LEE, Seismological Observations, *The Royal Society International Geophysical Year Antarctic Expedition, Halley Bay, 1955—1959*, Vol. III, The Royal Society, London, 1962, pp. 11—48.
Pays ayant la station en charge: Grande Bretagne (Crombie Seismological Laboratory, Cambridge)

Hallett : $\varphi = 72°19'$ S; $\lambda = 170°13'$ E;
soubassement:
appareils: un Willmore vertical

$T_0 = 1s$; $T_g = 2s$; amortissement critique; vitesse de déroulement: 30 mm
un Columbia vertical et deux Columbia horizontaux (NS et EW)
$T_0 = 15s$; $T_g = 75s$; amortissement (shunt): 100 ohms; vitesse de déroulement: 15 mm/mn

Pays ayant la station en charge: Nouvelle Zélande

Mawson: $\varphi = 67°36'21''$ S, $\lambda = 62°52'48''$ E; $h = 8$ m
soubassement: granite pré-cambrien
appareils: Séismomètres Leet-Blumberg à capacité variable, trois composantes

Pays ayant la station en charge: Australie.

Dumont d'Urville (Terre Adélie): $\varphi = 66°40'$ S, $\lambda = 140°01'$ E; $h = 40$ m
soubassement: roche métamorphique (migmatite)
appareils: un vertical courte période (Institut de Physique du Globe de Paris)
$T_0 = 1s5$
deux horizontaux Galitzine — Wilip Askania
$T_0 = 12s$

Pays ayant la station en charge: France

Mirny: $\varphi = 66°33'1$ S, $\lambda = 93°00'6$ E; $h = 35$ m
soubassement: cave creusée à une profondeur de 2,5 m dans une roche granodioritique
appareils: un Kirnos vertical (SVK)
$T_0 = 12s5$; $T_g = 1s1$; amortissement du pendule: 0,45; amortissement du galvanomètre: 6,5; amplification normale: 750; vitesse de déroulement: 30 mm/mn
un Kirnos horizontal (SHK) SN
$T_0 = 12s5$; $T_g = 1s1$; amortissement du pendule: 0,45; amortissement du galvanomètre: 6,0 à 6,5; amplification normale: 1300 à 1400; vitesse de déroulement: 30 mm/mn
un Kirnos horizontal (SHK) EW
$T_0 = 12s5$; $T_g = 1s1$; amortissement du pendule 0,45; amortissement du galvanomètre: 6,45; amplification normale: 1280 à 1470; vitesse de déroulement: 30 mm/mn
un Kirnos vertical modernisé (SVK-M):
$T_0 = 2s5$; $T_g = 1s0$; amortissement du pendule: 1,9; amortissement du galvanomètre: 3,0; amplification maxima: 12000 environ; vitesse de déroulement: 30 mm/mn

Pays ayant la station en charge: U.R.S.S.

Wilkes: $\varphi = 66°20'$ S, $\lambda = 110°31'$ E; $h = 10$ m
soubassement: gneiss avec intrusions pegmatitiques; la station est à 3 km de l'ice-cap
appareils: un séismographe Press-Ewing à trois composantes:
$T_0 = 15$s, $T_g = 90$s; vitesse de déroulement: 15 mm/mn
voir: PRESS, EWING, LEHNER, A long period seismograph system, *Trans. Am. Geophys. Union*, **39**, 106-108, 1958.
Pays ayant la station en charge: U.S.A. (Seismological Laboratory, Pasadena)

Oasis Bangera: $\varphi = 66°13'$ S, $\lambda = 100°43'$ E; $h = 25$ m
soubassement: roche métamorphique (magmatite)
appareils: un Kirnos vertical modernisé (SVK-M)
$T_0 = 2$s5; $T_g = 1$s2; amortissement du pendule: 2,0 à 4,9; amortissement du galvanomètre: 1,3 à 3,3; amplification maxima 43.000 pour $T = 1$s1; vitesse de déroulement: 30 mm/mn
deux vibrographes horizontaux électrodynamiques (VEGIK-M)
$T_0 = 2$s5; $T_g = 1$s2; amplification maxima environ 2000 dans une bande de 0s2—0s3 à 3s—4s; vitesse de déroulement 30 mm/mn
Pays ayant la station en charge: U.R.S.S.

Gabriel Gonzalez Videla: $\varphi = 64°49'$S, $\lambda = 62°51'$W, h = 3 m
soubassement: socle (bed-rock)
appareils: trois composantes Askania type Galitzine-Wilip
Vertical: $T_o = 5$s7; $T_g = 25$s0
N—S: $T_o = 14$s3; $T_g = 25$s0
E—W: $T_o = 4$s8; $T_g = 11$s7
Amplification maxima: environ 1200; vitesse de déroulement: 30 mm/mn; horloge à quartz
Pays ayant la station en charge: Chili

O'Higgins: $\varphi = 63°20'$S, $\lambda = 57°54'$W, h = 8 m
soubassement: socle (bed-rock)
appareil: Pendule conique orienté N—S; masse 2000 kg; $T_o = 11$s8; amplification maxima: environ 130; enregistrement sur noir de fumée; vitesse de déroulement: 30 mm/mn
Pays ayant la station en charge: Chili

Decepcion: $\varphi = 62°59'$ S, $\lambda = 60°43'$ W; $h = 8$ m
soubassement: Basalte
appareils: un Mainka NS; masse 75 kg; $T_0 = 5$s8;
$\varepsilon = 2{,}3$; grandissement $V = 67$
un Mainka EW; masse 75 kg; $T_0 = 5$s5;
$\varepsilon = 2{,}1$; grandissement $V = 77$
Pays ayant la station en charge: République Argentine

Macquarie Island: $\varphi = 54°29'55''$ S; $\lambda = 158°57'22''$ E; $h = 14$ m
soubassement: laves basaltiques altérées
appareils: un Grenet vertical
$T_0 = 1s3$, $T_g = 0s8$
deux Wood-Anderson horizontaux
$T_0 = 1s$; amortissement: 20:1; amplification: 2900
Pays ayant la station en charge: Australie

Kerguelen (Pointe Molloy): $\varphi = 49°21'39''$ S; $\lambda = 70°04'01''$ E; $h = 22$ m
soubassement: cave dans une coulée basaltique
appareils: un Grenet vertical
$T_0 = 1s08$; $T_g = 0s8$, puis $T_g = 0s3$; amplification maxima 3900 pour une période du sol de 0,7s; vitesse de déroulement: 60 mm/mn
un séismographe vertical à moyenne période
$T_0 = 9s75$, $T_g = 4s24$; amplification maxima: 322 pour une période du sol de 1s5
deux séismographes horizontaux (type Institut de Physique du Globe de Paris)
$T_0 = 15s$; $T_g = 4s25$; vitesse de déroulement: 60 mm/mn
une station tripartite pour l'étude des microséismes, comprenant le séismographe vertical à moyenne période décrit ci-dessus et deux autres séismographes verticaux ($T_0 = 8s9$ à $9s2$; $T_g = 4s20$ à $4s26$) placés respectivement à 1358 m et à 1083 m de la station principale
voir: BALTENBERGER, FLORENS, MOUROT et ATHIAS, La Station Séismologique de Pointe Molloy, *Revue des Terres Australes et Antarctiques Françaises*, fascicule 6, janvier–mars 1959 (La Documentation française), pp. 21–43, Paris 1959.
Pays ayant la station en charge: France

2.2. Liste des épicentres déterminés pendant l'AGI au Sud du 40e degré de latitude Sud

Pour dresser la liste des épicentres déterminés pendant l'AGI (du 1er juillet 1957 au 31 décembre 1958), nous avons utilisé principalement le Bulletin mensuel du Bureau central international de Séismologie dans lequel ont été publiées toutes les observations transmises par les différentes stations séismologiques ayant fonctionné pendant l'AGI (environ 10.000 dépouillements de séismes par mois). Nous avons utilisé également les listes d'épicentres publiées par KOGAN *et al*[9] et par LANDER[11].

Dans le Tableau 2 on a indiqué les coordonnées de 88 épicentres dont la po-

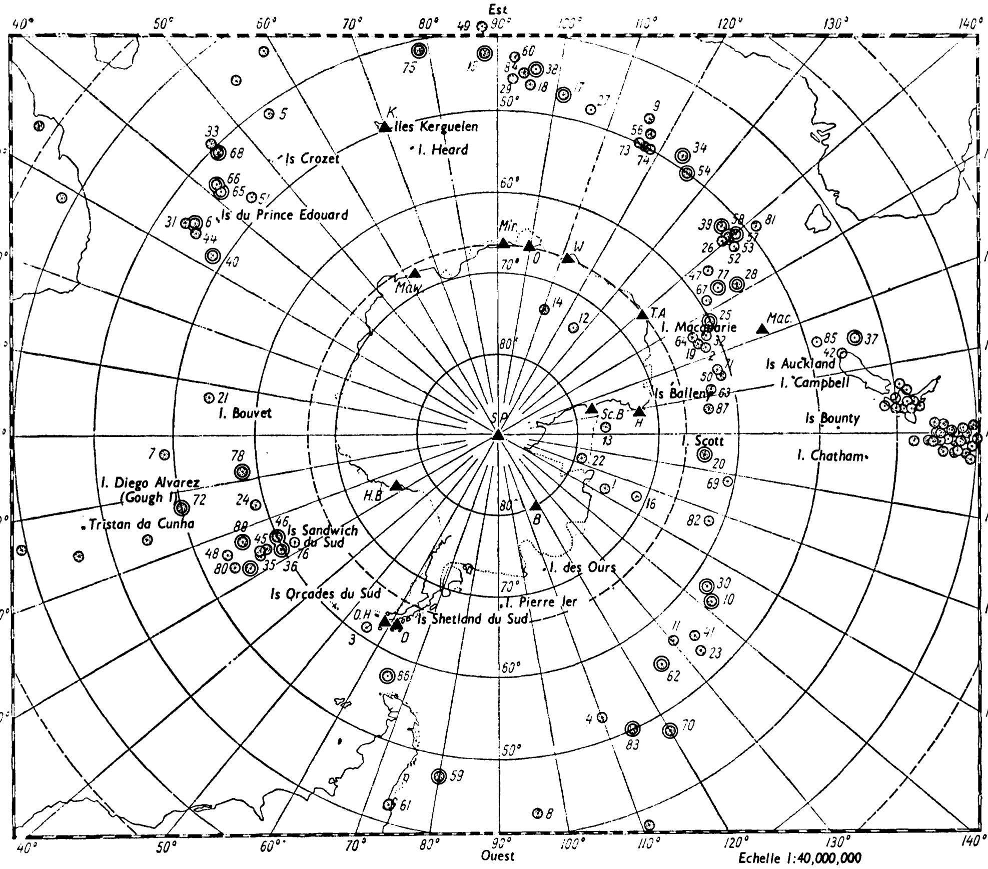

Fig. 1. Carte des épicentres (1er juillet 1957–31 décembre 1958). — Stations séismiques: B: Byrd; D: Decepcion; H: Hallett; H.B.: Halley Bay; K: Kerguelen; Mac: Macquarie; Maw: Mawson; Mir: Mirny; O: Oasis Bangera; O.H.: O'Higgins; Sc.B.: Scott Base; S.P.: South Pole; T.A.: Terre Adélie (Dumont d'Urville); W: Wilkes. ◌ 11 Epicentre et numéro d'ordre. ◎ 83 Epicentre d'un séisme enregistré par plus de 30 stations et numéro d'ordre.

sition accompagnée du numéro correspondant a été reportée sur la carte générale Fig. 1. Nous avons cru bon d'indiquer le nombre de stations ayant enregistré chacun des séismes étudiés; on a ainsi une indication grossière de la magnitude du séisme; lorsque cette magnitude a pu être déterminée exactement, sa valeur figure dans la dernière colonne du tableau. Les magnitudes calculées par la station de Mirny sont suivies de l'abréviation (M). Les épicentres des séismes enregistrés par plus de 30 stations (magnitude approximativement supérieure à $5\frac{1}{2}$) sont pointés sur la carte avec un signe spécial: ◎

TABLEAU 2

Liste des épicentres à partir du 40^{e} degré de latitude Sud (sauf Nouvelle Zélande) pour la période 1 juillet 1957—31 décembre 1958.

Numéro	1957 Date	Heure origine	Coordonnées		Nombre de stations	Magnitude; remarques
			Latitude Sud	Longitude		
—	4. 7	23 03 30	57°	28° W	6	
1	18. 7	19 45,2	76°	154° W	4	D'après J. F. Lander
2	21. 7	05 59 13	$62^{\circ 1}/_2$	156° E	13	6,0 (M); îles Balleny
—	24. 7	13 45	–	–	3	ressenti IV à Macquarie
3	25. 7	22 42,9	62°	55° W	9	inscrit à Decepcion
4	28. 7	02 51,1	$53^{\circ 1}/_2$	110° W	12	
5	4. 8	09 24 24	$40^{\circ 1}/_2$	$54^{\circ 1}/_2$ E	15	Crozet
6	4. 8	21 08 51	45°	35° E	119	7,1 (Quetta), 6 (Moscou), 5,7 (M), $6^3/_4$ (Matsushiro)
7	13. 8	12 19,8	$49^{\circ 1}/_2$	3° W	17	peu précis
8	16. 8	18 29,6	$41^{\circ 1}/_2$	$94^{\circ 1}/_4$ W	14	
9	17. 8	04 30 35	47°	116° E	5	4,4 (M)
10	18. 8	06 34 16	57°	$142^{\circ 1}/_2$ W	41	5,7 (M)
11	20. 8	04 48 08	$55^{\circ 1}/_2$	$131^{\circ 1}/_2$ W	11	5,4 (M)
12	22. 8	04 53,6	$73^{\circ 1}/_2$	$125^{\circ 1}/_2$ E	3	D'après J. F. Lander
13	24. 8	01 36,0	78°	175° E	4	D'après J. F. Lander; douteux
14	24. 8	15 09,2	$72^{\circ 1}/_2$	110° E	6	D'après J. F. Lander; douteux
15	4. 9	04 33 52	42°5	88°5 E	49	5,0 (M); $6\frac{1}{2}$ (Matsushiro)
16	4. 9	22 15,0	$70^{\circ 1}/_2$	$155^{\circ 1}/_2$ W	4	D'après J. F. Lander; douteux
17	9. 9	00 13 31	$47^{\circ 3}/_4$	101° E	89	5,4 (M); $6^3/_4$ (Matsushiro)
18	12. 9	08 45 50	46°	94° E	2	4,3 (M); douteux
19	14. 9	12 37,4	62°5	155° E	20	
20	29. 9	02 08 55	$64^{\circ 1}/_2$	$172^{\circ 1}/_2$ W	36	5,8 (M); 6 (Matsushiro)
21	2.10	20 42 53	$54^{\circ 1}/_2$	5° E	17	5,6 (M); Ile Bouvet
22	4.10	20 56,2	$79^{\circ 1}/_2$	164° W	3	D'après J. F. Lander
23	9.10	01 25 35	$53^{\circ 1}/_2$	$134^{\circ 1}/_2$ W	10	
24	12.10	16 46 30	59°	16° W	23	5,5 (M), 6 (Matsushiro)
25	13.10	20 33 01	60°	151° E	75	$6^3/_4$ (Strasbourg), 5,8 (M), $6^1/_2$–$6^3/_4$ (Matsushiro)
26	14.10	03 16 09	$53^{\circ 1}/_2$	140° E	11	
27	26.10	17 39 43	$48^{\circ 1}/_2$	107° E	13	
28	31.10	15 29 10	55°	148° E	30	5,3 (M), $5^3/_4$ (Matsushiro)
29	7.11	02 25 11	46°	94° E	2	épicentre avec 2 stations
30	7.11	06 21 56	$57^{\circ 1}/_2$	$143^{\circ 1}/_2$ W	44	6 (Matsushiro), 5,8 (M)
31	14.11	22 47 22	43°	35° E	13	BCIS donne 45°S, 35°E
32	19.11	02 34 18	61°5	153° E	11	Balleny
33	22.11	23 37,3	40°	45° E	6	
34	29.11	17 43 38	$48^{\circ 1}/_2$	$124^{\circ 1}/_2$ E	31	5,0 (M)
35	6.12	09 41 29	57°	28° W	21	

Numéro	1957 Date	Heure origine	Coordonnées Latitude	Longitude	Nombre de stations	Magnitude; remarques
36	18.12	20 44 53	60°	28° W	33	
37	31.12	14 28 15	45°	166°0 E	128	6,6 (Wellington), 6 1/2 (Kew, Matsushiro), 6,2 (Quetta)
38	31.12	21 16 03	45°	96° 1/2 E	30	URSS donne 48° S, 73° E
	1958					
39	17. 1	07 15 38	52°	139° 1/2 E	88	6 1/2–6 3/4 (Matsushiro)
40	24. 1	06 48 06	49°	32° E	32	
41	26. 1	03 35 17	54° 1/2	133° W	25	5,8 (Wellington)
42	30. 1	01 25 43	45°9	167°1 E	8	5,0 (Wellington); ressenti
43	25. 2	05 27,7	58°	27° W	5	épicentre peu précis
44	23. 3	02 22 06	45° 1/2	34° E	11	W île Prince Edouard
45	7. 4	03 28 52	57° 1/2	28° W	20	
46	20. 4	21 15 02	60°	25° W	54	
47	28. 4	06 10,7	57° 1/2	142° E	9	
48	8. 5	14 36,7	53° 1/2	25° W	7	
49	15. 5	05 47,8	40°	88° E	7	
50	16. 5	12 57,3	61°	165° E	5	
51	16. 5	22 31 50	48°	43° E	14	
52	24. 5	07 29 04	52° 1/2	141° E	15	
53	7. 6	12 55 01	53°	140° E	53	5,4 (M); 5 1/2–5 3/4 (Matsushiro)
54	13. 6	10 58 44	50°	126° E	38	5 1/4–5 1/2 (Matsushiro)
55	15. 6	18 47 00	46°	16° W	9	
56	18. 6	16 14 25	48° 1/2	116° 1/2 E	18	
57	19. 6	13 33 42	52° 1/2	140° E	15	
58	19. 6	18 02 15	52° 1/2	140° E	56	5 3/4–6 (Matsushiro)
59	24. 6	06 36,4	47°	80° W	53	
60	25. 6	20 56,7	43°	93° E	6	
61	25. 6	23 00,9	42°	73° W	17	
62	3. 7	10 23 02	55°	126° W	40	6 (Kew), 5,6 (M)
63	4. 7	13 05 37	63°	166° E	15	5,1 (M)
64	8. 7	19 28 24	62° 1/2	153° E	13	5,2 (M)
65	8. 7	22 48 36	43°	41° 1/2 E	93	6,2 (Roma, Pasadena), 5 3/4–6 (Matsushiro) 5,7 (Strasbourg), 6 (Kew) 5,1 (M)
66	9. 7	01 08 06	43°	41° 1/2 E	33	
67	11. 7	18 31 34	58° 3/4	147° 1/4 E	12	5,3 (M)
68	26. 7	06 13 50	40°	45° 1/2 E	97	5 3/4 (Moscou), 5 3/4 (Praha, Roma), 6 1/4 (Kew), 6–6 1/4 (Matsushiro)
69	26. 7	08 35 12	60° 1/2	168° 1/2 W	17	
70	30. 7	15 10.2	48°	120° W	40	
71	4. 8	21 01.7	61°	165° E	4	
72	9. 8	12 47 56	50°	13° W	57	
73	22. 8	00 01 21	50°	117° E	26	5 1/4 (Matsushiro)

Numéro	1957 Date	Heure origine	Coordonnées Latitude	Coordonnées Longitude	Nombre de stations	Magnitude; remarques
74	26. 8	13 27.9	50°	117° E	8	
75	12. 9	05 37 46	41°	78°½ E	37	5½–5¾ (Matsushiro), 5,5 (M)
—	17. 9	15 15	–	–	7	peut-être région des Balleny
76	18. 9	03 35 32	61°½	27°½ W	28	
77	1.10	09 29 43	56°9	148°0 E	118	6¼ (Pasadena), 6¼–6½ (Matsushiro), 6,2 (Uppsala), 6,4 (Kew), 5,9 (M)
78	2.10	04 24 27	58°	9°½ W	53	5¾–6 (Matsushiro, Lwiro), 6,1 (Kew), 6,1 (M)
79	9.10	11 20 17	55°½	27°½ W	116	6½–6¾ (Matsushiro), 6½ (Pasadena), 6,3 (Strasbourg), 6,2 (Uppsala), 6,2 (Kew)
80	18.10	19 01 08	53°	26° W	7	
81	23.10	17 50 29	48°½	141°½ E	10	
82	28.10	04 14 55	62°½	157° W	19	
83	4.11	22 54 46	50°	115° W	83	6,2 (Strasbourg), 6 (Pasadena, Matsushiro), 6,2 (Kew)
84	17.11	22 00.9	44°	94° E	3	
85	20.11	06 10 50	48°¾	164° E	8	5,2 (Wellington)
86	24.11	06 48 57	57°½	65°½ W	55	
87	27.11	13 41 47	64°¼	172°½ E	22	5,4 (Wellington): Balleny
88	13.12	04 07 30	55°½	22° W	44	5,6 (Wellington)

2.3. La séismicité de l'Antarctique et des régions subantarctiques

A l'examen de la carte Fig. 1, le fait le plus remarquable est la mise en évidence d'une zone séismique *à peu près continue* s'étendant au large de l'Antarctique orientale. On peut faire débuter cette zone dans l'Océan Indien où elle se sépare de la zone séismique médiane indo-atlantique vers 25° S, 70° E. Suivant une ligne de hauts-fonds jalonnés en particulier par les îles Saint-Paul et Amsterdam, cette zone séismique « indienne-antarctique » vient rejoindre le cercle séismique circumpacifique à l'ouest de l'île Macquarie. Les travaux antérieurs, en particulier ceux de GUTENBERG et RICHTER, laissaient entrevoir l'existence de cette zone, mais comme l'écrivaient ces auteurs[18]: « However, there are large unfilled gaps, even with the addition of a class c shock on 24 December, 1947, near 55° S, 115° E ». *On peut dire que, grâce aux observations séismologiques de l'Année Géophysique, ces lacunes sont aujourd'hui comblées.*

Cette même zone se poursuit par la région des îles Balleny et traverse le Sud-Est du Pacifique en remontant en direction de l'île de Pâques, coïncidant partout avec des lignes de hauts-fonds où la profondeur de la mer reste inférieure à 3000 mètres, parfois même à 2000 mètres. Ici encore les observations de l'Année Géophysique ont permis de préciser le tracé de cette zone: on comparera à ce sujet, la carte Fig. 1 que nous présentons et celle publiée par GUTENBERG et RICHTER (Ref. 18, Fig. 12, p. 44).

L'activité séismique de la boucle des Antilles du Sud dans la région des îles Sandwich est déjà bien connue. L'ouvrage de GUTENBERG et RICHTER contient une liste de 41 épicentres appartenant à cette zone. Plusieurs nouveaux épicentres, déterminés pendant l'AGI, viennent compléter cette liste.

L'activité séismique est moins forte dans la branche sud de la boucle jalonnée par les Orkney du Sud, les Shetland du Sud et l'archipel Palmer dont fait partie l'île Decepcion. Quelques épicentres avaient été signalés par GUTENBERG et RICHTER: nous les rappelons ci-dessous:

26 octobre	1946	00:21:03	60° S	35° W	7
2 mars	1950	18:39:45	60° S	35° W	6,9
24 janvier	1938	10:31:44	61° S	38° W	7,1
7 octobre	1937	07:51:45	$59^{\circ 1}/_2$ S	53° W	$6^1/_4$
27 décembre	1928	04:46:10	61° S	55° W	$6^1/_4$
2 avril	1938	06:02:00	$59^{\circ 1}/_2$ S	58° W	$6^1/_4$
18 novembre	1941	10:14:36	61° S	58° W	7,0
26 octobre	1933	12:07:02	60° S	60° W	$6^3/_4$

Le séisme du 25 juillet 1957 (épicentre N⁰ 3 de notre liste: 62° S, 55° W) appartient à cette zone. Un bon enregistrement de ce séisme a été obtenu par la station de Decepcion située seulement à 360 km de l'épicentre (Fig. 2).

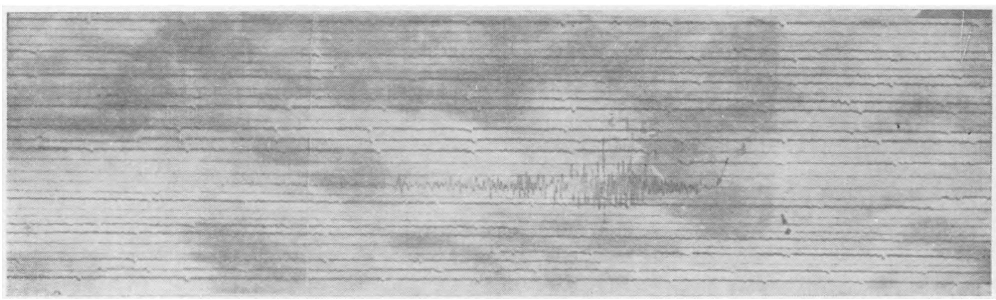

FIG. 2. Ile Decepcion: enregistrement du séisme du 25 juillet 1957 à 22h. 44 min (D = 360 km).

Au delà de l'archipel Palmer, dans la Péninsule antarctique, *l'activité séismique cesse complètement: il y a là un fait surprenant.* En effet, certains géologues, D. G. PANOV en particulier, pensent que la zone des plissements secondaires et tertiaires qui forme les Andes et la boucle des Antilles du Sud se poursuit vers l'Ouest par la Terre de Graham et la Terre de Marie Byrd; par les îles Balleny

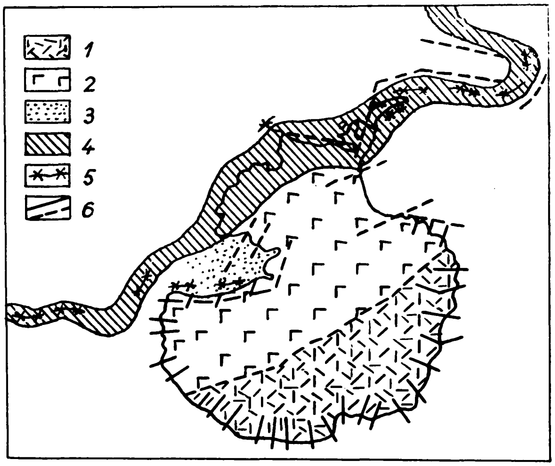

FIG. 3. Schéma tectonique de l'Antarctique (d'après PANOV). 1 — Socle ancien ; 2 — Plateforme; 3 — Zone d'affaissement dans la plateforme; 4 — Plissements secondaires et tertiaires; 5 — Lignes d'activité volcanique; 6 — Lignes de fractures principales.

cette zone viendrait ainsi rejoindre aux environs de l'île Macquarie le prolongement — séismiquement très actif — des plissements néo-zélandais (Fig. 3).

Plusieurs auteurs se sont demandés si la zone séismique des Antilles du Sud venait rejoindre à l'Est la grande zone séismique médiane « indo-atlantique ». Au point de vue structural, cette jonction est peu probable, car la boucle des Antilles du Sud — comme celle des Antilles du Nord — appartient tectoniquement à la zone andine et elle se prolonge, comme on l'a dit plus haut, vers la péninsule antarctique.

Les déterminations d'épicentres faites pendant l'AGI ne permettent pas de conclure de façon définitive.

La séismicité du continent antarctique lui-même est faible: les deux seuls épicentres qu'on ait pu y déterminer sont ceux du 22 août 1957 (N° 12) et du 24 août 1957 (N° 14); encore faut-il remarquer que les coordonnées de ces deux épicentres n'ont pu être calculées de façon précise, le nombre des données recueillies étant très faible.

Il convient cependant de noter que les stations séismologiques sont souvent distantes entre elles de plusieurs milliers de kilomètres: dans l'Antarctique occidentale, il y a respectivement 2200 km et 2600 km entre la station de Byrd et celles de Halley Bay et de Decepcion. Les distances de la station de South Pole aux autres stations antarctiques sont les suivantes:

Byrd	1110 km	Mirny	2600 km
Scott Base	1350 km	Terre Adélie	2600 km
Halley Bay	1610 km	Wilkes	2630 km
Hallett	1970 km	Decepcion	3000 km
Mawson	2500 km		

Il n'est pas douteux, par conséquent, que la faible densité des stations ne permet pas d'enregistrer tous les séismes susceptibles de se produire dans le continent antarctique.

2.4. Séismicité locale

La station australienne de l'île Macquarie, située à proximité d'une zone séismique active, connue depuis longtemps, enregistre fréquemment des séismes proches dont certains ont été ressentis dans l'île Macquarie. Le bulletin de la station contient les données de 82 séismes proches inscrits au cours des 18 mois de l'AGI.

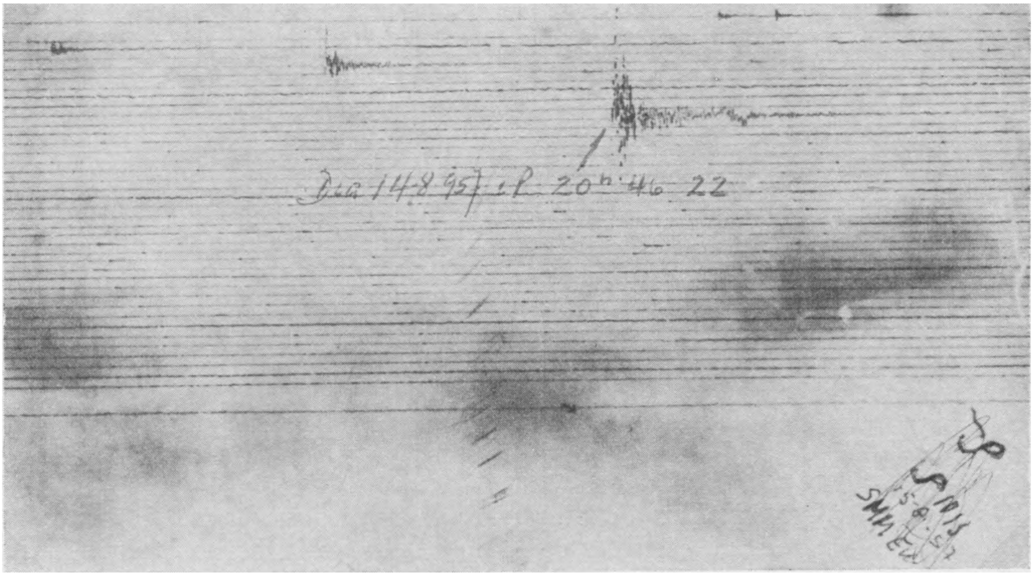

FIG. 4. Ile Decepcion: 14 août 1957, essaim de secousses volcaniques locales; secousse principale à 20 h. 46 mn 22s.

Le Directeur du Service météorologique national de la République Argentine, le Docteur F. L. FERNANDEZ, a communiqué la liste ci-dessous des secousses d'origine locale ayant été enregistrées par la station de l'île Decepcion.

16 janvier	1957	iP	23	47	48	local	
26 février	1957	iP	16	25	28	local	
3 avril	1957	iP	21	10	09	local	
22 avril	1957	iP	23	33	32	local	
14 août	1957	iP	10	47	12	local	
14 août	1957	iP	12	50	47	local	
14 août	1957	iP	20	46	22	local	(Fig. 4)
15 août	1957	iP	13	42	19	local	(Fig. 5)
16 août	1957	iP	08	12	22	local	
6 août	1958	iP	12	44	35	local	(Fig. 6)

Le séismographe utilisé ayant une faible amplification, il est probable, ajoute le Docteur Fr. Lucio FERNANDEZ, que la plupart de ces secousses correspondent

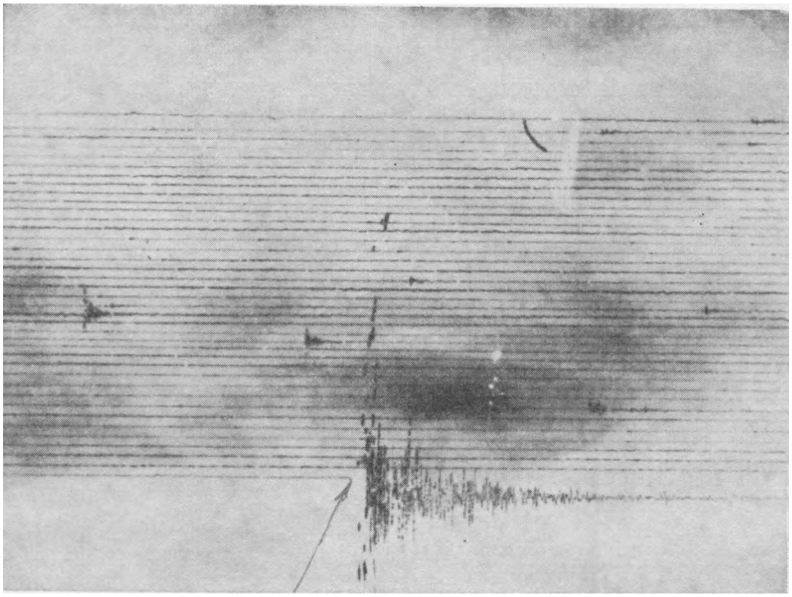

FIG. 5. Ile Decepcion: 15 août 1957, essaim de secousses volcaniques locales; secousse principale à 13h. 42 mn 19s.

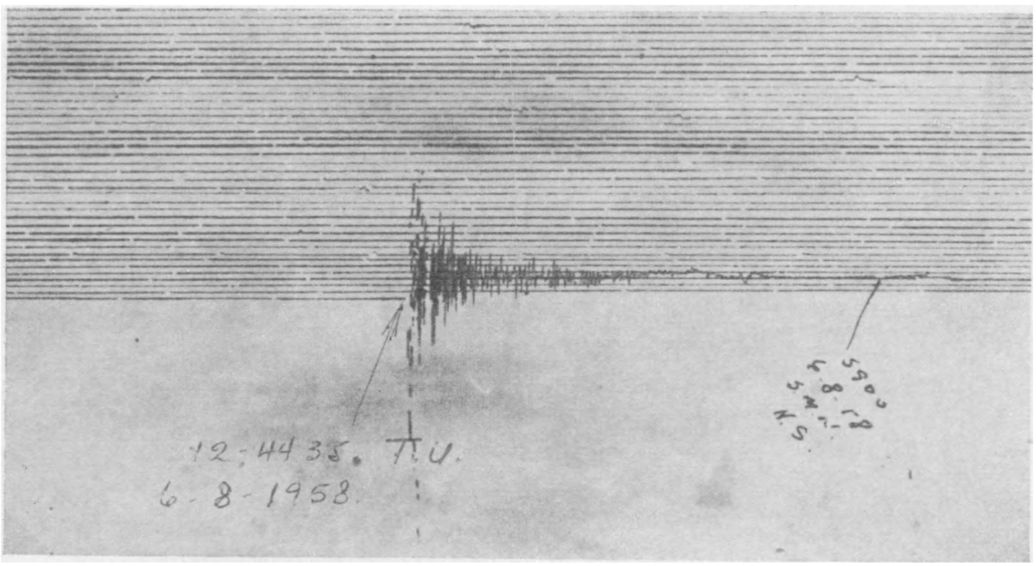

FIG. 6. Ile Decepcion: 6 août 1958 à 12 h. 44mn 38s., secousse locale d'origine volcanique.

à une activité locale limitée à l'île elle-même et qu'elles ont une origine volcanique.

Les bulletins de la station «Scott Base» contiennent un certain nombre de données se rapportant à des chocs locaux. D'après T. HATHERTON plus de 300 séismes proches dont la magnitude a toujours été inférieure à 3 ont été enregistrés à Scott Base pendant l'A. G. I. Pour 76 de ces séismes, la différence S—P est de l'ordre de 4 à 6 s; il pourrait s'agir de séismes liés à l'activité du Volcan Erebus situé à 30 km environ au nord de la station. On trouvera dans les articles de HATHERTON des reproductions de plusieurs enregistrements de séismes proches[21], [22].

D'après le rapport de I. P. PASSETCHNIK et S. D. KOGAN, aucun séisme proche n'a été enregistré dans les stations de Mirny et d'Oasis. Le seul cas douteux est celui du 29 août 1958 à 12h.42 mn: l'inscription (Fig. 7) ressemble à celle d'un séisme proche (distance épicentrale D = 20 km). Cependant, il est

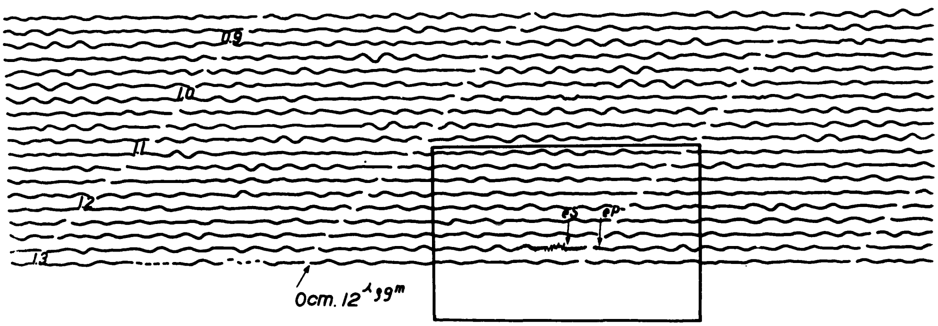

FIG. 7. Mirny, 29 août 1958, secousse locale (eP 12h. 42mn 27s., eS 12h. 42mn 33s., D = 20 km).

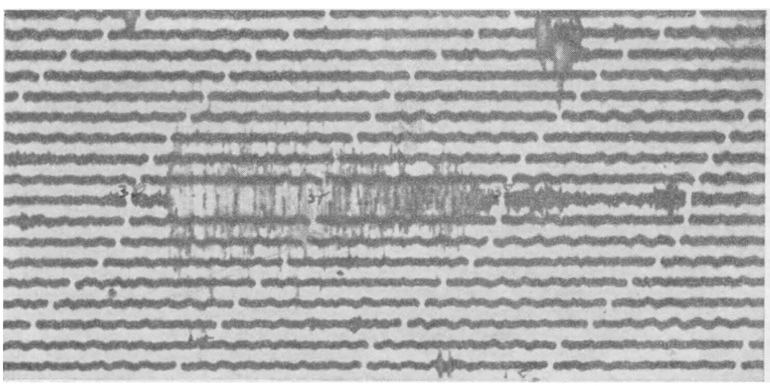

FIG. 8. Dumont d'Urville (Terre Adélie), 25 août 1959, séisme proche (iPg 13h. 35mn 41s., iSg 13h. 35mn 57s., D = 145 km).

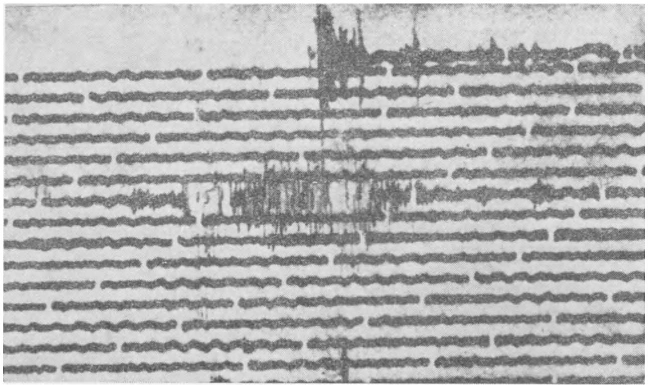

Fig. 9. Dumont d'Urville (Terre Adélie), 26 août 1959, séisme proche (iPg 05h. 14 mn 13s., iSg 05h. 14 mn 29s., D = 120 km).

possible que le mouvement enregistré dans ce cas soit dû à la formation de crevasses dans la glace ou au déplacement d'un iceberg raclant le fond dans une zone d'eau libre au voisinage de la station ou même au choc d'un iceberg sur le bord de l'ice-shelf.

La station française Dumont-D'Urville, en Terre Adélie, a enregistré — en 1959 — quelques séismes proches; les inscriptions du 25 et du 26 août 1959 sont reproduites Fig. 8 et 9.

2.5. Inscriptions indéterminables

En raison même de l'éloignement des stations antarctiques les unes des autres, de nombreuses inscriptions, obtenues isolément ou en deux ou trois stations seulement, restent indéterminables.

J. P. Passetchnik et S. D. Kogan, ont dressé une liste de 176 inscriptions de séismes obtenues isolément par les stations de Mirny ou d'Oasis pendant les 18 mois de l'Année Géophysique Internationale.

Par ailleurs, en utilisant le Bulletin mensuel du Bureau central international de séismologie et les données supplémentaires publiées par le Coast and Geodetic Survey (*Seismological Bulletin*, IGY Supplement) nous avons, pour les seuls mois de juillet et d'août 1957, relevé respectivement 116 et 191 séismes *inscrits uniquement par des stations antarctiques* et dont il n'est pas possible de fixer l'origine.

A titre d'exemple, on trouvera ci-dessous la liste de ces séismes pour le mois de juillet 1957.

Juillet 1957

Séismes inscrits uniquement par des stations antarctiques

Date	Station		Heure	
1	South Pole	iP	02 21 (35)	
1	Mirny	eP	17 42 43	
2	Terre Adélie	iP	07 53 49	
	Scott	eP	55 06	
2	Scott	e	14 56 54	
2	Halley Bay	e	20 33 26,7	i 20 37 26,3
3	Hallett	e	06 36 07	
3	Hallett	e	09 58 50	
3	Hallett	iP	10 05 01	
4	Byrd	eP	06 26 57	
5	Scott	e	00 34	proche
5	Terre Adélie	i	15 32 51	
5	Scott	e	16 59 29	
	Hallett	eP	17 01 43	
	Terre Adélie	i	02 42	
5	Scott	e	20 19 53	
6	Macquarie	iP	05 25 19	
6	Terre Adélie	e	21 43 48	
7	Scott	e	09 58 11	local
7	Byrd	eP	15 41 46	
7	South Pole	eP	17 23 44	i 17 23 55
9	Byrd	e(P)	03 28 08	
9	Byrd	eP	07 14 52	
9	Hallett	iP	07 57 06	e 07 58,2 P-S ou P-L about 1mn
9	Byrd	e(P)	18 49 52	
10	Hallett	eP	08 10 49	e (S) 08 12 05 local
10	Mirny	e	09 03 15	L 10 06,2
10	Halley Bay	e	17 08 38,5	
11	Terre Adélie	i	08 33 22	
11	Terre Adélie	iP	11 22 58	
11	Macquarie	iP	18 46 41	iS 18 47 25 (D = 4°2)
11	Hallett	iP	23 52 11	
12	Scott	iP	00 02 07	
12	Scott	iP	04 16 29	local
12	Mirny	iP	07 34 35	
12	Hallett	e	07 49,7	e 07 51,2
12	Oasis	i	20 52 43	
12	Scott	e	22 16 49	
13	Macquarie	i	00 58 10	
13	Scott	iP	01 50 38	
13	Halley Bay	i	16 07 15,3	
	Byrd	iP	09 57	
	Scott	iP	11 22	i 16 13 10
13	Terre Adélie	i	19 51 23	

Date	Station		Heure	
13	Scott	e	20 29 21	
14	Hallett	i	03 38 10	groupe S d'un séisme local
14	Scott	eP	07 02 52	
15	South Pole	eP	12 42 05	i 12 49 17
15	Scott	eP	12 48 02	
15	Terre Adélie	i	23 42 29	
16	Hallett	e(P)	16 27 59	
17	Oasis	e	01 50 45	
17	Scott	e	03 22 38	
17	South Pole	iP	04 33 42	
	Byrd	eP	34 15	
17	Scott	e	11 49 34	
17	Mirny	eP	13 07 40	
17	South Pole	iP	13 53 05	
17	Byrd	eP	22 02 29	
18	Scott	e	00 41 52	
18	Hallett	eP	02 34 55	iS 02 37 58
	Byrd	iP	35 58	
18	Scott	e	09 07 10	
	South Pole	eP	09 08 25	
18	Scott	e	10 09 01	
18	Macquarie	iP	14 09 57	i 14 10 00
	South Pole	eP	14 39	
18	Byrd	iP	16 51 20	
18	Scott	e	20 01 15	
18	Scott	e	21 52 14	
19	Mirny	iP	04 55 39	
19	Mirny	eP	09 13 21	
19	Scott	iP	13 38 43	
19	Terre Adélie	e	14 44 13	
	Scott	e	46 18	
19	Hallett	e	15 54 54	
19	Mirny	e	16 21 54	
20	Byrd	eP	11 23 29	
20	Scott	e	13 07 47	
20	Hallett	e	13 32 46	
20	Mirny	i	16 33 16	
20	Scott	e	19 03 40	
20	Scott	e	21 20 26	
21	Hallett	iP	02 08 42	
21	Mirny	i	03 54 25	
21	Hallett	e	15 10 14	
21	Mirny	i	16 09 10	
22	Scott	eP	10 45 35	
	Mirny	i	46 02	
23	Macquarie	i	02 00 04	
23	Mirny	e	07 21 28	

Date	Station		Heure	
23	Terre Adélie	iP	09 53 52	
23	Scott	e	22 37 33	
	Hallett	e	38 45	
24	Macquarie	iP	02 11 59	iS 02 12 09
24	Hallett	e	02 26 28	e 02 29,8
24	Scott	eP	02 36 42	
24	Oasis	e	03 52 51	
24	Scott	e	06 22 42	
24	Halley Bay	e	11 15 11.7	i 11 15 16.3
25	Scott	eP	01 26 25	local
25	Terre Adélie	iP	16 36 39	
26	Macquarie	iP	04 25 26	iS 04 25 31
26	Terre Adélie	iP	06 12 57	
27	Scott	e	06 48 50	
28	Scott	e	05 44 47	
28	Hallett	eP	09 10 01	
28	Oasis	P	15 23 26	
28	Macquarie	iP	22 28 58	iS 22 29 01
29	Scott	e	00 43 24	
29	Oasis	e	03 51 50	
29	Terre Adélie	iP	05 31 59	
29	Oasis	iP	09 22 47	
29	Macquarie	iP	11 06 30	iS 11 06 34
29	Scott	e	15 59 07	
29	Mirny	eP	19 35 13	
29	Hallett	eP	22 09 53	
29	Hallett	iP	22 19 22	
30	Hallett	iP	02 22 03	e(S) 02 24 48
	Scott	e	02 23 13	
30	Scott	e	02 38 08	
30	Hallett	e	04 16 56	
30	Scott	e	05 57 14	
30	Hallett	e	07 27 38	
30	Scott	e	07 50 20	e 07 51 09
	Hallett	eP	51 37	
30	Scott	e	08 11 39	
30	Hallett	iP	13 53 17	
30	Scott	e	17 05 51	
30	Scott	e	20 56 50	
30	Hallett	iP	21 51 14	local
31	Macquarie	iP	10 37 44	iS 10 37 46
31	Scott	e	10 54 00	

2.6. Travaux séismologiques concernant l'Antarctique

(A) *La structure de l'Antarctique*

En étudiant la propagation des ondes de Love et des ondes de Rayleigh et en comparant les résultats de cette étude avec ceux obtenus par la prospection séismique de l'inlandsis, plusieurs auteurs ont essayé de tirer des conclusions quant à la structure du continent antarctique[3,4,15,16,23,24].Ces recherches conduisent à penser que la partie orientale du continent antarctique présente une structure de type continental, avec une épaisseur crustale d'environ 35 km. Au contraire, à l'Ouest d'une ligne joignant la mer de Ross à la mer de Bellingshausen, le socle doit se morceler en groupes d'îles qui viennent se raccorder à la « péninsule antarctique » et à l'arc des Antilles du Sud. D'importantes parties de l'Antarctide se trouvent au dessous du niveau de la mer; le véritable continent antarctique est sans doute moins étendu qu'on ne pourrait le penser au premier examen de la carte.

Nous citerons ici la conclusion du récent mémoire de F. F. EVISON, C. E. INGHAM, R. H. ORR et J. H. LE FORT[4].

"It is concluded that eastern Antarctica as a whole is continental, for the present study has given a good coverage of this large area, and a crustal thickness of about 35 km has been consistently obtained. By contrast, Marie Byrd Land has an average crustal thickness of about 25 km and cannot be regarded as truly continental in structure. The data do not extend to the remainder of western Antarctica, nor has it been possible to study the important transition zone which runs from the Ross Sea to the Weddell Sea.

These findings are evidently related to the known geological contrast between eastern and western Antarctica. The result for the western sector, however, requires further elaboration, for the average crustal thickness of 25 km could be made up in a variety of ways. Beneath the high mountain ranges of Marie Byrd Land one would expect to find crustal roots giving locally a total thickness of at least 40 km. If this is so there must be other zones where the crust is much thinner than 25 km, implying that the rock surface lies well below sealevel. The offshore depth contour at 1000 fathoms generally lies much further from the coastline in western than in eastern Antarctica; this would account for some small fraction of the difference in average crustal thickness.

The picture of Marie Byrd Land as an archipelago made up of mountainous island chains finds support in recent seismic and gravity observations. A detailed survey of ice thickness has revealed extensive regions where the rock surface would undoubtedly be below sea-level if the ice were to melt. For example, along a traverse running close to latitude 77° S, between longitudes 87° W and 113° W, the rock surface was located consistently at between 1 km and 2 km below sea-level for a distance of some 400 km. (BENTLEY and OSTENSO)[2].

The conclusion that eastern Antarctica as a whole is continental and that the smaller average thickness of crust is confined to western Antarctica differs substantially from the conclusion of PRESS and DEWART[16]. For paths in all directions from Wilkes Station except to the south-east these authors obtained an average crustal thickness about as small as that indicated for Marie Byrd Land in the present study, and concluded that "at most only three-fourths of the Antarctic ice sheet is underlain by continent, the remaining area being oceanic in structure". The two exceptional paths, which passed close to Hallett Station, indicated a fully continental thickness. Six paths included approximately equal distances across eastern and western Antarctic, and traversed the Ross Sea–Weddell

Sea zone. If it happens that this zone included a wide belt of very thin crust, the average thickness obtained by PRESS and DEWART may prove to be consistent with the present results. Three further paths crossed eastern but not western Antarctica, and for these it is not easy to find agreement unless there is a large region of oceanic crust south-west of Wilkes Station, which seems unlikely since Wilkes itself is situated on rock and the value of gravity there is close to normal."

Des résultats analogues ont été obtenus par KOGAN, PASSETCHNIK et SULTANOV[23] qui ont également étudié, d'après les enregistrements de la station Mirny, la propagation des ondes superficielles de Love et de Rayleigh. Ces auteurs ajoutent: "Thus the obtained results show that the Eastern Antarctic is a really continent and the crustal structure between the antarctic continent and the belt of Alpian folding structures is typical for oceans . . . The oceanic structure of the crust between the Antarctic and the belt of the Alpian folding structures is confirmed also by the absence of the Lg waves on records of stations Mirny and Oasis. The parts of the Atlantic and Indian Oceans adjoining the Eastern Antarctic continent are probably not the part of the Eastern Antarctic platform."

Enfin nous citons les conclusions d'un récent travail de KOVACH et PRESS[24]:

"In the light of more recent theoretical calculations on a digital computer the Love and Rayleigh wave data of EVISON *et al.* from five earthquakes recorded at Hallett Station, Scott Base and Mirny have been reinterpreted. A mean crustal thickness of 40 km is indicated for Eastern Antarctica. Less certain is an indication of about 30 km for Western Antarctica. EVISON's determination of 10 km for the crustal thickness in the South Indian Ocean is unproven because of the inability of Love wave dispersion data for greater than 22 seconds period to distinguish between a 5 km or a 10 km oceanic crust."

(B) *Les microséismes dans l'Antarctique*

Les stations séismologiques côtières ont généralement enregistré de forts mouvements microséismiques. Les dépouillements détaillés ne sont pas encore achevés. Cependant, nous pouvons mentionner plusieurs publications intéressant l'agitation microséismique dans l'Antarctique.

T. HATHERTON et R. H. ORR ont étudié les microséismes à Scott Base[6]:

"The Benioff seismographs, installed at Scott Base in March 1957, have revealed a large seasonal variation in microseismic activity. Maximum activity occurs in the late summer and autumn months. On the other hand, in July–August–September, the level of activity is very low, being almost negligible on the short-period vertical component records. Then, in late spring, activity begins to increase again.

This seasonal microseismic variation results in a large seasonal variation in the number of readable earthquake records. In July–August–September 1957 an average of 60 *P* phases per month were registered; whilst in January–February–March 1958 the average per month was reduced to 12.

Microseisms of period 1–4 sec are considered to be due to conditions in the Ross Sea, which cause microseisms whenever ice conditions permit.

Microseisms of period 4–10 sec appear throughout the year. Their origin probably lies in the cyclones circulating round the oceans surrounding the Antarctic continent.

Long-period (30 sec) irregular microseisms occurring during the nights or early morning hours are thought to be due to cooling of the seismograph hut or the ground, during periods when the sun is at low altitude."

Citons encore les conclusions du mémoire de T. HATHERTON[20]:

"1. The seasonal variation of ice cover the Ross Sea and the oceans surrounding the Antarctic continent offers a partial control over the complex wind-sea-crust phenomena which generate and transmit microseisms, and this may be useful in the investigation of both microseisms and crustal structure.

2. Three classes of microseisms are recorded at Scott Base: short-period microseisms (1—3.5s) which are due to winds over the Ross Sea; long period microseisms (4—7s) which, it is thought, are created by swell from meridional and circular cyclones moving onto the continental shelf in the Ross Sea; and ultra long-period microseism (7 ½—10s) which are considered to be produced at or near the storm centre and transmitted through both oceanic and continental crusts to the Base.

3. The maximum amplitude of ground motion varies as the fourth power of the period of the microseisms. This relationship may allow a test of any quantitative theory of microseisms though at present the uncertainty in the mode of transmission and the lack of reliable attenuation measurements make this test difficult to apply. Confirmation of a mechanism of generation would, conversely, allow an investigation of transmission phenomena.

4. The complete absence of short-period microseisms when the Ross Sea is frozen confirms that sea roughness is essential to the generation of microseisms."

D'autre part, J. MAC DOWALL a publié une note sur la relation entre la vitesse du vent et l'amplitude des microséismes sur la côte de l'Antarctique, d'après les enregistrements obtenus à Halley Bay[12].

"Considerable microseismic activity was observed at the Royal Society Base, Halley Bay, during three summer months (December to February), particularly during on-shore winds. For the remaining eight or nine months microseisms became active only during prolonged spells of the strongest winds. It was therefore concluded that the microseisms originated at the ice-front and that the cover of sea-ice damped out this movement for three-quarters of the year and stifled microseismic activity."

Enfin A. D. SYTINSKY a fait une étude détaillée des relations entre les conditions météorologiques et l'agitation microséismique enregistrée à Mirny.[2] Il a présenté ses conclusions de la facon suivante:

Depending on the season there are registered in Mirny two types of microseisms which differ in period and place of formation.

1. The long-period microseisms with $T \simeq 5$—10 s are registered during the whole year; these microseisms have no direct connection with weather in Mirny and are mainly connected with the cyclonic activity in the region of Mawson, Marion, Kerguelen, Mirny.

The greatest increase of long-period microseisms, is observed with the meridional type of circulation. The zonal shift does not stimulate the increase of microseisms. The presence of the ice belt at the Antarctic coast in winter excludes the surf and other causes of the formation of microseisms near the coast.

2. The short-period microseisms with $T \simeq 2$—4 s appear in spring in October-November with the formation of a stationary polynia at the North-Western end of the Shackleton Shelf Glacier and disappear in April-May. Between the

increase in the wind velocity of cyclonic origin in Mirny and the amplitudes of microseisms mainly the synchronic connection is observed. When the so-called "down-slope" wind blows with any velocity the shortperiod microseisms do not originate."

2.7. Conclusions

Les enregistrements séismiques obtenus dans l'Antarctique à l'occasion de l'Année Géophysique Internationale ont permis de préciser très nettement la forme des principales zones séismiques subantarctiques. Certains problèmes restent à résoudre, comme par exemple la jonction de la zone séismique de la boucle des Antilles du Sud avec la zone séismique médiane indo-atlantique, ou encore l'absence de séismicité dans la péninsule antarctique et dans la zone plissée de l'Antarctide occidentale. La densité des stations séismologiques est encore trop faible pour apporter une solution définitive à ces questions. Il est à souhaiter que de nouvelles stations soient installées dans l'Antarctide occidentale et que leur appareillage soit aussi sensible que possible.

Le fait que de très nombreux séismes soient inscrits isolément par certaines stations antarctiques doit également attirer l'attention des observateurs qui poursuivraient après l'Année Géophysique des enregistrements dans ces stations: peut-être certaines de ces inscriptions sont-elles dues à des causes artificielles (chute de séracs, vêlage ou basculement d'icebergs, etc.) qu'il conviendrait de mettre en évidence avec soin.

Les premiers résultats obtenus par l'étude de la propagation des ondes superficielles sont très prometteurs. De nouvelles recherches doivent permettre de préciser la structure de la zone de contact entre l'Antarctide orientale et l'Antarctide occidentale.

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3. SEISMICITY OF THE ARCTIC

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3.1. Introduction

THE CSAGI Working Group on Seismology "approved the proposal of the U. S. S. R. to establish three seismological stations in the Arctic zone and recommended to other nations which maintain seismological stations in that zone that the equipment of their stations should be improved." We shall see in a later section that this recommendation was more than fulfilled. Three existing Arctic stations were re-instrumented and four new stations were installed; ten stations, lying so close to the Arctic that they contribute to our knowledge of its seismicity and structure, were installed or re-instrumented, and five additional Arctic stations were in such an advanced state of planning at the close of the IGY that their future operation is assured.

The working group also recommended that a study be made of the seismicity of inaccessible areas; the Arctic was specifically mentioned as one such area. It was clear to the working group that the study of seismicity cannot make very much progress in eighteen months. This report must therefore be regarded only as an interim report on this subject; the chief value of the IGY will have been in setting up the necessary seismic networks to permit a continued study of seismicity.

It should be borne in mind that this chapter deals with the *seismicity* of the Arctic, not with seismic studies of structure and the like. Microseisms, refraction studies, surface wave studies are all treated later separately in this volume. The Bibliography on pp. 41—45 does include known papers on Arctic microseisms and on the structure of the Arctic basin.

3.2. Knowledge before the IGY

The history of seismograph stations in and near the Arctic is given in Table 1. In preparing this table we have been faced with the problem of which stations to include. Very sensitive stations south of the Arctic circle may be of more value in the study of seismicity than insensitive stations within the circle; indeed in the case of very large earthquakes, stations throughout the entire northern hemisphere may contribute to the epicentral determination. We have made the arbitrary decision to include all stations north of latitude 55°.

It will be clear from the table that, prior to 1950, there were few sensitive seismograph stations actually within the Arctic circle. The early stations at Reykjavik, Disko Island, Advent Bay and Abisko were very insensitive by modern standards, and later stations, such as Ivigtut and Scoresbysund, were not much more sensitive in the short-period range. The situation was improved with the mounting of a short-period vertical in Scoresbysund in 1950, with the installation of Resolute and Kiruna in 1951 and the re-instrumentation of Reykjavik in 1952, but even with these stations knowledge of the seismicity of the Arctic has had to depend largely on sensitive stations south of the Arctic circle. Fortunately there have been enough good stations well placed for this purpose that no major earthquake has been overlooked for many years. Minor seismicity, however, is another matter. Coverage of the Arctic, either by historical records or by seismic stations has been very incomplete, and any list of minor earthquakes is certain to be prejudiced by this fact. No such list should be used for statistical studies.

With this limitation in mind we have produced, in Table 2, as complete as possible, a list of Arctic epicentres, but we have plotted in the map of Fig. 2 only those earthquakes with a magnitude 5 and greater. It seems probable that coverage for earthquakes of this magnitude is reasonably complete. The table is limited to epicentres north of the Arctic circle. This again is an arbitrary decision, for we may in this way have eliminated epicentres just south of the circle which are genetically related to epicentres north of the circle, but the adoption of any particular parallel would have been equally arbitrary.

The list produced in Table 2 has been taken from the International Seismological Summary (I. S. S.), from the Bulletin of the Bureau Central International de Séismologie (B. C. I. S.), from station bulletins and from a number of review papers. Notable among these are the papers by E. A. HODGSON[26] and by RAIKKO and LINDEN[50]. Both these summary papers quote a number of older individual references; these have all been included in the Bibliography, which we have attempted to make as complete as possible. The most important aid in the preparation of Table 2 has been "Seismicity of the Earth" by GUTENBERG and RICHTER[19]. This book treats Arctic epicentres with the same completeness and precision as those in the other parts of the world. Because it makes use of modern techniques and travel-time tables, and all available data, we have regarded it as more accurate than other sources, and have used its epicentral coordinates in case of any disagreement.

We shall return to a discussion of Fig. 2 in a later section.

3.3. Work during the IGY

Examination of Table 1 will show the extent of expansion in Arctic and sub-arctic stations and in the instrumentation of existing stations. Canada rebuilt the vault at Resolute and completely re-instrumented the station. Denmark installed the most northerly station in the world, at Nord, Greenland, and improved the instrumentation of its existing station at Ivigtut. Finland improved the instrumentation of its chief station at Helsinki and constructed 9 other stations, some of which did not come into operation until after the close

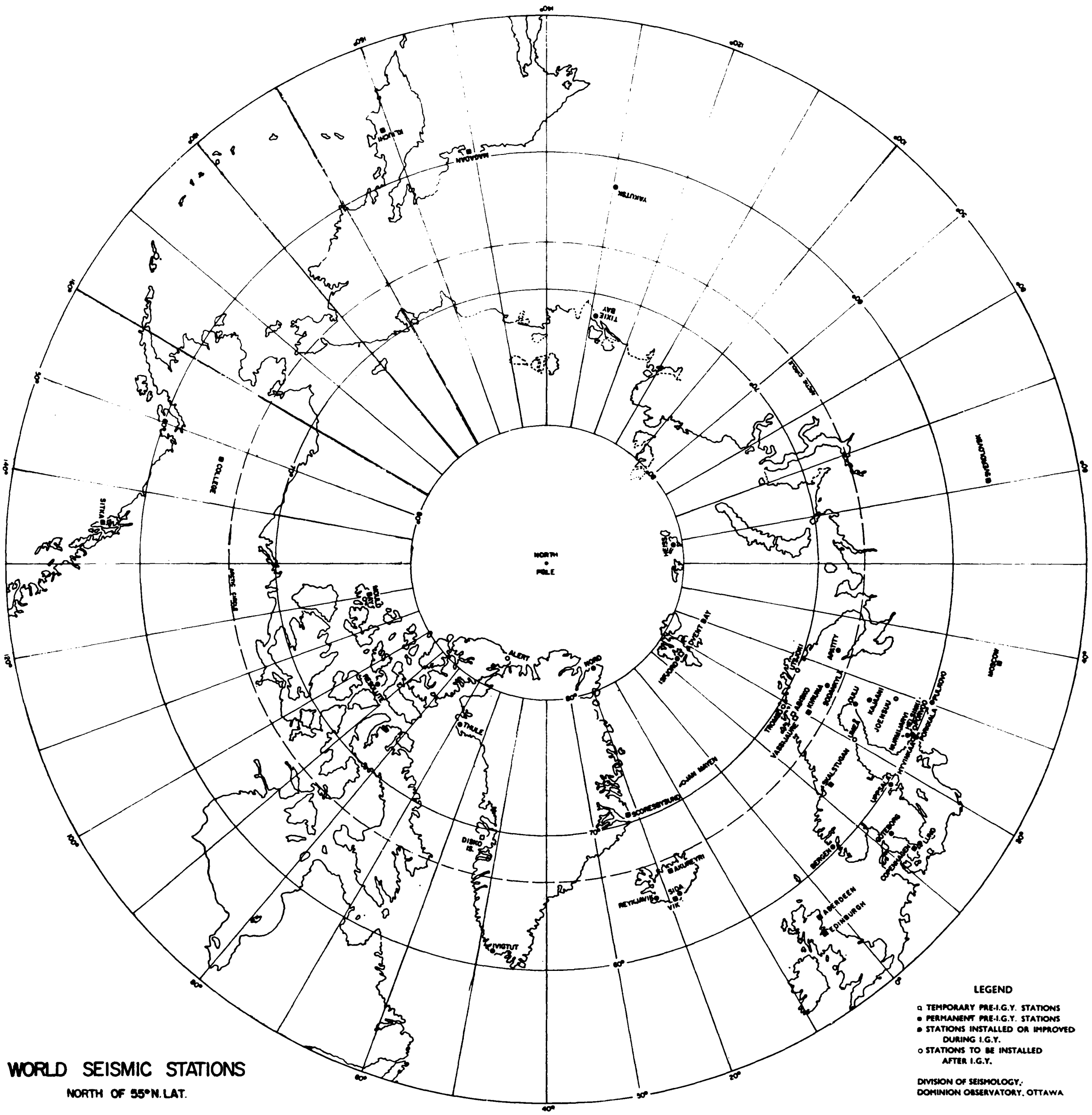

FIG. 1.

of IGC. Iceland improved the instrumentation of its Reykjavik station and installed a new station at Sida. Norway completed a new station at Isfjord, on Spitzbergen, improved the instrumentation of its central station at Bergen and was well advanced with the station at Tromso. Sweden completed stations

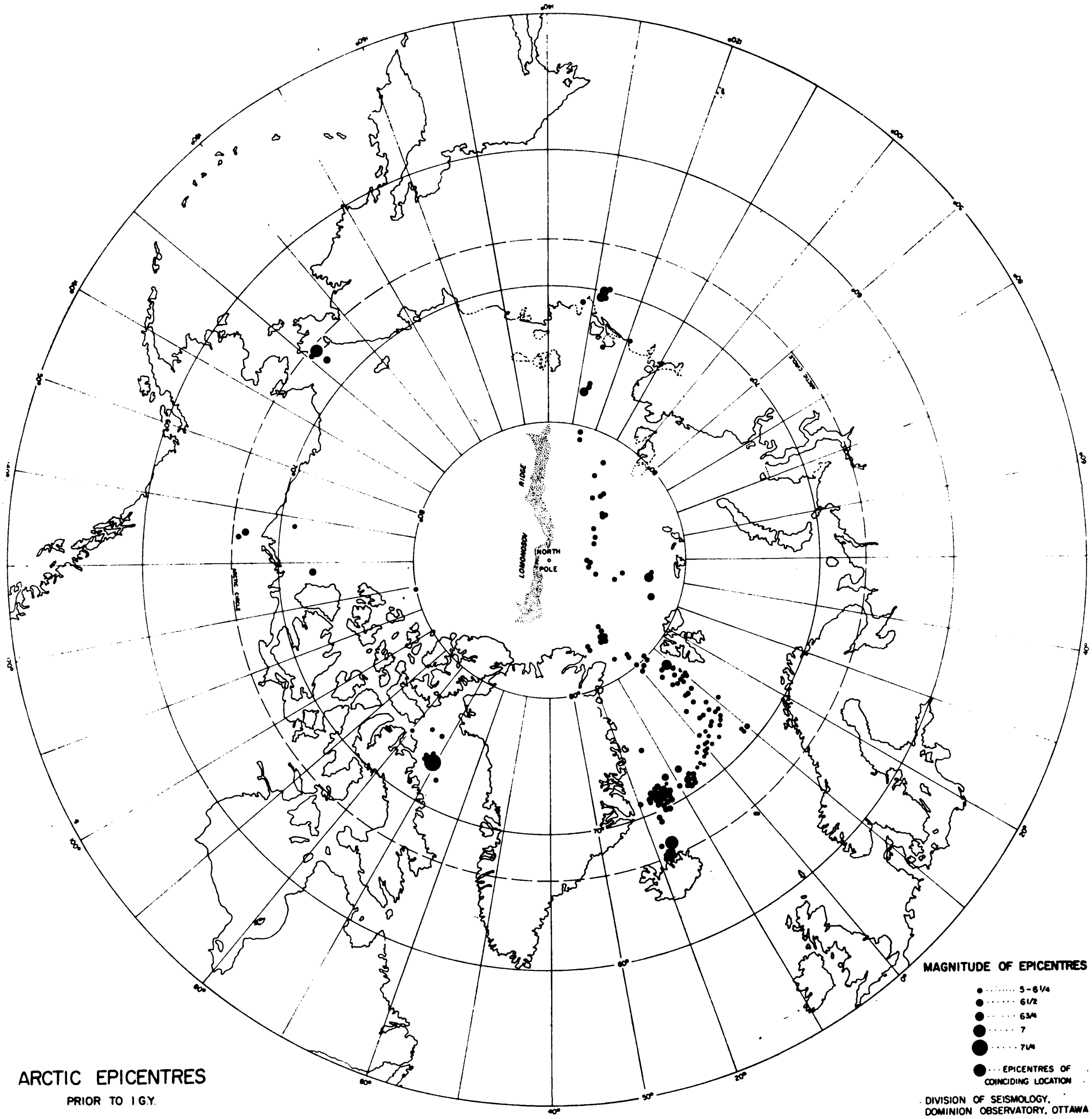

FIG. 2.

at Göteborg and Lund, and improved the instrumentation of its central station at Uppsala. The Soviet Union installed new stations at Apatity, Heiss, Tixie Bay and Yakutsk. The United States began operation of a new station at Thule, in Greenland. All these stations are shown in Fig. 1.

The standard of instrumentation in the new and re-instrumented stations was up to a level commensurate with the rest of the world. Three of the four Soviet stations had high-magnification instruments, with magnifications comparable to the Willmore or Benioff instruments which were installed almost universally in the other stations mentioned. In addition the Soviet stations were all equipped with the standard broad-band Kirnos seismographs. There were no instruments with ultra-long periods at the Soviet stations, but sets of these were operated at Resolute and at Uppsala in cooperation with the United States Committee for the IGY.

One of the most impressive results of all these new stations has been the increased numbers of epicentres which have been located in the Arctic. Table 3 lists all those epicentres for which it has been possible to give locations. In this table the accuracy of epicentral location has been indicated as A, B, C etc. in order of accuracy. For each of the earthquakes listed, a single epicentre has been given. In many instances the epicentres were determined independently by as many as five stations or organizations; we have contented ourselves with giving a single average epicentre. The duplication of effort does allow us to obtain some understanding of the accuracy of the entries. In almost all cases the epicentres as determined by the different groups agreed in latitude within 1° and in most instances the same order of agreement obtained in longitude. On a few occasions however there were very serious disagreements in longitude: in several instances by as much as 4° and in one instance by 7°. While a degree of longitude in these northern latitudes is not large, an error of 7° is quite serious, and illustrates the effect of having stations all exterior to the region. The fact that some of the earthquakes were too small to be recorded on all sides of the region also reduces the precision.

Earthquakes for which a magnitude has been determined are plotted in Fig. 3; we have limited the plotting in this way in order not to complicate the diagram with very small earthquakes, and with the thought that, where a magnitude could not be determined, the epicentral location was probably not reliable.

In addition to the earthquakes for which epicentres have been determined, most of the Arctic stations recorded a surprising number of small earthquakes At Resolute, for example, very few local earthquakes had been recorded before the sensitive short-period instruments were installed late in 1957, and it was thought that the area was relatively non-seismic. During 1958, the first complete year of operation of the Willmore instruments, a very large number of "locals" were recorded. W. E. T. SMITH, who is accustomed to reading the local earthquakes on the other Canadian records, discarded an appreciable number of these as not being certainly of seismic origin, but there remained 83 which he did not feel able to discard. Similarly in 1959 there were 64 disturbances which passed his critical examination. None of these earthquakes recorded at the nearest existing stations—Nord, Thule or College. There was no known source of blasts in the area but in order to eliminate this possibility the numbers of

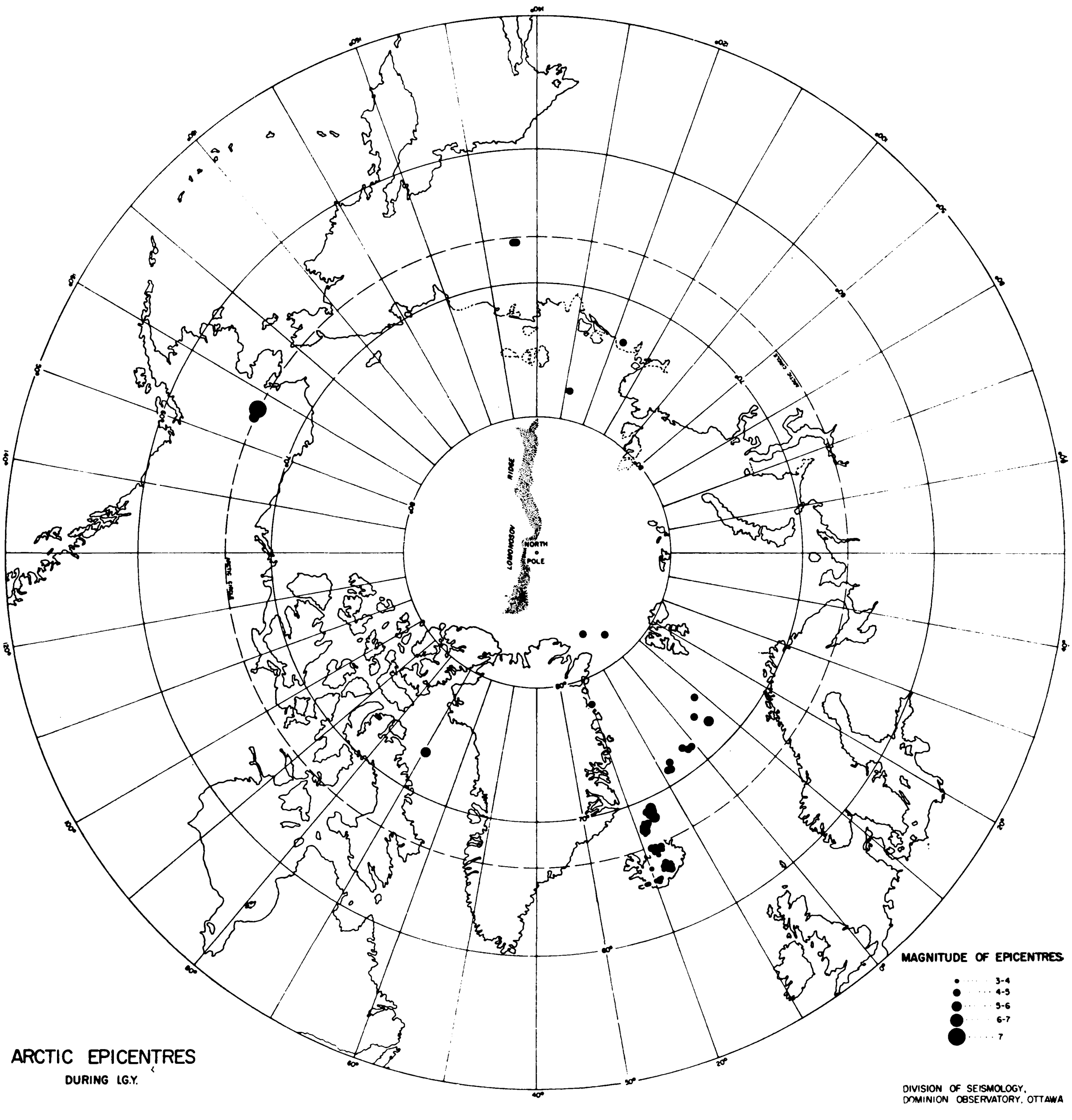

FIG. 3

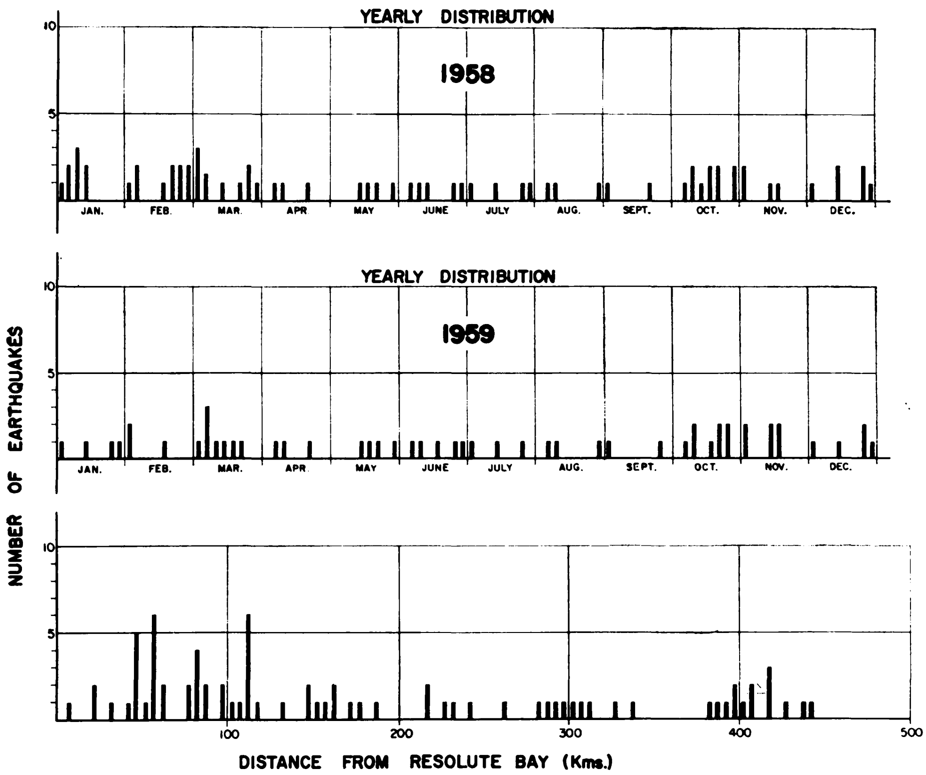

FIG. 4.

disturbances from different distant ranges were plotted as a function of distance (see the bottom diagram of Fig. 4). There is not a sufficient concentration at any particular distance to suggest a single source. There are more disturbances at the shorter distances, which only indicates that the smallest disturbances do not propagate very far. It was also thought that the disturbances might be associated in some way with ice movements. A plot of numbers of disturbances against calendar date (see upper part of Fig. 4) does not suggest any definite seasonal effect.

Resolute was not the only station which had large numbers of local disturbances. The following list gives numbers for various other stations for the year 1958. Complete tables giving the time of the earthquakes for each station have been filed with the World Data Centres.

Sodankyla	416
Nord	274
Tixie Bay	48
Apatity	60

Magadan	10
Isfjord	10 (in three months).

It is probable that some of the disturbances listed for Sodankyla are the *P* waves of distant earthquakes, not recorded at other, less sensitive stations. It is also possible that other disturbances may be blasts; nevertheless a large number of "local" earthquakes remain.

3.4 Discussion

The map of Fig. 2 has been deliberately limited to earthquakes of magnitude 5 and greater with the thought of confining this discussion to the major seismicity. A very much more complete plot has been given by LINDEN[35]. This has been reproduced by SAVARENSKY[54] and by HOPE[27]. The limited plotting of Fig. 2 shows that there is one main seismic zone within the Arctic. It begins at Iceland and follows a gently curved line, through Spitzbergen and Franz Josef land, passes north of Severnayla Zemlya, and reaches the Siberian mainland about at the mouth of the Lena River. Secondary, less well defined, zones exist in Baffin Bay, near the mouth of the Mackenzie River and to the north of the Bering Strait between Alaska and Siberia. Epicentres determined during the IGY confirm this picture except that activity in the main seismic zone has been limited to the two ends, particularly the western end.

It has been known for a very long time that the mid-Atlantic ridge has, associated with it, a narrow band of earthquake epicentres. Since the ridge, and its associated epicentres, continues up to Iceland, and since the epicentres continue beyond Iceland as shown in Fig. 2, it was initially assumed that the ridge also continued. This assumption was not supported by bathymetric findings. From the earliest work it was apparent that the ridge did not continue; the major structural feature in the Arctic is the Lomonosov ridge, which is aseismic and far displaced from the earthquake zone. There has been a great deal of discussion of this fact, and of the tectonics of the Arctic in general, particularly in the Soviet literature. It is beyond the scope of this report to review this literature and the reader is referred to the very comprehensive review and appraisal given by HOPE[27] as background material for his paper on the great Arctic magnetic anomaly.

Very recently a varient of the original theory has been put forward by EWING and HEEZEN (HEEZEN[25]). He has discovered from examination of bathymetric data obtained by the Woods Hole group that there is a deep canyon, or rift, down the middle of the Atlantic ridge. He postulates that the earthquakes are associated with this rift, not with the ridge in general, and he adopts the assumption that linear arrangements of earthquakes are diagnostic of rifts. He thus supposes that the main Arctic seismic zone is a continuation of the mid-Atlantic rift, if not of the mid-Atlantic ridge. He reports some confirmation of this from soundings made by the nuclear submarines *Nautilus* and *Skate* "which revealed a rugged formation along the epicentral belt . . . the profiles even showed a central valley to the ridge." Soviet oceanographic vessels also report[18] a trough lying in the seismic zone, further evidence in support of the Heezen-Ewing rift.

The other important matter is the minor seismicity suggested by the IGY data from all the Arctic stations. Until more stations exist within the Arctic, so that some at least of these minor earthquakes can be located, it is impossible to draw any conclusions about them, but we must at least be on guard against premature conclusions about seismic zones. At the present time there appears to be a good deal of seismic activity local to every existing Arctic station.

3.5 Prospects for the future

We have indicated in Table 1 those stations north of latitude 50° N which had not come into operation at the close of the IGY, but which were in a sufficiently advanced state of preparation for their future existence to be regarded as certain. Two of these stations, Alert and Mould Bay, facing on the Arctic Ocean, should be useful in locating the small earthquakes which have been recorded at Resolute. They also should prove of great value in the study of surface waves crossing the Arctic ocean as well as those traversing the length of the north American continent. Canada will also install additional stations to the south of Resolute, but it is yet too early to announce the exact position of these stations.

Further improvement in the knowledge of Arctic seismicity is to be expected because of the steady improvement in the instruments being employed in stations to the south. It is reasonable to expect that all earthquakes of tectonic significance will be recognized and located.

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TABLE 1. List of stations

Station name	Coordinates, elevation and foundation material	Instrument and component		Dates of operation	T_o secs	T_g secs	V_{max}	Paper speed mm/min
				Canada				
Alert	82° 30′ N	Willmore	NS	The station should be in-	1·0	0·3	200,000	60
	62° 20′ W		EW	stalled in the late summer	1·0	0·3	200,000	60
			Z	of 1961.	1·0	0·3	200,000	60
		Columbia	NS		20	90	600	15
			EW		20	90	600	15
			Z		20	90	600	15
Mould Bay	76° 14′ N	Willmore	NS	The station should be in-	1·0	0·3	200,000	60
	119° 20′ W		EW	stalled in the late summer	1·0	0·3	200,000	60
			Z	of 1961.	1·0	0·3	200,000	60
		Columbia	NS		20	90	600	15
			EW		20	90	600	15
			Z		20	90	600	15
Resolute	74° 41·2′ N	Sprengnether	NS	1951–1958	8	8	6,000	30
	94° 54·0′ W		EW	1951–1958	16	22	1,100	30
	h = 15 m.		Z	1951–1958	1·5	1·5	20,000	60
	Permanently	Columbia	Z	1954–1958	16	10	6,000	30
	frozen	Willmore	NS	Since 1957	1·0	0·25	200,000	60
	Palaeozoic		EW	Since 1957	1·0	0·25	200,000	60
	limestone		Z	Since 1957	1·0	0·25	200,000	60
		Columbia	NS	Since 1957	15	100	600	15
			EW	Since 1957	15	90	600	15
			Z	Since 1957	15	80	600	15

TABLE 1. List of stations *(continued)*

Station name	Coordinates, elevation and foundation material	Instrument and component		Dates of operation	T_o secs	T_g secs	V_{max}	Paper speed mm/min
				Denmark				
Copenhagen	55° 41′ N	Galitzin–Wilip	NS	Since 1927	12·5	12·5	1,000	30
	12° 26′ E		EW	Since 1927	12·5	12·5	1,000	30
			Z	Since 1927	12·5	12·5	1,000	30
		Wiechert 1000 kg	NS	Since 1927				12
			EW	Since 1927				12
		1300 kg	Z	Since 1927				10
		Benioff	Z	Since 1937				60
Disko Island	69° 15′ N	Bosche 100 kg	NS	Sept., 1910–May, 1912			70	18
	53° 27′ W		EW	Sept., 1910–May, 1912			70	18
	h = 1 m.							
Ivigtut	61° 12′ N	Wiechert 100 kg	NS	1929–1953	9	—	200	12
	48° 11′ W		EW	1929–1953	9	—	200	12
	h = 20 m.	1300 kg	Z	1929–1952	4	—	200	10
		Milne–Shaw	NS	1955–1960	12	—	300	30
		Willmore	Z	1957–1958	1·0	0·25	30,000	53
Nord	81° 36′ N	Strobach	NS	Since 1957	6		500	15
	16° 41′ W		EW	Since 1957	6		500	15
		Willmore	Z	Since 1957	1·0	0·25	30,000	53
Scoresbysund	70° 29′ N	Galitzin–Wilip	NS	Since 1928	12	—	1,000	30
	21° 57′ W		EW	Since 1928	12	—	1,000	30
			Z	Since 1928	9	—	600	30
		Coulomb–Grenet	Z	Since 1950	1·4	0·25	30,000	30

TABLE 1. List of stations (*continued*)

Station name	Coordinates, elevation and foundation material	Instrument and component		Dates of operation	T_o secs	T_g secs	V_{max}	Paper speed mm/min
				Finland				
Helsinki	60° 10·5′ N	Mainka 730 kg	NS	Since 1924	12	—	250	20
	24° 57·4′ E	730 kg	EW	Since 1924	12	—	250	20
	h = 20 m.	300 kg	Z	1926–1952	3·5	—		17
		Sprengnether	NS	1957–1958	12	15		30
			Z	1950–1955	2·9	3·0		25
		Nurmia	NS	1957–1958	0·5	1·1	36,000	30
			EW	1957–1958	0·5	1·1	36,000	30
			Z	1957–1958	0·5	0·2	54,000	60
		Nurmia	NS	Since 1959	10		1,300	30
			EW	Since 1959	10		1,300	30
			Z	Since 1959	1·0			120
Hyvinkää	60·7° N							
	24·8° E	Nurmia	Z	Since 1960	1·0			60
Joensuu	62·6° N	Nurmia	Z	Since 1960	1·0			60
	29·7° E							
Kajaani	64·1° N	Nurmia	NS	Since 1960	1·0			60
	27·7° E		EW	Since 1960	1·0			60
			Z	Since 1959	1·0			60
Nurmijärvi	60° 30·5′ N	Nurmia	NS	Since 1958	0·5	1·1		30
	24° 39·3′ E		EW	Since 1959	0·5	1·1		30
	h = 102 m.		Z	Since 1958	1·0	0·2		60

TABLE 1. List of stations (*continued*)

Station name	Coordinates, elevation and foundation material	Instrument and component		Dates of operation	T_o secs	T_g secs	V_{max}	Paper speed mm/min
Finland (continued)								
Oulu	65·0° N 25·5° E	Nurmia	Z	Since 1959	1·0			60
Porkkala	60·0° N 24·5° E	Nurmia		Since 1959	1·0			60
Porvoo	60·3° N 25·9° E	Nurmia						60
Sodankylä	67° 22·3′ N	Sprengnether	NS	1956–1957	15	15		30
	26° 37·7′ E	Nurmia	NS	Since 1957	0·5	1·1		30
	h = 181 m.		EW	Since 1957	0·5	1·1		30
		Benioff	Z	Since 1956	1·0	0·25		60
		Nurmia	NS	Since 1960	1·0			30
			EW	Since 1960	1·0			30
Utsjoki	69·8° N 27·0° E	Nurmia	Z	Since 1960	1·0			60

TABLE 1. List of stations *(continued)*

Station name	Coordinates, elevation and foundation material	Instrument and component	Dates of operation	T_o secs	T_g secs	V_{max}	Paper speed mm/min
			Germany				
Advent Bay	78° 13′ N 15° 36′ E	Mainka	7 months, 1911–1912				
			Iceland				
Akureyri	65° 40·3′ N 18° 06·0′ W h = 50 m.	Mainka 135 kg NS	Since July, 1954	3·5	—	100	27
Reykjavik	64° 08·9′ N 21° 57·2′ W h = 20 m.	Mainka 135 kg NS EW	1909–1914, 1925–1946 1911–1914, 1926–1946	(6) (6)	— —	(70) (70)	17 27
	64° 08·3′ N 21° 54·4′ W h = 44 m.	Mainka 135 kg NS EW Sprengnether NS EW Z Willmore Z	1946–1952 1952–1959 Since 1952 Since 1951 Since 1959	(6) (6) 1·5 1·5 1·5 1·0	— — 1·6 1·6 1·5 1·6	(70) (70) (2,500) (2,500) (4,000) —	17 27 30 30 30 30
Sida	63° 47·1′ N 18° 03·5′ W h = 26 m.	Willmore Z	Since 1958	1·0	0·25	10,000	30
Vik	63° 25·3′ N 19° 01·0′ W h = 19 m.	Mainka 135 kg NS	Since 1955	4·5	—	60	17

TABLE 1. List of stations *(continued)*

Station name	Coordinates, elevation and foundation material	Instrument and component		Dates of operation	T_o secs	T_g secs	V_{max}	Paper speed mm/min
				Norway				
Bergen	60° 23·3′ N	Bosch	NS	Since 1904				
	5° 18·3′ E		EW					
	h = 20 m.	Wiechert	NS	Since 1923	10	—	200	15
			EW	Since 1923	10	—	200	15
			Z	Since 1921	4	—	400	15
		Electromagnetic	NS	Since 1959	1·0	0·7		60
			EW	Since 1961	1·0	0·7		60
			Z	To be installed in 1962	1·0	0·7		60
Isfjord	78° 03·5′ N							
	13° 38·4′ E	Willmore	Z	Since Aug. 1, 1958	1·0	0·25	60,000	60
	h = 5 m.							
Jan Mayen	Approxi-							
	mately	Electromagnetic		To be installed in 1961				
	71° N, 8° W							
Tromsö	69° 40′ N	Benioff	NS	September, 1960	1·0	0·25		15
	19° E		EW	September, 1960	1·0	0·25		15
			Z	September, 1960	1·0	0·25		15

TABLE 1. List of stations (*continued*)

Station name	Coordinates, elevation and foundation material	Instrument and component		Dates of operation	T_o secs	T_g secs	V_{max}	Paper speed mm/min
Sweden								
Abisko	68° 20·5′ N	Wiechert (80kg)	NS	1915–1951	—	—	—	7
	18° 49·0′ E	Galitzin	NS	1919–1943	~12	~12	—	30
	h = 385 m.		EW	1919–1943	~12	~12	—	30
	Morainic material		Z	1919–1943	~12	~12	—	30
	on schist							
Göteborg	57° 41·9′ N	Grenet	Z	Since June, 1958	1·4	0·5	10,530	60
	11° 58·7′ E							
	h = 66 m.							
	Gneiss							
Kiruna	67° 50·4′ N	Galitzin	NS	Since July, 1951	12·8	11·9	910	30
	20° 25·0′ E		EW	Since July, 1951	11·8	11·8	780	30
	h = 390 m.		Z	Since July, 1951	9·6	11·6	740	30
	Porphyry	Grenet	Z	Since July, 1951	1·4	0·7	11,150	60
Lund	55° 41·9′ N	Wiechert (1000kg)	NS	1917–1953	—	—	—	15
	13° 11·2′ E		EW	1917–1953	—	—	—	15
	h = 32 m.	Grenet	Z	June to Oct., 1957	1·4	0·5	—	60
	Clay and sand							
Skalstugan	63° 34·8′ N	Grenet	Z	Since Jan., 1956	1·4	0·8	12,000	60
	12° 16·8′ E							
	h = 580 m.							
	Gneiss							

TABLE 1. List of stations *(continued)*

Station name	Coordinates, elevation and foundation material	Instrument and component		Dates of operation	T_o secs	T_g secs	V_{max}	Paper speed mm/min
				Sweden (continued)				
Umeå	63° 49·0′ N 20° 14·1′ E h = 20 m. Gneiss and pegmatite	Grenet	Z	Since Oct., 1960	1·4	0·7	--	60
Uppsala	59° 51·5′ N 17° 37·6′ E h = 14 m. Granite	Weichert (1000kg)	NS	Since Oct., 1904	9·5	—	257	15
			EW	Since Oct., 1904	10·6	—	244	15
		Benioff	NS	Since Jan., 1955	1·0	0·7	70,000	60
			EW	Since Jan., 1955	1·0	0·7	60,000	60
			Z	Since Jan., 1955	1·0	0·7	40,000	60
		Benioff	NS	Since Jan., 1955	1·0	87	2,600	30
			EW	Since Jan., 1955	1·0	74	2,810	30
			Z	Since Jan., 1955	1·0	76	1,660	30
		Press–Ewing	NS	Since Dec., 1957	15·0	81	2,200	15
			EW	Since Dec., 1957	15·0	87	2,200	15
		Sprengnether	Z	Since Dec., 1957	15·0	85	2,200	15
		Grenet	Z	1951–1957	1·4	0·5	10,530	60
Vassijaure	68° 25′ N 18° 11′ E h = 469 m. Porphyritic granite	Wiechert (80kg)	NS	1906–1915	—	—	—	7

TABLE 1. List of stations *(continued)*

Station name	Coordinates, elevation and foundation material	Instrument and component		Dates of operation	T_o secs	T_g secs	V_{max}	Paper speed mm/min
				Union of Soviet Socialist Republics				
Apatity	67° 35′ N	Kirnos	NS	June 30, 1956–Dec. 28, 1957	12·5	1·1	1,400	30
	33° 18′ E		EW	June 30, 1956–Dec. 28, 1957	12·5	1·1	1,400	30
	before June,							
	1957	Kirnos	NS	Dec. 28, 1957 to present	12·5	1·1	1,800	30
	67° 33′ N		N60°W	Dec. 28, 1957 to present	12·5	1·1	1,600	30
	33° 26′ E		N60°E	Dec. 28, 1957 to present	12·5	1·1	1,600	30
	after June,							
	1957	Kirnos	Z	June 30, 1956 to present	12·0	1·1	1,000	30
	h = 182 m.	Kharin	NS	June 30, 1956 to present	0·55	1·0	30,000	60
	Meta-gabro-		EW	June 30, 1956 to present	0·55	1·0	30,000	60
	diabase		Z		0·55	1·0	37,000	60
Heiss	80° 37′ N	Vegik	Z	Dec. 30, 1957–Feb. 12, 1958	1·2	1·2	21,000	30
	58° 03′ E		Z	Feb. 21, 1958–Apr. 13, 1958	1·7	0·4	40,000	30
	h = 30 m.		Z	Apr. 13, 1958–Apr. 16, 1959	1·2	0·4	30,000	30
	Frozen		Z	Apr. 16, 1959–Sept. 1, 1959	1·2	0·4	35,000	30
	gravel		Z	Sept. 1, 1959 to present	0·9	0·4	33,000	30
		Kirnos	NS	Dec. 30, 1957 to present	12·5	1·2	650	30
			EW	Dec. 30, 1957 to present	12·5	1·2	650	30
			Z	Dec. 30, 1957 to present	12·5	1·2	750	30
Kliuchi	56° 19′ N	Kirnos	NS	Since 1950	12·5	1·2	300	30
	160° 52′ E		EW	Since 1950	12·5	1·2	300	30
	h = 44 m.		Z	Since 1950	12·5	1·2	300	30

TABLE 1. List of stations *(continued)*

Station name	Coordinates, elevation and foundation material	Instrument and component	Dates of operation	T_o secs	T_g secs	V_{max}	Paper speed mm/min
Union of Soviet Socialist Republics (continued)							
Magadan	59° 33′ N	Kirnos NS	Since 1953	12·5	1·1	1,000	30
	150° 48′ E	EW	Since 1953	12·5	1·2	1,100	30
	h = 78 m.	Z	Since 1953	11·0	1·1	1,000	30
	Clay						
Moscow	55° 44′ N	Galitzin NS	Since 1935	9·5	9·5	595	30
	37° 38′ E	EW	Since 1935	9·5	9·5	540	30
	h = 124 m.	Z	Since 1935	9·5	9·5	650	30
	Sand	Kirnos N45°E	Since 1952	12·5	1·3	450	30
		N45°W	Since 1952	12·5	1·3	450	30
		Z	Since 1952	12·5	1·3	475	30
		Kirnos Z	July 1958–Oct. 1959	20	23	1,100	30
		Z	Oct. 1959 to present	30	23	2,000	30
		Kharin Z	Oct. 1958 to present	1·9	3·8	15,000	30
Pulkovo	59° 46′ N	Galitzin NS	Sept. 5, 1956 to present	9·7	9·7	750	30
	30° 19′ E	EW	Sept. 5, 1956 to present	9·5	9·7	750	30
	h = 65 m.	Z	Sept. 5, 1956 to present	9·5	9·4	730	30
	Clay	Kirnos N45°E	Apr. 1, 1959–July 25, 1959	25·0	25·0	2,000	15
		N45°W	Apr. 1, 1959–July 25, 1959	25·0	25·0	2,000	15

TABLE 1. List of stations *(continued)*

Station name	Coordinates, elevation and foundation material	Instrument and component		Dates of operation	T_o secs	T_g secs	V_{max}	Paper speed mm/min
Union of Soviet Socialist Republics (continued)								
		Kirnos	NS	Apr. 1, 1957–Apr. 1, 1959 and July 25, 1959 to present	25·0	25·0	2,000	15
			EW	As for NS	25·0	25·0	2,000	15
			Z	April 1, 1957 to present	25·0	25·0	2,000	15
		Kharin	NS	April 1, 1957 to present	2·0	4·0	4,000	30
			EW	April 1, 1957 to present	2·0	4·0	4,000	30
Sverdlovsk	56° 50′ N	Galitzin	NS	Since 1913	25	25	1,140	30
	60° 38′ E		EW	Since 1913	25	25	1,140	30
	h = 275 m.		Z	Since 1913	12	12	1,760	30
	Serpentine	Kharin	EW	Since 1950	1·7	1·5	29,000	30
Tixie Bay	71° 38′ N	Kirnos	NS	March 15, 1956 to present	12·5	1·2	1,450	30
	128° 52′ E		EW	March 15, 1956 to present	12·5	1·2	1,400	30
	h = 30 m.		Z	March 15, 1956–Apr. 15, 1958	12·5	1·2	1,400	30
	Permafrost		Z	April 15, 1958 to present	2·5	1·2	24,000	60
		Kharin	NS	May 15, 1957–Apr. 19, 1958	2·0	0·4	10,000	60
			NS	April 19, 1958 to present	2·0	0·4	13,000	60
Yakutsk	62° 01′ N	Kirnos	NS	March 1, 1958 to present	12·5	1·1	1,300	30
	129° 43′ E		EW	March 1, 1958 to present	12·5	1·1	1,100	30
	h = 495 m.		Z	March 1, 1958 to present	12·5	1·0	660	30
	Permafrost							

TABLE 1. List of stations *(continued)*

Station name	Coordinates, elevation and foundation material	Instrument and component		Dates of operation	T_o secs	T_g secs	V_{max}	Paper speed mm/min
United States of America								
College	64° 51.6′ N	Benioff	NS	Jan. 1948 to present	1·5	0·45	12,000	15 (film)
	147° 50·2′ W		EW	Jan. 1948 to present	1·5	0·45	12,000	15 (film)
	h = 159 m.		Z	Jan. 1948 to Sept. 1956	1·5	0·45	12,000	15 (film)
		Wilson–Lamison	Z	Sept. 1956 to present	1·5	0·45	12,000	15 (film)
		Wenner	NS	Jan. 1948 to present	9·3	16	1,000	15
	Outpost station							
	64° 53·1′ N	Benioff	Z	Feb. 1950 to present	1·5	0·53	425,000	60
	147° 48·1′ W							
Sitka	57° 03·4′ N	Wenner	NS	1932 to present	7·5	16·5	1,000	15
	135° 19·5′ W	Wenner	EW	1932 to present	7·5	13·2	1,000	15
	h = 19 m.	Wilson--Lamison	Z	June 1956 to present	1·12	1·5	24,000	60
	graywacke							
Thule	Before July 29, 1960							
	76° 33·4′ N	Wilson–Lamison	Z	Aug. 1957 to present	1·3	1·55	24,000	60 (since April, 1958)
	68° 40·5′ W	Sprengnether	NS	Aug. 1957 to Jan. 1959	6·0	7·32	Unknown	
	After July 29, 1960		EW	Aug. 1957 to Jan. 1959	6·0	7·6	Unknown	
	76° 25′ N							
	68° 18′ W							

TABLE 2. Arctic epicentres prior to I. G. Y.

Date			Time	Location				M
1908	Oct.	14	14 56.3	82	N	30	E	6.6
1909	April	10	18 46.9	77.5	N	128	E	$6\frac{3}{4}$
1910	Jan.	22	08 48.5	67.5	N	17	W	7.1
1912	April	13	02 39.7	80	N	100	E	
1912	May	6	18 59.9	64	N	20	W	7.0
1913	July	26	20 51.1	67.5	N	18.5	W	$5\frac{1}{4}$
1914	June	7	16 24	73	N	119	E	$5\frac{1}{4}$
1915	June	1	14 43.9	78.5	N	8	E	$6\frac{3}{4}$
1916	Dec.	6	22 17.2	87	N	48	E	$5\frac{3}{4}$
1917	May	14		72	N	3	W	
	July	9	00 21 35	64	N	20	W	
	Aug.	21	10 44 10	72	N	3	W	
1918	Nov.	30	06 48 40	71	N	132	E	$6\frac{1}{4}$
1919	Feb.	2	20 02 54	72.3	N	9.5	W	
	Feb.	15	02 17.3	68	N	13	W	
	Setp.	12	14 26.6	72	N	3	W	
1920	Nov.	16	08 30 57	72.5	N	128	W	$6\frac{1}{2}$
1921	July	20	05 25 40	72	N	3.0	E	
	Aug.	23	20 17 28	67	N	18	W	$6\frac{1}{4}$
1922	April	8	20 42 21	72	N	8.5	W	$6\frac{1}{4}$
1923	May	30	08 30 40	77	N	127	E	6
	May	30	17 56 42	77	N	127	E	6
	Oct.	10	07 11 18	72	N	12	W	$6\frac{1}{2}$
1924	March	12	13 52 48	73	N	2.5	E	$5\frac{3}{4}$
	July	19	02 50 09	73.5	N	4	E	
	July	25	19 36 22	72.5	N	16	E	
	Oct.	10	09 21 17	71	N	16	W	
1925	Nov.	28	08 14 53	69	N	18	W	
1926	April	9	10 04 50	74	N	125	E	$5\frac{3}{4}$
	April	24	08 56 26	82	N	3	E	
	July	14	22 22 25	66	N	163	W	
	Aug.	6	05 23 58	89	N	85	E	6
	Dec.	25	05 13 20	75	N	5	E	
1927	Jan.	7	10 43 12	80	N	117	E	
	July	16	01 27 10	71	N	14	W	$5\frac{1}{4}$
	July	16	01 35 03	71	N	17	W	$5\frac{1}{2}$
	July	16	02 16 03	71	N	17	W	$5\frac{1}{4}$
	Aug.	7	23 57 05	74	N	4	E	$5\frac{1}{4}$
	Aug.	8	00 25 28	75	N	2	E	$5\frac{1}{4}$
	Aug.	8	03 44 14	74	N	5	E	$5\frac{1}{4}$
	Sept.	6	07 16 09	77	N	10	E	$5\frac{1}{4}$
	Oct.	30	03 09 04	71.5	N	14	W	$5\frac{1}{4}$
	Nov.	14	00 12 05	70.5	N	128	E	$6\frac{1}{2}$
	Nov.	14	04 56 29	70	N	128	E	$6\frac{3}{4}$
	Nov.	15	21 48 56	70.5	N	128	E	6
1928	Feb.	3	13 47 35	70.5	N	128	E	$6\frac{1}{4}$

TABLE 2 *(continued)*

Date		Time	Location	M
1928	Feb. 21	19 49 04	67 N 172 W	6.9
	Feb. 24	14 10 23	67 N 171 W	$6\frac{1}{4}$
	Feb. 26	01 19 10	68 N 172 W	$6\frac{1}{2}$
	May 1	18 54 41	67 N 172 W	$6\frac{1}{4}$
	Aug. 16	07 36 37	70 N 127 E	$5\frac{1}{2}$
1929	June 10	23 03 14	71.0 N 10 E	$6\frac{1}{4}$
	June 27	22 39 07	71.0 N 6.0 W	$5\frac{1}{2}$
	Aug. 6	01 30 13	72 N 8 W	$5\frac{1}{2}$
	Aug. 16	23 29 02	80.5 N 5 E	
1930	March 15	09 13 30	78.5 N 4 E	
	Oct. 11	03 06 22	71 N 13 W	6
1931	June 20	15 05 15	86.2 N 79 E	6
1932	June 20	15 39 16	74 N 4 E	
	Nov. 29	08 34 38	71 N 8 W	
1933	April 25	22 37 54	71 N 19 W	$5\frac{1}{4}$
	Nov. 20	23 21 32	73 N 70.7 W	$7\frac{1}{4}$
	Dec. 19	05 39 58	86.5 N 35 E	$5\frac{1}{2}$
	Dec. 19	17 48 20	75 N 72 W	$5\frac{3}{4}$
1934	Feb. 13	09 51 45	70.5 N 14.5 W	$5\frac{3}{4}$
	Feb. 24	00 49 03	73.5 N 71.5 W	$5\frac{1}{2}$
	May 21	10 07 19	71.7 N 1.5 W	$5\frac{3}{4}$
	June 2	13 42 40	66 N 18.5 W	$6\frac{1}{4}$
	Aug. 31	05 02 45	73 N 71 W	$6\frac{1}{2}$
1935	Jan. 26	17 41 34	85 N 35 E	$5\frac{1}{2}$
	April 18	22 15 28	70.5 N 73 W	$5\frac{1}{4}$
	May 11	19 14 58	77.2 N 4.5 E	$5\frac{1}{4}$
	Aug. 22	20 30 52	73.2 N 71.5 W	$5\frac{3}{4}$
	Aug. 25	05 07 49	78.2 N 5 E	6
	Sept. 30	19 00 42	84 N 2.5 W	6
	Oct. 15	20 41 30	80 N 5 E	$5\frac{1}{4}$
1936	Jan. 2	00 37 07	79.7 N 2 E	$5\frac{3}{4}$
	June 7	03 58 41	73 N 5 E	$5\frac{1}{2}$
	June 7	04 38 13	72.5 N 4 E	$5\frac{3}{4}$
	June 14	10 04 17	74 N 5 E	$5\frac{1}{4}$
	Oct. 22	23 49 28	66.8 N 17.4 W	$5\frac{1}{2}$
	Oct. 23	00 00 24	66.8 N 17.4 W	$5\frac{3}{4}$
	Oct. 26	23 05 40	70.5 N 8 W	$6\frac{1}{4}$
1937	July 5	01 41 04	71 N 138 W	$5\frac{1}{2}$
	Nov. 27	20 10 30	71 N 9 E	$5\frac{1}{2}$
1938	April 25	09 04 11	79 N 1.5 E	$5\frac{1}{2}$
	June 25	23 45 08	77 N 9 E	$5\frac{3}{4}$
	July 2	07 40 32	77.5 N 10 E	$5\frac{1}{4}$
	July 27	19 51 22	77 N 8 E	$5\frac{1}{2}$
	Aug. 8	13 06 56	71 N 15 W	$5\frac{1}{4}$
	Aug. 8	13 14 43	71 N 15 W	$5\frac{1}{4}$
	Aug. 8	15 35 10	71 N 15 W	$5\frac{1}{2}$

TABLE 2 *(continued)*

Date		Time	Location				M
1938	Aug. 8	16 49 41	71	N	15	W	$5\frac{1}{4}$
	Aug. 28	21 03 18	69 2	N	16.5	W	$5\frac{1}{4}$
	Dec. 6	09 13 40	71.5	N	12	W	$5\frac{1}{4}$
1939	Jan. 16	00 11 16	80	N	5	E	
1940	May 29	01 57 52	67	N	135	W	$6\frac{1}{4}$
	June 2	23 23 27	71.5	N	15	W	$5\frac{1}{4}$
	June 5	11 01 10	67.5	N	136	W	$6\frac{1}{2}$
	June 23	06 55 38	74.7	N	14	W	$5\frac{1}{2}$
1941	June 6	21 02 24	72	N	0.5	W	$5\frac{1}{2}$
	July 17	22 08 49	78.5	N	8	E	$5\frac{1}{2}$
	Sept. 7	00 50 51	71.2	N	2.5	W	$5\frac{1}{2}$
1942	May 31	02 42 33	81	N	0		$5\frac{1}{4}$
	June 1	12 13 15	73	N	1	E	$5\frac{1}{4}$
1943	Nov. 5	10 32 17	72	N	0		$5\frac{1}{2}$
	Nov. 8	06 59 19	80	N	5	E	$5\frac{3}{4}$
1944	Feb. 4	23 45 13	71.5	N	3	W	$5\frac{1}{4}$
	May 21	00 15 39	71	N	10	W	$5\frac{1}{2}$
	Aug. 24	15 58 55	78	N	9	E	$5\frac{3}{4}$
1945	Jan. 1	01 20 42	73	N	70	W	$6\frac{1}{2}$
	Oct. 15	18 24 42	72.5	N	2.5	E	$5\frac{1}{2}$
	Nov. 8	09 05 23	83	N	15	W	6
	Nov. 8	10 02 37	83	N	15	W	6
1946	April 9	20 40 07	74	N	11	E	$5\frac{1}{4}$
	May 11	16 25 30	66	N	0.5	W	$5\frac{1}{4}$
	May 11	18 39 27	66	N	0.5	W	$5\frac{1}{2}$
1947	July 10	10 48 45	73	N	71	W	6
	Dec. 9	09 45 47	83	N	5	W	$5\frac{1}{4}$
1948	Feb 18	20 29 47	82.5	N	40.0	E	$6\frac{3}{4}$
	May 20	07 13 46	71	N	7	W	$5\frac{1}{4}$
	May 31	14 56 44	84.5	N	105	E	$5\frac{1}{4}$
	July 8	12 34 36	71.5	N	4.0	W	6
	Aug. 23	11 50 39	72	N	4	E	$5\frac{1}{4}$
	Aug. 30	01 38 35	66.5	N	18	W	$5\frac{1}{4}$
	Sept. 26	05 51 12	84.5	N	40	E	$5\frac{1}{4}$
	Nov. 22	23 32 48	82.5	N	42	E	$5\frac{1}{2}$
1949	March 2	06 54 30	72.2	N	2.2	W	$5\frac{1}{2}$
	April 14	23 28 40	72.2	N	0.2	E	5
	May 13	07 15 46	69.5	N	16.5	W	$4\frac{3}{4}$
	July 8	18 18 04	72.2	N	0.5	E	$5\frac{1}{2}$
1950	June 5	11 16 10	87.2	N	50.0	E	$5\frac{1}{2}$
	June 20	14 11 42	74.5	N	8.0	E	$5\frac{1}{4}$
	Sept. 13	11 59 35	77	N	8	E	5
1951	Jan. 9	16 00 24	81	N	125	E	$5\frac{1}{2}$
	April 22	12 36 16	75	N	75	W	$5\frac{1}{4}$
	April 29	07 35 46	80.5	N	126	E	$5\frac{1}{2}$
	June 6	16 10 49	71	N	8	W	$6\frac{1}{2}$

TABLE 2 *(continued)*

Date		Time	Location				M
1951	Oct. 16	06 54 30	76	N	6	E	$5\frac{1}{2}$
1952	Feb. 10	06 10 03	72.7	N	3.0	E	$5\frac{1}{2}$
	March 8	11 36 53	71.0	N	14.2	W	$5\frac{1}{4}$
	March 9	05 44 27	70.9	N	14.5	W	$5\frac{1}{4}$
	April 28	01 15 15	74	N	7	E	5
	Dec. 10	05 58 05	70.8	N	6.8	W	$6\frac{1}{4}$
	Dec. 10	12 47 44	84	N	100	E	$5\frac{1}{4}$
	Dec. 10	14 06 40	84	N	100	E	$5\frac{1}{2}$
1953	Feb. 22	02 19 00	87	N	40	E	$5\frac{1}{4}$
	Feb. 28	05 50 48	72.5	N	1	W	5
	April 24	02 09 41	77	N	8	E	$5\frac{1}{2}$
	Nov. 4	01 35 57	71.2	N	9	E	5
	Nov. 8	06 37 05	68.8	N	15.5	W	$4\frac{1}{2}$
	Nov. 8	07 18 10	68.8	N	15.5	W	$4\frac{1}{2}$
1954	Jan. 6	15 53 59	76	N	6	E	$5\frac{1}{2}$
	Feb. 1	14 00 44	83.0	N	5	W	5
	Feb. 1	15 46 00	83.0	N	5	W	$5\frac{1}{2}$
	Feb. 2	17 45 44	83.0	N	5	W	$5\frac{3}{4}$
	Feb. 14	22 56 21	71.1	N	6.1	W	5
	May 4	17 51 22	74	N	80	W	5
	May 9	14 14 32	71	N	13	W	$5\frac{1}{2}$
	May 11	04 26 45	81.5	N	6	W	$5\frac{1}{4}$
	May 27	24 27 30	69	N	18	W	$4\frac{1}{2}$
	June 10	23 33 57	71.0	N	7.7	W	5
	June 25	05 20 11	73.5	N	8	E	5
	Aug. 20	19 21 33	70.5	N	15.0	W	5
	Aug. 20	20 24 13	70.5	N	14.7	W	$4\frac{3}{4}$
	Aug. 20	22 59 16	70.5	N	15.0	W	$5\frac{1}{4}$
	Aug. 21	00 25 35	70.7	N	14.7	W	$5\frac{1}{2}$
	Aug. 21	00 26 06	70.8	N	15.2	W	$5\frac{1}{2}$
	Aug 21	04 13 14	70.7	N	15.3	W	5
	Aug. 21	07 19 48	70.7	N	15.3	W	$5\frac{1}{2}$
	Aug. 21	13 05 05	70.8	N	14.9	W	5
	Aug 21	17 40 06	70.8	N	14.8	W	5
	Aug. 21	22 50 54	70.8	N	14.9	W	$4\frac{3}{4}$
	Aug. 21	22 51 04	70.8	N	14.9	W	$5\frac{1}{2}$
	Aug. 22	02 51 42	70.9	N	14.7	W	$4\frac{3}{4}$
	Aug. 22	10 08 02	70.7	N	15.3	W	$5\frac{1}{4}$
	Aug. 22	12 39 39	70.6	N	14.9	W	5
	Aug. 22	18 21 14	70.7	N	15.3	W	$4\frac{3}{4}$
	Aug. 22	23 52 09	70.6	N	15.7	W	$4\frac{3}{4}$
	Aug. 23	09 32 39	70.8	N	14.9	W	$4\frac{4}{4}$
	Aug. 23	11 39 18	70.8	N	14.4	W	$4\frac{3}{4}$
	Aug. 24	06 18 10	70.8	N	14.7	W	$6\frac{3}{4}$
	Aug. 27	12 21 28	70.9	N	14.4	W	$5\frac{1}{4}$
	Sept. 9	18 27 23	74.5	N	9	E	$4\frac{3}{4}$

TABLE 2 *(continued)*

Date			Time	Location				M
1954	Oct.	16	00 28 11	70.9	N	13.8	W	5
	Oct.	16	20 15 32	70.9	N	13.8	W	5
	Nov.	23	04 25 33	71.8	N	1.0	W	5
1955	Jan.	25	14 50 05	81.0	N	0	W	$5\frac{3}{4}$
	Feb.	4	09 04 39	69.9	N	15.0	W	5
	Feb.	6	00 55 32	70.7	N	14.6	W	$5\frac{1}{2}$
	Feb.	6	02 27 53	70.7	N	14.6	W	$6\frac{1}{4}$
	Feb.	27	07 47 00	66.2	N	16.3	W	$4\frac{3}{4}$
	Feb.	27	08 28 31	66.2	N	16.3	W	$4\frac{3}{4}$
	March	1	00 33 48	71.0	N	8	W	$4\frac{3}{4}$
	March	3	20 47 21	71.4	N	4.2	W	$5\frac{1}{4}$
	May	19	03 11 17	66.5	N	17.5	W	5
	May	22	18 32 40	77.0	N	6	E	$5\frac{1}{4}$
	June	28	04 28 08	86.5	N	70	E	$5\frac{3}{4}$
	July	14	01 50 39	68	N	19	W	$4\frac{1}{2}$
	Sept.	14	17 32 10	82	N	110	E	5
	Nov.	16	23 54 04	83.5	N	3	W	5
1956	Jan.	5	20 27 04	72.0	N	1.5	E	5
	Feb.	21	22 59 24	73.5	N	8	E	5
	Feb.	22	00 07 37	73.5	N	8	E	$5\frac{1}{4}$
	March	4	03 18 10	83.0	N	112	E	5
	May	8	10 46 22	75.0	N	6	E	5
	May	10	18 12 01	80.0	N	0		5
	May	13	04 27 15	85.0	N	90	E	5
	May	13	08 56 36	85.0	N	90	E	5
	May	13	14 34 00	85.0	N	90	E	$5\frac{1}{4}$
	June	3	05 19 22	80.0	N	118	W	$5\frac{1}{2}$
	Sept.	8	18 08 15	77.5	N	8	E	$5\frac{1}{4}$
	Sept.	28	15 01 34	78.5	N	7	E	5
	Sept.	29	23 01 25	69.3	N	13.1	E	$4\frac{3}{4}$
	Oct.	29	13 48 25	66.7	N	17.5	W	$4\frac{1}{2}$
	Oct.	29	16 21 00	66.7	N	17.5	W	$4\frac{3}{4}$
	Oct.	29	16 31 56	66.7	N	17.5	W	$4\frac{1}{2}$
	Oct.	30	00 11 05	66.7	N	17.5	W	5
	Nov.	12	20 57 35	73	N	8	E	$4\frac{3}{4}$
	Nov.	13	02 58 37	73	N	8	E	5
	Dec.	31	04 42 29	71	N	14	E	$4\frac{3}{4}$
1957	Jan.	25	23 26 09	80.5	N	8	W	$4\frac{1}{2}$
	May	2	03 55 34	72.0	N	67.5	W	6
	May	22	18 32 35	77.0	N	10	E	$5\frac{1}{4}$

TABLE 3. Arctic epicentres during I. G. Y.

1957

Date	Origin Time	Epicentre	Magnitude	Accuracy
July 9	20 20 38	68.2 N 18.3 W	4	B
July 9	20 35 02	68.2 N 18.3 W	$4\frac{1}{2}$	A
July 9	21 04 46	68.2 N 18.3 W	$3\frac{3}{4}$	B
July 9	21 20 22	68.2 N 18.3 W	$4\frac{3}{4}$	A
July 10	06 05 48	68.2 N 18.3 W	$4\frac{3}{4}$	A
July 24	17 52 50	64.5 N 17.5 W	$3\frac{3}{4}$	B
July 28	06 17 17	68 N 33.5 E	—	—
Aug. 14	18 48 02	64.5 N 22.5 E	—	B
Sept. 2	07 01 10	82.0 N 0	$4\frac{3}{4}$	B
Sept. 2	07 02 18	70 N 4 W	—	—
Sept. 3	05 33 30	63 N 30 E	—	B
Sept. 8	01 21 23	77.8 N 128.4 E	$4\frac{1}{4}$	C
Sept. 11	21 11 37	66.2 N 18.5 W	$3\frac{1}{4}$	B
Sept. 16	01 34 36	82 N 120 E	—	—
Sept. 19	17 29 02	79.5 N 3 W	5	—
Sept. 20	06 28 20	83 N 10 W	$4\frac{3}{4}$	C
Sept. 20	06 31 15	64.5 N 30.0 E	—	C
Sept. 27	04 58 55	63.5 N 178.0 E	$5\frac{1}{2}$	B
Sept. 30	09 09 15	73.5 N 38 E	—	C
Oct. 1	03 10 13	71 N 8 W	$4\frac{1}{2}$	C
Oct. 10	06 54 44	71 N 52.5 E	—	—
Oct. 11	04 53 35	64.7 N 16.5 W	$3\frac{3}{4}$	B
Nov. 1	20 08 15	65.4 N 30.0 E	—	C
Nov. 4	11 56 20	64.0 N 19.0 W	$3\frac{3}{4}$	B
Nov. 9	22 13 10	71.5 N 2 W	$4\frac{3}{4}$	B
Nov. 18	05 54 17	64.0 N 19.0 W	$3\frac{3}{4}$	B
Nov. 24	09 43 36	78.0 N 20 W	$4\frac{3}{4}$	B
Nov. 29	11 34 52	65.0 N 19.8 W	$3\frac{3}{4}$	B
Nov. 30	15 50 02	64.0 N 19.0 W	$3\frac{1}{2}$	B
Nov. 30	17 41 15	83.8 N 116.5 E	—	B
Dec. 2	23 58 58	83 N 25 W	(h=100)	—
Dec. 5	14 04 30	72.0 N 6 E	$5\frac{1}{4}$	B
Dec. 9	08 02 18	64.8 N 17.3 W	$4\frac{3}{4}$	A
Dec. 9	22 07 40	65.0 N 133.5 W	$5\frac{1}{4}$	B
Dec. 9	22 39 53	66.0 N 18.5 W	3	B
Dec. 10	08 15 58	65.0 N 133.5 W	$5\frac{3}{4}$	B
Dec. 12	11 52 54	68 N 14 E	—	—
Dec. 27	05 08 46	72 N 3 E	—	—
Dec. 31	00 24 02	71 N 05 W	—	C
		1958		
Jan. 9	21 03 26	65.5 N 80 W	$4\frac{1}{2}$	B
Jan. 9	21 14 34	65.5 N 80 W	5	B
Jan. 12	11 56 34	81 N 09 E	—	C

TABLE 3 *(continued)*

1957

Date	Origin Time	Epicentre	Magnitude	Accuracy
Jan. 19	19 45 01	Near Nilwara Sweden	—	—
Jan. 23	13 35 08	64.8 N 7.5 E	$5\frac{1}{4}$	B
Jan. 25	09 09 17	73.0 N 8 E	$4\frac{3}{4}$	B
Jan. 27	15 07 00	64.0 N 21.5 W	$3\frac{3}{4}$	B
Jan. 30	18 31 44	69 N 122.5 W	—	B
Feb. 5	13 53 20	64.0 N 21.5 W	$3\frac{1}{2}$	B
Feb. 7	00 47 33	74 N 52 W	—	B
Feb. 8	07 53 20	64.0 N 21.5 W	$3\frac{3}{4}$	B
Feb. 15	15 14 21	65.7 N 15.5 E	—	—
Feb. 16	22 58 03	67.8 N 18.4 W	$3\frac{3}{4}$	B
Feb. 16	23 01 57	67.8 N 18.4 W	$4\frac{3}{4}$	A
Feb. 19	14 25 03	65 N 145 W	—	—
Feb. 24	14 40 41	66.5 N 13 E	—	—
March 2	17 17 47	67.0 N 144.0 E	$4\frac{1}{2}$	B
March 2	18 14 34	67.0 N 144.0 E	$4\frac{1}{4}$	B
March 16	10 09 12	73.2 N 117.6 E	$4\frac{1}{4}$	C
March 24	00 41 47	64.5 N 17.0 W	4	B
March 25	09 39 12	66 N 122.5 W	—	B
March 30	01 45 17	66.5 N 05 E	—	C
March 30	17 56 44	79 N 30 W	—	B
March 31	15 48 39	66.5 N 157 W	—	C
April 2	03 22 31	64.0 N 19.5 W	$3\frac{1}{2}$	B
April 4	08 37	Arctic Ocean, West of Svalbard	—	—
April 6	16 00 15	83.5 N 18 E	—	B
April 7	04 56	East of Jan Mayen	—	—
April 7	15 30 38	66.5 N 157 W	7	—
April 7	16 06 07	66.5 N 157 W	—	—
April 7	16 38 32	66.5 N 157 W	—	—
April 8	00 14 20	66.5 N 155.5 W	$5\frac{1}{2}$	—
April 10	23 08 08	66 N 151 W	—	—
April 12	15 26 07	67 N 155 W	—	—
April 13	01 48 35	65.2 N 155.0 W	—	B
April 13	09 07 24	65.5 N 155.0 W	6	B
April 14	03 12 25	66 N 155 W	—	—
April 25	22 09 02	64.8 N 17.5 W	$3\frac{1}{2}$	B
April 30	20 10 33	63.5 N 129 W	—	B
May 8	07 25 36	64.6 N 17.1 W	$3\frac{3}{4}$	A
May 8	13 07 37	72.5 N 07 E	—	C
May 9	17 45 59	82.5 N 17 W	—	B
May 10	22 54 37	64.0 N 152.0 W	6	B
May 10	23 13 19	64.5 N 152.5 W	—	—
May 11	05 23 54	64 N 152.0 W	$5\frac{1}{2}$	B
May 11	05 37 01	65 N 151.5 W	—	—
May 11	09 08 43	65 N 152.5 W	—	—

TABLE 3 *(continued)*
1958

Date		Origin Time	Epicentre				Magnitude	Accuracy
May	11	12 11 22	65	N	153.5	W	—	—
May	11	13 22 14	Alaska				—	—
May	13	03 33 58	86.5	N	52	E	—	B
May	13	11 17 20	66	N	151.5	W	—	—
May	16	16 29 41	63.5	N	26	E	—	B
May	24	20 16 10	63.5	N	28	E	—	B
May	29	02 42	72	N	7	E	—	—
June	7	19 42 22	64	N	25	E	—	B
June	8	12 01 05	66.2	N	17.0	W	3	B
June	14	07 58 25	74.0	N	8	E	$4\frac{3}{4}$	B
June	15	03 01 03	64.5	N	22.5	E	—	B
June	18	01 15 01	68.8	N	16.5	W	$5\frac{1}{2}$	A
June	18	02 23 24	68.8	N	16.5	W	$5\frac{1}{4}$	A
June	18	02 54 34	68.8	N	16.5	W	4	B
June	18	04 34 00	68.8	N	16.5	W	$5\frac{1}{2}$	A
June	18	19 44 14	68.8	N	16.5	W	$4\frac{1}{2}$	B
June	30	11 29 53	63	N	27	E	—	B
June	30	14 02 08	73.0	N	69.5	W	5	B
June	30	21 01 34	65	N	24	W	—	B
July	1	01 00 41	65.5	N	17	W	—	B
July	5	14 12 59	64.5	N	22.5	E	—	B
July	6	16 03 14	65	N	155.0	W	5	B
July	8	03 47 36	63.5	N	27	E	—	B
July	18	13 33 47	67.7	N	13	E	—	—
July	18	18 51 12	64	N	25.3	E	—	B
July	19	16 00 44	66.5	N	18.0	W	$3\frac{1}{4}$	B
July	21	15 52 39	66.5	N	9.5	E	—	—
July	22	09 35 51	64	N	18	W	—	B
July	24	02 44 09	66.5	N	9.5	E	—	—
July	26	18 31 05	64	N	19	E	—	B
Aug.	28	03 12 05	66.5	N	21.0	E	—	—
Aug.	31	23 00 13	62.0	N	143.5	W	$5\frac{1}{2}$	B
Sept	3	01 46 30	63	N	26	E	—	B
Sept.	13	04 04	Arctic Ocean				—	—
Sept.	16	03 52 59	61.3	N	136.4	E	$4\frac{1}{2}$	B
Sept.	21	01 30 22	64.7	N	16.6	W	$3\frac{1}{2}$	A
Sept.	21	22 33 56	81	N	04	E	—	C
Sept.	24	19 45 32	64.6	N	17.2	W	$3\frac{1}{2}$	A
Sept.	27	10 41 30	66.1	N	17.8	W	$4\frac{1}{2}$	A
Sept.	27	11 20 10	66.1	N	17.8	W	$3\frac{3}{4}$	B
Sept.	27	21 00 11	66.1	N	17.8	W	$3\frac{1}{4}$	B
Oct.	1	16 43 39	71.8	N	3.0	W	$4\frac{3}{4}$	B
Oct.	2	14 29 59	71.7	N	1	W	—	—
Oct.	2	15 21 27	71.5	N	1	W	$4\frac{1}{2}$	B
Oct.	2	22 21 04	71.7	N	0		—	—

TABLE 3 *(continned)*

1958

Date	Origin Time	Epicentre	Magnitude	Accuracy
Oct. 4	00 33 08	64.9 N 19.3 E	—	—
Oct. 4	12 58 24	Alaska	—	—
Oct. 11	00 41 35	65.5 N 132.5 W	—	—
Oct. 26	15 24 13	65.5 N 133 W	—	—
Nov. 2	03 21 32	79 N 20 W	—	C
Nov. 4	20 29 14	72.5 N 04 W	—	B
Nov. 8	00 01 35	64.8 N 17.4 W	$3\frac{3}{4}$	A
Nov. 8	07 07 12	66.1 N 17.8 W	$3\frac{1}{4}$	B
Nov. 8	08 30 25	66.1 N 17.8 W	$3\frac{3}{4}$	B
Nov. 8	13 42 44	64.8 N 17.4 W	4	A
Nov. 10	03 42 17	77.5 N 01 W	—	B
Nov. 13	10 39 00	71 N 08.5 W	$4\frac{1}{2}$	C
Nov. 14	13 29 11	64.8 N 17.4 W	$3\frac{3}{4}$	B
Nov. 18	07 53 35	69.5 N 10 W	—	B
Nov. 22	20 57 16	71 N 16 W	—	B
Nov. 26	22 05 35	81 N 121 E	—	C
Dec. 6	00 55 01	66.4 N 18.5 W	$3\frac{1}{4}$	B
Dec. 6	00 59 16	66.4 N 18.5 W	$3\frac{1}{2}$	B
Dec. 6	09 43 28	66.4 N 18.5 W	$3\frac{3}{4}$	B
Dec. 6	09 44 32	66.4 N 18.5 W	$3\frac{3}{4}$	B
Dec. 6	09 51 54	66.4 N 18.5 W	3	B
Dec. 6	11 12 39	66.4 N 18.5 W	$4\frac{3}{4}$	B
Dec. 6	11 17 40	66.4 N 18.5 W	$3\frac{1}{4}$	B
Dec. 6	11 25 45	66.4 N 18.5 W	$3\frac{1}{2}$	B
Dec. 6	11 46 40	66.4 N 18.5 W	$3\frac{1}{4}$	B
Dec. 6	13 04 54	66.4 N 18.5 W	$3\frac{1}{4}$	B
Dec. 6	13 07 43	66.4 N 18.5 W	$3\frac{1}{4}$	B
Dec. 6	15 31 32	66.4 N 18.5 W	$4\frac{1}{2}$	B
Dec. 6	15 33 19	66.4 N 18.5 W	$4\frac{1}{2}$	B
Dec. 6	15 48 40	66.4 N 18.5 W	$3\frac{1}{2}$	B
Dec. 6	19 00 23	66.4 N 18.5 W	$3\frac{3}{4}$	B
Dec. 7	01 17 23	83 N 162 E	—	B
Dec. 7	20 38 17	66.4 N 18.5 W	$3\frac{3}{4}$	B
Dec. 7	20 40 26	66.4 N 18.5 W	$3\frac{1}{2}$	B
Dec. 8	04 48 15	66.4 N 18.5 W	$3\frac{3}{4}$	B
Dec. 8	09 20 16	66.4 N 18.5 W	3	—
Dec. 8	09 21 02	66.4 N 18.5 W	$3\frac{1}{4}$	B
Dec. 8	15 14 55	66.4 N 18.5 W	$3\frac{3}{4}$	B
Dec. 8	16 08 09	66.4 N 18.5 W	3	B
Dec. 9	09 10 40	87.5 N 45 W	—	B
Dec. 10	02 49 11	83 N 16 W	—	B
Dec. 11	04 00 58	66.4 N 18.5 W	$3\frac{1}{2}$	B
Dec. 11	05 34 18	66.4 N 18.5 W	3	B

4. MICROSEISMS

4.1. Microseisms in the Western Atlantic Region

by ROBERT STONELEY

4.1.1. *Introduction*

MANY of the introductory comments made in the Report on Microseisms in the Eastern Atlantic Region apply also to the Western Atlantic Region, and it is unnecessary to repeat them here.

There is, however, one notable meteorological difference between the two regions, and that is the prevalence of hurricanes of tropical type off the eastern coasts of North America; these storms sometimes move inland, causing extensive damage, and seem to be associated with microseismic activity over and above the activity attributable to low pressure systems in the Atlantic Ocean. The earlier success of a tripartite system of seismographic stations in locating tropical storms led to an intensive study of the possibility of tracking hurricanes seismologically before they reached the coast. A twelve years' programme of observational work covering the years 1944–55 was carried out by the U.S. Coast and Geodetic Survey on behalf of the U.S. Navy, and the discussion of the material and the preparation of the report, which was published in April, 1959, represents the main body of microseismic investigations in this region associated with the IGY. Systematic readings of microseisms were made by a number of stations in the Western Atlantic Region, including Canada and the West Indies network, but no intimation has been received of "special studies" other than those cited at the end of the present report: nor has any material been received from South American stations.

4.1.2. *Investigations in the United States, 1957–60*

A detailed report has been received of progress in microseismic research in the United States for the triennium 1957–60. The earlier data have been critically reviewed in the 1959 report mentioned above, which reproduced complete microseismic amplitude data from stations operated in the Atlantic–Carribbean hurricane area and in the south-west Pacific typhoon area during 1944–55. Storm-tracks, amplitude — distance — intensity graphs from selected hurricanes and typhoons, and microseismic vs. wave data were likewise reproduced. It was concluded that hurricanes, typhoons and other storms at sea are responsible for the generation of storm microseisms, and that such microseisms indicate the presence of a disturbance at sea which may be a tropical storm — if in season. However, there is often a time lag, of as much as 36 hours, between the closest passage of the storm and the maximum microseismic amplitudes; such a lag would not be expected if the microseisms were generated in the vicinity of the storm, and may be explained in terms of the travel-time of the swells from the continental boundary or coast near the station. The reflexion of these swells or opposing swells generated by other means would create the proper condition for the wave interference theory of microseismic generation formulated by Longuet-Higgins.

A comparison of microseismic amplitude and period at certain stations with data from wave gauges situated near the stations has shown a high degree of correlation between microseisms and swell. Although small microseisms may be generated directly from the storm centre, the study concludes that the major microseismic generating mechanism at coastal and inland stations is associated with the interference of swells near the stations. The U.S. Navy experience of attempting to track tropical storms by the use of tripartite stations was not satisfactory, and other attempts using relative amplitudes at several stations proved partially satisfactory; additional research, using arrays more extensive than a tripartite array, has been recommended. It was found that about one tropical storm a year could be sensed by the microseismic technique as early as 24 hours prior to sensing by other techniques.

4.1.3. *Collection and evaluation of raw data from IGY stations*

The U.S. report deals with data collected at stations in regions other than the Western Atlantic. It would be inappropriate to refer to them in detail in the present report; but it ought to be pointed out that the classification into microseismic regions is one of convenience only, and that it may be expected that at some future date the observations made in these other regions, Antarctica, Arctica and the Pacific Islands, will be integrated with those of the Atlantic regions into a comprehensive world-wide study.

Station bulletins are available from the four stations in and around North America that are known to have measured microseisms at six-hourly intervals; these are San Juan, Porto Rico; Morgantown, West Virginia; Sitka, Alaska; Honolulu. These readings have still to be evaluated. During the last decade the effort in the United States has changed its emphasis from the mere collection of data to a study of case histories.

4.1.4. *Special studies*

(i) P. POMEROY gave a paper at the Americal Geophysical Union meeting in May, 1959 entitled "Background and Storm Microseisms in the Period Range 11-22 sec.". Detection of these microseisms is possible because of the filter system which removes the shorter periods from the records. The amplitudes have magnitudes from 0·01 to 0·1 those of the 6 to 8 sec. microseisms. They are believed to be caused by gravity wave action near the shore.

(ii) C. A. NANNY in *Nature*, 22 March 1958, reported that there is a close correlation between microseismic frequencies and amplitudes and earthquakes.

(iii) J. E. DINGER in *Earthquake Notes, 27*, Nos. 2–3, reported a lag in microseismic amplitudes observed on Barbados which would allow travel time for the swell to be reflected from the South American coast and return to Barbados.

(iv) R. F. HEINRICH in *Earthquake Notes, 27*, Nos 2–3, using new instrumentation in a tripartite net, found that 0·3 and 0·6 sec. microseisms were moving waves, but came to no conclusion as to their origin.

(v) *The Microseismic Program of the U.S. Navy.* A terminal report by D. S. CARDER and R. A. EPPLEY. April, 1959, p. 196. U.S.C.G.S., Washington 25, D.C.

4.2. Microseisms in the Eastern Atlantic Region

by ROBERT STONELEY

4.2.1. *Introduction*

The classification of studies of microseisms by regions is necessarily only a rough attempt to make a systematic attack on the nature and origin of microseisms, their relation to other branches of earth science (such as oceanography and meteorology) and their use in attempting to solve some of the problems of the structure of the crust of the Earth. The subject, despite the vast amount of effort expended over many years, is still without a firmly established basis; evidently more than one mode of generation is involved, and one hope of the IGY investigations is that the differing causes may gradually be sorted out.

The Atlantic region is specially important for this work. One major cause of microseisms seems to be the transmission to the ocean floor of pressure fluctuations produced by standing waves, which in their turn arise from storms at sea. This particular mechanism has been shown to be quantitatively competent to generate the 4–8 sec. period microseisms observed at stations on the extreme western regions of Europe. This example, however, at once shows up the difficulty that microseisms observed in any one region may have originated far away, perhaps in another region, or on the other hand relatively near to the station; their path (if the notion of a 'path' is a correct one) may be continental or oceanic, and their amplitude has been shown fairly convincingly to depend on the rocks underlying the observatory. Progress in confirming that this mechanism is actually responsible for observed long-period microseisms has been made by considering cases where the weather situation in the Atlantic is fairly simple, but generally the weather situation is far from simple, and much labour will be required to examine this hypothesis on a day-to-day basis. There is good evidence for other ways of generating microseisms: if any one way can be established quantitatively there is hope that the systematic removal of that type of microseism from the observations, by frequency-analysis or otherwise, would make it easier to separate out other types. It is clear, however, that progress must be slow and that scientists must avoid prejudice. If the spectrum of microseismic waves should ultimately permit a reliable construction of the relevant part of the dispersion curve (of group velocity against period) we should have a valuable technique of investigating both continental and oceanic structure.

4.2.2. *Observations*

Within the above-mentioned limitations, the observations for the Eastern Atlantic will for the most part relate to the data recorded in European observatories. Readings of seismograms from a large number of stations are stored at the three World Data Centres; their systematic analysis is presumably not to be carried out immediately, and for many purposes a detailed examination of the records, made at a number of stations, of the microseisms recorded in a short interval may be more fruitful than the study of averages of period and amplitude over longer intervals. The two kinds of research are complementary, and will naturally follow the ideas and plans of the various investigators. Thus, this report will necessarily deal with a limited number of researches rather than a résumé of all the work carried out during the IGY on microseisms. Nor would it be reasonable to confine attention merely to what has been done during the 18 months of the IGY. Some projects were already in being before this period, and the stimulus of the IGY will have started investigations for which at most a progress report only can be expected. Thus, no list of station bulletins of microseisms will be attempted in this report.

The information that is embodied in the present report may be taken as relating to the North Atlantic region, but no line of demarcation can (or should) be drawn between the Atlantic and the polar regions. No material relating to the Southern Atlantic has been received. It will be seen that patient and systematic studies are revealing quantitative relationships; but the extremely irregular coast-lines (at which refraction no doubt introduces a complication both in direction of approach of microseisms and in intensity through partial focusing of the wave fronts), the variations in crustal thickness and composition, and the rapid fluctuations of meteorological conditions all combine to prevent any simple resolution of the problem.

A. *Czechoslovakia*

A very detailed investigation of the nature, and origin of microseisms recorded at a number of European stations has been made by A. ZÁTOPEK, who has furnished the following summary:

"The general characteristics of European microseisms in the period-range from 3 to 9 sec were the subject of investigation. A continuous analysis of amplitudes and periods of microseisms in relation to synoptical factors at 0h, 6h, 12h and 18h GMT, and additional parameters, has been carried out at the seismological station of Praha (50°04′.2 N, 14°26′.0 E) since 1948. Comparisons of results obtained were made with those of more than 15 European stations; the periods of activity falling into the IGY and IGC were specially considered. Earlier results of GUTENBERG, BÅTH and others were confirmed, completed and generalized.

The common character of microseisms observed at the seismological stations in Europe appears to be dominated by the cyclonic activity in the eastern part of the frontal zone between the eastern coast of North America and the western coast of Europe, mainly in the area situated north of 50° N and east of 40° W. In this region the most important sources of microseisms occur in connexion with the development and movement of intense cyclones passing over the open ocean roughly from west to east. Microseisms provoked by these sources "of first order" are propagated through the European continent and can still be noted in records of seismological stations in S. E. Europe, e.g. that of Athens. These microseisms are often superimposed by oscillations coming from nearer sources "of second order" which are responsible for regional particularities of microseisms and appear rather in the vicinity of the coast. The latter effects, due to secondary barometric lows, coastal winds, cold front passages from the sea to the land and the surf, may be very strong at coastal stations and are observed at varying distances from the source and from the coast; geological conditions are of a great importance. In this way some discrepancies in results of various stations can be explained. On the contrary, the continental disturbances in well-protected vaults are negligible.

Using idealized sum curves of amplitudes the general character of microseismic activity may be derived for various active periods and for individual stations. For a given period of activity the fundamental shape of adjusted amplitude curves remains similar for different stations. This is clear evidence for a close relation of microseismic activity with the atmospheric circulation in the frontal zone mentioned above. An index number connected with the west-east component of circulation in the isobaric surface of 500 mb. shows a similar general course like that of microseismic amplitudes.

The complexity of microseisms indicates that they are generated in different ways. Their origin in the open ocean may be explained by pressure variations transferred to the sea bottom from the surface according to the theory of Longuet-Higgins. This conclusion seems to be supported by the passive behaviour of most stationary cyclones. Standing waves may also produce microseisms in closed basins, e.g. in the Baltic Sea. Some of the microseisms with their origin in coastal areas can be immediately attributed to the surf effect.

Correlating positions of barometric low centres in the Atlantic with amplitude spanning amplitudes in Central Europe, where the coastal effects may be neglected, one obtains, by higher frequency numbers, statistical information on the position of the origin regions. For the station of Praha it was found that there exists a preference for oceanic depths in the range of 1000 to 3000 m.

B. *Denmark*

Coastal regions are situated favourably for attempts to discriminate among the various theories of the origin of microseisms.

A study has been made at Copenhagen, Denmark, by H. JENSEN, of the daily determination of the direction of approach of microseisms at that station. These are of a complicated nature, as might be expected from the situation of the station in relation to the North Sea and the Baltic. The instruments used were the homogeneous Galitzin-Wilip long-period installation.

The procedure adopted was the "method of the empty half-plane", according to which the north and east velocities are read at the instants when Z is a maximum; the mean of these horizontal velocity-vectors will be opposite to the direction of approach of the waves. The interval 23h 59m to 00h 01m was used for every day of the year 1958, so that no bias in favour of a particular type of microseism might be introduced. A suitable 'measure of reliability' (p) was devised for the accuracy of the direction of arrival (α); on this basis it seemed justifiable to classify the material into octants of the horizontal. The values of α and p have been tabulated for each day along with the period of a full oscillation on the vertical instrument and the corresponding maximum single amplitude in Z; these last two were taken from the station bulletins.

Some interesting general results emerged. The winter months all give a pronounced maximum in the northern octant of arrival: during the summer there is a secondary maximum in the neighbourhood of the south-east; this annual variation is, of course, a statistical one. There are annual variations in the period and the amplitude of Z, and when analysed according to direction the great periods are seen to be associated with microseisms from the north; this result is not merely the effect of storms. The author points out the need for extended studies over longer periods of time.

The results from the co-operative studies in the Danish programme microseismic readings for every 6 hours, and for IGY days and periods, have been made and published by the Danish stations except Ivigtut. The following instrumental notes may be of interest.

Kobenhavn. This is a well equipped station, and at present has north, east, Z medium period Galitzin-Wilip seismographs, Wiechert north, east, Z medium-period seismographs and a Benioff Z short-period instrument.

Ivigtut. Since 1953 the station has been only partly in operation; at present it is equipped with a Milne-Shaw north medium-period seismograph and a Willmore short-period vertical instrument. A special microseismic bulletin will be issued for the IGY period.

Scoresbysund. The present equipment consists of Galitzin-Wilip north, east, Z seismographs and a Grenet short-period vertical.

Nord. This station, located at 81° 36′ north, 16° 41′ west has been in operation since 1 September, 1957. It is equipped with Strobach north and east seismographs of 6 sec. period and a Willmore short-period vertical.

The situations of these last three stations are important in providing a link between the phenomena of continental Europe and the Arctic and North West Atlantic regions.

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C. *German Democratic Republic*

The following report has been received from Professor Dr. H. Martin, Director of the Institute of Earthquake Research at Jena:

"Die Auswertung der mikroseismischen Bodenunruhe wurde von der Station Jena und den Nebenstationen Halle und Potsdam laufend durchgeführt und während des IGJ den Weltzentralen und anderen interessierten Stellen zugeleitet. Spezialuntersuchungen sind nicht vorgenommen worden, da das Institut mit anderen Aufgaben zu sehr in Anspruch genommen war.

Den Auswertungen lagen in Jena die Seismogramme des Wiechert-Horizontalseismometers ($T_0 = 8$ sec., $V = 200$) bei den Nebenstationen in Halle die Seismogramme vom Krumbach-Horizontalseismometer ($T_0'' = 7$ sec., $V = 150$) und in Potsdam die vom Galitzin–Wilip-Horizontalseismometer ($T_s =$ ca. 12 sec., $T_G = ca.$ 12 sec., $V_{\max} =$ ca. 1000) zugrunde. Ein homogener Satz elektrodynamischer Seismographen ($T_0 = 20$ sec.) wird demnächst in der Station Jena in Betrieb genommen, der günstigere Aufzeichnungen der Mikroseismik erwarten lässt. Die geplante Inbetriebnahme von Vektor-Seismographen konnte bisher wegen technischer Schwierigkeiten nicht verwirklicht werden.

An den Arbeiten sind beteiligt:

Herr Dr. Friedrich GERECKE und Fräulein Dipl. Geophys. Dorothea GÜTH, Jena, und Fräulein Dr. Gertrand RICHTER, Halle."

D. *U.S.S.R.*

Extensive studies of microseisms were carried out during the I.G.Y. The following report by N. V. VESHNYAKOV is reproduced in full as submitted:

During the period preceding the IGY, observations of microseisms were carried out in Moscow State University (MGU) and in the Institute of Earth Physics, U.S.S.R. Academy of Sciences (IFZ). Special apparatus was designed for observations based on the tripartite-station system.[1]

During the IGY, three microseismic stations of Head Management of Hydrometeorological Service of the U.S.S.R (GUGMS) were in continuous operation:

Barentsburg	78°39'N	16°23'E	granite rock.
Murmansk	68°58	33°03	granite rock.
Vyborg	60°43	28°48	granite rock.

The following tripartite microseismic stations of MGU functioned intermittently:

Pulkovo	59°46' N	30°19' E
Simferopol	44°57	34°07'
Yalta	44°30	34°10'

In addition, microseismic records were processed on seismograms of 19 general-type seismic stations, namely:

Apahty	67°33' N	33°20' E
Askhabad	37 57	58 21
Goris	39 30	46 20
Irkutsk	52 16	104 19
Khorog	37 29	71 32
Kishinev	47 01	28 52
Kurilsk	45 14	147 52
Magadan	59 33	150 48
Makhach-Kala	42 56	47 28
Moscow	55 44	37 38
Petropavlovsk on Kamchatka	53 01	158 39
Pulkovo	59 46	30 19
Southern Sakhalinsk	47 01	142 43
Semipalatinsk	50 24	80 15
Simferopol	44 57	34 07
Sverdlovsk	56 48	60 38
Tixie	71 38	128 52
Uglegorsk	49 05	142 04
Vladivostok	43 07	131 54

The observational procedure at tripartite stations is well known and will not be described here.[1] The apparatus used at these stations consisted of vertical electrodynamic seismographs with galvanometric registration and a recording apparatus that permitted recording on photographic film at a rate of from 2·5 to 15 mm/sec.

The parameters of the instruments of the tripartite stations were*

Station	T_1 sec	T_2 sec	D_1	D_2	σ^2	V
Barentsburg	8·0	4·2	0·6	1·4	0·38	8000 in 3–4 sec.
Murmansk	7·0	4·2	1·5	1·0	0·25	
Vyborg	7·0	4·2	1·5	1·0	0·30	

In 1958, these constants were changed at the stations Murmansk and Vyborg:

	T_1 sec	T_2 sec	D_1	D_2	σ^2	V	
Murmansk	7·0	4·3	0·6	2·0	0·45	7700	as of 20/IV/58
Vyborg	7·0	4·3	1·0	2·0	0·52		as of 15/VIII/58

At the intermittently functioning stations of Yalta, Simferopol, Pulkovo, the constants had the following values in 1955–57:

Station	T_1 sec	T_2 sec	D_1	D_2	σ^2	V
Pulkovo	7·0	4·0	0·3–0·4	0·7	0·2	15000 for 5·5 sec.
Yalta	7·0	4·0	0·3–0·4	0·7	0·2	15000 for 5·5 sec.
			In 1958–59			
Pulkovo	7·0	4·0	1·5	2·0	0·46	7000 for 4·5 sec.
Simferopol	7·0	4·0	1·5	2·0	0·46	7000 for 4·5 sec.

The position of seismographs in the vertices of the triangles at the microseismic stations is shown in Fig. 1.

Processing procedures of observations (1). Microseismic recordings at ordinary seismic stations were processed in accordance with procedures[2] recommended by the Special Committee of the IGY. Measurements were made four times daily, during microseismic storms, eight times a day, and during World Days and Special World Intervals, every hour. When processing recordings in intervals of several minutes, characteristic sections were chosen, for which selections were made of amplitude A in microns, period T in seconds, and an index K which characterizes the type of microseisms according to the international code.

The results of these observations were sent to the World Data Centres in the form of a summary bulletin of microseismic observations made at the seismic stations of the U.S.S.R. This bulletin was compiled in MGU on the basis of the data of the separate stations.

At tripartite stations, observations were ordinarily carried out four times a day, and, during periods when there was a considerable increase in microseism intensity, eight times a day.

Initial processing of observations was carried out at the stations. It consisted

* T_1 is the pendulum period, D_1, the damping of the pendulum σ^2 the coupling coefficient, T_2, the period of the galvanometer, D_2, the damping of the galvanometer, V, the maximum increase.

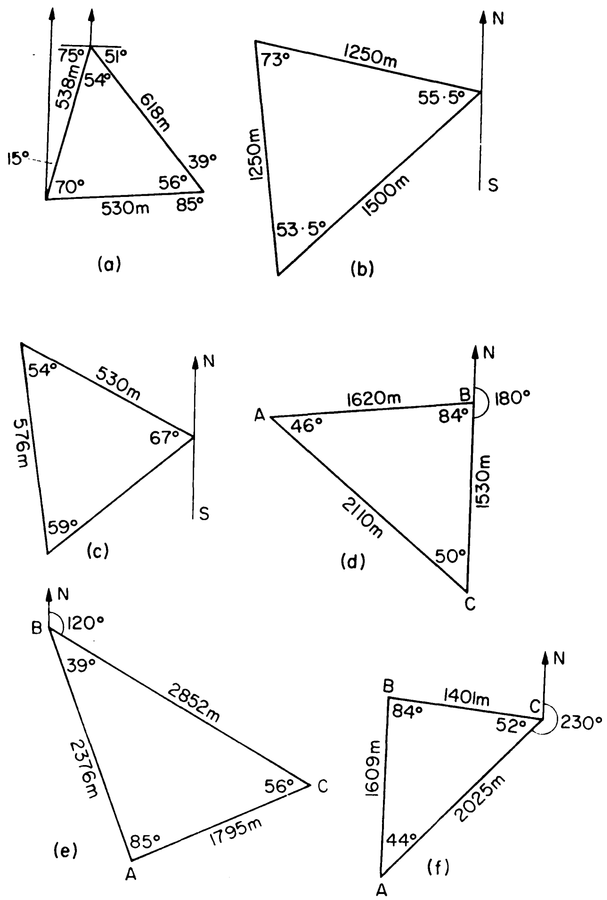

FIG. 1. Positions of vertices of triangles at microseismic stations: (a) Yalta, (b) Simferopol, (c) Pulkovo, (d) Barentsburg, (e) Vyborg, (f) Murmansk.

of determining the periods, amplitudes and azimuths of directions in which the microseisms were propagated. To measure these quantities, sections were selected on the seismograms with regular and identical microseisms in all three channels. Within the limits of these selected sections measurements were made of periods, amplitudes and displacements of vertices of the microseisms. From the vertex displacement, determinations were made (with the aid of special tables compiled for each tripartite station) of the azimuths of directions to the microseism sources. The data obtained were averaged and recorded in the bulletin of the station according to the form given in the instructions.[2]

In addition to the chief quantities — period, amplitude and azimuth — calculations were made also of phase velocity of the microseisms. The phase velocity sometimes helps to detect groups of oscillations that form as a result of waves arriving from essentially different directions. In such cases, the phase velocity is greater than normal (i.e., greater than 3 km/sec.). Such groups of oscillations were excluded.

The observations of the stations Barentsburg, Vyborg and Murmansk are summarized in GUGMS, where seismograms, bulletins and vector diagrams of the stations are sent every ten days. On the basis of these data, a monthly summary bulletin was compiled according to a form standardized for the IGY. The summary bulletin was sent to the World Data Centres and to other institutions.

Summary bulletin of tripartite microseismic stations

Date	Observational period	Name of station	K	T sec	A	Drum speed mm/min	St_1 sec	St_2 sec	Coordinates of sources

In addition to the apparatus mentioned above, other instruments and apparatus were used for determining the direction of propagation of microseisms, namely:

The Gamburtsev–Galperin azimuthal apparatus,[3-5] which consists of 6–8 vertical electrodynamic seismographs situated at one place. The seismographs are established according to different azimuths at intervals of equal numbers of degrees relative to each other and inclined at an angle of 45° to the horizon so that the axes of their maximum sensitivity form a conical surface. The recording takes place on a single tape. From the phase shift between recordings and also from the relationship of the amplitudes it is possible to determine the type of wave and to find the direction of propagation.

Observations were carried out with a vector apparatus,[3] in which the oscillations of two paired perpendicular horizontal pendulums are combined. An apparatus of this type was first used by GUTENBERG.

F. I. MONAKHOV proposed and tested a method of position-phase correlation.[4] From 3 to 5, or more, vertical seismographs are arranged in profile at intervals of several kilometres from each other. The total length of the profile is 15 to 20 km. Synchronization of time is done by radio from a special transmitter. According to the author, this method is most reliable for determinations of the direction of propagation of microseisms.

Propagation conditions of microseisms and location of their sources. The method of tripartite stations was used at all the microseismic stations mentioned above.

The processing of observational materials of the Yalta, Simferopol, and Pulkovo stations during the period from 1954 to 1959 yielded positive results. In the overwhelming majority of cases, the directions to microseism sources in different parts of the Atlantic and Arctic oceans, and also in the Mediterranean indicated a cold cyclone front. In the cases (unfortunately, few) when it was possible to determine the azimuths to microseism sources at two stations simultaneously, for example, Yalta and Pulkovo or Simferopol and Pulkovo, the intersection of these azimuths occurred invariably in a region of the closest (beyond the centre of the cyclone) section of the cold front.[6]

An analysis of the observational results of the microseismic stations of Murmansk and Barentsburg has shown the absence of sufficiently stable directions of propagation of microseisms. The spread of equally probable directions at times reached 90° and even 180°.

To determine the conditions of propagation of microseisms and to elaborate additional methods for locating their origins, special investigations were carried out.

L. N. RYKUNOV and V. M. PROSVIRNIN worked out a method of reproducing the trajectory of a microseism, with the microseismic station as the point of departure[7]. This work is a development of the studies of STONELEY,[8] who showed that in the propagation of Rayleigh surface waves along the ocean floor, the phase velocity depends on the relationship between the wavelength and the thickness of the layer of water at the point under study. For this reason, the transition of a wave from one isobath to another may be accompanied by refraction described by the ordinary sine law. On the basis of these ideas suggested by STONELEY, DARBYSHIRE explained a number of amplitude anomalies in the records of microseisms in the British Isles and the Bermuda Islands.[9–10] The construction carried out by RYKUNOV and PROSVIRIN showed that neglect of the influence of the varying depth of the ocean can lead to gross errors when determining the position of the region of origin of microseisms.

Applying a procedure similar to that proposed by GILMORE,[11] MONAKHOV established a correlation between the position of cyclones in the Atlantic ocean and the nature of amplitude and period variations recorded at specific seismic stations. He compared such correlational characteristics with the amplitudes and periods of the appropriate stations and succeeded in solving the inverse problem as well — that of determining the position of a cyclone in the Atlantic ocean.[12] This method yielded an encouraging result, but it necessitated further verification and development.

A different approach to this problem was proposed by V. N. TABULEVICH:[13] Let a source of microseisms be studied at a point $M(x, y)$, with observation stations at M_1 and M_2 (see Fig. 2). The distances between these points and the sources are d_1 and d_2, respectively. Let us suppose that the relationship between amplitudes at points M_1 and M_2 and distance may be represented by the formula

$$\frac{d_1}{d_2} = \frac{A_2^{n_2}}{A_1^{n_1}}$$

where A_1 and A_2 are the amplitudes of microseisms at points M_1 and M_2, and n_1 and n_2 are exponents that depend upon the shape of the wave front and the

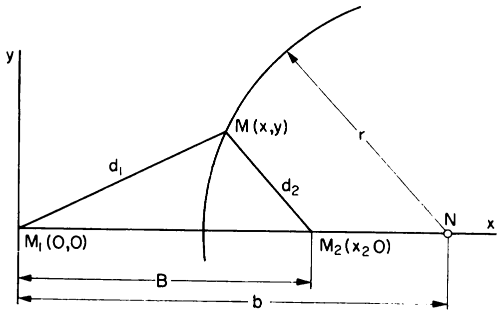

FIG. 2. Positions of source of microseisms $M(x, y)$ relative to the seismic stations M_1 and M_2.

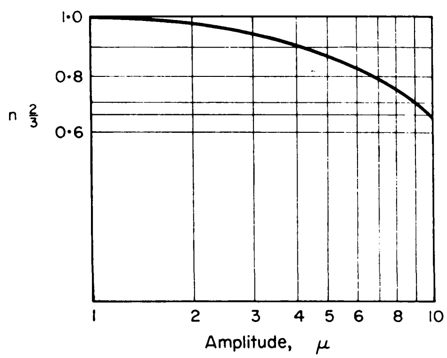

FIG. 3. The exponent n as a function of amplitude of microseisms.

presence of dissipative forces (Fig. 3). We find the locus satisfying the given condition:

The locus will be a circle of radius $z = \frac{K}{K^2-1} B$ with its centre at a point N $(\frac{K^2}{K^2-1} B, 0)$. Performing this construction for three stations (Fig. 3), it is possible to find the region of origin of microseisms. A synchronous variation of microseism periods at all three stations shows that the microseisms are propagated from a single source of disturbance. This method was tested for the Caspian Sea, and the Pacific and Atlantic oceans and yielded satisfactory results.

To determine the effectiveness of excitation of microseisms as a function of cyclone position, MONAKHOV correlated the energy of a microseism with cyclone energy.[14] This relation may be given the form:

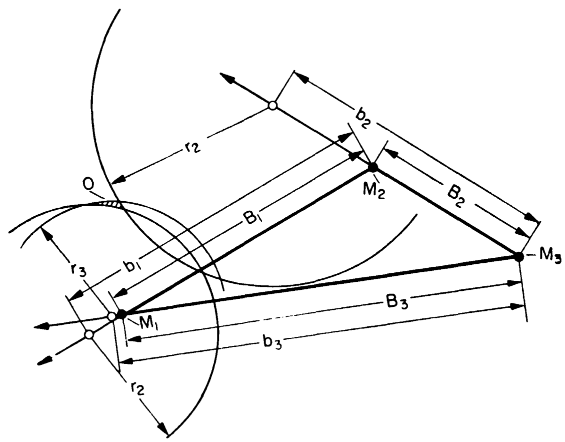

FIG. 4. Fixing the region of microseism excitation by Tabulevich method.

$$K = \frac{E_m}{E_c} = \left(\frac{A}{T}\right)^2 : \frac{L^2}{S}$$

where: A is the amplitude of the microseisms, T is the microseism period, S is the area of water surface encompassed by the cyclone, L is the total length of isobars within the limits of this area.

On the basis of such correlations (about 80 measurements) the following table was compiled:

Region of cyclonic activity	K
Coast of Scandinavia	5·0
Coast of Europe	1·4
Norway and Greenland seas	0·8
Southern coast of Greenland and Iceland	0·2
South of Greenland and Iceland (Northern Atlantic Ocean)	0·0

On the basis of this table MONAKHOV came to the conclusion that microseisms are not excited just at any spot on the sea floor, but only at certain spots that are favourable; while in other regions microseisms are strongly absorbed. In the main, the latter are parts of the Atlantic ocean with west longitudes. If a cyclone is found in these unfavourable parts, the production of microseisms could be explained by the swell of the sea. When the swell reaches the steep shores of Scandinavia, it is reflected from the shores, producing standing sea waves, which give rise to microseisms (in accordance with the Longuet–Higgins theory).

Further study of the field of periods and amplitudes of microseisms originating near the north-western shores of the Scandinavian peninsula showed a predominant decrease in the amplitudes of these microseisms if their path crossed the Scandinavian mountains (which extend in a meridional direction) at angles close to the normal. An explanation of this phenomenon was sought in the studies of Soviet and foreign scientists. These investigations establish the variation of wave forms and the intensity of Rayleigh waves when propagated over a rough surface. The effect proved significant even for dimensions of unevenness that are small compared with the wavelength. This phenomenon was studied from the point of view of the propagation of microseisms using a model.

A three-dimensional model was constructed which reflected the peculiarities of the Scandinavian relief. The source of microseisms was modelled by means of a piezo-emitter at a fixed position on one side of the peninsula. A piezo-receiver, which modelled the seismic station, was situated on the other side of the model and was moved from point to point along the arc of a circle, in the centre of which was the emitter (Fig. 5). The measurements showed qualitative agreement between observations and model results.[15]

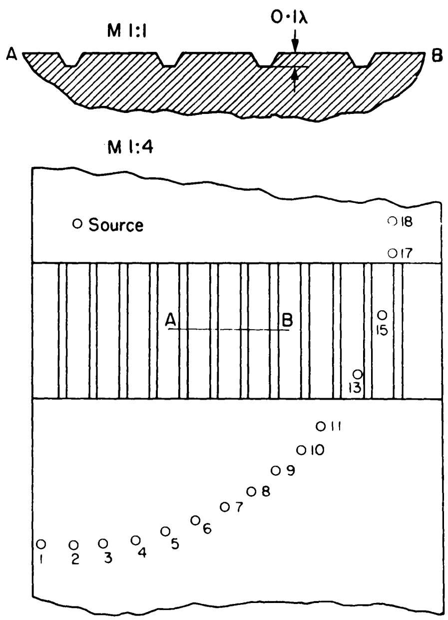

FIG. 5. Model of Scandinavian peninsula with positions of source and receivers indicated.

By way of further study of this problem, routes were considered from the source ($\psi = 15°$ W, $\varphi = 69°20$ N) to Moscow, Makhach-Kala, Warsaw, Ashkhabad, Trieste, Goris, and Semipalatinsk. The models of these routes were made of plates of organic glass on the faces of which the relief was cut (on a scale of 1 :1,000,000). The distance between the emitter and receiver remained constant for all models of this type (Fig. 6).

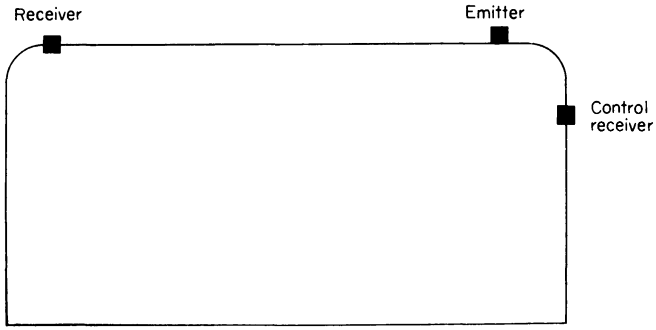

FIG. 6. Route model from source of microseisms to Moscow and other points.

The intensity of the microseisms measured by the above-mentioned stations were reduced to a single distance to the source measured in wavelengths. It was assumed that for considerable distances, the amplitude diminishes as $d^{-1/2}$ where d is the distance and d is the coefficient of absorption.

The results of the models and of observations are compared in the following table:

Station	Ashkhabad	Moscow	Makhach-Kala	Warsaw	Goris	Trieste	Semipalatinsk
Nature	1	0·75	0·75	1	0·13	0·07	0·25
Model	1	0·78	0·78	1	0·38	0·11	0·68

Figures denote quantities proportional to the microseism amplitudes at one and the same (for all stations) distance from the source.

This comparison shows that for certain routes, there are other causes giving rise to reduced microseismic intensity, in addition to geometrical discrepancies in the front of microseismic waves, absorption, and scattering on surface irregularities (Goris, Trieste, Semipalatinsk). Investigations on appropriate models showed that these causes may be irregularities in the lithological composition of the rocks of mountainous areas. Allowance for these irregularities on the routes under study led to satisfactory qualitative agreement between the observational and model results.

These investigations have shown that the ground relief is a factor which produces a significant effect on the propagation of microseisms and on intensity reductions. In addition to relief, a significant role in the reduction of microseism intensity is played by the lithological inhomogeneity of the medium along the route of propagation (the presence of nearly vertical boundaries dividing rocks of unlike properties).[16]

Suggestions for the future. (1) It is necessary to continue detailed observations both by the method of tripartite and multi-partite stations with the aim of establishing conditions of their applicability and, when unsatisfactory results are obtained, to determine the reasons for such results.

(2) It is necessary to develop amplitude methods for locating the original microseisms.

(3) It is useful to utilize more widely the model method for solving a series of problems connected with the conditions of propagation of microseisms.

(4) It is desirable to carry out direct measurements of oscillations of the ocean floor in the region of cyclone movements.

The following institutions participated in microseismic studies during the IGY:

1. Moscow State University (Physics Department).
2. Central Institute of Weather Forecasting, Head Management of the Hydrometeorological Service, U.S.S.R.
3. Institute of Earth Physics, Academy of Sciences, U.S.S.R.
4. Sakhalin Scientific Research Institute, Academy of Sciences, U.S.S.R.

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E. *United Kingdom*

For a number of years the relation between storms in the Atlantic Ocean and microseisms recorded in S. England has been studied intensively at the National Institute of Oceanography. The general success of the Deacon–Longuet–Higgins theory of the generation of microseisms from standing oscillations of the sea has concentrated attention on the distribution of frequency and the correlation between the horizontal and vertical movements. The researches of J. and M. DARBYSHIRE showed that considerable Love-wave motion is present in addition to the expected Rayleigh waves, and by studying the correlation of the vertical records with the horizontal records it was shown that, despite the admixture of the two types of wave, the direction of approach could be estimated and a storm successfully tracked. The tracking of a storm has been achieved also by A. E. M. GEDDES, using the horizontal component seismograms of the University of Aberdeen, who found that during the rapid movement of a deep depression across the north of England microseisms of very large amplitude were recorded. On the Longuet–Higgins theory these disturbances arose from standing waves, behind the low centre, produced by reflexion from the coast. They appeared to be a mixture of Rayleigh and Love waves, approaching the station from the south–south-west.

This preliminary work, further developed by H. M. IYER, has led to the setting up of a sensitive three-component installation at the National Institute of Oceanography; the seismographs have a magnification of up to 18,000 with a flat response for periods from 1 to 10 sec, and their outputs are fed into an analyser which will determine automatically the root mean square amplitude, the period and the correlation coefficients between the components. The intention is to develop a storm-warning system in areas at present not catered for. Already some interesting relationships are indicated: for instance, a narrow spectral band at the maximum of the spectrum tends to follow the storm centre more closely than the total spectrum, so that microseismic sources of different periods may be detected independently.

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4.3. Microseisms in the North-Western Pacific Region

by F. I. MONAKHOV and O. A. KORCHAGINA

4.3.1. *Introduction*

This review paper contains an analysis of microseismic observations carried out during the International Geophysical Year in the north-western part of the Pacific Ocean. The principal aim of these observations was to study the conditions of formation and propagation of microseisms. The observations were carried out both at conventional seismic and at special tripartite microseismic stations. Figure 7 gives a scheme of the distribution of the observation stations. The tripartite microseismic stations were set up by the Head Management of the Hydrometeorological Service of the U.S.S.R. (GUGMS).

Microseisms were studied on a small scale in this area prior to the International Geophysical Year. In 1952, a tripartite station was established on Sakhalin Island. The observations of this station, together with those of conventional seismic stations, made it possible to obtain certain results concerning the relationship between microseisms and cyclones over a water surface.[1-3] In these papers it is shown that enhancement of microseisms in the eastern part of the Asiatic continent and adjacent islands is always associated with cyclonic activity over the Japan and Okhotsk Seas, and also over the north-western part of the Pacific Ocean. The observed directions of arrival of microseisms varied, in time, in accordance with the direction of motion of a cyclone, yet in many cases these directions did not correspond to the directions towards the central parts of the cyclones. MONAKHOV[1] has shown that one of the reasons for this disparity is the time lag in the development of microseisms with respect to cyclone development. For this reason, in certain cases the angle between the directions towards the microseism source and the cyclone centre can amount to 180°.

MONAKHOV[3] notes an azimuthal relationship in the phase velocities of microseisms in the region of Yuzhno-Sakhalinsk. It was established that during the propagation of microseisms (and also surface waves of the earthquakes) from the east, the phase velocity exceeds the normal value, at times is close to infinity, and even has a negative value. In the meridional direction, the waves are propagated with the normal phase velocity. This circumstance greatly

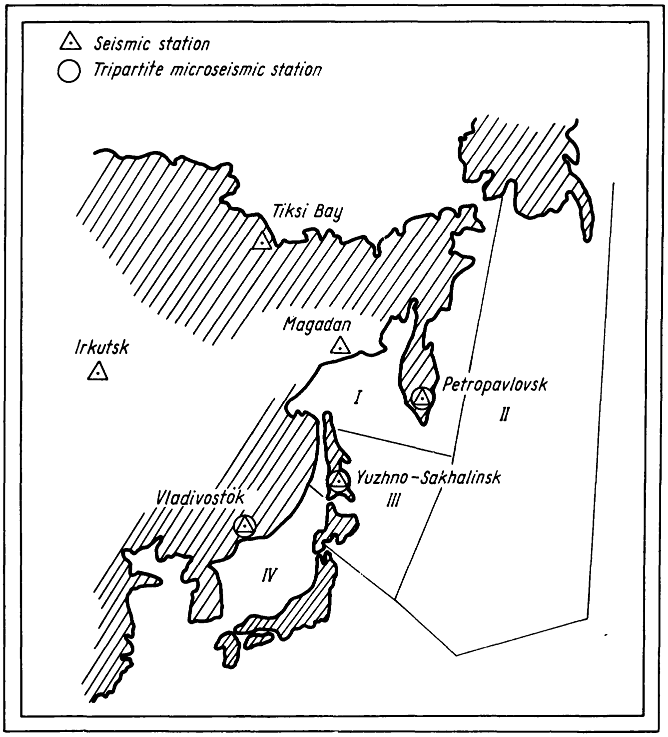

FIG. 7. Distribution of seismic and tripartite microseismic stations and zones of cyclonic activity.

complicates the determination of the directions of the microseisms origins based on observations in the area of Yuzhno-Sakhalinsk.

This review utilizes in part the data of the Sakhalin Research Institute of the U.S.S.R. Academy of Sciences (SKNII) (T. A. ANDREYEV and V. G. BUKHTEYEV) and the Vladivostok Hydrometeorological Observatory (O. V. SOBOLEVA).

4.3.2 *Apparatus and Observational Procedure*

The seismic stations are equipped with standard seismic recording apparatus (SVK and SGK seismographs*, and GK-VI galvanometers) described by Savarensky and Kirnos.[4] The tripartite microseismic stations were equipped with modernized SVK seismographs, M-21 galvanometers and RS-2 recorders with panel control. The records of the microseisms at the tripartite stations were accompanied by time marking with a scale division of 0·1 sec. The recording apparatus permitted a wide range of speed adjustment.

The main characteristics of the microseismic stations are given in Table 1.

* SVK and SGK are the abbreviations for the Kirnos vertical and horizontal seismographs.

T_1, D_1 and T_2, D_2 are the periods and damping of the seismographs and galvanometers, respectively; σ^2 is the coefficient of electrodynamical coupling between seismographs and galvanometres.

The stations operated on the first regime up to the summer of 1958, after which they were switched to the second regime so as to displace the frequency characteristic maximum towards longer-period microseisms.

The comparison of the instrument parameters served as the basic criterion of normal functioning of the microseismic stations. At the beginning of the IGY, an accuracy capable of ensuring microseism-source azimuths with errors not exceeding 10°–15° was (judging by the state of microseisms studies at that time) considered to be sufficiently accurate for practical work. In terms of microseism phase shifts, when the distance between the seismographs comes to about 2000 metres (the mean distances at U.S.S.R. stations), this corresponds to roughly 0·1 sec.

Simultaneous recording of microseisms at one spot by all instruments serves as a control of the precision of the instruments. Figure 8 shows such a record obtained at the Yuzhno-Sakhalinsk station. From this record it follows that possible phase shifts do not exceed 0·1 sec. For comparison, Fig. 9 gives a microseism record at the same station by separated seismographs. In this case, the phase shifts are readily traced and considerably exceed 0·1 sec.

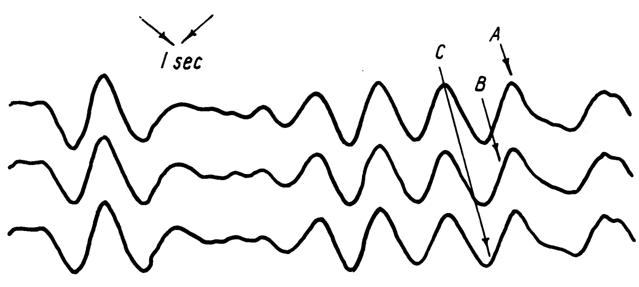

FIG. 8. Simultaneous recording of microseisms at one site by three seismographs.

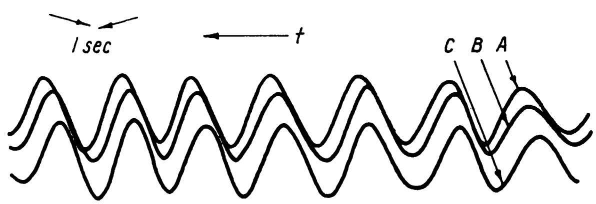

FIG. 9. An example of storm microseisms recorded by separated seismographs.

The seismographs of the microseismic stations were mounted in specially built underground and ground-level buildings. Recording were made at one site. The seismographs were connected with the galvanometers by cables with chloro-vinyl insulation suspended in the air on supports.

The microseismic records of conventional seismic stations were processed by a method recommended by the Special Committee of the IGY. Ordinarily, measurements were made at 0, 6, 12, 18h, but during microseismic storms measurements were made eight times a day. Characteristic groups of microseisms were selected on the seismograms and determinations made of the average amplitudes and periods.

Observations at the microseismic stations were conducted four times every 24 hr, and eight times per 24 hr in the case of enhanced microseismic activity. The observations were processed as follows: on the seismograms, sections were selected that had regular, more or less identical microseisms in all three channels.

TABLE 1

Station	Distances between seismographs, metres	T_1 sec	First Regime				Second Regime			
			D_1	T_2	D_2	σ^2	D_1	T_2	D_2	σ^2
Petropavlosk	1820; 1780; 1855	7·0	1·5	4·2	1·00	0·25	1·00	4·1	2·00	0·35
Yuzhno-Sakhalinsk	3240; 3020; 2810	7·0	1·5	3·9	1·90	0·23	0·82	4·0	2·00	0·40
Vladivostok	3280; 2850; 1625	7·0	1·5	4·0	1·60	0·28	0·84	4·2	2·00	0·38

Within the limits of the selected sections, determinations were made of the periods, amplitudes and peak shifts of the microseims. Using special tables compiled for each station, and on the basis of the peak shift, determinations were made of the azimuths directions to the microseism sources. The data obtained were averaged. In addition, the phase velocity of each microseism was determined. At times the phase velocity helped to detect groups of oscillations produced as a result of the interference of waves arriving from essentially different directions; in these cases, the phase velocity is greater than usual. We excluded groups with anomalously high velocity and those, which in a 20-min observation interval exhibited irregular changes in the directions of propagation of the microseisms, and those for which it was impossible to select a predominant direction.

In addition to the three microseismic stations operating according to the IGY programme, there was a similar microseismic station of the Academy of Sciences (SKNII) functioning near Yuzhno-Sakhalinsk. The parameters of this station were:

$T_1 = 8 \cdot 0$ sec; $D_1 = 1 \cdot 5$; $T_2 = 4 \cdot 0$ sec; $D_2 = 1 \cdot 0$ $\sigma^2 = 0 \cdot 28$
$U_{max} = 6000$; $a = 2165$ m; $b = 2575$ m; $c = 3215$ m
The SKNII station was situated 10 km to the north of the GUGMS station.

4.3.3. *Directions of propagation of microseisms*

On the basis of the observations of tripartite microseismic stations, the directions of propagation of the microseisms were studied from the point of view of their general characteristics and their agreement with the directions towards cyclones.

Before giving the results, let us consider in brief the relationship between the observed directions of propagation of the microseisms and the true directions towards their sources. We may point to three principal causes for discrepancies between the observed and true directions towards the microseism sources:

1. Unstable instrument parameters and errors in measurement of phase shifts.
2. Interference of microseisms propagated in various directions.
3. Curvature of the path of propagation of microseismic waves, due to refraction in the earth's crust.

An analytical discussion of the errors associated with the first cause is given by Veshnyakov.[5] As applied to our microseismic stations, these errors do not exceed 10°–15°. An analytical and experimental discussion of the second cause associated with the interference of microseisms is given by MONAKHOV *et al.*,[6] in which it is shown that, given interference, the directions determined by the method of tripartite stations do not, generally speaking, coincide in any of the true directions towards the microseism sources, and are, at best, intermediate. The reality of the third cause has been noted by a number of investigators.[7-8] The degree of influence of this cause may be evaluated only if the structure of the earth's crust over the path of propagation of the microseismic waves is sufficiently well known. Wherever it is necessary we shall, in future, bear these factors in mind.

At different stations, the nature of the microseisms and the general pattern of their directions of propagation are different. Thus, for example, in Vladivostok the microseisms have a more regular shape, while the sector of their propagation during a 20-min interval of time is narrower than at Petropavlovsk and in Yuzhno-Sakhalinsk. The records of these latter two stations frequently exhibit such scattered directions towards the microseism origin that it is impossible to isolate any preferential directions. Such a picture is observed in particular in cases when the cyclones are situated close to the observation stations. This fact can not be explained in terms of observational errors. It is undoubtedly a case of formation of microseisms in various directions from the stations.

The observational data of the Yuzhno-Sakhalinsk GUGMS microseismic station confirm the earlier established fact[3] of anomalous values of the phase velocities of microseisms in the vicinity of Yuzhno-Sakhalinsk. Figure 4 gives the position of a deep cyclone to the east of Yuzhno-Sakhalinsk. This cyclone began to form on 24 December. In 26 Dec., Yuzhno-Sakhalinsk detected the appearance of microseisms associated with the action of this cyclone. They had an enlarged period, a small amplitude and sufficiently regular oscillations in the form of groups. During the period of time under consideration there were no other essential cyclonic formations in the vicinity of Yuzhno-Sakhalinsk bounded by the approximate azimuths 90°–150°, which corresponded to the position of the rear part of the cyclone.

The mean value of the phase velocity of the microseisms amounted to about 6000 m/sec. Velocity measurement errors did not exceed 500 m/sec. Since under normal conditions microseisms are propagated with a velocity of about 3000 m/sec. we may consider that in the area of Yuzhno-Sakhalinsk the apparent velocities of the microseisms, propagated from the east, are anomalously high. This conclusion is corroborated by numerous other examples. At the same time, the microseisms propagated from the Tatar Strait, that is, from west to northwest, have phase velocities close to and less than 3000 m/sec.

Another peculiarity of the propagation of microseisms in the area of Yuzhno-Sakhalinsk is that the preferential directions of propagation of microseisms towards Yuzhno-Sakhalinsk are west and northwest and the periods of these microseisms ordinarily amount to 3 to 4 sec. Apparently, either the conditions for exciting microseisms are more favourable or the attenuation is lower in the Tatar Strait than in the Sea of Okhotsk. Westerly directions towards the microseism sources are frequent when the cyclones are mainly over the Sea of Okhotsk. With the purpose of verifying this idea, averaged directions of the propagation of microseisms were compared over a series of intervals based on observations of the SKNII and GUGMS stations, which are situated 10 km apart. The results of this correlation are given in Table 2.

TABLE 2

No.	GUGMS A_z°	SKNII A_z°	No.	GUGMS A_z°	SKNII A_z°
1.	270	255	8.	240	235
2.	270	245	9.	270	260
3.	275	260	10.	290	275
4.	255	240	11.	115	120
5.	225	240	12.	280	260
6.	250	245	13.	280	245
7.	265	240			

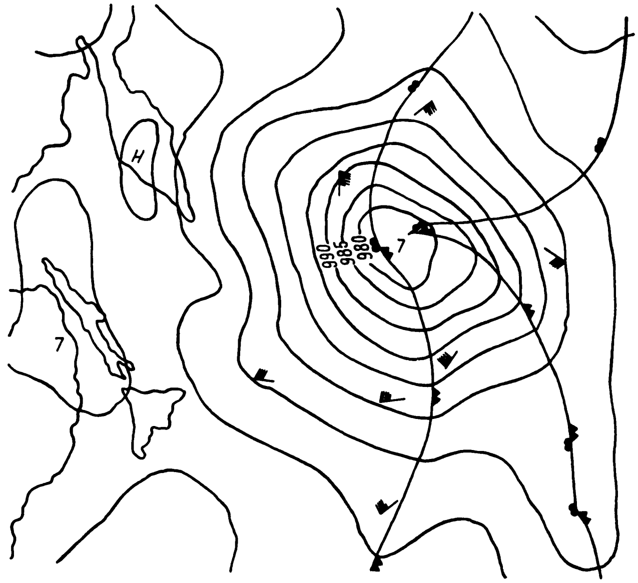

FIG. 10. Position of cyclone at 12 hr, 24 Dec. 1957.

The data given in this table show that, judging by the observations of both stations, the directions of propagation of microseisms are correlated within certain limits. The discrepancies may be explained by the small number of individual measurements and the large scatter of actual directions towards the microseism sources.

An attempt was made to give a phase correlation of microseisms based on the recordings of the GUGMS and SKNII stations, for which purpose time marks were transmitted by radio to both stations. Phase correlation of microseisms at these stations proved to be practically impossible. Apparently, interference of microseisms is one of the essential reasons.

The preferential directions in which microseisms with a period of up to 4 sec. are propagated towards Petropavlovsk are west and northwest. The longer-period microseisms ordinarily arrive from the east. Both types of microseisms are frequently superimposed and it then becomes rather difficult to determine the directions to the microseisms sources.

The directions of propagation of microseisms obtained from the observations of tripartite stations during 1957–58 were correlated with the positions of cyclones during this period. The results are given in Table 3. It was assumed that the region of excitation of the microseisms was in the same direction as the cyclone if the mean direction of propagation of the microseisms was included in the sector embracing the central part of the cyclone and the cold front. The analysis was based on the regions of cyclonic activity.

TABLE 3

Area of cyclonic activity	Vladivostok			Yuzhno-Sakhalinsk			Petropavlovsk		
	No. of cyclones	No. of coincidences	Per cent of coincidences	No. of cyclones	No. of coincidences	Per cent of coincidences	No. of cyclones	No. of coincidences	Per cent of coincidences
Sea of Japan	278	74	27	254	11	4	80	0	0
Sea of Okhotsk	328	20	6	393	27	7	205	9	4
Northwestern part of Pacific Ocean	954	46	5	1042	10	1	265	8	3

The data in Table 3 were obtained without taking account of the factors that affect the convergence of directions, namely: the lag of the region of excitation of microseisms behind the centre of the cyclone, regional peculiarities in the sense of the conditions of microseisms excitation; and the azimuthal dependence of phase velocities. This, apparently, explains the exceedingly small percentage of coincidence between the directions of microseism sources and cyclones. However, there are other factors which will be dealt with later on.

It should be borne in mind that in certain cases coincidence in the direction of propagation of microseisms with the direction towards a cyclone may be accidental when a microseism source, not connected with a cyclone, is in the range of the cyclone. In order to determine whether an active cyclone is the cause of the observed microseisms, the variations in amplitude and periods of

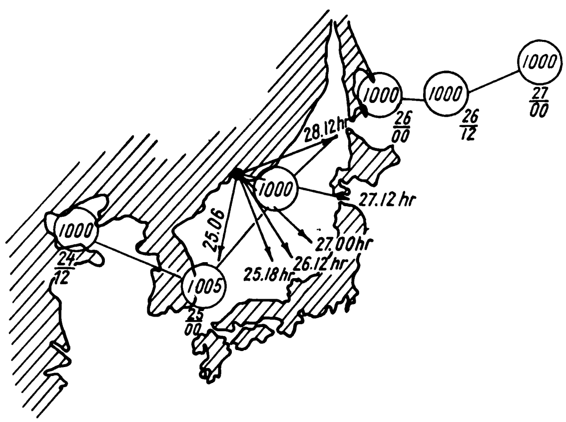

FIG. 11. Trajectory of cyclone and directions towards region of microseism origin.

the microseisms during the development of the cyclone should be analysed. A still more cogent conclusion may be drawn in cases when (with the passage of time) the observed directions of arrival of the microseisms vary in accordance with the cyclonic movements. Figure 11 illustrates such a case.

The most general conclusion to be drawn from the data of Table 3 is that the observed directions of arrival of microseisms to the Vladivostok, Yuzhno-Sakhalinsk and Petropavlovsk stations rarely correspond to cyclone positions. A particularly poor correspondence is observed when the cyclones are far from the shores that are closest to the station. A detailed analysis of individual cases likewise yields an unsatisfactory result, particularly when the cyclones pass at a distance from the recording station. These facts cannot be attributed to measurement errors, since the angle between these direction is often from 90° to 180°.

4.3.4. *The conditions of microseism formation*

In addition to the data on directions of propagation of microseisms, let us consider the microseism intensity as a function of the intensity and position of the cyclones. For this purpose we take advantage of the approximate relation[9] $\left(\frac{A}{\mathrm{T}}\right)^2 = K\frac{L^2}{\mathrm{S}}$, where A and T are the amplitudes and periods of the microseisms, L is the total length of closed isobars in the region of the cyclone, and S is the area covered by the cyclone. To simplify construction and analysis, the values of A, T and L^2/S were plotted on the graphs separately.

The northwestern part of the Pacific Ocean with contiguous seas under consideration is arbitrarily divided into 4 zones (Fig. 7). The graphs (A, T, L^2/S) were constructed for each zone for a number of months in 1957–59. The values of A and T were taken from the bulletins of seismic stations, and L^2/S were calculated on the basis of synoptic maps. Some of the graphs are given in Fig. 12.

Certain peculiarities in the formation of microseisms may be seen from these graphs. First, let us consider examples when only one of the zones exhibited cyclonic activity. In zone I there are the following such cases: 10–12, 18–19 Oct. and 12–13, 21–22 Nov. 1957; 10–11, 24–25 April, 27 Oct. 1958. During each of these periods, with the exception of 27 Oct. 1958, the observations in Petropavlovsk of microseisms were observed to be the strongest. A more detailed analysis shows that if the cyclone is located over the northern part of the Sea of Okhotsk enhancement of the microseisms is detected simultaneously in Petropavlovsk, Magadan and Tixie. When the cold front of the cyclone passes close to the eastern shores of Kamchatka, a microseismic storm is observed only in Petropavlovsk.

Increases in the amplitude of microseisms in Southern Sakhalinsk are, as a rule, associated with the presence of cyclones in zone III. However, it does not always happen that cyclones in zone III give rise to enhanced microseisms in Yuzhno-Sakhalinsk, as was the case, for example, on 13–14 Feb., 19–20

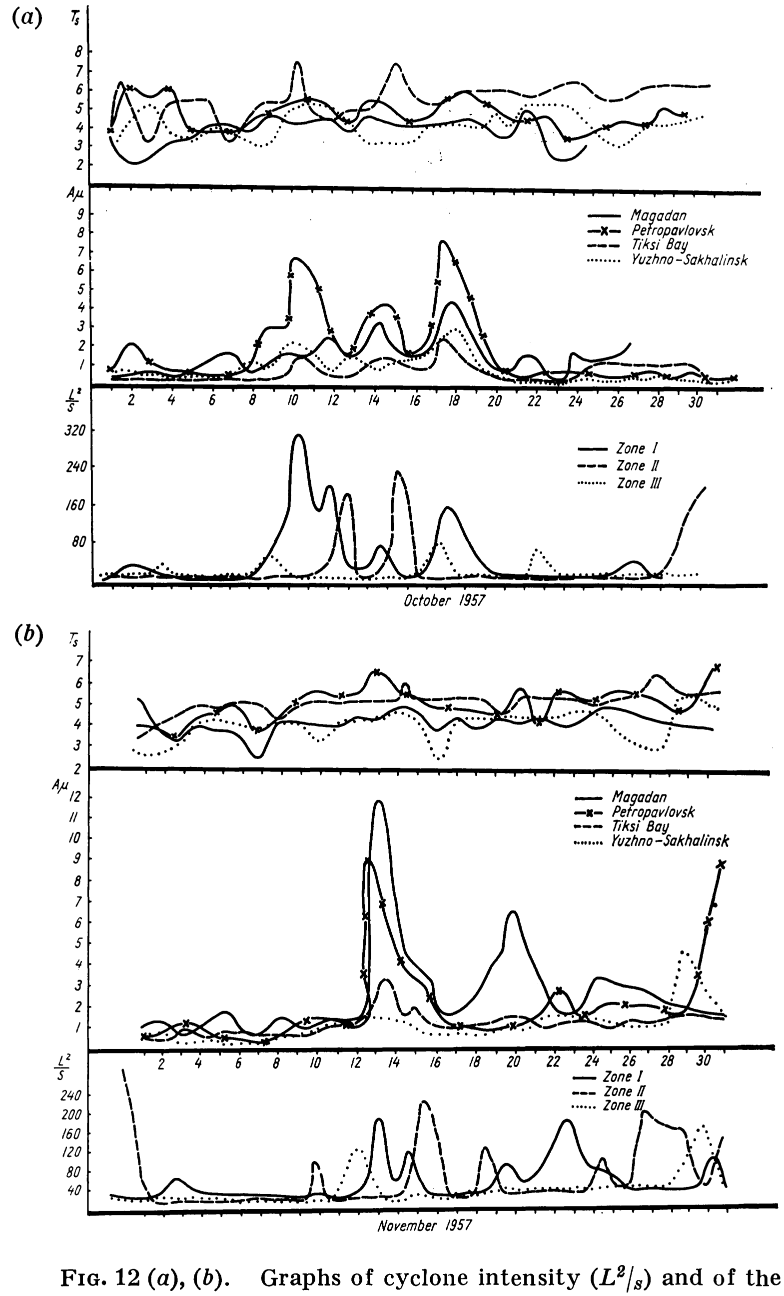

FIG. 12 (*a*), (*b*). Graphs of cyclone intensity (L^2/s) and of the amplitudes A and periods T of microseisms.

April, 1958, and 13–14 April 1959. It usually occurs when the centre and cold front of the cyclone are situated over the eastern part of zone III.

The intensification of microseisms in Vladivostok is usually connected with cyclones in zone IV, as for example, on 1–3 Oct. 1958 and 2–4 April, 1959.

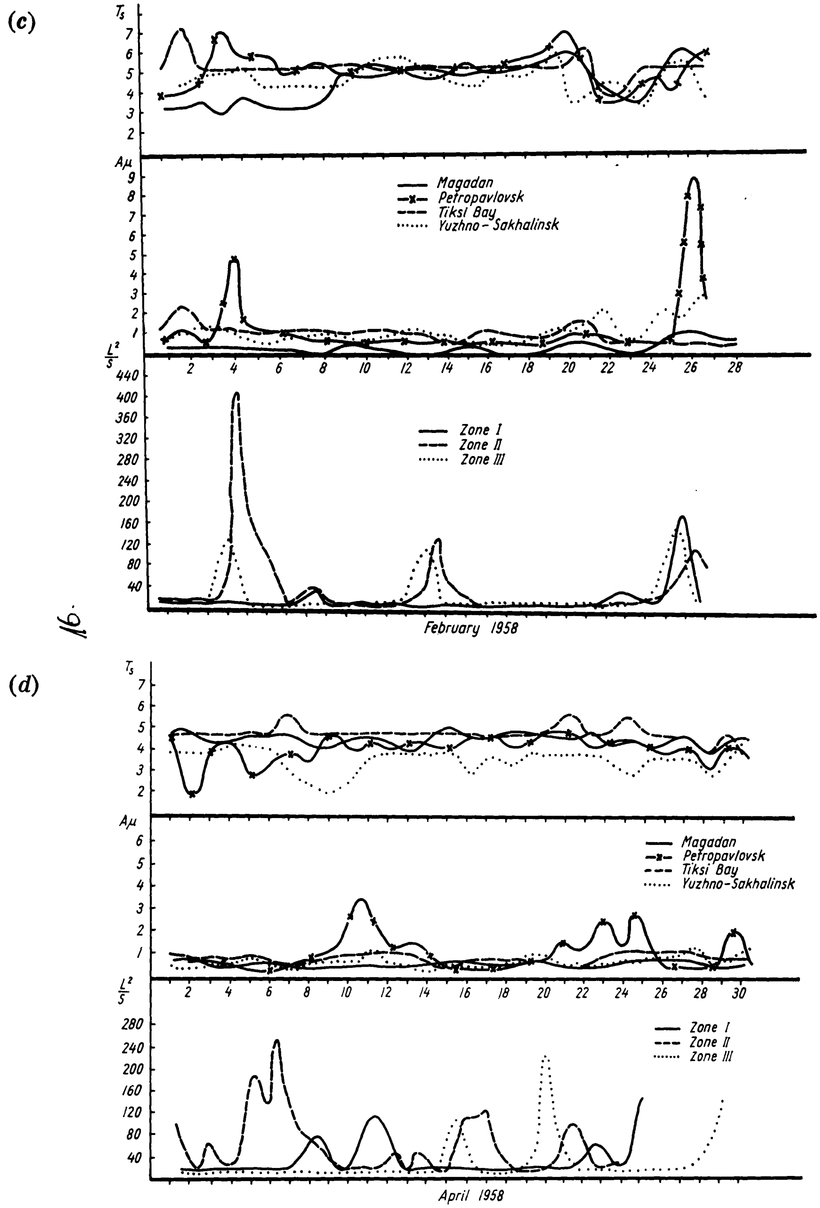

FIG. 12 (*c*), (*d*). Graphs of cyclone intensity ($L^2/_S$) and of the amplitudes A and periods T of microseisms.

Judging by the L^2/S curve, the strongest cyclones occurred over zone II. At the same time, during cyclonic activity in zone II not a single station recorded any amplification of the microseisms. This may be seen from the foregoing graphs for the periods 29 Oct.–1 Nov. 1957, 4–7 April, 1–2, 13–14 and 17–19

(e)

(f)

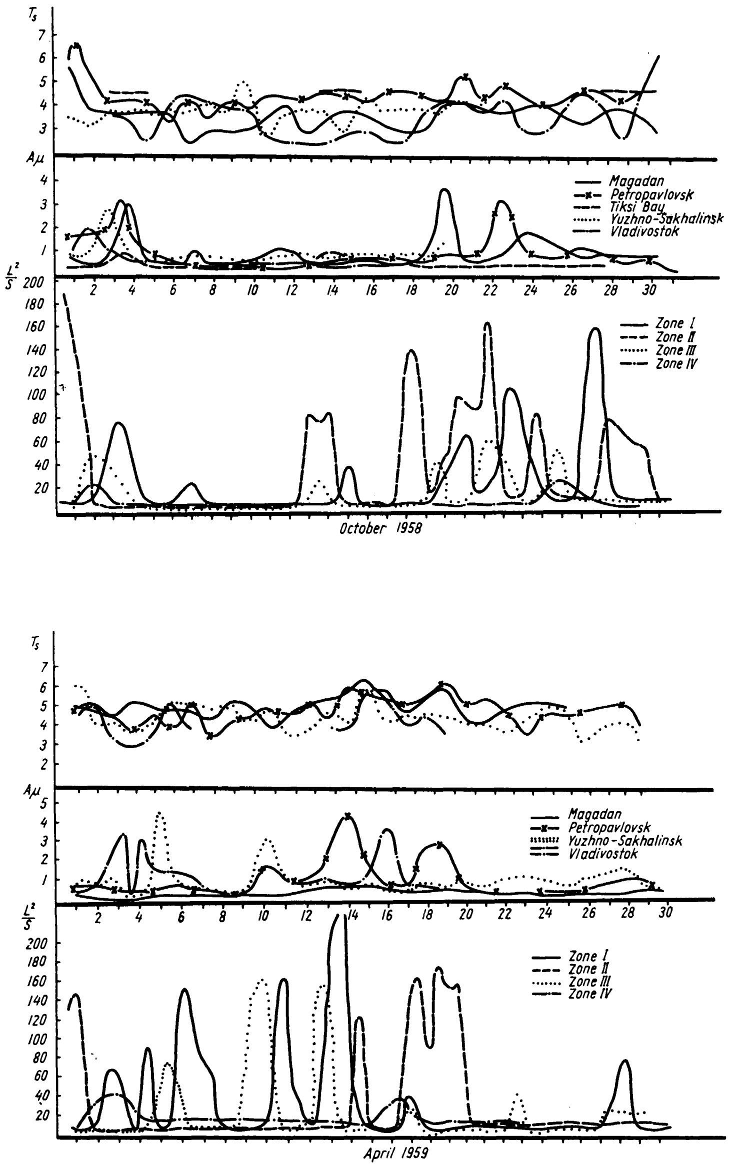

FIG. 12.(*e*), (*f*). Graphs of cyclone intensity (L^2/s) and of the amplitudes A and periods T of microseisms.

Oct. 1958, and 1–2 April 1959. In certain cases, an intensification of microseisms in Petropavlovsk is observed which is connected with cyclonic activity in zone II. This occurs when the cyclones occupy a large area and their periphery extends to the shores of Kamchatka.

(a)

(b)

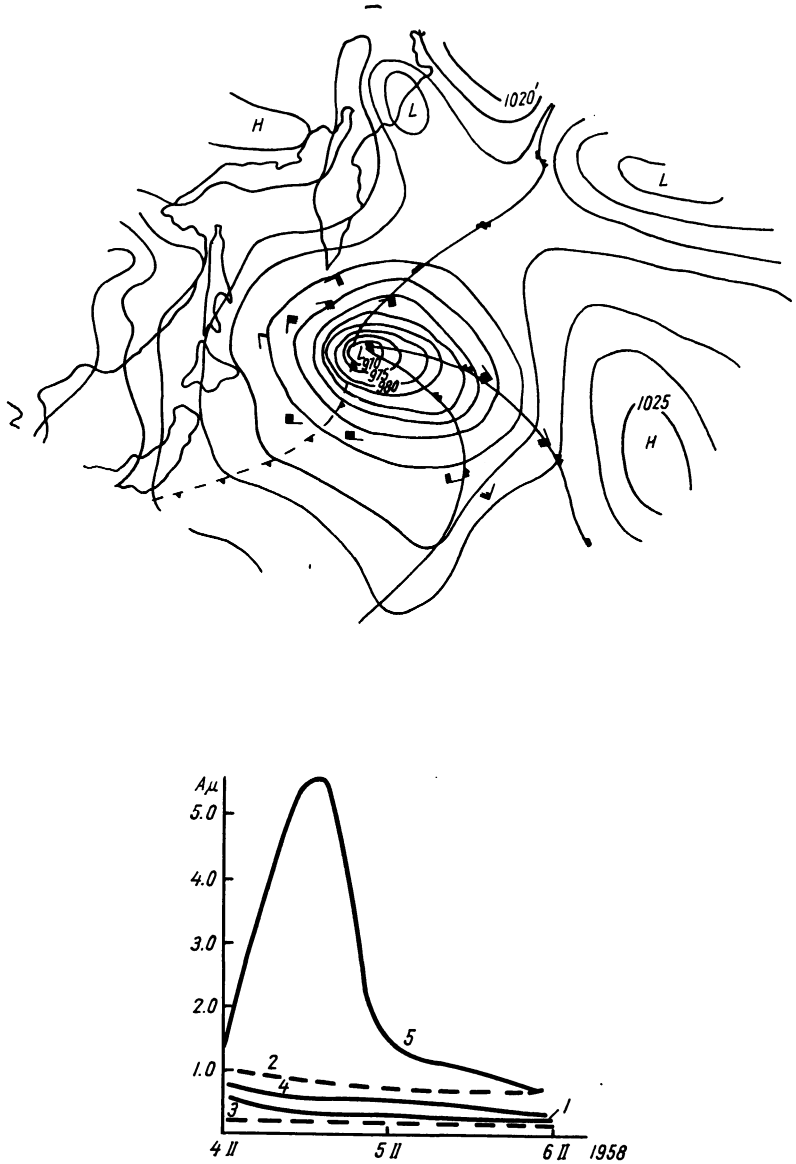

FIG. 13 (a). Position of cyclone at 00 hr. 4 Feb. 1958. (b) Graph of microseism amplitudes: 1 — Irkutsk, 2 — Tixie, 3 — Magadan, 4 — Yuzhno-Sakhalinsk, 5 — Petropavlovsk.

The foregoing consideration of the relationship between intensity of microseisms and the intensity and position of cyclones is rather general. Nevertheless, these data show us convincingly that the microseismic enhancement observed at seismic stations is, as a rule, associated with cyclonic activity near the shores

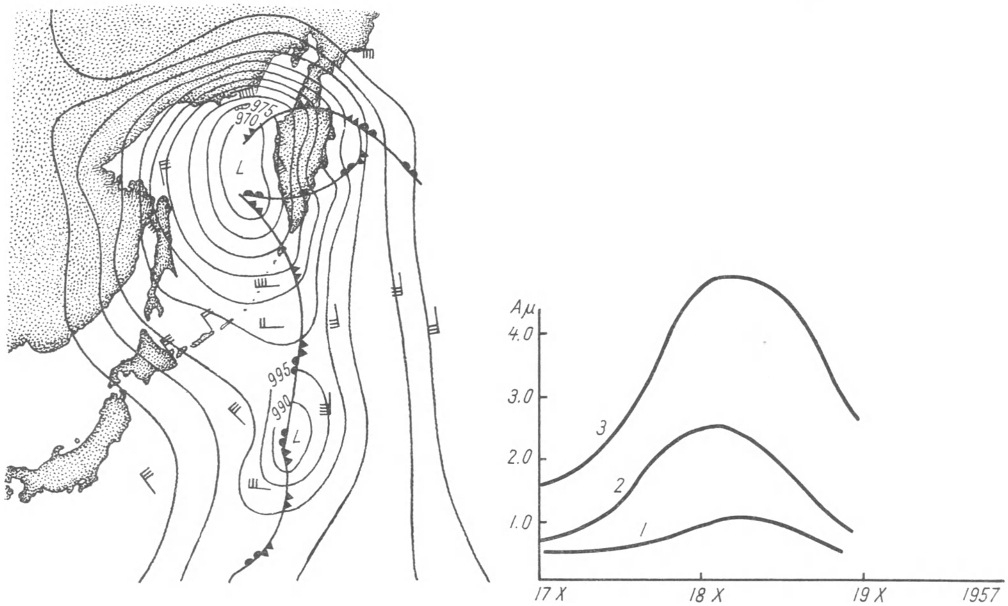

FIG. 14. (*a*) Position of cyclone at 00 hr, 18 Oct. 1957. (*b*) Graph of microseism amplitudes: 1 — Irkutsk, 2 — Tixie, 3 — Magadan.

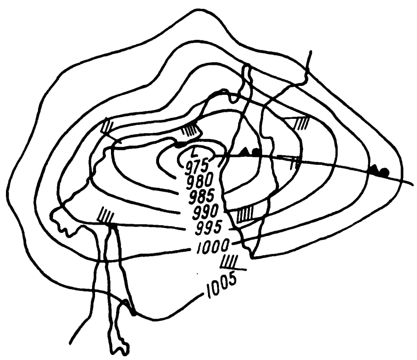

FIG. 15. Position of cyclone at 12 hr, 7 April, 1959.

of that part of the continent on which the stations are located. For a more detailed understanding of this relationship let us consider individual cases.

(1) Figure 13a shows the position of a cyclone at the instant of maximum intensity. During 4 and 5 Feb. the cyclone did not move considerably. It caused an increase in the amplitudes and periods of microseisms in Petropavlovsk, but at the other stations no change in the state of the microseisms occurred (Fig. 13b). It is interesting to compare this case with another case that occurred on 17–20 Oct. 1957. At this time, a cyclone with an intensity lower than that of the

preceding one was located over the northern part of the Sea of Okhotsk (Fig. 14a). It gave rise to microseismic enhancement not only at the closest stations but at the relatively distant stations, too (Fig. 14b). From a comparison of these two examples we can draw the conclusion that microseismic waves propagated along the ocean bottom are damped to a greater extent than those propagated along continental paths.

(2) Figure 15 shows the position of a cyclone in zone II at 00 hr. on 15 March 1959. The cyclone had then reached maximum development, and it remained in this place. The characteristic peculiarity of this case is that in Yuzhno-Sakhalinsk and Petropavlovsk the alteration in the character of the microseisms (brought about by the presence of the cyclone under consideration) took place only 24 hours after the cyclone had reached maximum intensity. On 26 March both stations began to observe weak but long-period microseisms. Such a considerable lag of the microseisms excludes any suggestion of their being formed in the region of the cyclone itself. If we assume that this microseism formation was connected with the approach of a swell to the shores of Kamchatka and Sakhalin, then both the lag in the microseisms and their increased period are satisfactorily explained.

(3) On 7–8 April 1959, a deep cyclone became stationary over the northern part of the Sea of Okhotsk (Fig. 15). However, in contrast to the above-considered and analogous case of 17–20 Oct. 1957, this time not a single station recorded a perceptible increase in the microseisms (Fig. 12). A probable reason is the presence, in the northern part of the Sea of Okhotsk, of an ice cover which in March attains maximum extent.

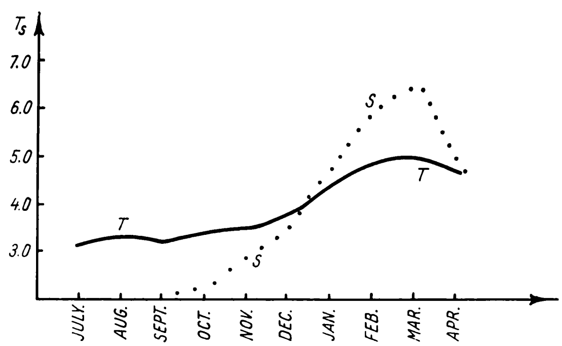

FIG. 16. Graph of the variations in the microseismic period T in Magadan as a function of the relative distance S from the ice edge.

Theoretically, microseism formation in regions of ice cover is impossible. This is illustrated by the foregoing example and also by other data. Figure 16 is a graph of the seasonal variations of microseism periods in Magadan; it is compared with the seasonal position of the ice edge in the northern part of the

Sea of Okhotsk. Table 4 gives the statistical data on microseism periods in Magadan and Uglegorsk during 1956–57.[9] From the graph and the table, it follows that as the edge of the ice recedes from the shores seawards, the prevailing microseism periods increase. This is undoubtedly due to the fact that the ice cover in the most shallow littoral strip excludes the possibility of the formation of seisms in this zone.

TABLE 4

Station	Autumn		Winter	
	$T<4$ sec.	$T>4$ sec.	$T<4$ sec.	$T>4$ sec.
Magadan	48%	52%	0%	100%
Uglegorsk	45%	55%	20%	80%

T. A. ANDREYEV and V. G. BUKHTEYEV[10] have correlated the microseism intensity in Kurilsk and Petropavlovsk with the heights of sea waves near the shores. The results show that, on the whole, the phenomena can be correlated, but that complete correspondence is lacking. It may be that the microseisms do not originate only in the immediate vicinity of the coast; yet it is possible that the incomplete correspondence between the height of the sea waves and the amplitudes of microseisms can be explained by the fact that the analysis of the observations is not sufficiently detailed.

4.3.5. *Conclusions*

During the period of the International Geophysical Year, extensive observational material of microseisms was obtained in the region of the northwestern part of the Pacific Ocean. This material has not yet been subjected to detailed analysis, but the studies already carried out permit us to formulate the following conclusions:

(1) Microseismic oscillations recorded by seismic stations are formed in littoral zones. Sea waves are the immediate causative agent of microseisms.

(2) During cyclonic action over the northwestern part of the Pacific Ocean, seismic stations in the Far East do not observe intensification of microseismic activity. The most probable reason for this is the considerable damping of microseisms during propagation along the bottom of the sea.

(3) The great scattering in the directions of propagation of microseisms, based on observations of seismic stations located close to the coastline, is due to the proximity of the region of origin of the microseisms, in accordance with item 1 of this résumé.

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4.4. Microseismic observations in India during the IGY

BY A. N, TANDON

In India, observations of microseisms are carried out mainly by the India Meteorological Department of the Government of India. This department maintained microseismograph stations at Bombay, Madras, Shillong and Port Blair (Andaman Islands) during the IGY. The three former stations were equipped with Sprengnether microseismographs, using recording drums with a speed of 30 mm per min. One Sprengnether microseismograph was also installed at Port Blair but its operation had to be discontinued after some time due to successive galvanometer failures caused by climatic conditions. The microseismic observations were continued at that station with the help of a Milne–Shaw Seismograph. The constants of the instruments operating at these observatories are given in Table 1. All the Sprengnether instruments were regularly standardized according to the method employed by GILMORE of the U.S. Navy. In each case the periods of the seismometer and of the galvanometer were kept equal at about 7 seconds. The galvanometer was always kept critically damped and the seismometer damping was so adjusted that in the operating condition

*In Russian

a sudden impulse of the seismometer boom gave a deflection to the galvanometer such that the ratio of the first two consecutive deflections (φ_1/φ_2) was $2{\cdot}294 \pm 0{\cdot}15$.

TABLE 1

S. No.	Station	Type of Seismograph	Galv. period	Galv. Damp.	Seismo. period	Seismo. damping (φ_1/φ_2)	Drum speed mm/min.	Peak Magnification
1.	Bombay	Sprengnether	7 sec	Critical	7 sec.	2·2	30	5000
2.	Madras	Sprengnether	7·5 sec	-do-	7·5 sec.	2·35	30	5000
3.	Port Blair	Sprengnether	7·3 sec	-do-	7·3 sec.	2·3	30	3000
		Milne–Shaw	—	—	12 sec.	20·1	8	250
4.	Shillong	Sprengnether	6·6 sec	-do-	6·6 sec.	2·2	30	4000

In addition to the above stations microseismic observations were also recorded during the IGY by the Geophysical Observatory of the Bengal Engineering College, Howrah, and the Naval Physical Observatory, Cochin. The former maintained a set of three component Benioff Seismographs and the latter a three component electromagnetic Seismograph with electronic magnification and pen recording system.

Microseismic data giving the average amplitude and period at the Synoptic hours 0000, 0600, 1200 and 1800 hrs GMT were tabulated for all the six stations mentioned above and their monthly tabulations were sent to the World Data Centre "C" at Strasbourg. For specified "World Days" the tabulations were done hourly. The observations were also continued during the IGC. The amplitudes and periods were measured for five largest regular groups within 10 minutes of the hour of observation and the average taken.

The day to day amplitudes and periods of microseisms for the stations Bombay, Shillong, Madras and Port Blair have been plotted and are given in Figs. 1 to 36 for the whole of the IGY period. The amplitudes for all the stations have been plotted in millimeters as read from the seismograms except for Port Blair, for the period 30$^{\text{th}}$ August 1957 to 22$^{\text{nd}}$ July 1958, when they were measured from the records of a Milne–Shaw Seismograph. For this period the ground motion is given in microns. While plotting the amplitudes in millimeters, they were all reduced to a standard magnification of 5000. Out of the set of observations a special study has been made in the following pages of microseisms which were recorded at the stations mentioned above (functioning under the India Meteorological Department) for those occasions when a depression or a cyclonic storm existed in the Indian Seas.

During the IGY period, 17 depressions and storms formed in the Bay of Bengal and the Arabian Sea, tracks of which are shown in Fig. 53. Brief descriptions of the history of thirteen of these depressions and cyclones along with the microseismic activity observed at these four stations is given below. No microseisms were recorded for the remaining four depressions at any of the stations. Positions of centres of storms and brief description of the synoptic situation are also indicated in Figs. 17–52.

4.4.1. *7–12 August, 1957*

According to the synoptic situation, weather became unsettled in the Northwest Bay of Bengal on the 8th August, and a shallow depression was formed which crossed coast before the morning of the 9th when it lay as a low pressure area over Orissa and adjoining Madhya Pradesh. The microseismic amplitudes started rising rather steeply at Madras from the morning of the 7th reaching a peak at 1200 GMT of 8th August after which they started declining. Microseismic amplitudes started rising at Bombay even before the 7th. They reached a peak about the 8th morning, and then started decreasing gradually. There was no significant rise in the amplitudes recorded at Shillong. The period of microseisms rose slightly as the amplitudes increased at all the stations. The peak of microseismic activity was recorded at Madras at about the time the storm was crossing the coast.

4.4.2. *19–28 August, 1957*

Unsettled conditions at the head of the Bay of Bengal concentrated into a depression which had its centre at 20·5° N and 90·5° E on the morning of the 19th. The depression started moving in a westerly direction, intensified into a cyclonic storm and according to weather reports crossed the coast on the morning of the 22nd. The microseismic amplitudes at Madras and Shillong started rising steeply from about 1200 hr GMT of 19th August, reaching their maximum at 1200 hr GMT on 21st and thereafter decreased gradually, becoming normal after several days. The amplitudes at Bombay also followed a similar pattern except that the increase in amplitudes started about 12 hr earlier. It will be seen from the graph of amplitudes and periods that the maxima of the amplitudes at all the stations were recorded at about 1200 hr GMT on 21st while according to weather reports, the storm crossed the coast on the morning of the 22nd. This would mean that before crossing the coast, the storm had already started weakening. Periods of microseisms showed an increase with amplitudes at all the three stations, more conspicuously at Madras.

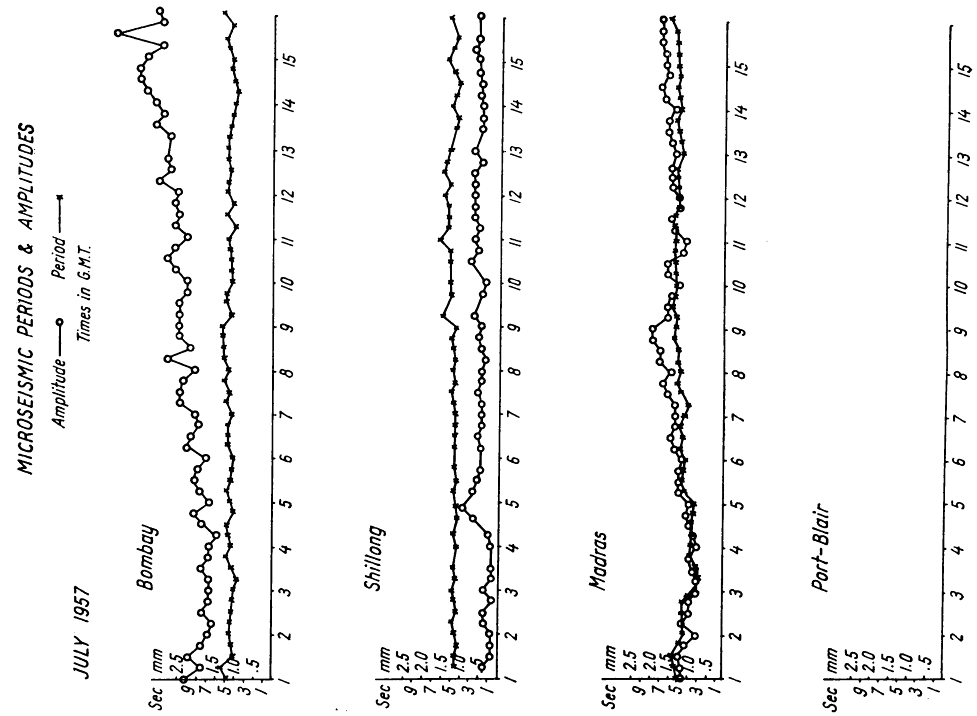

MICROSEISMIC PERIODS & AMPLITUDES
Amplitude
Period
Times in G.M.T.
JULY 1957
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 17.

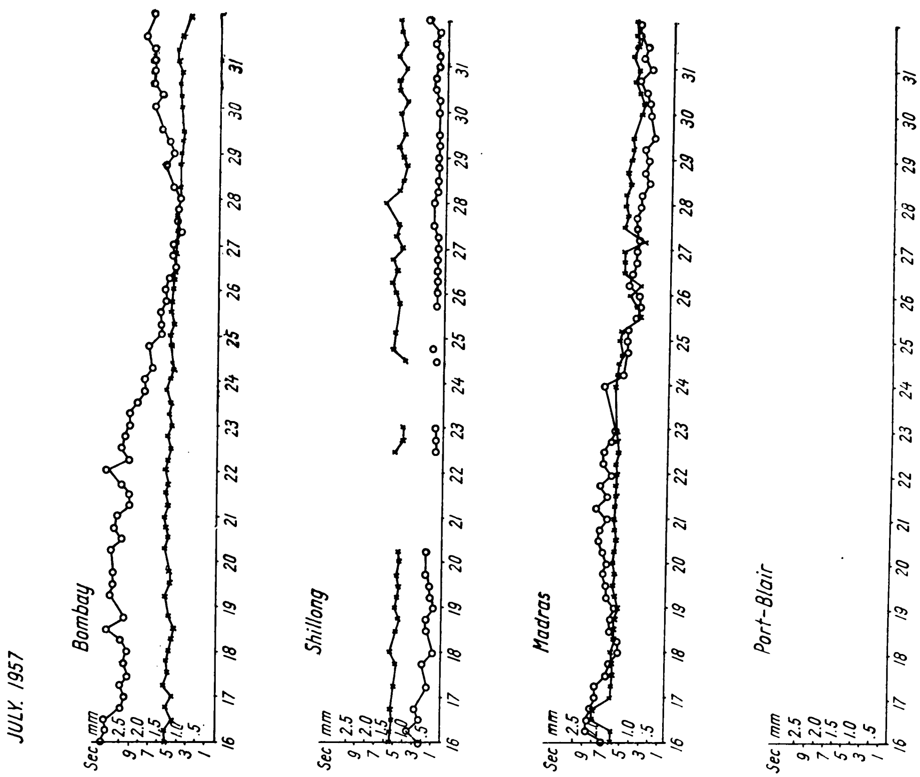

JULY 1957
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 18

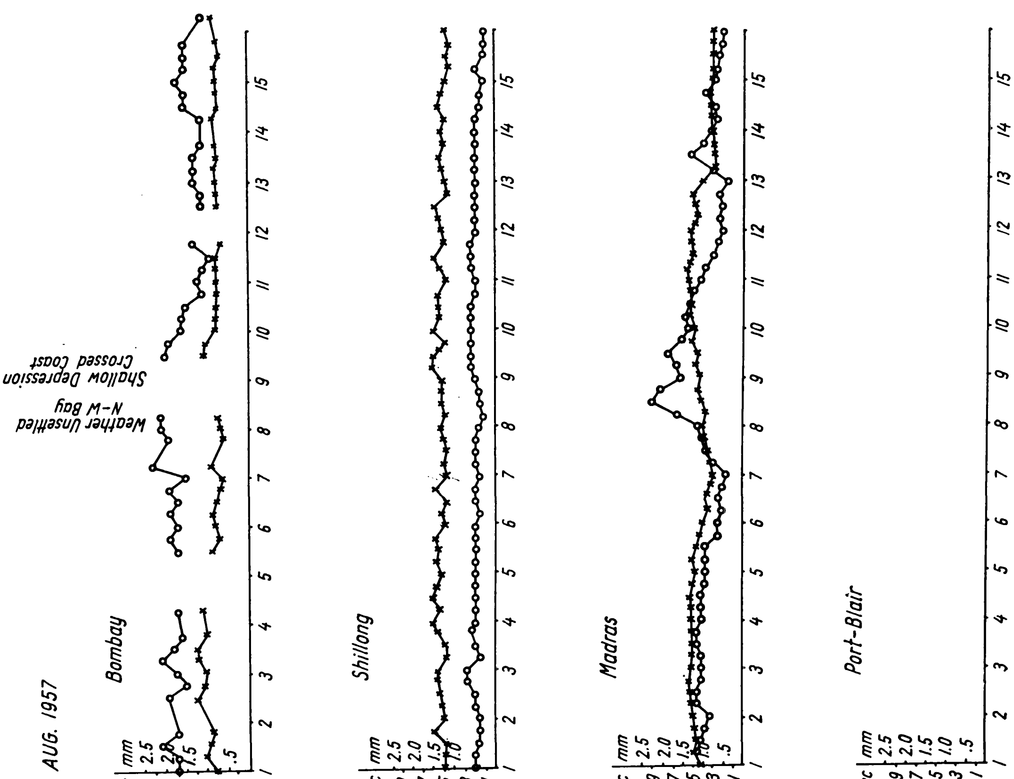

AUG. 1957
Bombay
Shillong
Madras
Port-Blair
Sec
mm
Weather Unsettled N-W Bay
Shallow Depression Crossed Coast

FIG. 19.

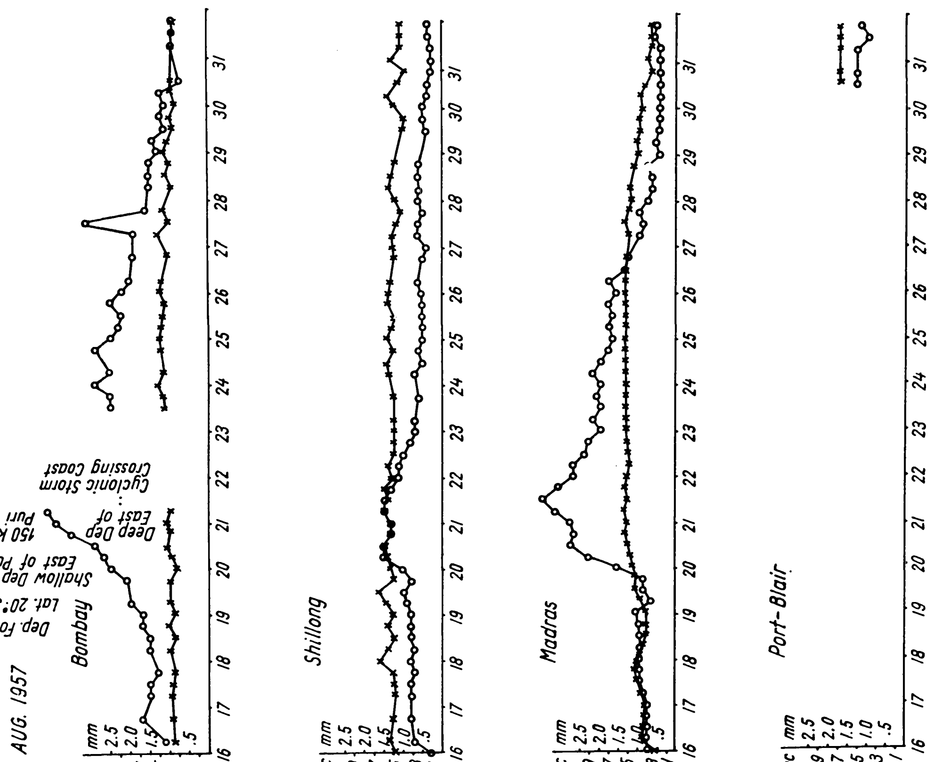

AUG. 1957
Bombay
Shillong
Madras
Port-Blair
Sec
mm
Dep. Forming Lat. 20°5N, 90°5E
Shallow Dep 300 km s East of Puri
150 k Puri
Deep Dep East of
Cyclonic Storm Crossing Coast

FIG. 20.

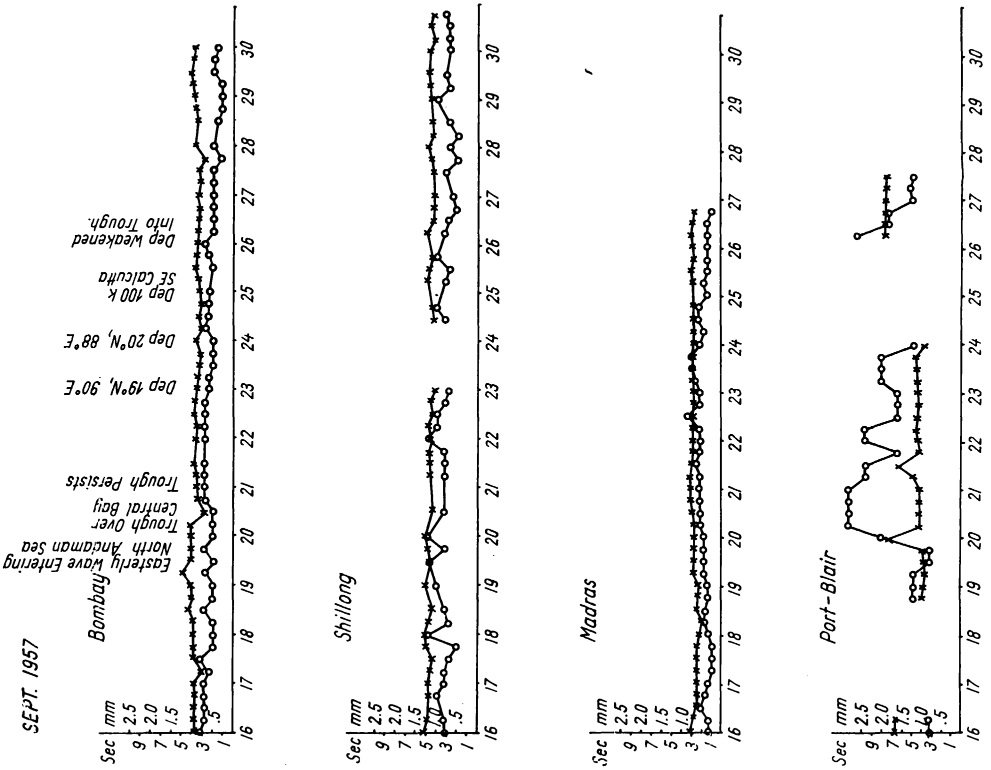

SEPT. 1957
Bombay
Shillong
Madras
Port-Blair
Sec
mm
Easterly Wave Entering North Andaman Sea
Trough Over Central Bay
Trough Persists
Dep 19°N, 90°E
Dep 20°N, 88°E
Dep 100 k SE Calcutta
Dep Weakened into Trough.

FIG. 22.

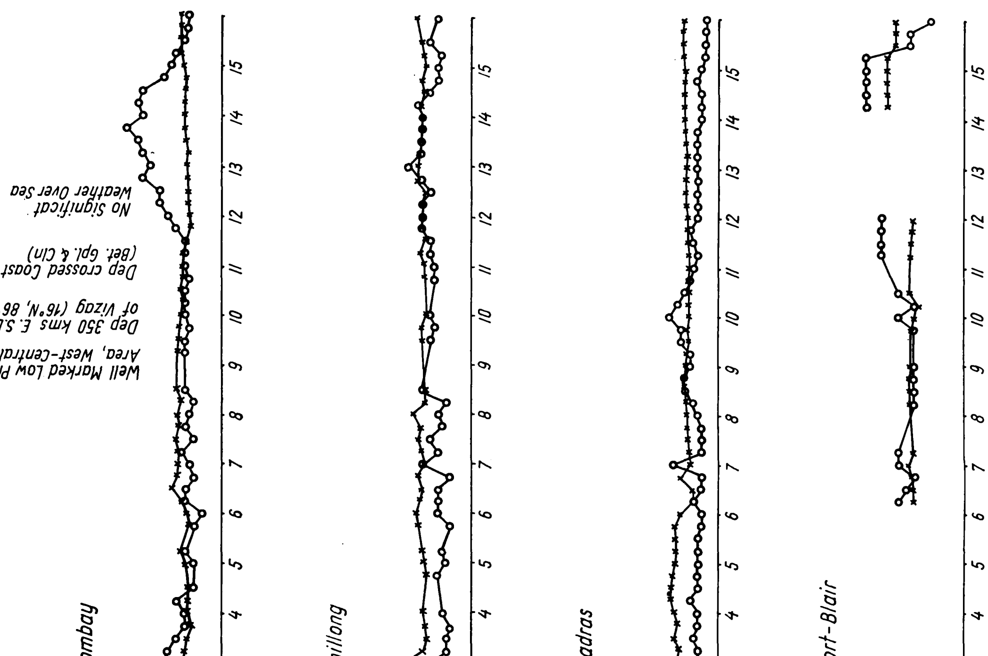

SEPT. 1957
Bombay
Shillong
Madras
Port-Blair
Sec
mm
Well Marked Low Pressure Area, West-Central BAY
Dep 350 kms E.S.E of Vizag (16°N, 86°E)
Dep crossed Coast (Bet. Gpl. & Cln)
No Significant Weather Over Sea

FIG 21.

4.4.3. *8—13 September 1957*

On the 9th of September, a well marked low pressure area was observed in the west central Bay of Bengal. On the 10th it became a depression with its centre near latitude 16° N 86° E. This depression crossed the coast between Gopalpur and Kalingapatam in the early hours of the 11th morning. There was no significant rise of amplitudes at Bombay and Shillong. Madras seismograms showed a slight increase in the amplitude of microseisms which reached its peak on the 10th morning and thereafter started declining and becoming normal on the 11th. The periods at all the stations remained nearly steady. It will be seen that the maximum was reached about 24 hours before the storm crossed the coast showing that the depression must have started weakening much before it reached the coast. The amplitudes of microseisms show a significant rise from 1200 hr GMT of 11th to 1400 hr GMT of 13th at Bombay alone in spite of the fact that there was no significant weather either over the Arabian Sea or the Bay of Bengal. During this period, the amplitude of microseisms at Madras remained almost steady and at Shillong only a slight increase was noticed. This phenomenon could only be attributed to some unusual local activity.

4.4.4. *19—25 September, 1957*

An easterly wave moving across the North Andaman Sea on the 19th formed a depression with its centre near 20° N, 88° E on 23rd night. On the morning of the 25th this depression had its centre about 100 km south-east of Calcutta and on the 26th it weakened into a trough of low pressure. The microseismic amplitudes and periods at Madras, Bombay and Shillong do not show any significant change during this interval. The amplitudes at Port Blair however, show a sudden rise at about 1800 hr GMT on the 19th, reaching a peak at 1200 hr GMT on the 20th and then declined.

4.4.5. *23—27 October, 1957*

A depression was formed on the 22nd with its centre near 12° N 72° E in the Arabian Sea. The depression travelled in a N/NNW direction and weakened into a trough of low pressure by the 26th. The microseismic amplitudes at Madras and Shillong do not show any significant rise. The amplitudes at Bombay started rising slowly from 12hr GMT of the 23rd, reaching a peak at about 1800 hr GMT of the 25th and then declined. The periods of microseisms at Bombay also show a slight rise between the 24th and 25th.

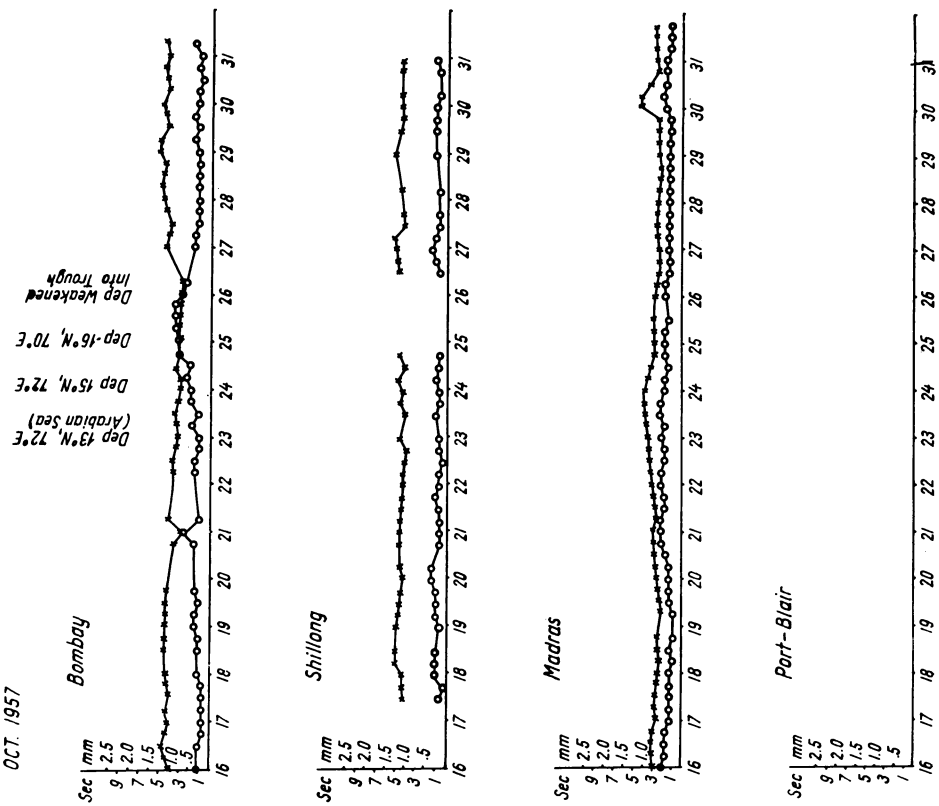

OCT. 1957
Bombay
Shillong
Madras
Port-Blair
Sec
mm
Dep 13°N, 72°E (Arabian Sea)
Dep 15°N, 72°E
Dep-16°N, 70°E
Dep Weakened Into Trough

FIG. 24.

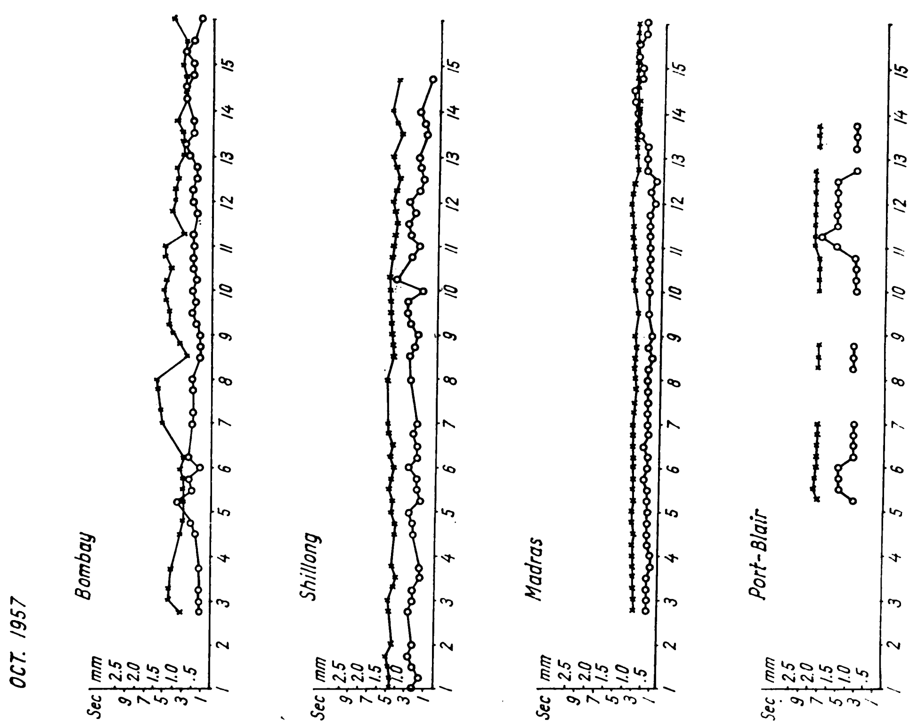

OCT. 1957
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 23.

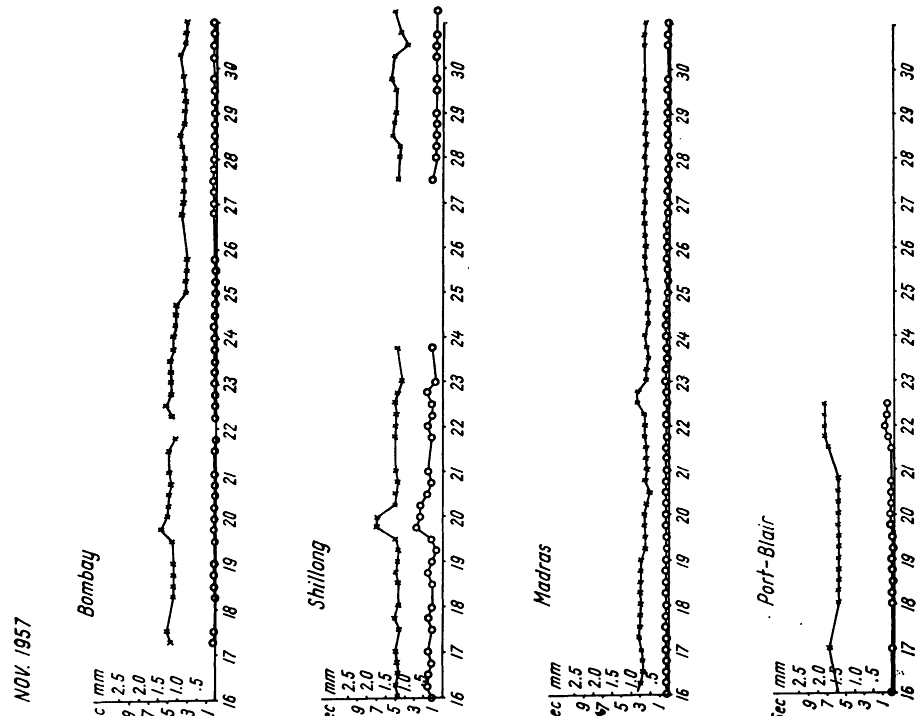

NOV. 1957
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 26.

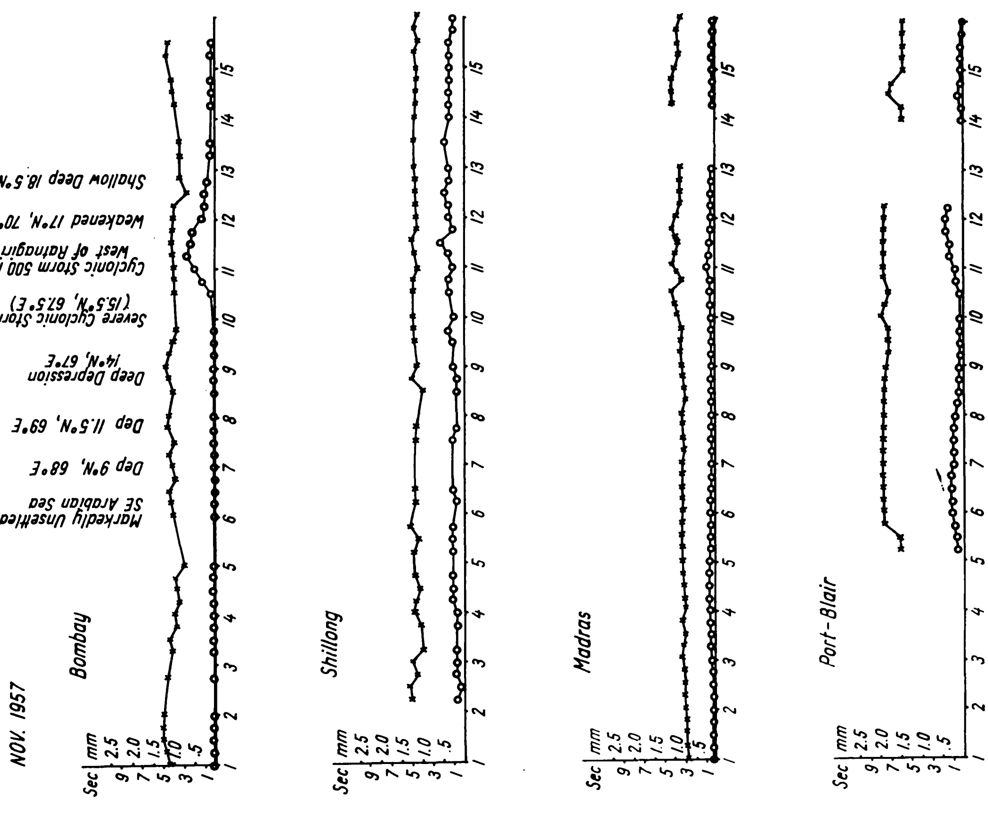

NOV. 1957
Bombay
Shillong
Madras
Port-Blair
Sec
mm
Markedly Unsettled SE Arabian Sea
Dep 9°N, 68°E
Dep 11.5°N, 69°E
Deep Depression 14°N, 67°E
Severe Cyclonic Storm (15.5°N, 67.5°E)
Cyclonic Storm 500 kms West of Ratnagiri
Weakened 17°N, 70°E
Shallow Deep 18.5°N, 70°E

FIG. 25.

4.4.6. *10—13 November, 1957*

A depression formed in the Arabian Sea on the 7th with its centre near 9° N and 68° E. It travelled in a north-westerly direction, intensifying into a deep depression on the 9th. By the 10th morning, it further intensified and became a severe cyclonic storm with its centre at 15·5° N and 67·5° E. It travelled in a north-easterly direction and its centre on the 11th was about 500 km west-south-west of Ratnagiri. From the 12th onwards, the storm weakened and travelled as a depression in a northerly direction and finally weakened into a trough of low pressure by the 14th. The microseismic amplitudes at Madras and Shillong show a very slight rise just before the 11th. The amplitudes at Bombay and Port Blair show a significant rise from 1200 hr GMT of the 10th. At Bombay the maximum was reached at 0600 hr GMT of 11th and at Port Blair at 1800 hr GMT of the 11th. The amplitudes at these stations also returned to normal soon after the 13th. The period of the microseisms at Port Blair remained almost steady and at Bombay showed a very slight rise. It is significant to note that the amplitudes of microseisms at Bombay remained small until the 10th even after the depression had concentrated into a severe cyclonic storm.

4.4.7. *25–28 December, 1957*

Conditions became markedly unsettled in the extreme south-west Bay of Bengal on the 24th. On the morning of the 25th, a depression was formed with its centre near 7·5° N and 82° E. The depression concentrated into a cyclonic storm on the 26th with its centre about 200 km east of Pamban and finally became a trough of low pressure on the 28th. The amplitudes of microseisms at Port Blair and Madras were almost steady except for a very slight rise at Madras on the 25th which reached its peak at about 0600 hr GMT of the 26th and then declined gradually. It is curious to note that the amplitudes at Shillong registered a significant increase at 1800 hr GMT on the 24th. After registering a peak at 0600 hr GMT on the 25th they declined to their normal value by the morning of the 26th. The rise in amplitudes at Shillong apparently does not correlate with the presence of this cyclonic storm on the 26th.

4.4.8. *17–19 May, 1958*

A trough of low pressure in the central Bay of Bengal concentrated into a depression during the night of the 16th. It further intensified into a cyclonic storm of small extent with its centre on the morning of the 17th at 18½° N and 88° E. The storm moved in a north easterly direction and was centred on the morning of the 18th at 20½° N and 89½° E. It crossed the coast near Sunderbans on the night of the 18th. The amplitudes of microseisms started rising slightly at Madras and Shillong from the morning of the 17th. From 1800 hr

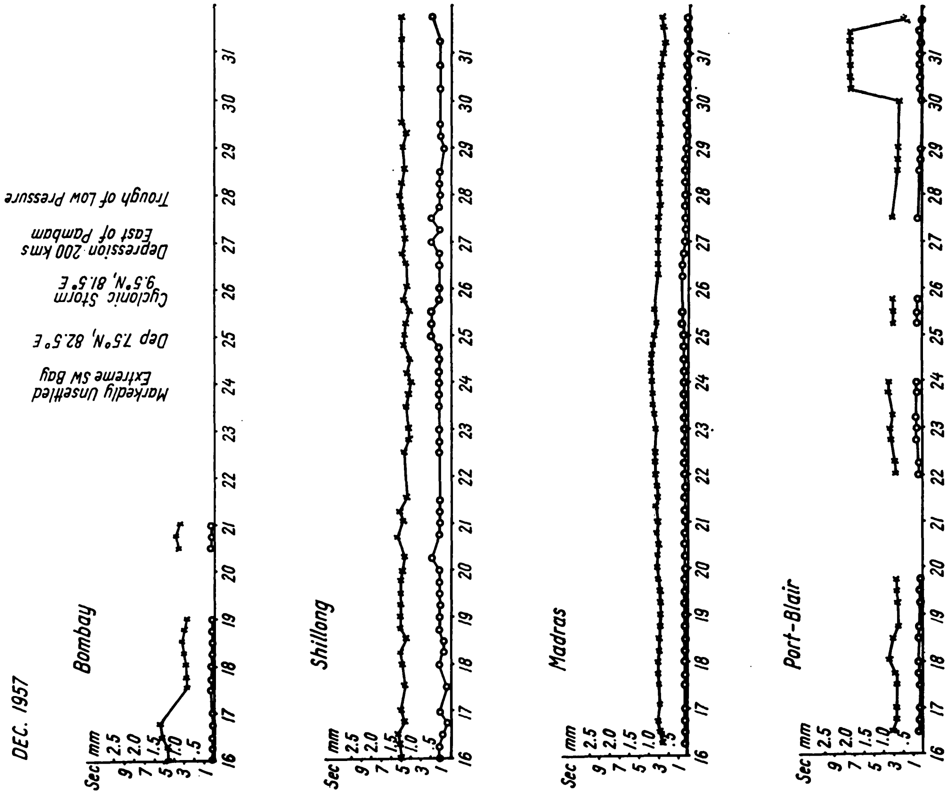

DEC. 1957
Bombay
Shillong
Madras
Port-Blair
Sec
mm
Markedly Unsettled Extreme SW Bay
Dep 7.5°N, 82.5°E
Cyclonic Storm 9.5°N, 81.5°E
Depression 200 kms East of Pambam
Trough of Low Pressure

FIG. 28.

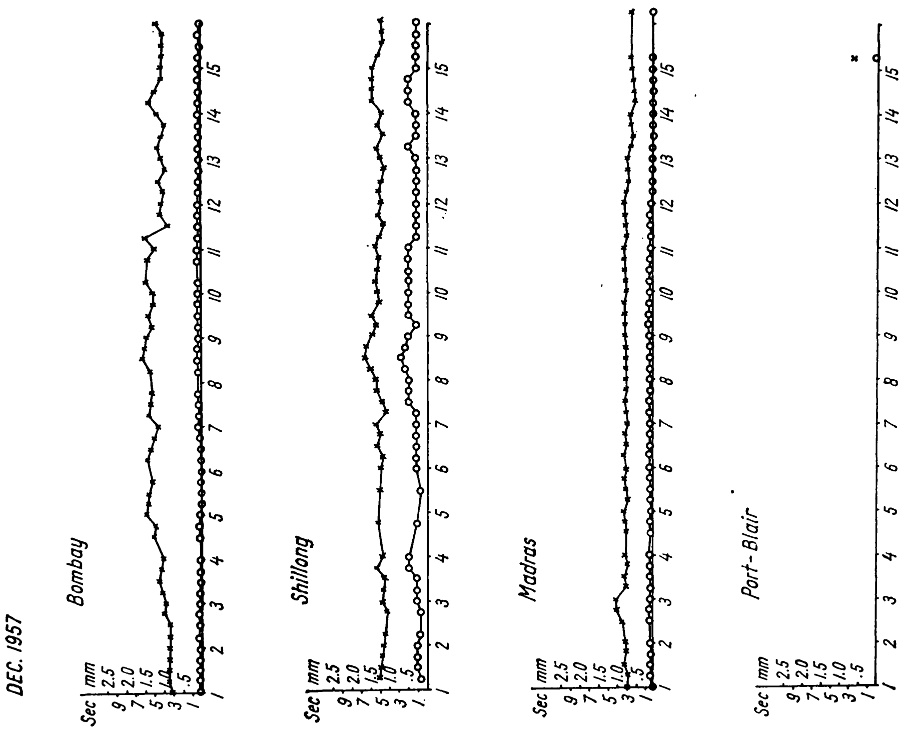
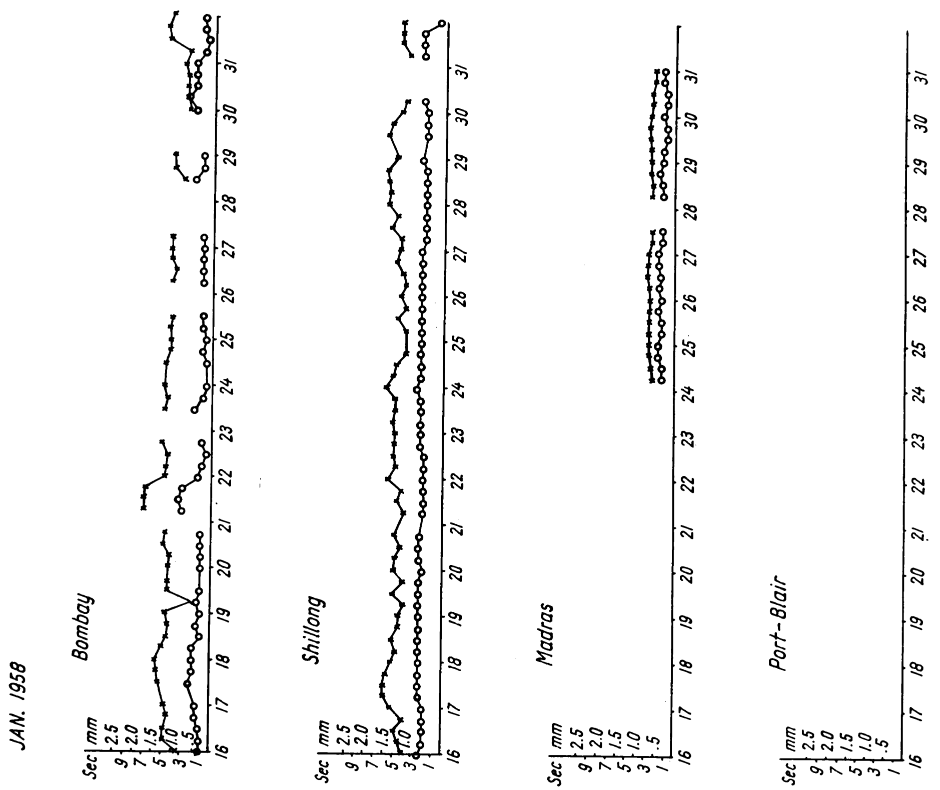
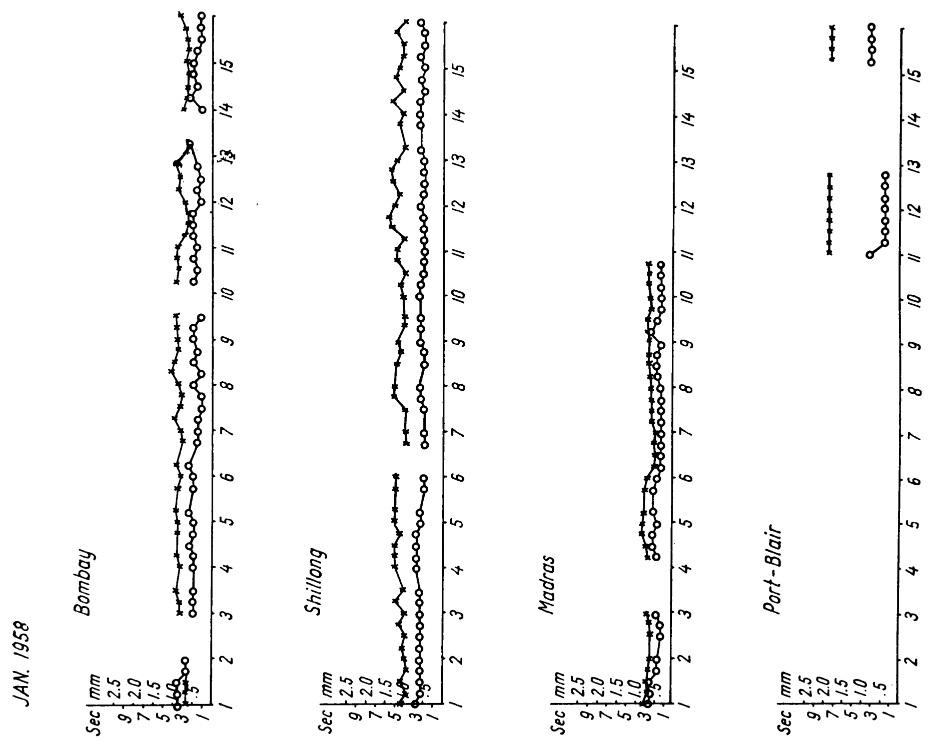

DEC. 1957
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 27.



JAN. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 30.



JAN. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 29.

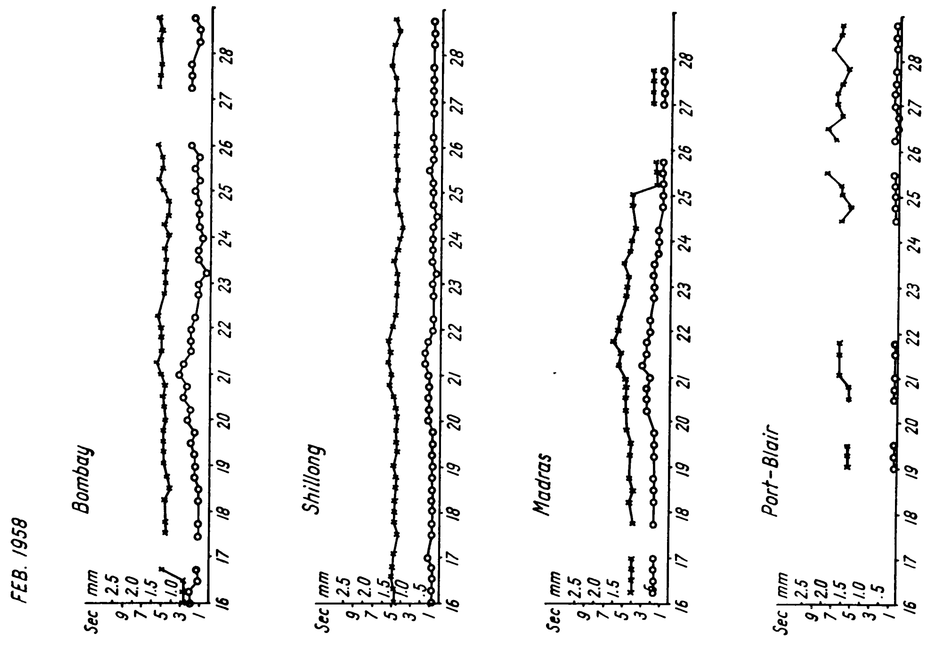

FEB. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 32.

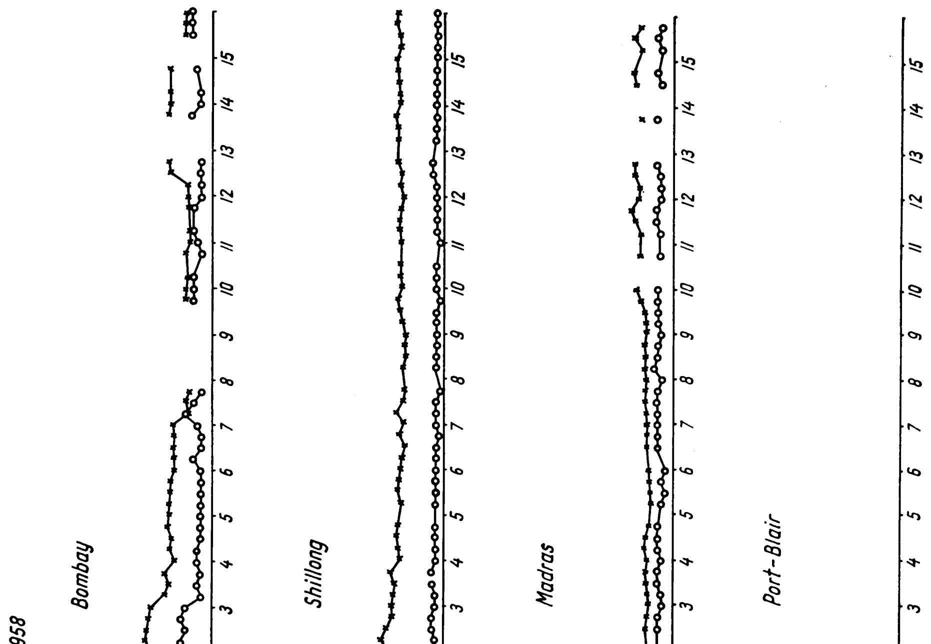

FEB. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 31.

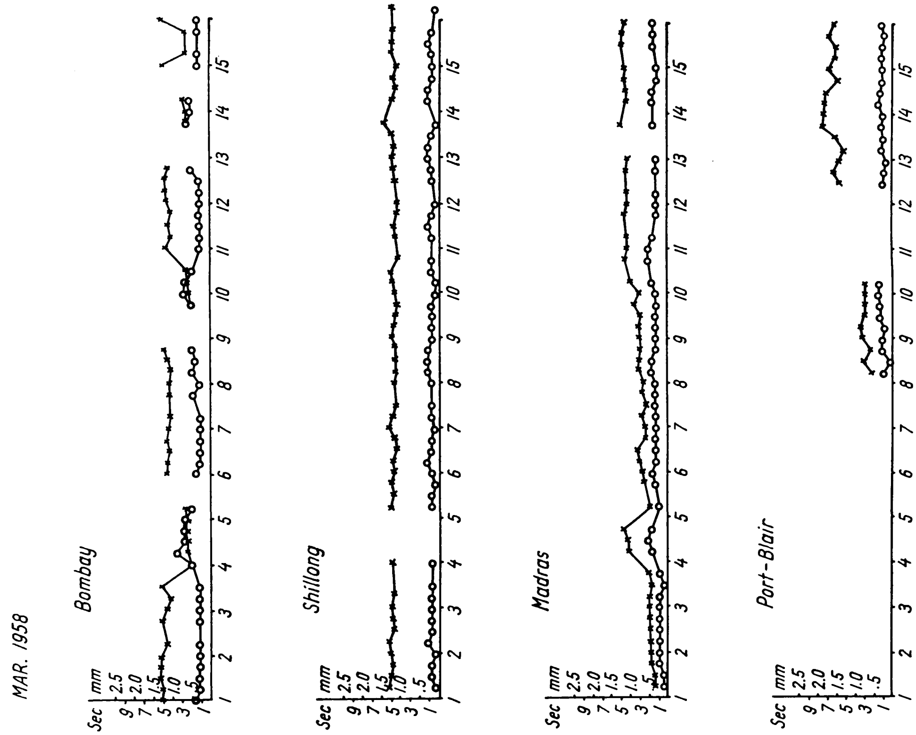

MAR. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 33.

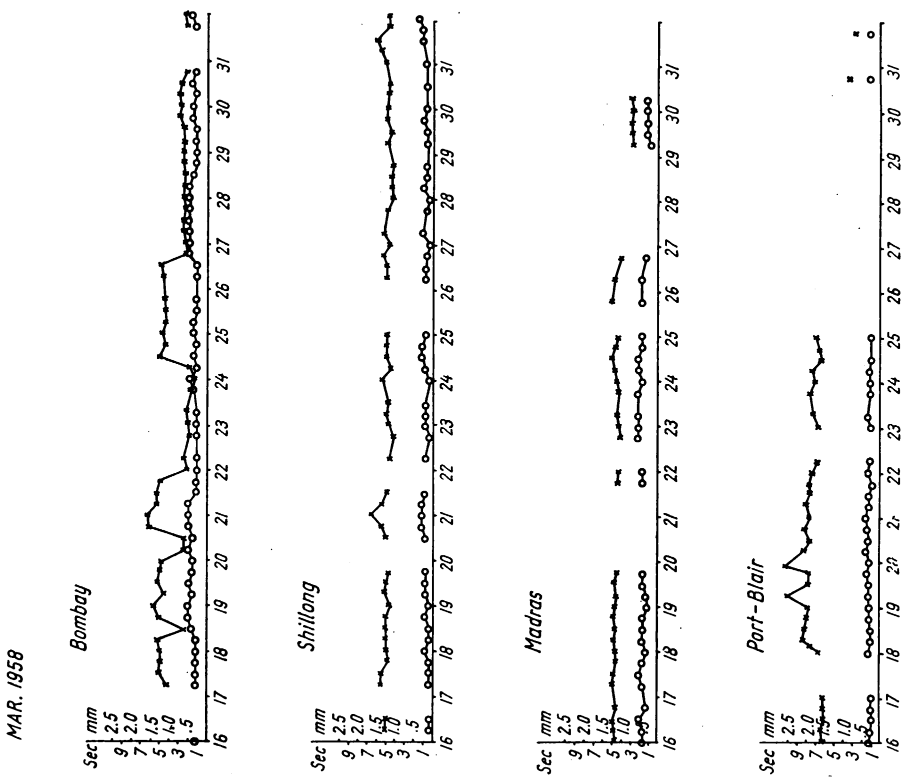

MAR. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 34.

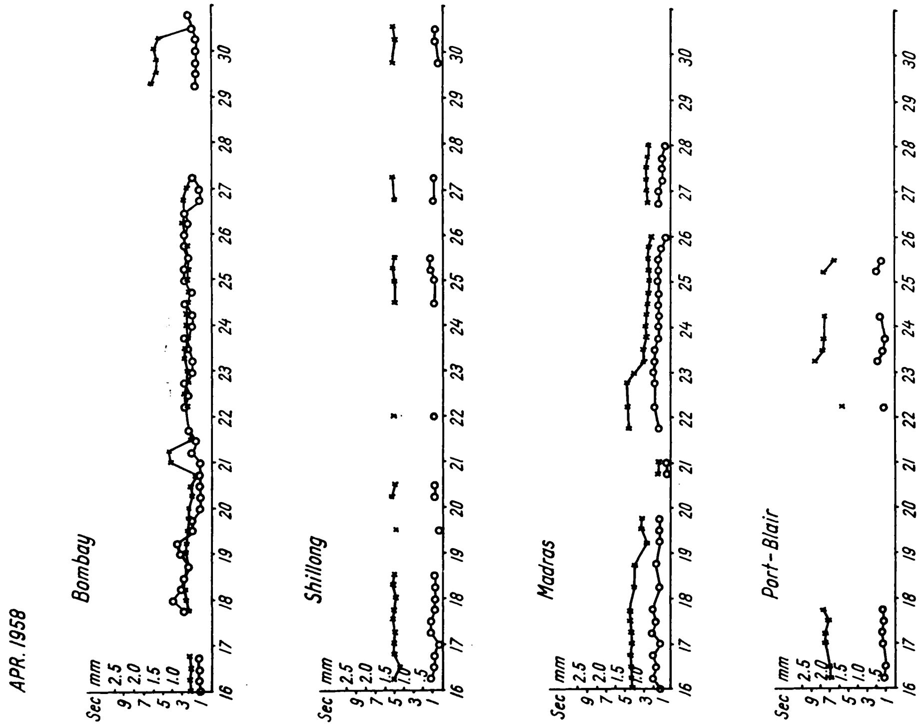

APR. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 36.

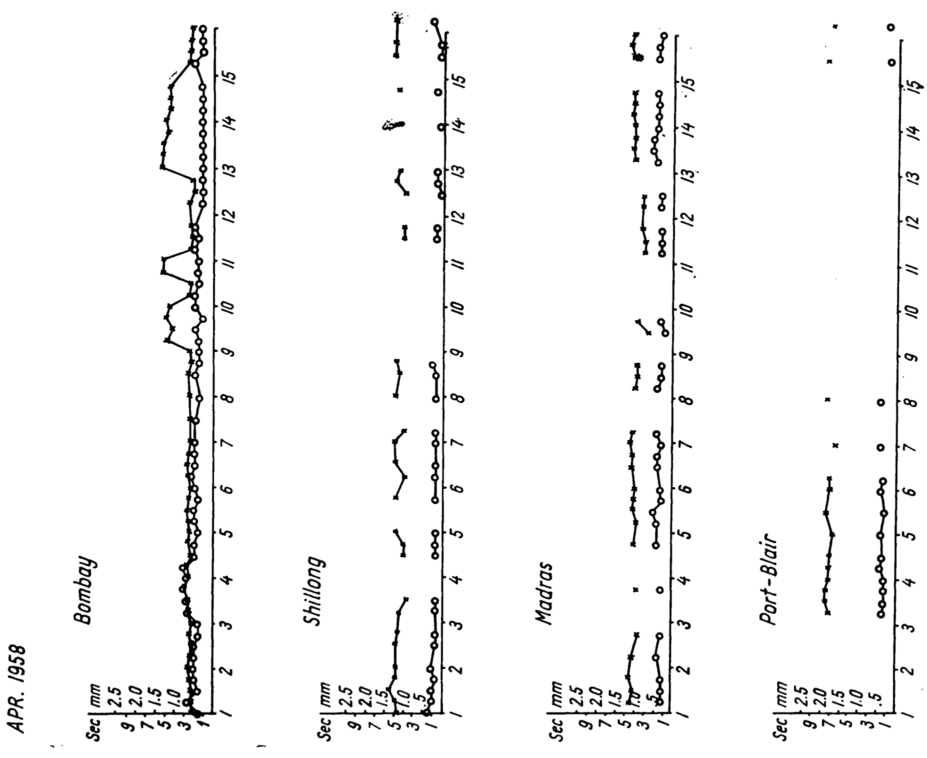

APR. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 35.

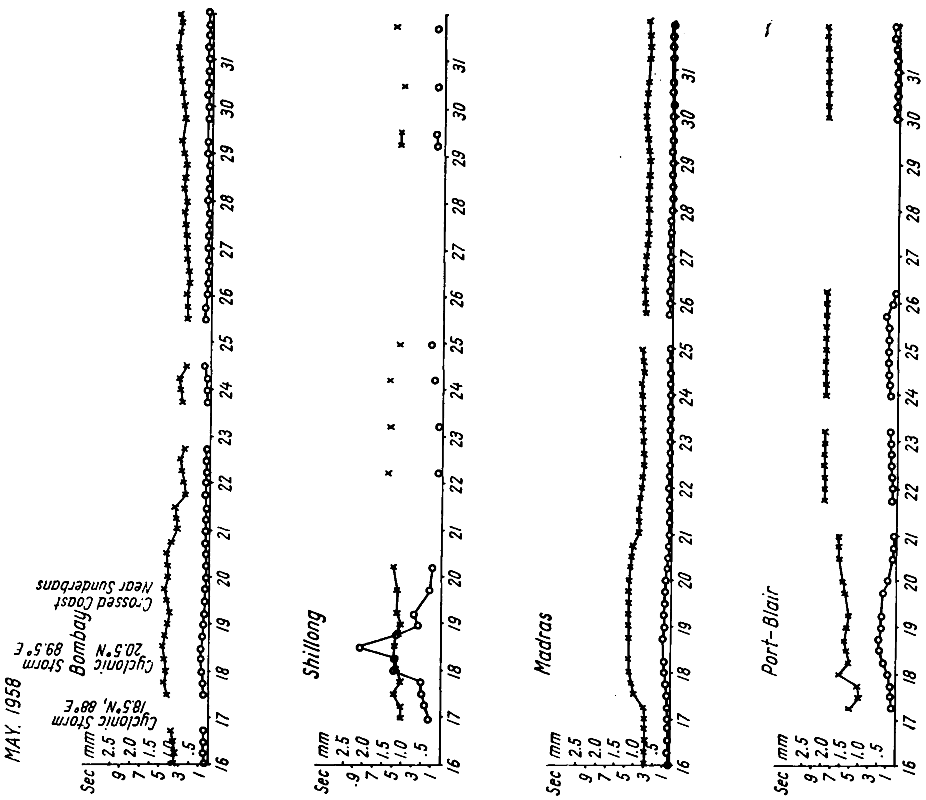

MAY. 1958
Bombay
Cyclonic Storm 18.5°N, 88°E
Cyclonic Storm 20.5°N 89.5°E
Crossed Coast Near Sunderbans
Shillong
Madras
Port-Blair
Sec
mm

FIG. 38.

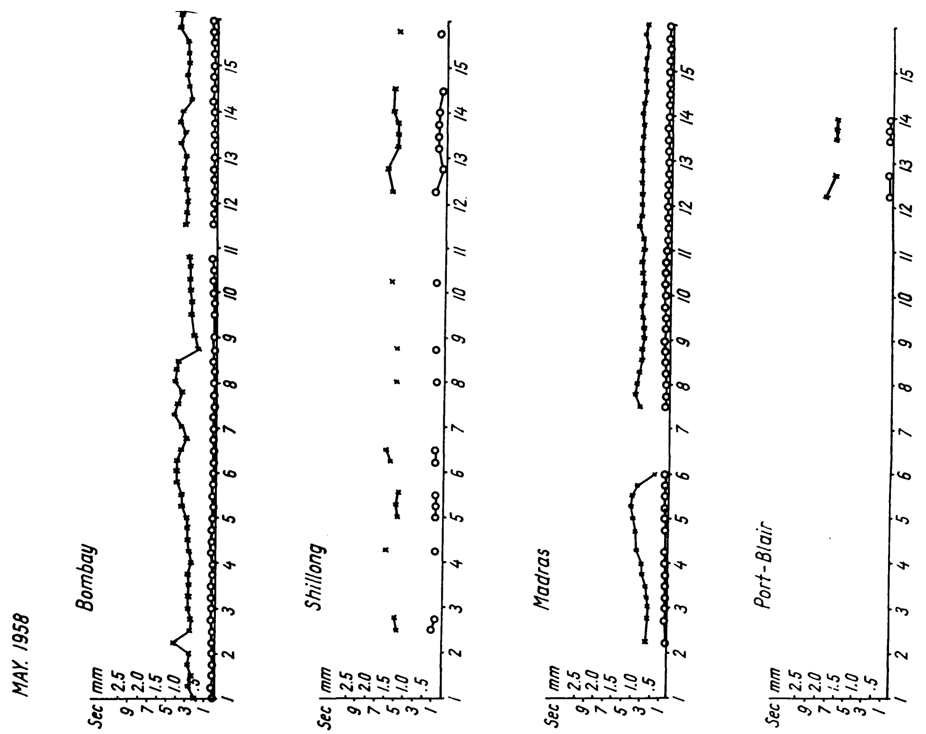

MAY. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 37.

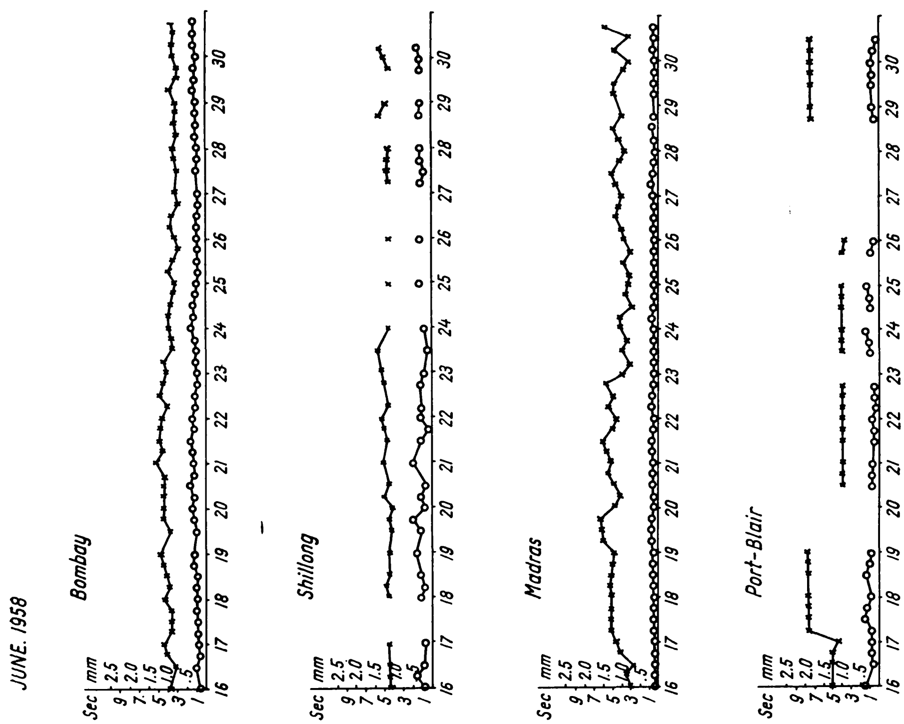

JUNE 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 40.

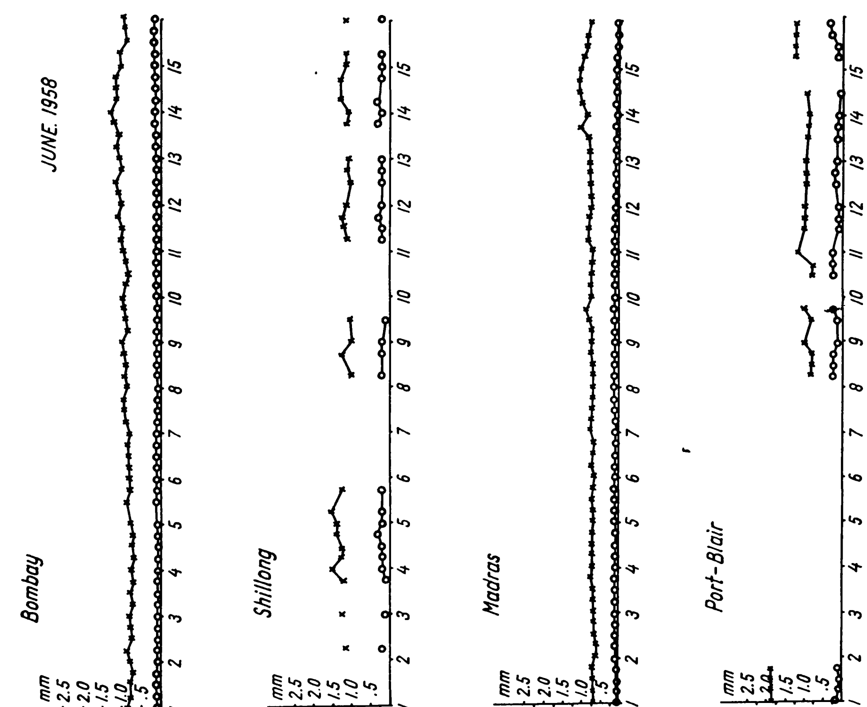

JUNE 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 39.

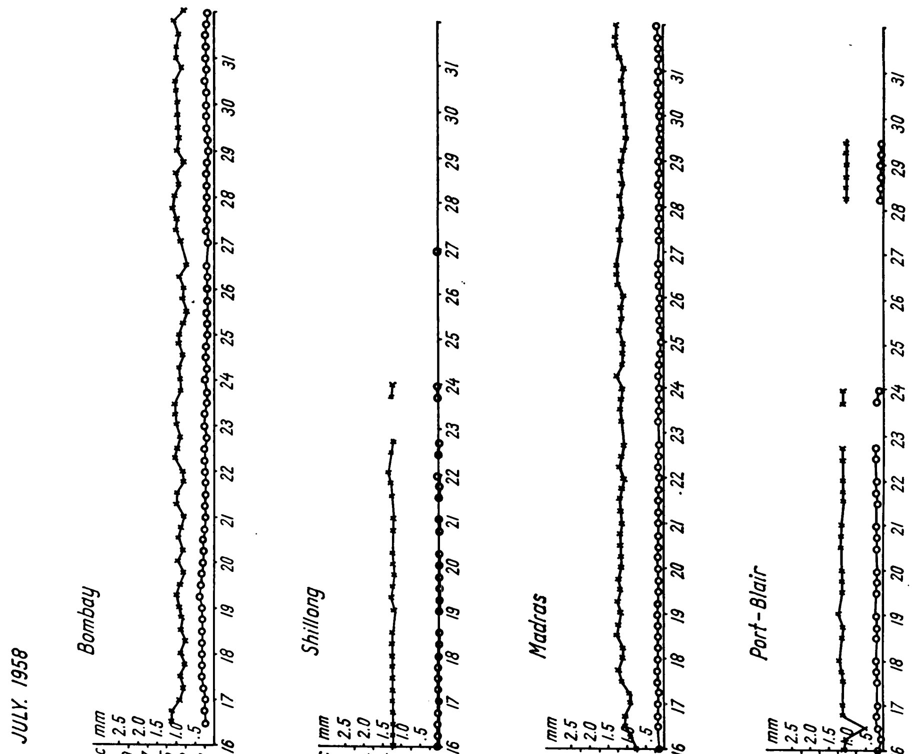

JULY. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 42.

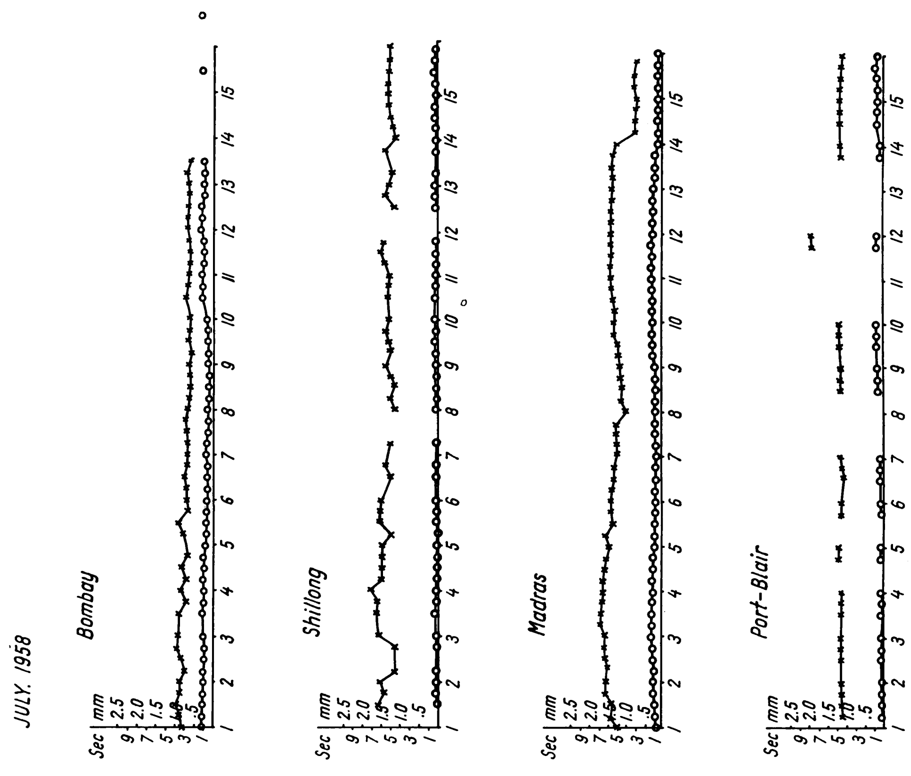

JULY. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 41.

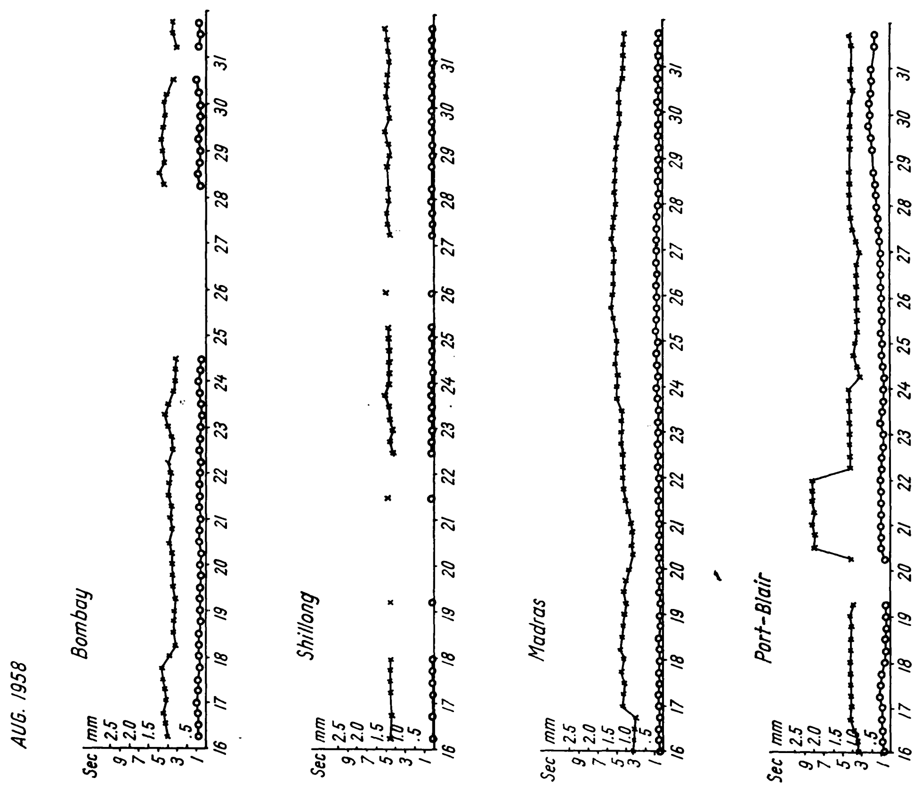

AUG. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 44.

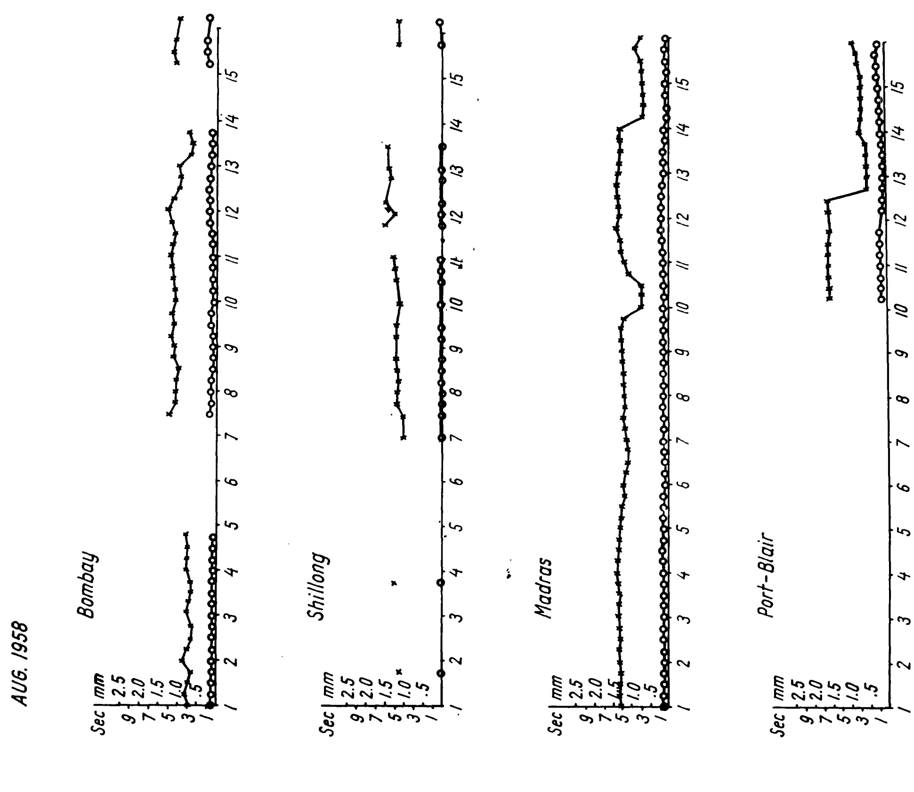

AUG. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 43.

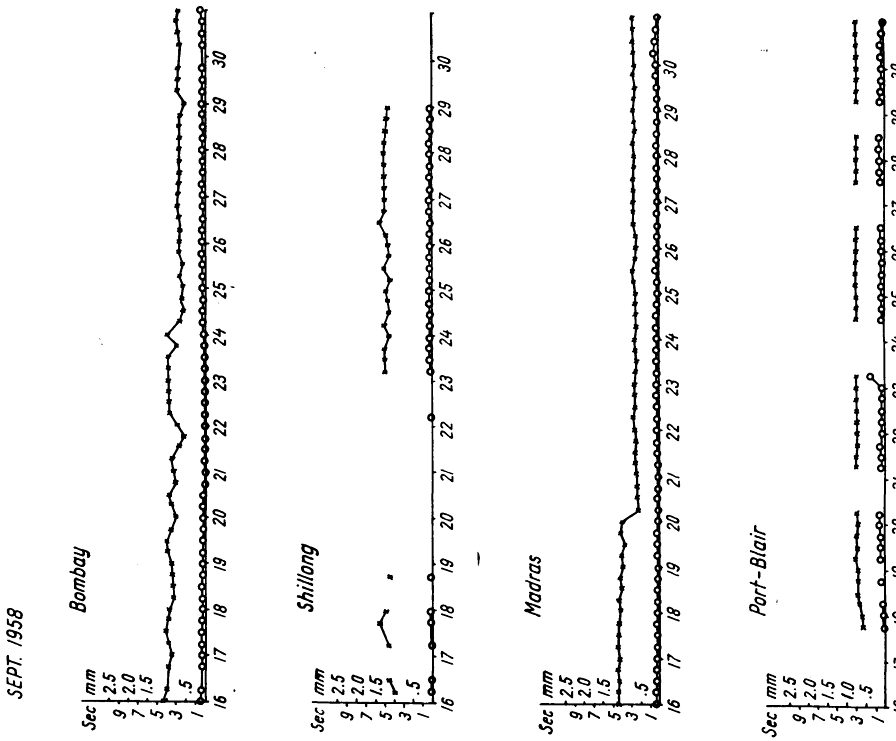

SEPT. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm
9 7 5 3 1
2.5 2.0 1.5 1.0 .5
16 17 18 19 20 21 22 23 24 25 26 27 28 29 30

FIG. 46.

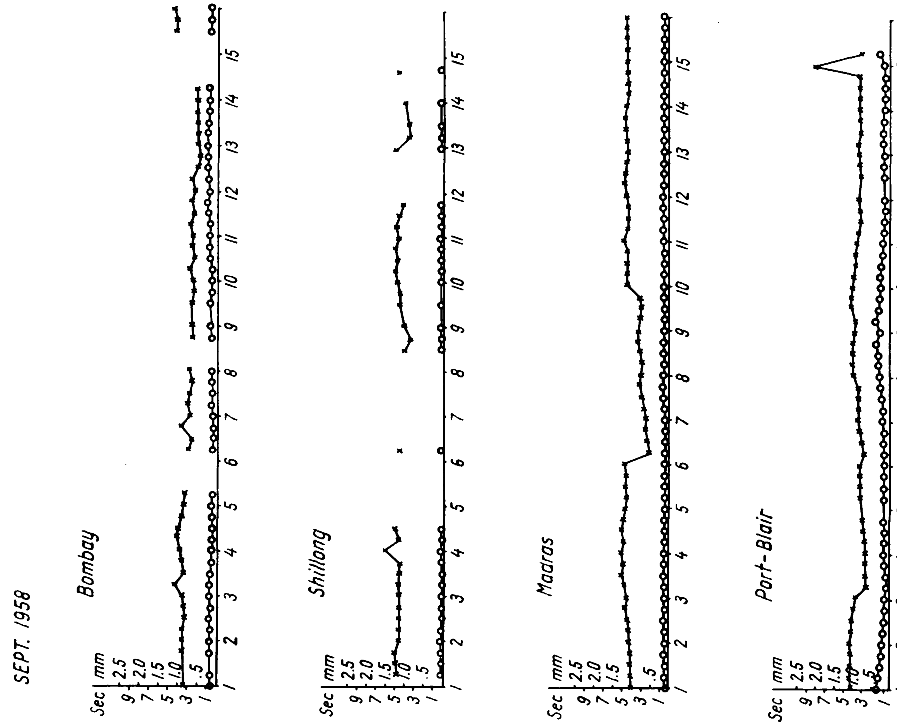

SEPT. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm
9 7 5 3 1
2.5 2.0 1.5 1.0 .5
1 2 3 4 5 6 7 8 9 10 11 12 13 14 15

FIG. 45.

of the 17th, the amplitude rose rather steeply at Shillong and Port Blair and continued to rise slightly at Madras and Bombay. The peak of microseismic amplitudes was reached at 12 hr GMT of the 18th. The amplitude at all the stations declined considerably by the morning of the 19th except at Port Blair where the decrease was more gradual. The periods at all the stations show an increase with increase in amplitude. It will be seen that the maximum microseismic activity was recorded a few hours before the storm crossed the coast.

4.4.9. *6–9 October 1958*

The synoptic chart on the 6th morning showed a depression with its centre at 10° N and 87° E. This moved in a north-westerly direction and crossed the coast near Negapatnam about 1200 hr GMT of 7th. None of the stations recorded any significant rise in amplitudes.

4.4.10. *9–15 October, 1958*

A depression formed in the Arabian Sea on the 9th October with its centre near 16° N and 71° E. By the morning of the 10th it had intensified into a cyclonic storm and had its centre at 17° N and 68·5° E. The storm continued to travel in the north-westerly direction until the 13th when it recurved and travelled in a north-easterly direction and finally weakened by the 15th. The microseisms at all the Indian stations did not show any significant rise although the amplitude of microseisms at Bombay remained slightly above normal during this period. The periods of microseisms at Bombay and Madras increased from 9th to 12th after which they became almost steady.

4.4.11. *21–23 October, 1958*

A depression was formed on the 22nd October with its centre at about 19·5° N and 90° E. By the morning of the 23rd it intensified into a cyclonic storm and crossed the coast the same night. Except for a very slight rise in amplitudes of microseisms, at Shillong, there was no significant rise of amplitudes at any of the Indian stations.

4.4.12. *6–10 November, 1958*

A depression formed in the Bay of Bengal on the 6th with its centre at latitude 16° N and longitude 85° E and started moving in a north easterly direction. It intensified into a cyclonic storm by the morning of the 8th, travelled eastwards and weakened on the morning of the 9th. Microseismic amplitudes and periods at Bombay, Madras and Shillong started rising very slightly from 1800 hr GMT of the 6th reaching a peak at 0600 hr GMT of 8th and thereafter declining slightly.

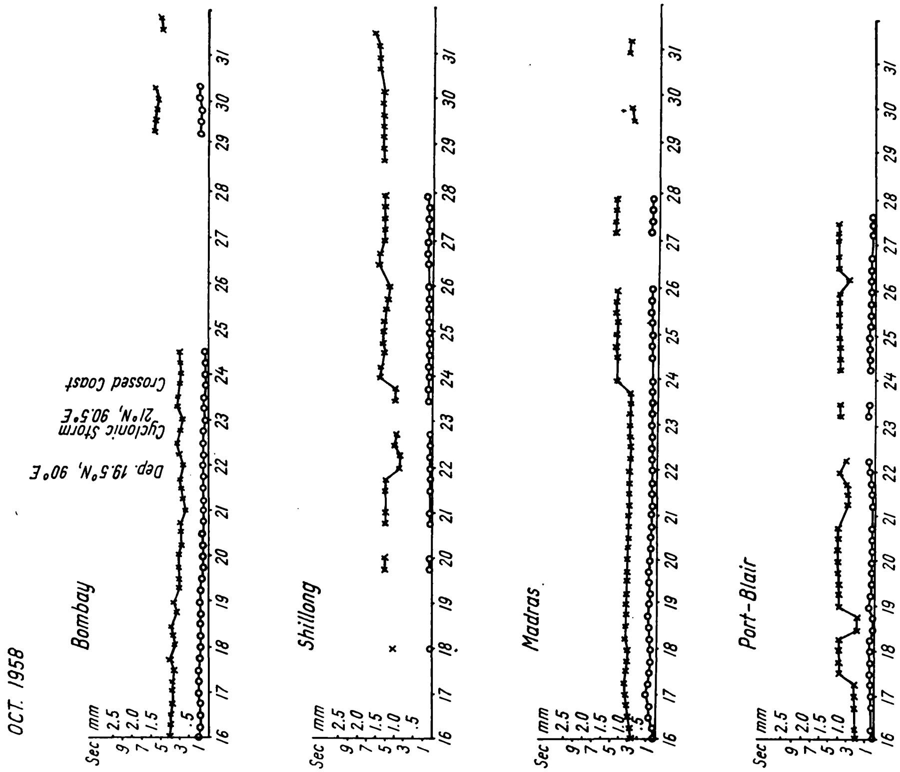

OCT. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm
Dep. 19.5°N, 90°E
Cyclonic Storm 21°N, 90.5°E
Crossed Coast

FIG. 48.

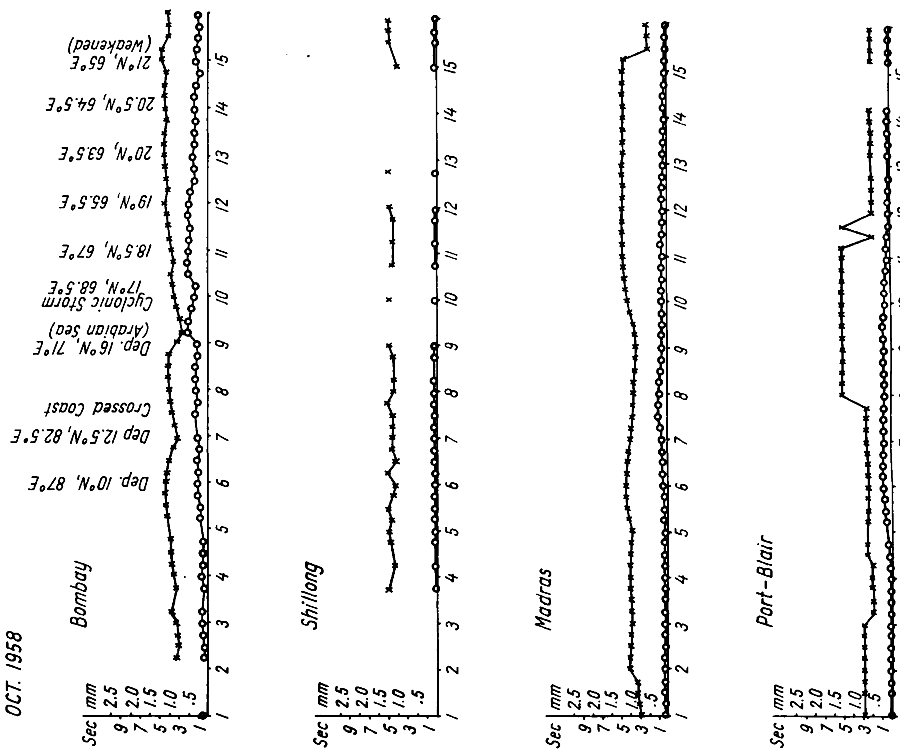

OCT. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm
Dep. 10°N, 87°E
Dep 12.5°N, 82.5°E
Crossed Coast
Dep. 16°N, 71°E (Arabian Sea)
Cyclonic Storm 17°N, 68.5°E
18.5°N, 67°E
19°N, 65.5°E
20°N, 63.5°E
20.5°N, 64.5°E
21°N, 65°E (Weakened)

FIG. 47.

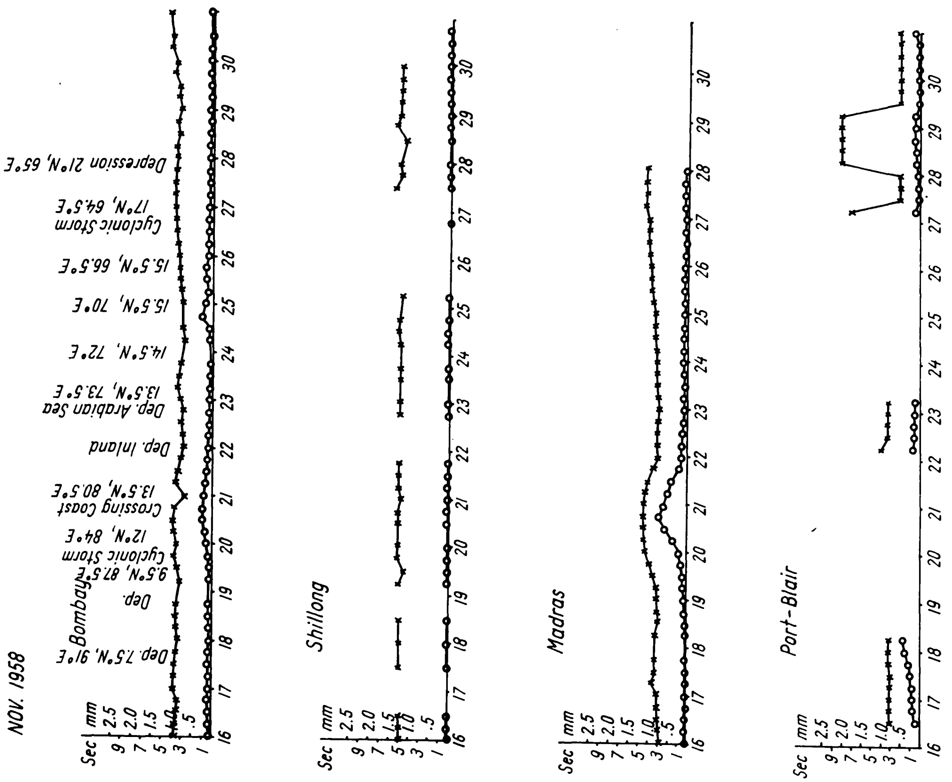

NOV. 1958
Bombay
Shillong
Madras
Port-Blair
Dep. 7.5°N, 91°E
Dep. 9.5°N, 87.5°E
Cyclonic Storm 12°N, 84°E
Crossing Coast 13.5°N, 80.5°E
Dep. Inland
Dep. Arabian Sea 13.5°N, 73.5°E
14.5°N, 72°E
15.5°N, 70°E
15.5°N, 66.5°E
Cyclonic Storm 17°N, 64.5°E
Depression 21°N, 65°E
Sec
mm

FIG. 50.

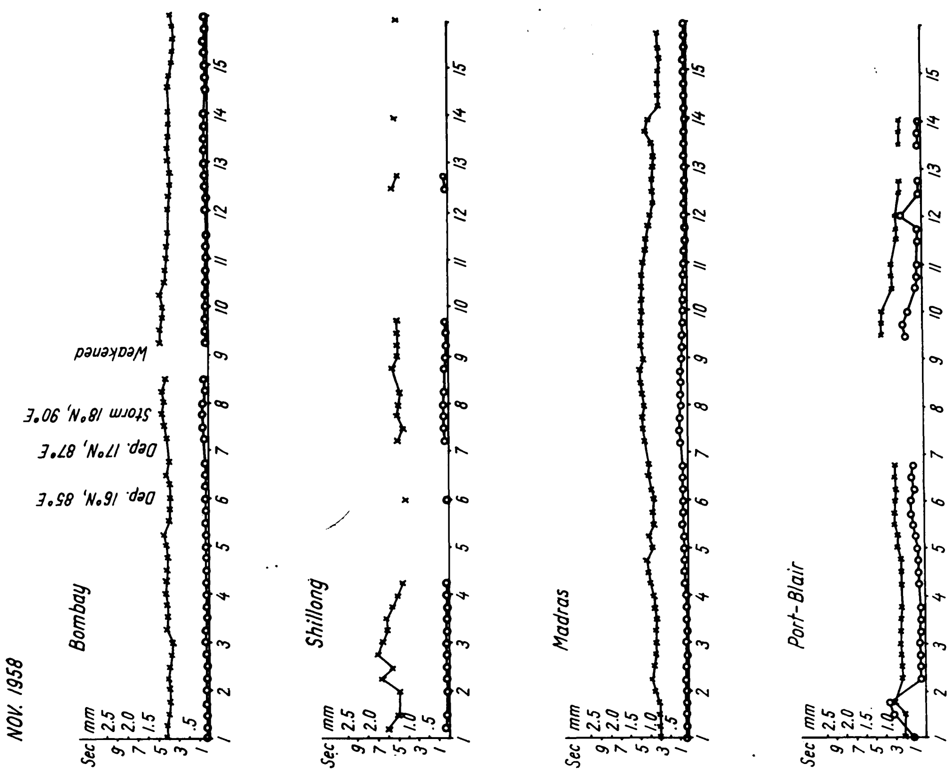

NOV. 1958
Bombay
Shillong
Madras
Port-Blair
Dep. 16°N, 85°E
Dep. 17°N, 87°E
Storm 18°N, 90°E
Weakened
Sec
mm

FIG. 49.

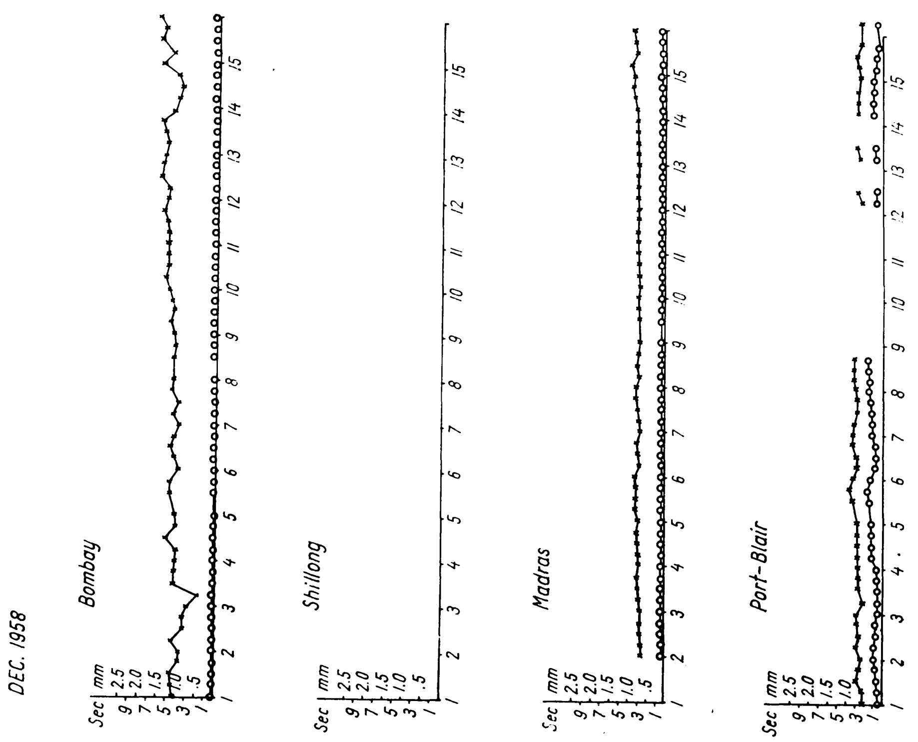

DEC. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 51.

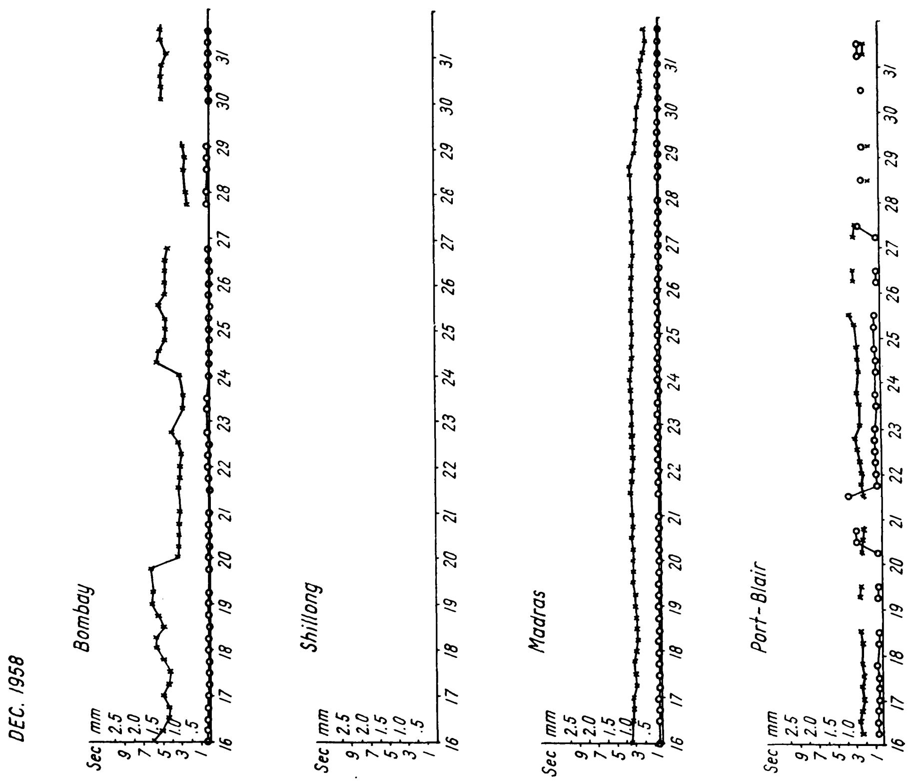

DEC. 1958
Bombay
Shillong
Madras
Port-Blair
Sec
mm

FIG. 52.

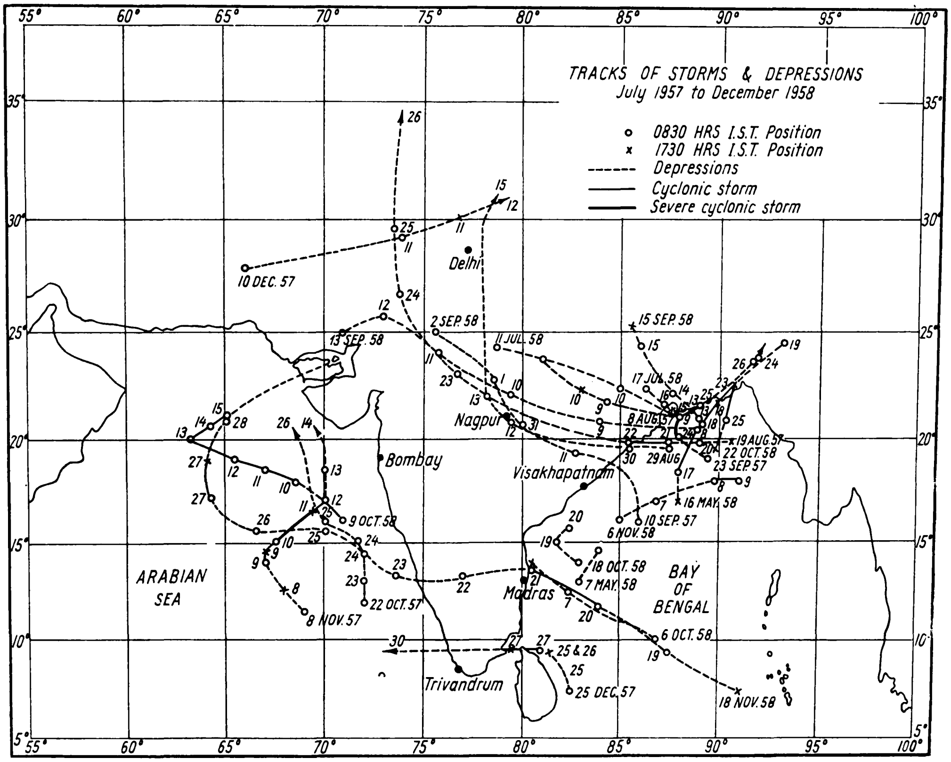


FIG. 53. Depressions and storms formed in Bay of Bengal and Arabian Sea during IGY.

4.4.13. *18–28 November, 1958*

A depression formed in the south Bay of Bengal on the 18th with centre at 7·5° N and 91° E. It moved in a north westerly direction and on 20th morning it lay as a cyclonic storm with its centre at 12° N and 84° E. The storm crossed the coast on the morning of the 21st near 13·5° N and 80·5° E. It travelled inland, emerged again as a depression in the Arabian Sea and on the 23rd had its centre at 13·5° N and 73·5° E. The depression moved in a westerly direction from 25th to 26th and then again north westerly from 26th to 27th when it intensified into a cyclone and recurved to north, then to north east and finally weakened after the 28th. There was a significant rise in the amplitude of microseisms at Madras, Bombay and Shillong near about 00 hr GMT of 20th. The peak of this activity was recorded the same day at 1800 hr after which the activity started declining, and returned to normal by the morning of the 22nd. During its passage in the Arabian Sea, there was no significant rise in the microseismic activity at the Indian stations except for a slight rise in amplitudes from 00 hr GMT to 1000 hr on the 24th at Bombay. The periods of microseisms at Madras, Bombay and Shillong show slight increase from 20th to 21st and at Madras and Bombay from 24th to 27th.

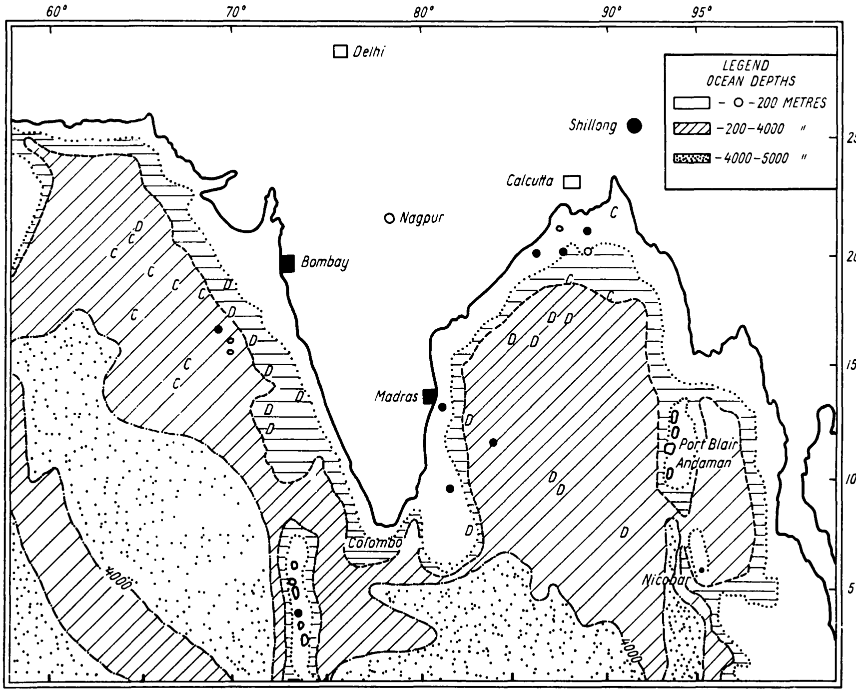


FIG. 54. Position of centres of cyclonic storms for which record of microseisms are plotted in Figs. 16 to 52.

4.4.14. *Discussion*

The map in Fig. 54 gives the position of centres of cyclonic storms and depression for which records of microseisms have been collected during the IGY period and plotted in Fig. 17-52. Depressions have been indicated by circles and letter *D* and cyclonic storms by thick circles and letter *C*. The letters *C* and *D* indicate those cyclones and depressions for which no significant microseisms recorded at any of the Indian stations. The map also shows contours for depths 0 to 200 metres, 200 to 4000 metres and over 4000 meters.

Although the number of positions of storms and depressions centres is not very adequate, yet a few important inferences of a qualitative nature are obvious from an inspection of this map. It will be seen that nearly all the depressions and cyclonic storms located within the depth contours 0–200 metres have given rise to good microseismic build-ups at all the Indian stations. No significant microseisms due to depressions outside this belt were recorded at any of the Indian Stations. Cyclonic storms having their centres at other locations in the Bay of Bengal if they are sufficiently strong do generate microseisms which can be recorded at the Indian stations. This is not very evident from the map given in Fig. 54 as the number of cyclonic storms in the Bay of Bengal during the period under study was not sufficient. In an earlier paper (Tandon 1957), the case of a severe cyclonic storms which had its track right through the central Bay of Bengal from South to North was discussed, and it was found that microseisms with good build-ups and amplitudes were recorded at all the coastal and inland seismographs. Figure 54 however, indicates that there are certain regions in the Bay of Bengal which attenuate transmission of microseisms. One such region appears to be the extreme north east angle of the Bay of Bengal and the other consists of the belt of deep water just outside the 0 to 200 metres depth contours.

In the Arabian Sea, the transmission of microseisms appears to take place only from locations confined to the 0–200 metres depth contour, except for a small area about 500 km to the west of Ratnagiri from where microseisms with attenuated amplitudes can be recorded at Bombay. No storm in the Arabian Sea within the belt 0–200 km was located during the present study, but previous data indicate that microseisms from storms located within this belt are recorded at all the Indian stations. The data collected during the IGY period seem to support the view that microseisms generated by storms located over deep waters can only be of operational value if the storm is fairly severe. It appears that a deep sea bottom is not a good medium for transmission of microseisms.

Figure 17 shows that, in nearly all cases when microseisms of significant amplitudes have been recorded, the amplitudes rise and fall almost simultaneously at all the stations. The microseismic maximum is recorded a few hours before the storm crosses the coast.

The effect of distance of the storm centre from the recording stations on the amplitude of microseisms does not appear to be a very straightforward. The amplitudes, as a rule, are not higher at stations nearest to the storm. It appears that the path of seismic rays from the storm to the station, and the geology on which the station is located, play an important role.

The variation of periods of microseisms has been attributed by various workers to a number of causes which include the distance of the recording station from the storm, the intensity of the storm, the depth of water below the storm, the geology of the path and the recording station and the type of seismograph used. The data presented in the present paper show that the periods recorded at any station are not dependent upon the distance of the storm from the recording station. There is a distinct evidence from all the records presented here that the periods depend on the intensity of the storm. In nearly all the cases studied the periods show a rise with increase in amplitudes. The higher the period recorded, the more severe has been the storm. Periods of microseisms from cyclonic storms are seldom found to be below 4 sec and in most cases are higher. Periods higher than 4 sec are transmitted better through the portion of the sea where the depth is large. Coastal stations like Madras and Bombay very often record periods of the order of 2 to 3 sec. These periods cannot be ascribed to microseisms due to cyclonic storms. Local surf activity is perhaps responsible for the generation of these periods. Whenever significant microseisms due to cyclonic storms have been recorded, it is found that the recorded periods are generally higher than 4 sec. Sometimes, even during the presence of a depression or storm near the station, the lower periods are recorded at coastal stations and the higher periods are eclipsed by them. Even in such cases the generation of short periods should be attributed to the increase in local surf activity caused by the neighbouring storm.

Microseismic periods recorded at Shillong are always higher than those recorded at Madras and Bombay, although the constants of seismographs operating at all the three stations are the same. The cause for these variations can only be attributed to local geology. The higher periods at Shillong may be due to greater depth of sediments as pointed out by NAG (1959).

4.4.15. *Conclusion*

Data collected at the microseismic stations during the IGY period indicate that operational use of microseisms for locating and tracking storms in the Indian Seas is feasible but with certain limitations. Microseisms generated by cyclonic storms of sufficient intensity in the Bay of Bengal are usually recorded at all coastal and inland stations. Microseismic storms generated in the Arabian Sea outside the 0–200 metres depth are not transmitted with sufficient amplitudes to seismographs to be of any operational value. Detection of cyclonic storms in the Arabian Sea can be done more expeditiously by conventional methods used in Meteorology. In order to use microseismic data for operational use, it may be necessary to use highly sensitive seismographs having magnification of the order of 20,000 for the period range of the order of 4 to 6 sec. At coastal stations records of microseisms are vitiated by local surf activity which generates periods of the order of 2 to 3 seconds. In order to make these stations more effective for storm detection and location, it will be necessary to use some technique to filter out these local microseisms. Technique suggested and used by DARBYSHIRE and IYER (1958) may be useful for this purpose.

Empirical techniques such as those used by Gilmore can also be profitably used for storm tracking but for this purpose data from standardised instruments will have to be collected for a large number of years. Collection of more data is also necessary to earmark areas in the sea from where microseisms are either not recorded or the amplitudes suffer considerable attenuation.

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4.5. Microseismic observations in Japan

Regular microseismic readings were made four times a day (0, 6, 12, 18 hr GMT), and in the case of microseismic storms four measurements were added at 3, 9, 15, 21hr GMT. On regular International Days (Regular World Days) and in International Periods (World Meteorological Intervals), microseismic readings were made hourly at each due hour. The data have been sent to the World Data Centers by way of the National Center.

To promote the investigation of microseisms in the north-western Pacific Region, nine supplementary microseismic stations were added to the planned IGY stations. These stations made the measurements four times per day (0, 6, 12, 18hr GMT) throughout the IGY period, regardless of the occurrence of microseismic storms, or the Regular International Days and International Periods.

Three tripartite stations made occasionally microseismic observations, especially when special weather events such as typhoons took place in the Western Pacific region. They reported the amplitudes, periods, character of the microseisms and the velocity and direction of propagation of microseisms.

A Research Group for Microseisms including more than 50 Japanese seismologists was organized. The representative of the group is T. MATZUZAWA and the secretaries are KIZAWA and T. ASADA. This group compiled the microseismic data and summarized them in two volumes under the title "Report of Microseismic and Sea Wave Observations in Japan during the International Geophysical Year, 1957–58", published by the National Committee of IGY, Science Council of Japan. The contents of these books are as follows:

(1) Descriptions of microseismic stations, instruments, and methods of measurements.

(2) Graphs showing the variations of daily mean value (or reading value four times a day) of the amplitude and the period of microseisms.

(3) Graphs showing the variations of the period and height of wind waves and swells observed visually at sea wave stations along the coast of Japan.

(4) Graphs showing monthly frequency distributions of the amplitude and the period of microseisms.

(5) Surface weather map near Japan at 0000 GMT every day during IGY period.

(6) Graphs showing the correlations between microseisms and wind waves.

(7) Tables of the observational values of the period and height of sea waves.

The numerical data obtained during microseismic storms and from tripartite observations have been published in another book under the title "Data of Microseismic Observations during Microseismic Storms and Tripartite Microseismic Observations".

4.6. Microseism investigations in Australia

by K. E. BULLEN

Microseism investigations in Australia are carried out by Owen JONES and P. S. UPTON of the University of Queensland at two stations on the east coast of Australia: Brisbane (27°29′ S., 153°02′ E.) and Townsville (19°19′ S., 146°44′ E.). The instruments used for this purpose are Sprengnether horizontal-component seismometers with periods of 7 seconds, giving a peak response around 4 seconds, with a magnification of about 5000 at that period.

The following are some of the results of the work of JONES and UPTON:

(i) Cyclones have been detected at distances of the order of 1000 miles from Brisbane.

(ii) The microseism amplitude (here taken as the mean of all waves of 2 mm occurring in a 20-min interval) was inversely proportional to the square root of the distance from the cyclone centre, during two recorded cyclones.

(iii) Microseism periods associated with tropical cyclones are 4 to 5½ sec.

(iv) Storms outside the tropics are associated with microseism periods of 7 to 8 sec. This applies especially when there is a low-pressure area over the ocean to the south of Australia. In general, there is good correlation between microseisms in Brisbane and Townsville, and meteorological conditions in eastern and southern Australia.

(v) Ragged short-period microseisms of periods 2 to 4 seconds are associated with cold fronts or strong easterlies.

Publications

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UPTON, P. S. 1960. The generation of microseismic storms in the Coral Sea. Univ. of Queensland Papers, Dept. of Geology.

5. THE STRUCTURE OF THE EARTH'S CRUST

5.1. Summary of special seismic crustal studies during the IGY

by P. S. VEITSMAN and I. P. KOSMINSKAYA

AT the beginning of the International Geophysical Year, reliable data were available, showing a difference in the crustal structure of the oceans and the continents, the presence of mountain roots, as well as certain regular relations between the crustal structure of various regions and the history of their geological development. Other, complimentary studies were, however, required and these were based on the principles of detailed seismic observations on land [6–8, 17, 18] and at sea [31–33, 36, 37, 39, 40, 42, 44–48, 49, 60] which had been perfected prior to the IGY.

During the preparatory period, and during the IGY itself, crustal studies were carried out by many countries. Some idea of the volume of the investigation made during this time can be gained from the map given in Fig. 1. This map indicates, schematically, the lines of special observations carried out with the aim of studying the earth's crust by means of seismic procedures using large industrial explosions, comparatively small experimental blasts, or special earthquake recording procedures which yield sufficiently complete data to allow the determination of the thickness and the mean values of the physical parameters of the crust, and also to clarify its structure.

According to the problems posed and the observational procedure, all crustal studies carried out during the IGY and the IGC may be divided, schematically, into three groups.

(1) *Studying the earth's crust over large areas by infrequent reconnaissance observations to establish general regularities of the crustal structure of the globe.*

Studies belonging to this group have been carried out mainly in the oceans by the United States. Observations cover areas in the Pacific and Atlantic oceans, west and east of South America, and in the Mediterranean and Red seas.[61] Studies on land made with large blasts should be also included in this group: in the Alps (carried out by France, Italy, Belgium, etc.), in Canada,[60] Chile, the United States, Finland,[48] Iceland,[25] and also some studies making use of earthquakes carried out by the U.S.S.R. in certain areas of Central Asia.[2,3]

(2) *Studies of definite regions to reveal the peculiarities of crustal structure in various tectonic zones, such as inland seas, zones of transition from platforms to geosynclines, or from continents to oceans.*

These include investigations by the United States in the Caribbean Sea,[61] and by the U.S.S.R. in the Black Sea,[20,21] the Caspian Sea,[5] the Okhotsk Sea and

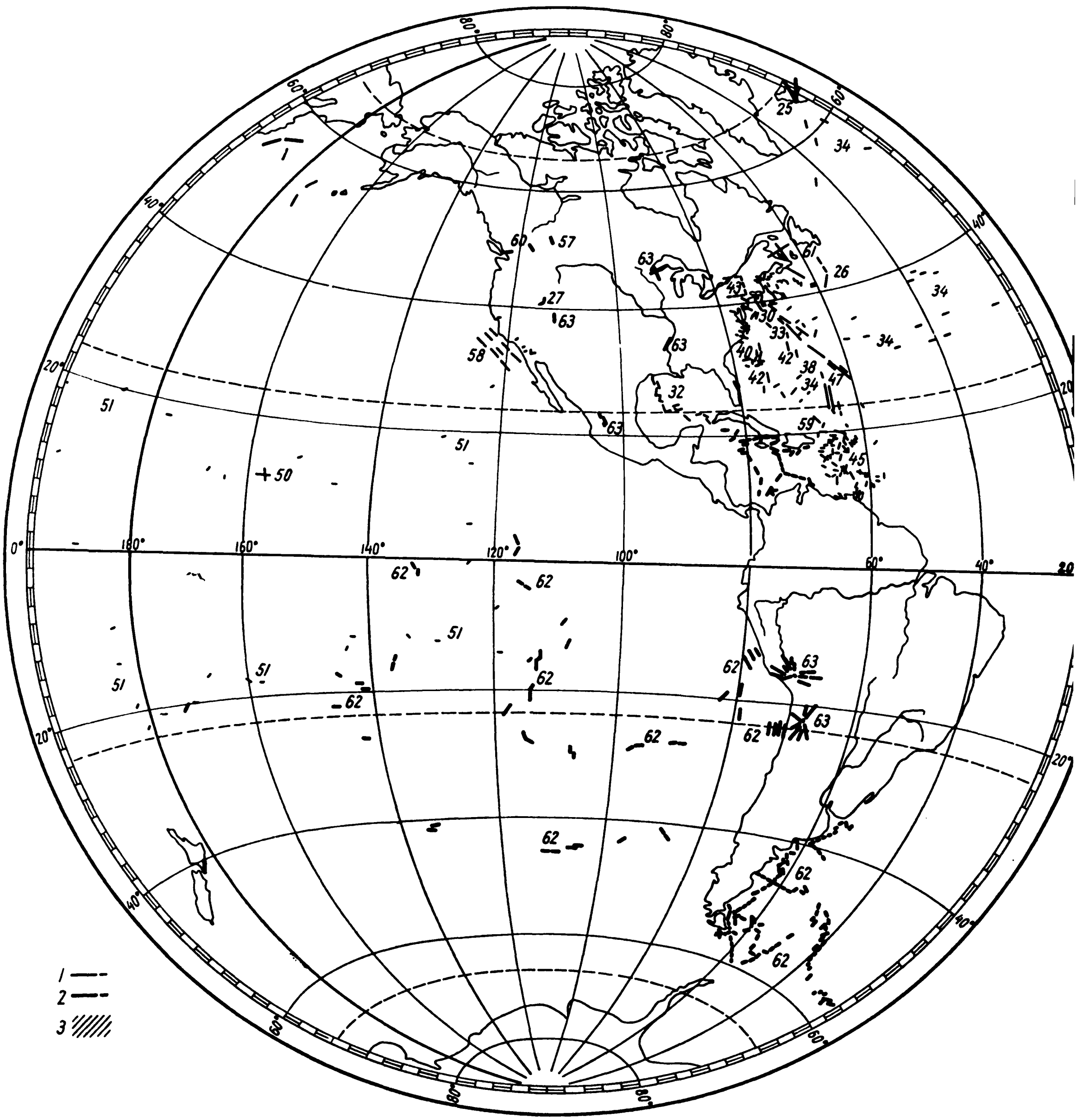

FIG. 1. Regions of special seismic crustal studies.
1. Observations carried out prior to the IGY 2. Observations with experimental explosions carried out during the IGY and the IGC (1956–59) 3. Observations carried out during the IGY and the IGC using earthquake data. Numbers by profiles are those of the list of references.

in the adjacent parts of the Pacific.[1,4] Such studies are usually more detailed than those of the first, because use is made of comparatively small blasts and specialized systems of profiles, intersecting areas which are known to have a different crustal structure. On land, similar studies have been carried out in Hungary.[35,36]

(3) *Detailed studies of the crustal structure and of the sedimentary series in separate areas having a different structure as well as a different history of geological development.*

These studies were carried out in coordination with other geophysical methods of regional investigation during the period of the IGY, in the following areas of the Soviet Union: the Baltic Shield,[19] south-east of the Russian platform,[9,10] Turkmenia,[14] Bukhara,[12] Kazakhstan,[15,16] Fergana.[11]

These results have not yet been fully analysed. Only some preliminary data have been published, and these corroborate the difference in crustal structure of the continents and oceans. They have also revealed areas in which the crust differs from the typically continental and the typically ocean types. It has also been found that the crustal structure of each of these types varies considerably. In the zones of transition from ocean to continent the alternation and transition of the earth's crust from one type to another were found to be extremely complex.

The published results of the extensive research carried out on continents confirm the substantial difference in seismic properties of the upper and lower parts of the crust. Detailed investigations of the U.S.S.R. have made it possible to establish several seismic boundaries with different velocities increasing with depth in the earth's crust.

New laboratory data on the behaviour of rocks of different composition, at pressures close to those occurring in the crust can, at present, be correlated with the results of field observations in essentially different areas, thus permitting us to increase our knowledge of the possible composition of rocks forming the earth's crust and to extend the limited and arbitrary "granite" and "basaltic" layer classification.

Crustal studies of various sections of oceans and continents carried out during the IGY will undoubtedly provide rich material essential for the solution of a variety of geophysical problems, including that of establishing and clarifying the relations between the thickness and structure of the crust, the relief and the gravitational field. All this should play an important part in solving one of the most important problems of modern times: that of the formation of continents and oceans.

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5.2. Crustal thickness in Australia

by K. E. BULLEN

PER area, Australia is less affected by earthquakes than any other country of the world, and the problem of investigating local crustal structure has been handicapped both by the paucity of suitable earthquakes and also by the consequent small number of observing stations. With stimulus from the IGY and other sources, the number of first-class stations has been raised during the past two years from 3 to 10. But there is still little natural-earthquake data from which reasonably precise inferences can be made on local P and S velocities and crustal thicknesses.

5.2.1. *Work on crustal structure of Australian mainland*

(i) The most noted work carried out on Australian crustal structure during the IGY was the analysis of seismograms of atom bomb explosions at Maralinga, South Australia. Studies were made by groups drawn from the Australian National University, the Universities of Sydney and Adelaide, and the Australian Commonwealth Bureau of Mineral Resources (Geology and Geophysics Division).

This work gave the first precise information on crustal structure in any part of Australia. All observations, except that at Adelaide, were taken close to a line running to ten degrees west from Maralinga: this region includes the Nullabor plain. (The geology of this plain is described as horizontally bedded limestones, known to be 900 ft thick in places, with the Yilgarn pre-Cambrian shield to the west and possibly under the limestones.) The three most distant stations were on the Yilgarn shield.

The inferred travel-times gave P_n and S_n velocities of 8·21 and 4·75 km/sec. There was also evidence of P and S phases with velocities of 6·03 and 3·55 km/sec.

The observations were consistent with the presence of a single crustal layer having constant P and S velocities. The crustal thickness formally indicated on the single-layer hypothesis was 32 km from the P observations, and 39 km from the S observations.

(ii) In the University of Queensland investigations were carried out of records, taken at three stations in the Brisbane region, of a detonation of 2000 kg of explosives in the Toowoomba City Council Quarry on 10 April 1959. Evidence was found of a P phase of velocity 5·5 km/sec.

(iii) Preliminary studies of waves from quarry blasts have also been carried out in the Australian National University and the University of Sydney.

(iv) A study of surface seismic waves across Australia was carried out in the University of Sydney. The mean Lg and Rg group velocities were found to be $3{\cdot}50 \pm 0{\cdot}06$ and $3{\cdot}03 \pm 0{\cdot}07$ km/sec, in striking agreement with results for North America and Eurasia.

5.2.2. *Waves from hydrogen-bomb explosions*

Seismic records from four hydrogen-bomb explosions near Bikini Island were studied by members of Riverview Observatory and the University of Sydney.

The analysis gave negative P residuals against standard travel time-tables, and gave results compatible with a structure in the Bikini region in which the P velocities are somewhat higher, depth for depth, than in the average continental crust.

The analysis showed further that the P travel-times from Bikini to Australia and from Bikini to North America differ by less than a second for the same distances.

5.2.3. *Further Australian seismic work during the IGY*

Theoretical work on travel-times, with some special relevance to the Earth's upper mantle, as well as to deeper regions, was carried out in the University of Sydney.

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5.3. The Canadian IGY programme of crustal studies

by P. L. WILLMORE

Canadian crustal studies during the IGY were all related to the explosion of 2,750,000 lb of high explosive, which was detonated under Ripple Rock on 5th April, 1958. Although the primary purpose of the explosion was to clear the navigation channel between Vancouver Island and the mainland of British Columbia, it was realized from the outset that the explosion would provide a unique opportunity for a study of the large-scale structure of the Cordilleran region.

The work was organized in three parts — first, a programme for recording the explosions of depth charges in the waters around Vancouver Island, second, the observation of the seismic waves from the Ripple Rock explosion and, third, observations along a refraction profile in Alberta.

In planning long-range refraction experiments in Canada, the minimum of reliance is being placed on the conception of plane stratification in the crust, which would lead to the expectation that travel-times should be linear functions of distance. Instead, it is assumed only that one will be able to recognize waves refracted through underground layers, the boundaries of which may undulate in all directions. "Time terms" a_i for each station and b_j for each shot point are introduced, such that the travel-time t_{ij} between the i^{th} shot and the j^{th} station is

$$t_{ij} = a_i + b_j + \Delta_{ij}/v \tag{1}$$

where Δ_{ii} is the horizontal distance and v is the velocity of the seismic waves in the refracting layer. To a first approximation, a_i and b_j are independent of the form of the boundary or of the azimuth of the station relative to the shot point, and corrections can be introduced to allow for some of the effects of large gradients in the structure.

Equation (1) is ambiguous, in that an arbitrary quantity can be added to the time term of each station in a connected survey, and subtracted from the time term of each shot, without altering the pattern of travel-times. A complete solution for the time terms and v requires that waves from several shot points should be recorded by several stations, and that data should be available for determining α. The latter requirement may be met by making at least one station coincide with a shot point, or by having one station or shot point in a location at which the time term can be estimated independently.

The depth-charge study was organized by the Dominion Astrophysical Observatory and the Pacific Naval Laboratory. Most of the recording was confined to the permanent stations at Alberni, Horseshoe Bay and Victoria, although a few shots were observed by portable records elsewhere in Vancouver Island. The shots were fired by Naval vessels, and were distributed along two lines running along the Northeast and Southwest shores of the island, and also in an area on Constance Bank (see Fig. 2).

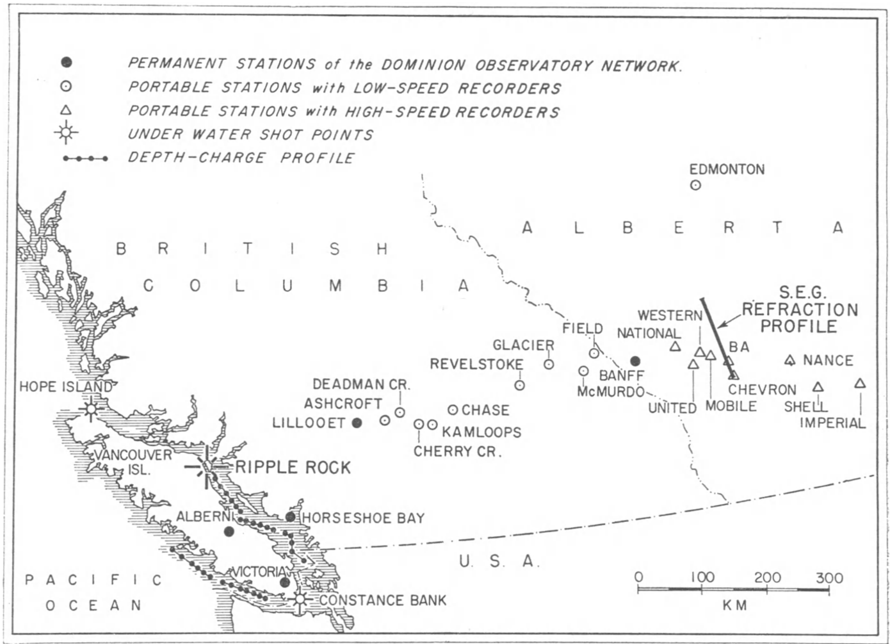

FIG. 2.

Separate solutions were derived for the three groups of shots, using the P_1 waves which are transmitted through the upper part of the crust. As the three recording stations are all on exposed basement rocks, the arbitrary constant α was determined by setting the mean of the station time terms equal to zero. The time terms for the shots then represent the time required for the waves to penetrate the sediments on the sea bottom.

The data for the two long lines yielded independent estimates of the crustal velocity, which came out to be $6{\cdot}28 \pm 0{\cdot}06$ and $6{\cdot}29 \pm 0{\cdot}07$ km/sec. respectively, and the distribution of time terms generally fitted in with the known geology of the area. The mechanics of the solution are such that the velocities and their standard deviations are weighted in favour of regions in which shot points and stations form over-lapping networks, which in this case is the southeastern tip of Vancouver Island, and the nearby part of the Strait of Georgia. At the northwestern end of the Strait of Georgia near Ripple Rock, the time terms run negative. This may indicate a locally high value for seismic waves in the crust.

The seismic waves from the Ripple Rock explosion were recorded in Canada by Dominion Observatory seismographs as far away as Banff, by commercial crews operating geophysical prospecting equipment in Alberta, and by a portable seismograph operated by the University of Edmonton. Records were also obtained by a number of stations in the United States. Time control was provided by including special time signals in a news programme which was sent out by the Canadian Broadcasting Corporation. In some mountain locations radio reception was unsatisfactory, and stations in these positions were provided with time signals over the telegraph lines of the Canadian Pacific Railway.

The distance from Ripple Rock to most of the recording stations was such that the first onsets were P_n, whereas the waves in the depth-charge study had all been P_1. It was therefore necessary to have additional shot points to provide data on the P_n velocity and on the arbitrary term, and for this purpose the Navy fired charges of high explosive near Hope Island and on Constance Bank. Unfortunately conditions for energy transmission were less favourable than had been hoped, and the ideal condition, which was to record P_n from either or both of the auxiliary shots at some stations which also recorded P_n from Ripple Rock, was not satisfied. Nevertheless, the available data do give a fair degree of control over the crustal thickness in the general area of Vancouver Island, and this has been used as a basis for estimating the arbitrary term for the Ripple Rock observations.

The chief remaining uncertainty was in the velocity of P_n under the long profile into Alberta, for a slight error in estimating this parameter would distort the whole time term picture. In order to overcome this difficulty, the IGY committee of the Canadian Branch of the Society of Exploration Geophysicists organized a field experiment in the Alberta plains.* Charges were fired at the ends of a line, 81 miles long, the southern end of which passed near two of the stations which had recorded waves from Ripple Rock. Fifteen seismic parties, equipped with commercial refraction apparatus, recorded the propagation of seismic waves along the line. The results showed that the Mohorovicić Discontinuity occurred at a depth of 43 km under the profile, and also indicated the presence of an intermediate layer with a velocity of about 7·2 km/sec at a minimum depth of 29 km.

By combining estimates of the time terms in Alberta and under Ripple Rock with the observed travel time over the whole length of the Ripple Rock profile, an estimate of the P_n velocity was obtained. These data enable time terms to be determined for all the other stations which observed the Ripple Rock tremors. The results indicate crustal thicknesses of 65 km or more in the vicinity of Banff, with lower values further west. Under Ripple Rock itself, the crust was estimated to be 32 km thick. These values were derived on the assumption that the mean crustal velocity was equal to the P_1 velocity of 6·28 km/sec which was derived from the depth-charge experiment, and would have to be increased if the intermediate layer, which was found in Alberta, extended to the West Coast. The Banff figure indicates the existence of deep sides. It is possible that the estimated thickness refers to the length of an oblique ray through the feature, rather than to the depth vertically below the station.

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5.4. Seismic crustal studies in Finland

by E. VESANEN

The explosion seismic investigations carried out in Finland are listed below. The map (Fig. 3) explains the regions of work. The main results are also given with the corresponding literary notes.

5.4.1. *Explosion seismic investigations*

Pori 1958

$Pg_1 = 5{\cdot}72$ km/sec $\qquad Sg_1 = 3{\cdot}34$ km/sec

(Vesanen, Metzger, Nurmia and Porkka[1])

Reflexion method:

Upper layer (Conrad) 18 km

(Penttilä and Nurmia[2])

Porkkala 1958

$Pg_1 = 5{\cdot}73$ km/sec

$Pg_2 = 5{\cdot}95$ km/sec $\qquad Sg_2 = 3{\cdot}52$ km/sec

$Pb = 6{\cdot}37$ km/sec $\qquad Sb = 3{\cdot}72$ km/sec

$Pn = 8{\cdot}23$ km/sec $\qquad Sn = 4{\cdot}67$ km/sec

Refraction method:

Hanko–Kotka

Upper layer (Conrad)	21 km
Intermediate layer	8 km
Mohorovičić	29 km

Reflexion method:

1. Hanko–Porkkala

Upper layer (Conrad)	18·5 km
Intermediate layer	8·0 km
Mohorovičić	26·5 km

2. Helsinki–Kotka:

Upper layer (Conrad)	21·0 km
Intermediate layer	8·0 km
Mohorovičić	29·0 km

(Penttilä, Karras, Nurmia and Vesanen[3])

Kotka 1960

$Pg_1 = 5{\cdot}89$ km/sec

$Pg_2 = 6{\cdot}21$ km/sec $Sg_2 = 3{\cdot}55$ km/sec

$Pb = 6{\cdot}65$ km/sec $Sb = 3{\cdot}96$ km/sec

$Pn = 8{\cdot}25$ km/sec $Sn = 4{\cdot}67$ km/sec

Refraction method:

Upper layer (Conrad)	21 km	
Intermediate layer	19 km	
Mohorovičić	40 km	(calculated from *P*-impulses)
Upper layer (Conrad)	19 km	
Intermediate layer	18 km	
Mohorovičić	37 km	(calculated from *S*-impulses)

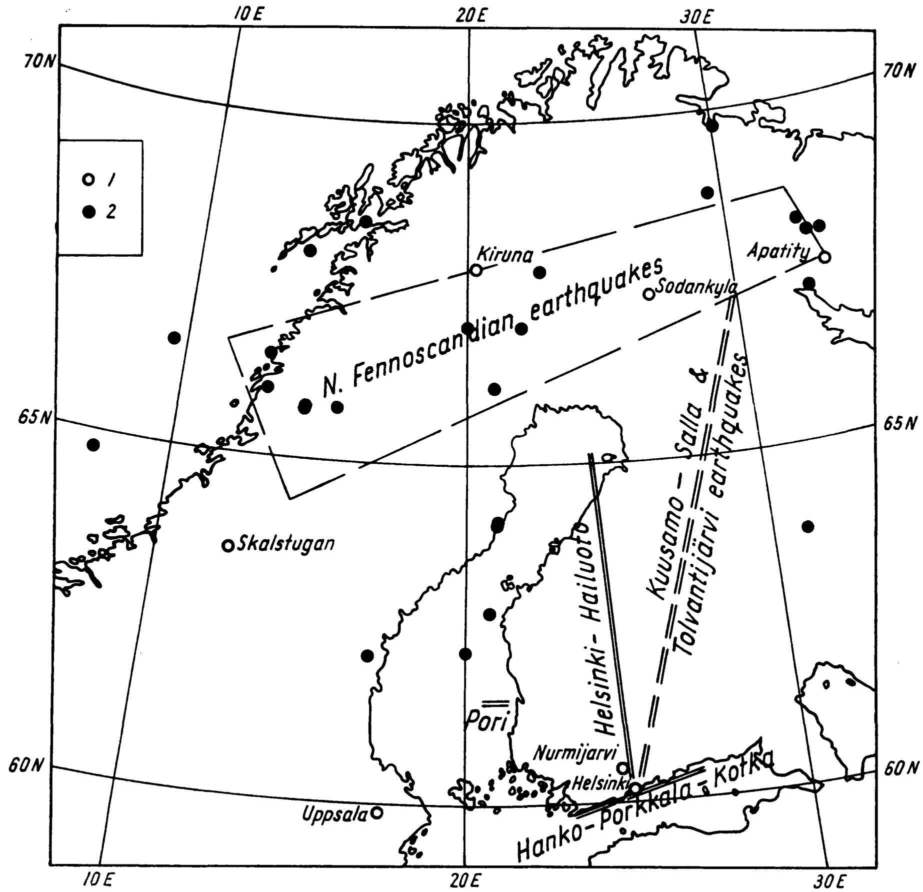

FIG. 3. Regions of crustal studies in Finland: 1 — explosions; 2 — epicentres of earthquakes.

Reflexion method:

Upper layer (Conrad)	21 km
(Penttilä and Vesanen[4])	

Hailuoto 1960

$Pg_1 = 5{\cdot}80$ km/sec	$Sg_1 = 3{\cdot}28$ km/sec
$Pg_2 = 6{\cdot}17$ km/sec	$Sg_2 = 3{\cdot}52$ km/sec
$Pb = 6{\cdot}77$ km/sec	$Sb = 3{\cdot}73$ km/sec
$Pn = 8{\cdot}39$ km/sec	$Sn = 4{\cdot}69$ km/sec

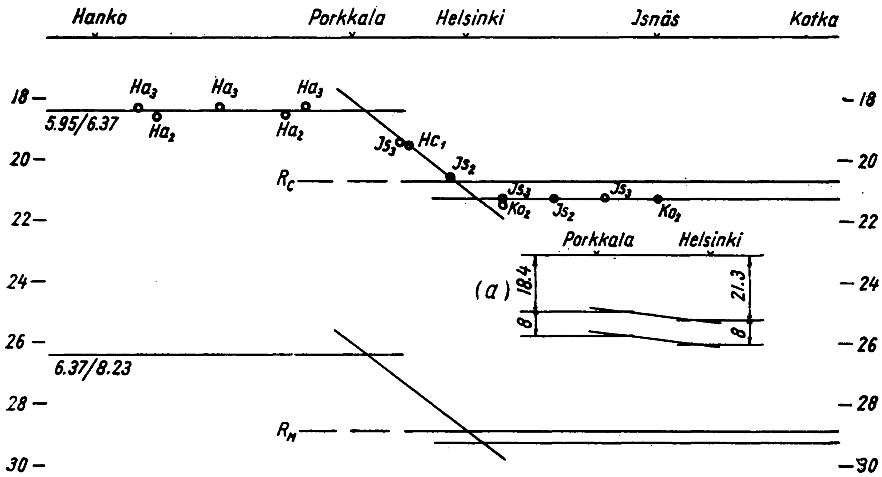

FIG. 4. Depth profile of the Moho and Conrad surfaces and points obtained by the reflexion method along the line Hanko–Kotka. The vertical unit is 10 × the horizontal unit; in insert (*a*) the scale is the same. R_C and R_M — refraction results.

Refraction method:

Mohorovičić	36 km	(calculated from *P*-impulses)
Mohorovičić	34 km	(calculated from *S*-impulses)
(Penttilä and Vesanen[4])		

5. 4. 2. *Investigations using Northern Fennoscandian earthquakes*

N–Fennoscandia

$Pg = 5{\cdot}79$ km/sec	$Sg = 3{\cdot}41$ km/sec
$Pb = 6{\cdot}60$ km/sec	$Sb = 3{\cdot}77$ km/sec
$Ph = 8{\cdot}12$ km/sec	$Sn = 4{\cdot}53$ km/sec

Refraction method:		
Mohorovičić	34·6 km	(calculated from *P*-impulses)
Mohorovičić	33·5 km	(calculated from *S*-impulses)
(Porkka[5])		
Kuusamo–Salla		
Pg_1 = 5·72 km/sec		Sg_1 = 3·38 km/sec
Pg_2 = 6·00 km/sec		Sg_2 = 3·53 km/sec
Pb = 6·70 km/sec		Sb = 3·90 km/sec
Pn = 8·10 km/sec		Sn = 4·59 km/sec
Refraction method:		
Mohorovičić	31 km	(calculated from *P*-impulses)
Mohorovičić	29 km	(calculated from *S*-impulses)
(Penttilä[6])		

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5.5. Etudes de la croûte terrestre faites pendant l'AGI par les séismologues français. Rapport sur les expériences d'enregistrements séismiques de la Sous-Commission des explosions alpines

par Mme Y. HENRI LABROUSTE

Les expériences d'enregistrements séismiques dans les Alpes occidentales ont été entreprises comme suite au vœu émis, à Rome, en séance plénière de l'Union géodésique et géophysique internationale:

« L'U.G.G.I., constatant l'intérêt présenté par les résultats déjà obtenus à l'occasion de grandes explosions en ce qui concerne la structure profonde de l'écorce, recommande aux gouvernements des pays intéressés par la zone alpine de participer à l'organisation de nouvelles expériences systématiques mettant en jeu de fortes charges explosives. La Commission séismologique européenne est chargée de l'organisation pratique des expériences et de la coordination des recherches. L'U.G.G.I. prie les gouvernements de contribuer à cette étude ».

Une Sous-Commission fut créée à cet effet au sein de la Commission séismologique européenne et, en 1955, le Comité national français de géodésie et géophysique prit l'initiative de réaliser en France les premières expériences, d'abord sous la présidence du R. P. LEJAY, puis sous celle du Prof. P. TARDI.

5.5.1 *Organisation*

L'organisation de ces expériences a constitué une innovation dans le domaine de la séismologie expérimentale. Une étroite collaboration internationale a, en effet, réuni sur le terrain des équipes allemandes, anglaises, françaises, italiennes et suisses, et s'est traduite également par la publication en commun des résultats.

Au point de vue géologique, trois régions principales entraient en ligne de compte pour l'exécution des tirs: en France, la zone plissée briançonnaise et l'arc des massifs cristallins, extérieur au précédent; en Italie, la zone d'anomalie positive de la gravité.

Trois expériences ont ainsi été réalisées, suivant un programme qui a fait l'objet de discussions au cours des réunions de la Sous-Commission des Explosions alpines et qui a été adopté par la Commission séismologique européenne: le premier à Vienne le 4 avril 1956, le deuxième à Utrecht le 8 avril 1958 et le troisième à Helsinki le 25 juillet 1960.

En 1956, l'exécution des explosions avait été assurée par la section technique du Génie militaire, sous la direction du Colonel Tessier; en 1958, l'organisation dans son ensemble a été prise directement en charge par la Section de séismologie, sous la direction de Mme Labrouste. Mais, dans les deux cas, les expériences n'ont pu être menées à bien que grâce au concours apporté au Comité national français par différents services officiels ou organismes privés (Centre national de la Recherche scientifique, Comité d'Action scientifique et Services de défense nationale, Bureau international de l'heure et Services radio-électriques, Electricité de France, Institut géographique national, . . .).

Les expériences de 1960 ont été réalisées en Italie par l'Institut de géophysique appliquée de Milan, sous la direction du Prof. L. SOLAINI et avec le concours de la Société des Pétroles A.G.I.P. Mineraria.

5.5.2 *Equipes séismiques et appareils*

Les équipes séismiques disposaient en 1956 de 19 stations mobiles ainsi réparties:

Les équipes allemandes, dirigées par le Dr. Closs, comprenaient deux équipes de l'« Amt für Bodenforschung » de Hanovre, une équipe de l'Institut de géophysique de Hambourg et une équipe de l'Institut de géophysique appliquée de Munich. Cette dernière utilisait un séismographe mécanique: les trois autres disposaient d'appareils de réfraction à amplification électronique et à plusieurs traces.

Les équipes de l'Observatoire géophysique de Trieste étaient dirigées par le Prof. Morelli. Les appareils utilisés étaient, d'une part, un séismographe type Héligoland modifié pour l'enregistrement sur le terrain et un appareil de réfraction à 12 traces du « Houston Technical Laboratory » muni de géophones ayant une fréquence de 2 Hz et, d'autre part, un laboratoire de réflexion.

Parmi les équipes françaises, celle de la Compagnie générale de géophysique, dirigée par MM. Geneslay et Richard, disposait d'un laboratoire de réflexion à 24 traces; les trois équipes de l'Institut de Physique du Globe de Strasbourg, sous la direction du Prof. Rothé, utilisaient deux séismographes Askania et un laboratoire de prospection de la Texas Industrial Company enregistrant en réfraction à courtes distances; enfin, les huit équipes de l'Institut de Physique du Globe de Paris que dirigeait Mme Labrouste disposaient de deux séismographes Mintrop à trois composantes, d'une station électromagnétique à trois composantes et de cinq stations de séismographes électromagnétiques à quatre traces.

En 1958, le nombre total des équipes d'observation s'est élevé à 34: 3 d'entre elles enregistraient en réflexion et 31 en réfraction.

Parmi les trois équipes de « Réflexion », deux opéraient en collaboration sur la piste d'accès au lac; (Compagnie générale de géophysique et Institut français du pétrole); la troisième, de l'Observatoire géophysique de Trieste, enregistrait sur le versant Italien.

Les 31 équipes de réfraction se répartissaient comme suit:

1°. dix équipes allemandes dépendant de sept organismes différents: (Amt für Bodenforschung, Hanovre: 3 équipes; Institut de géophysique de la Bergakademie Clausthal; Université de Hambourg; Université de Munich: 2 équipes; Ecole supérieure technique d'Aix-la-Chapelle; Service séismologique de Bade-Wurttemberg, Stuttgart; Société de prospection «Prakla», Hanovre);

2°. une équipe anglaise (British Petroleum);

3°. quatre équipes italiennes: (Institut de géophysique appliquée de Milan; Observatoire géophysique de Trieste: deux équipes A.G.I.P. Mineraria, Milan);

4°. une équipe suisse: (M. Susstrünck);

5°. dix équipes de l'Institut de Physique du Globe de Paris (avec le concours de divers organismes: Centre d'Etudes géophysiques et laboratoire de Bellevue du C.N.R.S.; Bureau de Recherches géologiques et minières; Centre scientifique et technique du bâtiment; Compagnie française des pétroles; Compagnie française de prospection séismique);

6°. quatre équipes de l'Institut de Physique du Globe de Strasbourg;

7°. une équipe du Service central hydrographique de la Marine.

En 1960, 42 équipes ont enregistré en Italie, dont 5 en réflexion (Institut de géophysique appliquée de Milan; A. G. I. P. Mineraria, Compagnie générale de géophysique, Observatoire géophysique de Trieste, Bergakademie Clausthal) et 37 en réfraction, à savoir 29 équipes allemandes appartenant aux Instituts énumérés ci-dessus, 6 équipes de l'Institut de Physique du Globe de Paris et 3 équipes de celui de Strasbourg.

La plupart des séismographes de réfraction utilisés en 1958 et en 1960 étaient munis d'amplification électromagnétique ou électronique et avaient des amplifications maxima comprises entre 300.000 et 5.000.000.

5.5.3 *Dispositif expérimental*

1. *Explosions.* Les tirs ont été exécutés conformément au programme résumé dans le Tableau 1.

TABLEAU 1

Lieu	Méthode de tir	Date	Charge (tonnes)
1956			
Lac Rond des Rochilles	Au fond du lac sous	25 août	
(Briançonnais)	10m d'eau	27 „	1
Altitude 2500m	Tir en nappe	29 „	2
Profondeur 10m	Charges unitaires	31 „	2
	de 80 à 90kg	3 sept.	5
		4 „	5
		6 „	10
1958			
Lac Négre (Mercantour)	Au fond du lac	4 sept.	0,09
Altitude: 2300m	charge concentrée	6 „	1
Profondeur maximum: 30m		9 „	5
		11 „	5
		15 „	10
		17 „	10
		20 „	25
1960			
Près de Levone	Forages de 45 à 60m	18 sept.	0,114
	Charges empilées	20 „	0,30
	dans les forages	23 „	0,45
Monte Bavarione	Au fond d'une galerie de 60m de profondeur; charge concentrée	26 „	2

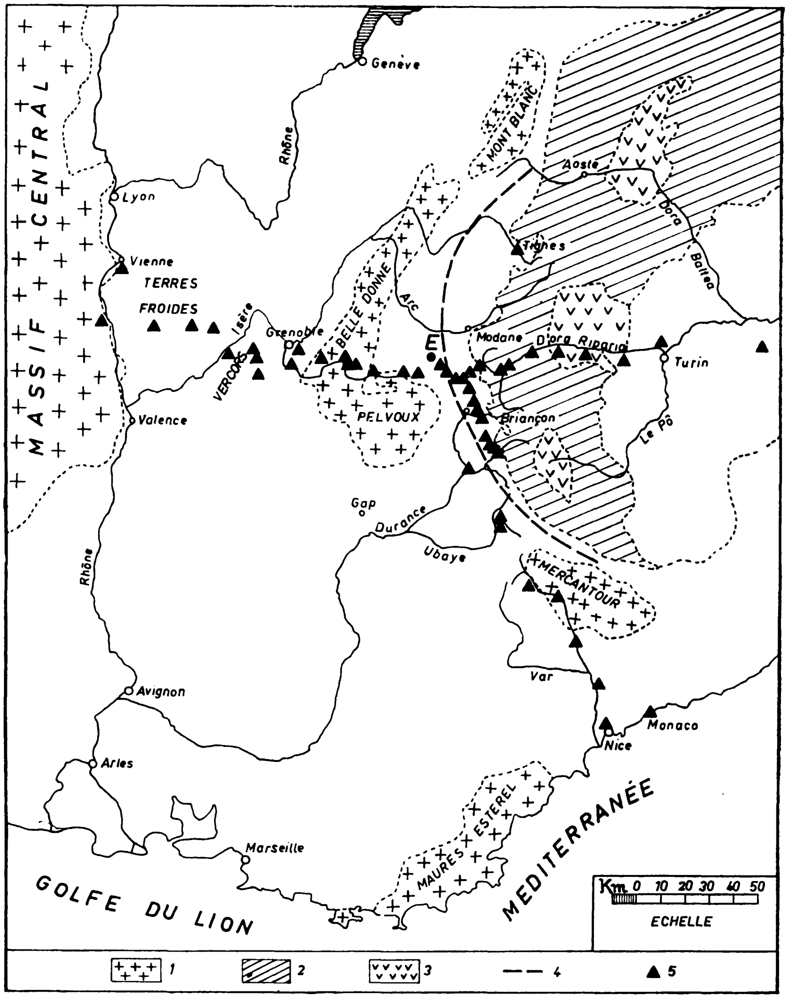


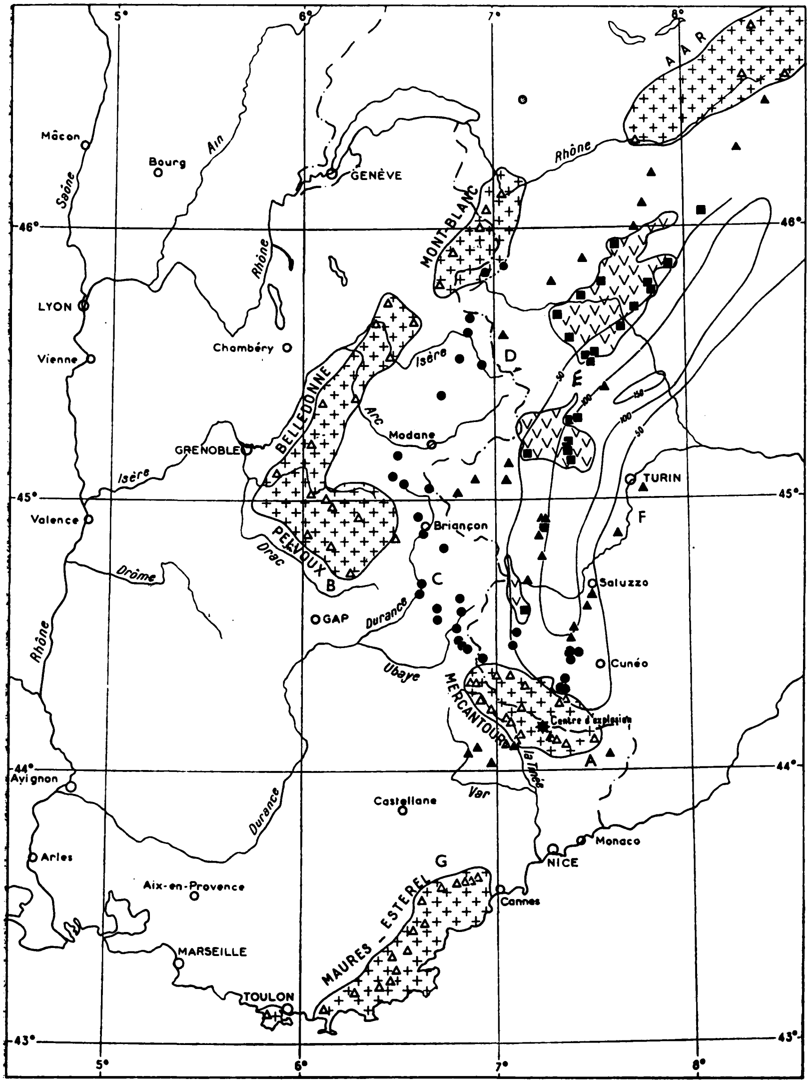
FIG. 5. 1. Massifs cristallins externes. — 2. Zone des schistes lustrés et massifs cristallins internes. — 3. Roches basiques. — 4. Axe de la zone briançonnaise. — 5. Stations séismographiques.

2. *Signal d'explosion et signaux horaires.* L'explosion et les enregistrements étaient repérés par rapport aux signaux horaires de Pontoise. Le signal d'explosion était en outre transmis par fil aux stations de prospection.

Des postes radio émetteurs-récepteurs, assuraient les liaisons entre le poste de tir et les stations d'observation.

3. *Stations d'observation.* En 1956, des enregistrements séismiques ont été obtenus en 61 stations réparties sur deux profils (Fig. 5).

L'un en direction sud-sud-est, suivait dans sa première partie l'axe de la zone briançonnaise, jusqu'à la haute vallée de l'Ubaye, et se prolongeait, à l'ouest du Mercantour, jusqu'à Nice. Ce profil (29 stations) était occupé par les trois équipes de Strasbourg et par sept équipes de Paris.



Massifs cristallins externes — Roches basiques

△ Stations du profil de référence MAURES-ESTEREL et des massifs cristallins externes

▲ Stations sur du permien

● Stations de la zone interne plissée

▲ Stations de la zone interne des massifs cristallins et schistes lustrés

■ Stations de la zone d'anomalie positive de la gravité sur des roches basiques

FIG. 6.

L'autre profil, transversal, s'étendait à l'est jusqu'à Casale Monferrato, dans la vallée du Pô, et vers l'ouest jusqu'à Peaugres, sur la rive droite du Rhône. Il était occupé dans la partie est (9 stations) par les équipes de Trieste et, dans sa partie ouest (23 stations) par les équipes allemandes et par une équipe de Paris.

En 1958, les 149 stations de réfraction ont été réparties le long de sept profils (Fig. 6):

A. un court profil vers le sud-est dans le massif cristallin du Mercantour;

B. vers le nord-ouest, un profil suivant la zone externe des Alpes et s'appuyant sur les massifs cristallins du Mercantour et du Pelvoux. Ce profil était prolongé par un alignement vers le nord-est, dans les massifs de Belledonne, du Mont-Blanc et de l'Aar;

C. dans la direction nord-nord-ouest, un profil suivant la zone plissée et constituant, dans sa partie comprise entre le lac Nègre et le camp des Rochilles, sur 125 km, un profil inverse de celui de 1956;

D. sur le versant italien, un alignement de stations vers le nord, dans la zone des massifs cristallins internes;

E. un autre alignement, dirigé vers le nord-nord-est et longeant la zone d'anomalie positive de la gravité;

F. un profil en direction de Turin;

G. enfin un profil de référence, extérieur au massif alpin et traversant les massifs cristallins de l'Estérel et des Maures.

En 1960, les tirs de Levone ont été enregistrés, d'une part, sur un profil orienté vers le sud-sud-ouest et suivant sensiblement le profil E de 1958 et, d'autre part, dans la direction nord-est sur le profil Levone-Monte-Bavarione. Le dernier tir a été enregistré, à partir de Monte-Bavarione, sur le profil sud-ouest inverse du précédent ainsi que sur un profil ouest-sud-ouest, en direction du Petit-Saint-Bernard.

Dans les trois expériences, les équipes de réflexion ont exécuté des profils de plusieurs kilomètres à proximité du point d'explosion.

5.5.4 *Publication des résultats*

Les résultats des expériences ont été discutés par le Groupe d'Etudes des explosions alpines au cours de différentes réunions tenues respectivement à Strasbourg les 9 et 10 décembre 1956, à Gif du 2 au 4 juin 1957 et du 1er au 3 avril 1959, à Münich les 22 et 23 mai 1960, et au cours de deux réunions restreintes à Strasbourg (8 au 12 juin 1960 et 8 au 12 mars 1961).

Des communications[1,2,3] à l'Académie des Sciences ont rendu compte des expériences de 1956 et de 1958 ainsi que des premiers résultats obtenus en 1956. Des rapports ont, en outre, été présentés à la Commission séismologique européenne (Utrecht 1958, Alicante 1959 et Helsinki 1960) ainsi qu'à l'Association internationale de séismologie[4] (Toronto 1957 et Helsinki 1960).

Il a été décidé, dans les réunions de Strasbourg (décembre 1956), Briançon (septembre 1958) et Gif (avril 1959), de préparer un mémoire collectif dans lequel seront réunis les résultats des expériences réalisées depuis 1956 par la Sous-Commission des explosions alpines.

Le manuscrit qui est en cours de préparation doit comprendre les chapitres suivants:

(1) Historique, organisation et mise en oeuvre des expériences.
(2) Aperçu gravimétrique et commentaires géologiques.
(3) Description du matériel d'observation: liste des stations comportant les coordonnées géographiques et une courte description géologique; reproductions des enregistrements.
(4) Exposé des résultats. A. — Réflexion. B. — Réfraction: tableau des temps de propagation bruts et corrigés, hodochrones.
(5) Essai d'interprétation.

5.5.5 *Commentaire des résultats*

A. *Ondes réfléchies.* — Les enregistrements obtenus en 1956 dans la zone plissée sont très complexes. Ils montrent un grand nombre de trains d'ondes se succédant jusqu'à 15s et plus; certains de ces trains pourraient être interprétés comme des réflexions multiples entre la surface et un socle à 2 km de profondeur. On peut toutefois se demander si les séries réfléchissantes de 9,8 à 10,7s ne proviennent pas de miroirs profonds.

En 1958, deux bases ont été occupées dans le Mercantour: la base CGG-IFP longue de 3900 mètres et la base O.G.S. Trieste beaucoup plus courte. La qualité de films après composition est très satisfaisante; leur aspect présente une analogie frappante avec les documents classiques de prospection séismique en terrains sédimentaires. La plage la plus riche en séries réfléchissantes se place entre 5,2 et 9,2 secondes sur l'un et l'autre documents, les principales réflexions se situant l'une entre 8,2 et 8, 6s et l'autre vers 6,5s.

Dans la zone d'Ivrée, un profil de 6400 m, a été jalonné par 101 traces, près de Levone.

Les films d'aspect complexe présentent de nombreuses séries d'ondes réfléchies dont les principales ont été relevées entre 3 et 3, 5s., et entre 7 et 8s. Les enregistrements de Monte-Bavarione sont en cours d'étude.

En résumé, l'examen de ces divers documents fait apparaître le besoin impérieux d'exploiter des profils continus encore plus longs pour obtenir une vue plus claire des phénomènes profonds.

B. *Ondes réfractées.* Nous nous limiterons aux résultats concernant directement le massif Alpin (profils de A à E et zone d'Ivrée).

Hodochrones (Fig. 7) *Premières arrivées.*

Phase Pg. Sur tous les profils, le début de l'enregistrement est constitué, à partir d'une distance de 5 à 10 km, par des ondes qui se sont propagées avec une vitesse apparente voisine de 6 km/s. Cette onde arrive en tête jusqu'à des distances au moins égales à 150 km, sauf en Italie, sur le profil E où une autre onde la précède à partir de 73 km.

Les observations relatives à la phase Pg définissent les hodochrones représentées par les équations

$$t = t_0 + \frac{\triangle}{v}$$

dont les coefficients t_0 et v sont donnés dans le Tableau 2.

Deux profils inverses (zone plissée et zone d'Ivrée) ont permis de déterminer la vitesse vraie, soit 6,06 km/s. En adoptant, pour représenter la phase *Pg*, l'équation moyenne

$$t_g = 0{,}42 + \frac{\triangle}{6{,}06}$$

on constate que, pour la plupart des stations du secteur nord (profils de B à E), les écarts entre les temps observés et les temps calculés restent compris entre plus ou moins 0,1s, ce qui met en évidence l'homogénéité des résultats concernant la phase Pg sur ces divers profils.

Anomalie. Par contre, les expériences de 1960 dans la zone d'Ivrée ont mis en évidence une remontée de la couche à vitesse 6,06 km/s, l'ordonnée à l'origine étant inférieure à 0,2s. La même anomalie s'observe en cinq stations de la zone plissée situées de part et d'autre de la frontière franco-italienne, entre le Val d'Isère et le Petit-St. Bernard.

TABLEAU 2

Profil	t_0 s	v km/s	$\triangle$km	observations entre et $\triangle$km
A	0,22	6,05	5,9	29,9
B	0,4	6,0	13,0	34,5
B	0,53	6,11	97,8	212,2
C	0,38	6,06	44,4	119,4
	0,20	6,06	119,4	169,9
D et E	0,63	6,03		
Sud 1956	0,50	6,07	11,9	124,7
Levone Sud	0,18	6,06	12,1	62,6 vers SW
	0,04	6,08	16,1	67,9 vers SSW
Levone N	0,18	6,06	2,4	23,5
Monte Bav. S	0,18	6,06	16,9	53,5

Amplitude. On doit remarquer que l'amplitude de la phase Pg décroît rapidement avec la distance.

Phase Pb. Au delà de 70 km apparaît en Italie, l'importante anomalie qui a été découverte au cours des expériences de 1958: alors que, pour l'ensemble des stations situées dans le secteur nord-ouest par rapport au lac Nègre (profils B, C et D), les temps correspondant aux premières arrivées se placent encore sur l'hodochrone Pg, on constate qu'entre 73 et 86 km, les premières ondes enregistrées dans la direction d'Ivrée (profil E) s'alignent sur une hodochrone indiquant une vitesse apparente de 7,0 km/s (t_0 = 2,0s). Au delà de 110 km, ces ondes arrivent avec une avance importante et une vitesse apparente plus grande

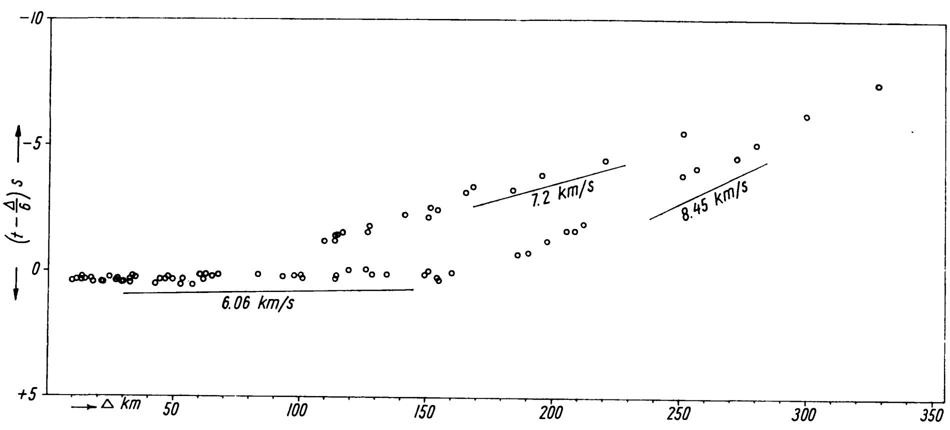


FIG. 7.

(entre 175 et 275 km, v = 7,2 km/s et t_0 = 0,3s), marquant une forte dénivellation de la surface de discontinuité entre 86 et 175 km.

Les expériences réalisées en 1960 dans la zone d'Ivrée étaient spécialement destinées à l'étude de cette phase. En raison de la faible énergie des explosions, les ondes Pb n'ont pu être enregistrées en première arrivée qu'en trois stations sur le profil SSW à partir de Monte–Bavarione.

Phase Pn. Les heures de début aux stations situées au-delà de 170 km dans les massifs de Belledonne et du Mont-Blanc et au-delà de 270 km en Italie et en Suisse s'alignent sur la droite (v = 8,4 km/s, t_0 = 8,3s). Dans l'hypothèse d'une vitesse vraie de 8,1 à 8,2 km/s, la vitesse apparente trouvée indique une pente des surfaces de discontinuité remontant dans la direction NE à partir du massif de Belledonne vers le massif de l'Aar.

Anomalies. En résumé les hodochrones des premières arrivées d'ondes permettent de délimiter les zones d'anomalie.

La zone de grande anomalie d'Ivrée, s'étend dans la direction SW-NE de l'ouest de Turin à la frontière suisse. Une zone de plus faible anomalie la prolonge vers le sud.

Une autre zone, définie par l'anomalie des Pg, a déjà été signalée plus haut. Ces résultats sont en bon accord avec ceux de la gravimétrie.

5.5.6 *Corrélations. Phase Pb en seconde arrivée*

Entre 90 et 130 km, on observe dans la zone plissée deux alignements parallèles (v = 6,7 km/s, t_0 = 3,6s et t_0 = 4,2s) mettant en évidence une forte dénivellation de la couche à vitesse voisine de 7 km/s, d'ouest à l'est, entre la zone plissée et la zone d'Ivrée.

Corrélations des grandes ondes. L'interprétation plus complète des observations, notamment en ce qui concerne la surface de Mohorovicic est difficile, d'une part, en raison de la présence de la grande anomalie mise en évidence

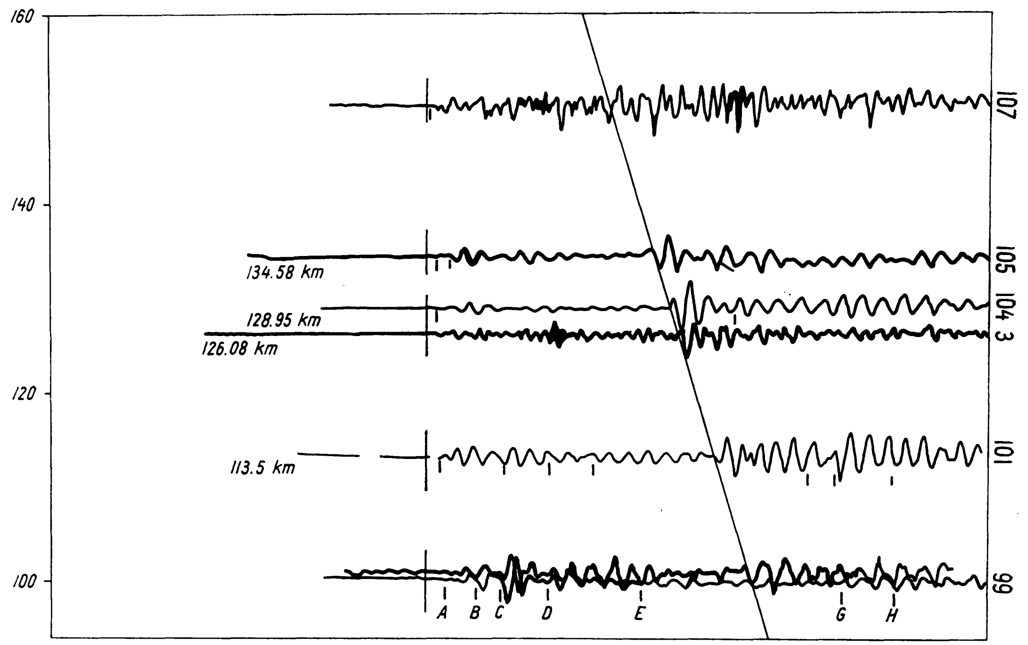


FIG. 8.

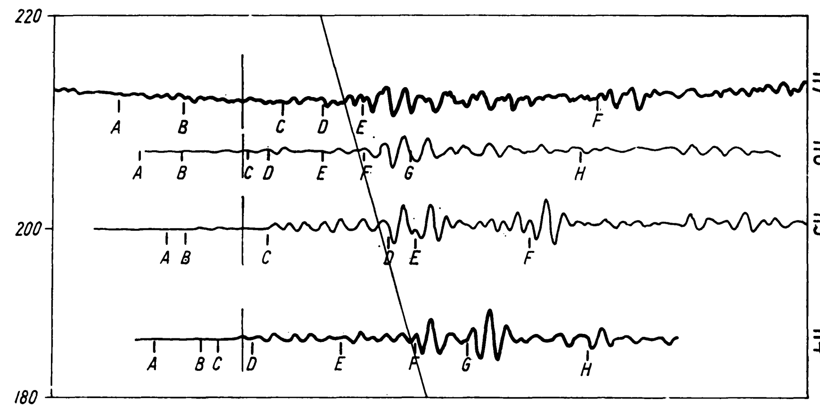


FIG. 9.

dans la région d'Ivrée et, d'autre part, du fait que les premières arrivées d'ondes enregistrées aux stations éloignées avaient une énergie très faible et qu'elles se sont souvent trouvées noyées dans le bruit de fond, ce qui rendait les observations douteuses. Il était donc important de compléter les résultats obtenus d'après ces premières arrivées d'ondes en essayant d'interpréter les plus grandes phases des enregistrements (dans la partie de ces derniers qui précède l'arrivée des ondes transversales).

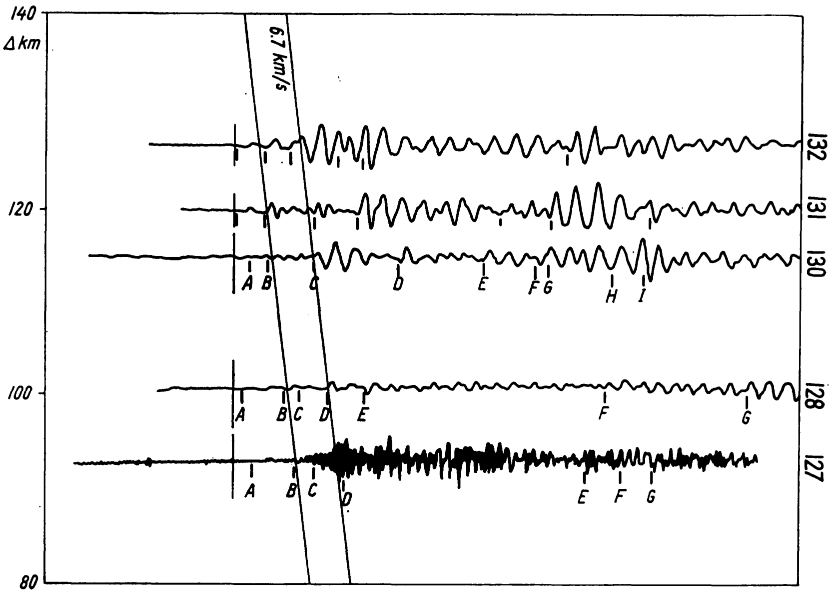


FIG. 10.

L'interprétation des grandes phases pose avant tout un problème de corrélations rendu difficile par la structure complexe du massif alpin ainsi que par l'hétérogénéité du matériel d'observation et le nombre encore insuffisant des stations pour une étude détaillée des structures.

Il a été fait appel à différentes méthodes pour essayer de résoudre ce problème de corrélation: assemblages de séismogrammes après réduction des échelles de temps à une valeur commune et modèles à trois dimensions. De tels modèles ont été réalisés et étudiés au cours des réunions de travail tenues à Strasbourg: les uns utilisant une représentation conventionnelle des phases, les autres des reproductions sur plexiglass des enregistrements.

En ce qui concerne les grandes ondes qui paraissent liées à la surface de Mohorovičić, les corrélations les plus satisfaisantes concernent la zone externe des Alpes (assemblages suivant les directions d'azimut 315° (Fig. 8) et 353° (Fig. 9).

Dans la zone plissée (Fig. 10), les interférences entre trains rendent déjà ces corrélations précaires.

Il paraît cependant possible de faire état du retard de 0,5s, que présentent, dans cette zone, les premières grandes ondes, par rapport à celles de la zone externe.

Dans la zone d'Ivrée, les corrélations deviennent très douteuses: étant donnée leur faible vitesse apparente dans cette zone, les grandes ondes pourraient aussi bien se rapporter à la surface de Conrad qu'à celle de Mohorovičić (Fig. 11).

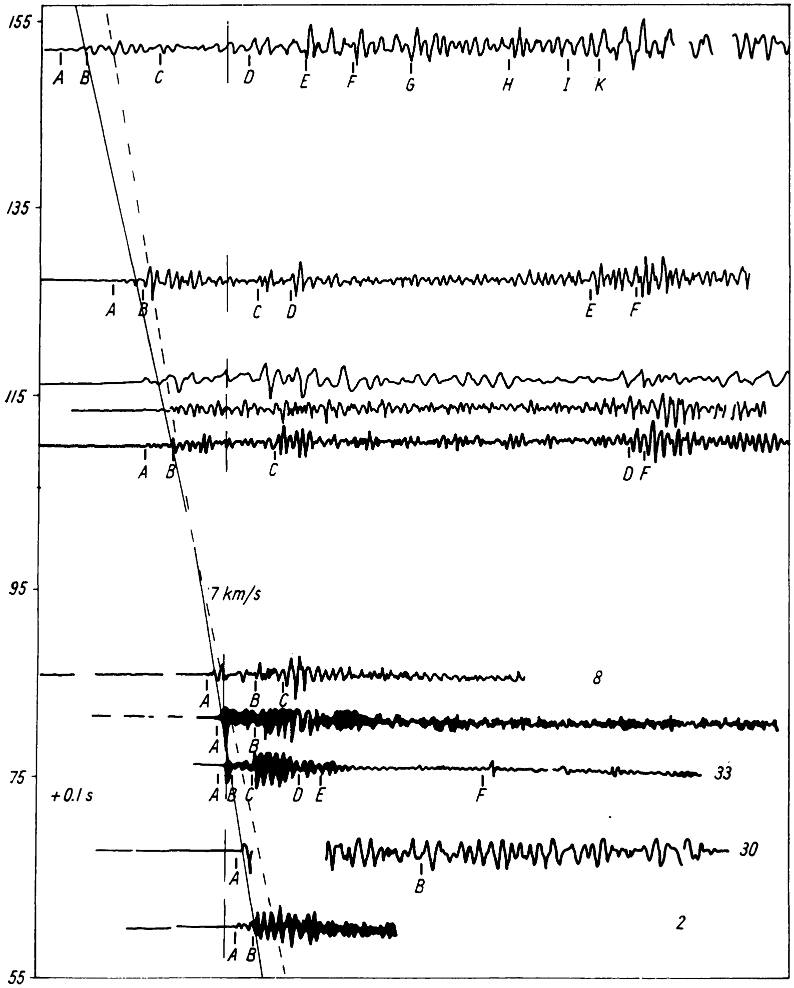


FIG. 11.

5.5.7 *Conclusions*

Des calculs sur modèles qui ont été effectués par le groupe de Strasbourg ont permis de préciser l'état du problème, d'autres calculs sont encore en cours.

Le résultat le mieux acquis est l'existence d'une remontée importante de la couche à vitesse voisine de 7 km/s dans la zone d'Ivrée, avec un fort gradient à l'ouest, remontée qui l'amène à une profondeur très faible dans la région du maximum d'anomalie.

De nouvelles expériences sont vivement souhaitables pour préciser la forme de la surface de discontinuité dans cette zone.

L'autre résultat important, pour lequel des compléments d'information sont également souhaités, est l'approfondissement de la surface de Mohorovicić, de la zone externe vers la zone interne, montrant l'existence d'une « racine » des Alpes.

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5.6. Seismic investigation of the earth's crust in the Western part of Germany

by A. Stein

Large explosions are being effected, occasionally, in various quarries. Most of them are chamber blasts, in a few cases they are drillhole explosions. Such explosions have been used for a long time for the seismic investigation of the structure of the earth's crust[2]. Since the beginning of 1958, they have been recorded with seismic refraction instruments within the programmes of research of university geophysical institutes and of the Geological Surveys. The recorded refraction profiles were in general so long that the refraction arrivals of the Moho–discontinuity were recorded as first arrivals. The objective of such research is to ascertain the structure of the earth's crust, and the relief of the main discontinuities down to the Moho–discontinuity.

Profiles recorded so far have been included in a general map (Fig. 12). The following table shows the coordinates of the relevant explosions. The last column gives the number n of the refraction observation crews.

Location	Date	charge tons	n
Eschenlohe/Obb. $\lambda = 11°08',8$ $\varphi = 47°37',9$	15. 2.58	12	18
	27. 6.59	16	18
	17.10.59	8,8	12
	30. 1.60	8,3	20
	30. 4.60	11	37
	2. 7.60	7,3	37
	3. 9.60	7,2	11
	19.11.60	12	28

Location	Date	charge tons	n
Seiferts/Rhön $\lambda = 10°02',4$ $\varphi = 50°32',6$	28. 3.58 18. 9.58 11. 3.60 20. 5.60	10 10 7,6 8,6	10 10 17 22
Gersfeld/Rhön $\lambda = 9°53,0$ $\varphi = 50°26,6$	12. 6.58 3. 7.59	4,6 14	7 20
Lenggries/Obb. $\lambda = 11°34'$ $\varphi = 47°43'$	22.11.58	2	26
Großenritte b. Kassel $\lambda = 9°23'$ $\varphi = 51°16'$	27. 2.59	5	13
Birkenau/Odenwald $\lambda = 8°43'$ $\varphi = 49°33'$	3. 4.59	7,3	15
Adelebsen/Solling $\lambda = 9°44',5$ $\varphi = 51°36',7$	12. 6.59	8,2	16
Kirchheimbolanden/Pfalz $\lambda = 7°58'$ $\varphi = 49°40'$	13. 4.60	2,6	21
Böhmisch-Bruck $\lambda = 12°21',3$ $\varphi = 49°34',1$	9. 7.60	3,6	22

Especially large quarry blasts at Eschenlohe made possible radial recording, with profiles in all directions. The evalution of these profiles is partly completed, part is still in progress.

On the NE profile proceeding from Eschenlohe were recorded, in the range 125–220 km, first arrivals of the Moho-discontinuity which occur with an apparent velocity of 7·9 km/s. The basement appears to have a velocity of 6·1 km/s. Evidence of the Conrad-discontinuity is not yet available. Even the profile to the North does not indicate, so far, any signals from the Conrad discontinuity.

From the reversed profiles Seiferts–Eschenlohe we obtain a true velocity of about 8·4 km/s for the Moho-discontinuity. One profile from Eschenlohe to the WNW serves as reversed profile to the Haslach profile of 1948[3].

A SW profile continues from Eschenlohe via the Gotthard-massif to a distance of about 300 km and is being carried out in conjuction with investigations made in 1956 and 1958 in the Western Alps in cooperation with several European countries. An S profile across the Alps has only recently been started.

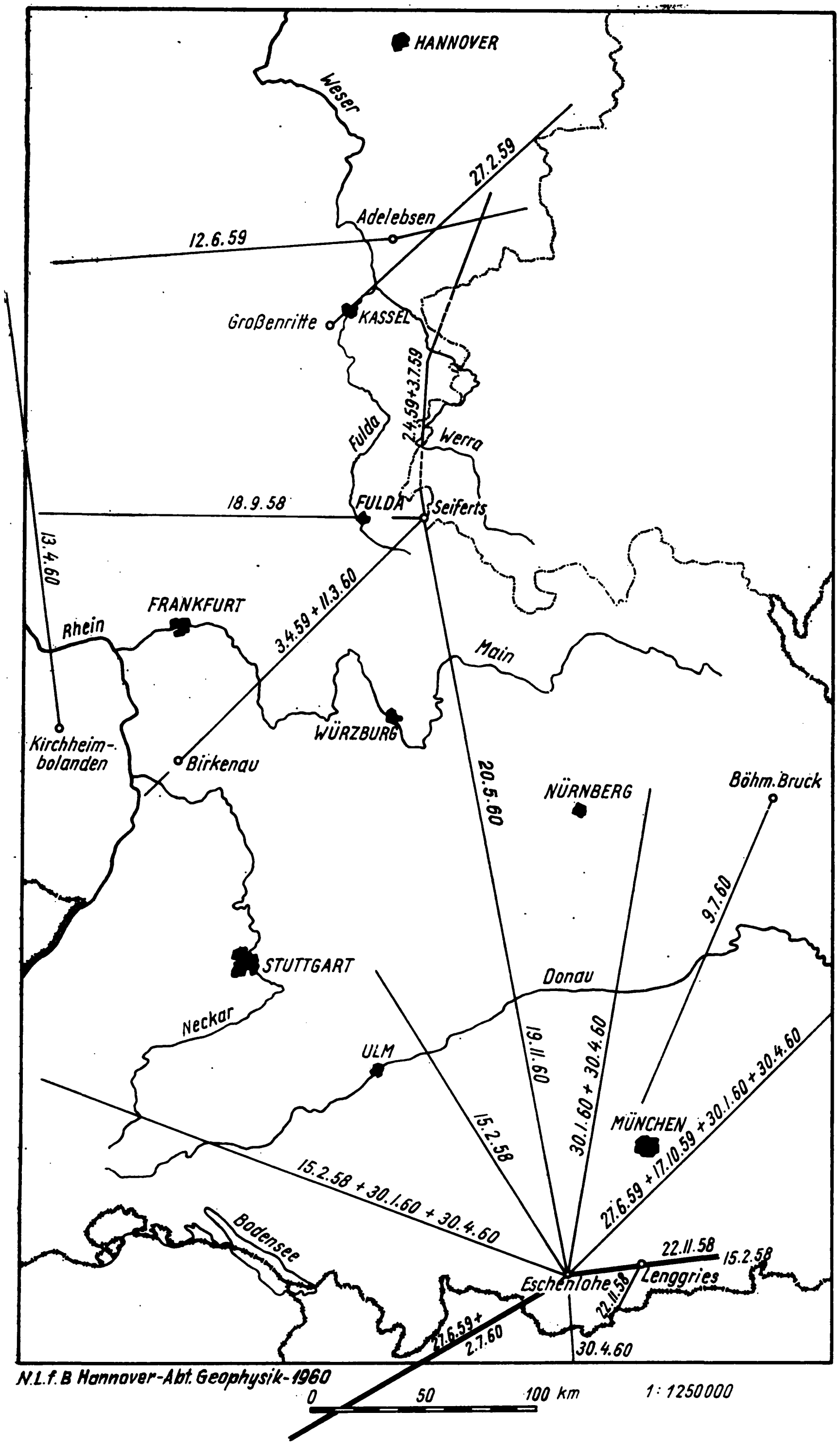
HANNOVER
Weser
27.2.59
Adelebsen
12.6.59
KASSEL
Großenritte
24.4.59+3.7.59
Fulda
Werra
18.9.58
FULDA
Seiferts
13.4.60
FRANKFURT
Rhein
3.4.59 + 11.3.60
Main
WÜRZBURG
Kirchheim-bolanden
Birkenau
20.5.60
NÜRNBERG
Böhm. Bruck
9.7.60
STUTTGART
Donau
Neckar
ULM
19.11.60
30.1.60 + 30.4.60
MÜNCHEN
15.2.58
27.6.59 + 17.10.59 + 30.1.60 + 30.4.60
15.2.58 + 30.1.60 + 30.4.60
Bodensee
22.11.58
15.2.58
Eschenlohe
Lenggries
22.11.58
27.6.59 + 2.7.60
30.4.60
N.L.f.B Hannover-Abt. Geophysik-1960
0 50 100 km
1: 1250000

FIG. 12.

The profile Birkenau–Seiferts has been observed in both directions. On the remaining profiles, some of the distances between points of observation are so large as to render necessary auxiliary recordings. It is, however, expected that we shall succeed in combining all these seismic data into one uniform picture of the crust's structure.

In connection with reflection-seismic oil prospecting methods we obtained, in some parts, reflections down to the Moho-discontinuity by increasing the recording-time.

Together with recordings of large explosions, special gravimetric and magnetic regional measurements were made with a view to supplementing seismic evaluation data.

During the investigations, a number of different types of instruments were employed, some of which were especially constructed for the research program by the respective institutes. It was found that in spite of these differences a correlation of the later refraction signals is possible, provided the distance between stations is limited to 5–6 km. A large part of the equipment consists of geophones (natural frequency 1–5 cps), low-frequency amplifiers and oscillographs. Even with instruments without electronic amplification good results have been obtained by using sensitive galvanometers. The maximum amplification at 5 cps reaches 600,000 for most of the instruments. There are some three component-stations. Recording speeds vary between 3 cm/s and 8 cm/s. Some of the instruments are portable and can be set up even in an inaccessible area. This proved to be of considerable advantage, especially during investigations in the Alps.

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5.7. Research on crustal structure in Hungary

by L. STEGENA

Research on the structure of the Earth's crust was continued during the International Geophysical Year. The present paper summarizes the results of the Hungarian investigations. In order to determine crustal structure, the following techniques were used:

1. Reflection techniques
2. Refraction techniques
3. Analysis of waves due to splitting of waves on interfaces
4. Joint evaluation of regional gravity anomalies and seismic data
5. Analysis of earthquake travel-time graphs
6. Study of dispersion of surface waves

(1) At present, deep soundings with the reflection method have yielded useful results at nine points. We used 23 to 500 kg of explosive, in boreholes. Recordings were made with seismometers of 7 cps period, located from 500 to 2000 metres from the shot point.

For 8 of the above 9 points, two reflections were obtained, which can be related to the Conrad, and the Mohorovičić interfaces. At one point, only the reflection was observed. In some points, weak reflections or traces of reflections were obtained from shallower depths, which could correspond to the Förtsch interface.

(2) The time data obtained by the reflection method were recomputed in terms of interface depth by utilizing the velocity data yielded by refraction. The charges used for refraction shooting amounted to 1–500 kg, shot in holes. The recordings were effected at distances of 1 to 142 kilometres, with seismometers of 7 cps frequency.

At present, two profiles have been made in Hungary for the purpose of crustal research. The interface velocities thus determined do not differ significantly from velocities observed elsewhere in Europe. The interface depths obtained by refraction seem to fit quite well the data obtained in the neighbourhood by reflection methods. This suggests that neither the reflection, nor the refraction results were significantly modified by velocity anisotropy, multiple reflections or the like.

Refraction measurements have shown that the intensity maximum of the seismic wave in the case of great distances of observation is situated at considerably lower frequencies than hitherto supposed. To illustrate this point, two refraction records are shown (Fig. 13); these are records of one and the same shooting, with two different apparatuses. One of them was an ordinary apparatus, unable to transmit low frequencies. The other was an apparatus expressly built for crustal research, fully transistorized, and with good low-frequency

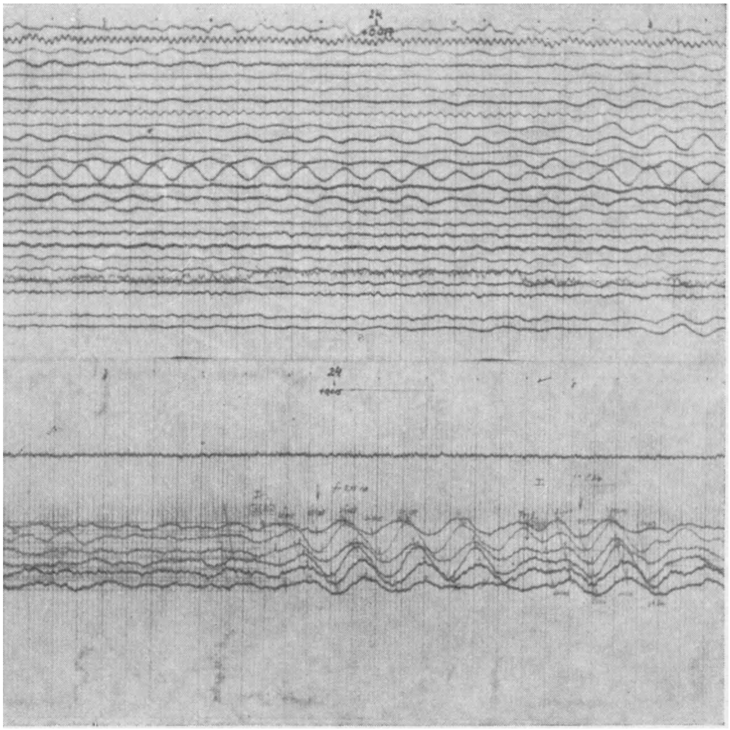

FIG. 13.

transmission properties. The first of the apparatuses yielded no useful record, while the other recorded fair arrivals of 8 cps quasi-frequency.

(3) Crustal structure can be determined from earthquake records in the following way: a *P* wave arriving at the Mohorovičić interface from a given epicenter generates a *P* and an *S* wave at that interface. From the time difference of the arrivals of these two waves, it is possible to compute the thickness of the crust. For the purposes of this computation, the velocity data obtained from the refraction measurements were utilized.

In Hungary four seismological stations (Budapest, Kecskemét, Szeged, Kalocsa) exist. The calculation was performed for all of these stations, giving the thickness of the crust below each station. The results obtained appear to agree with the data obtained in other ways.

(4) By utilizing gravity data, crustal research was performed based on the following premise: within the Carpathian Basin, the average density of the sediments is 2·23 g cm^{-3}, while that of the sediments basement is 2·67 g cm^{-3}. These two formations offer a rather sharp density contrast. Subtracting from the Bouguer anomaly the anomaly due to the sediments, we obtain the residual anomaly characteristic of the crustal structure. The relation between the residual anomaly and the thickness of the crust can be determined if the grav-

ity, sediment thickness and crustal thickness data are known at a sufficient number of points.

By this method, an attempt was made to determine in detail the crustal structure of West Hungary. The crust seems to present here a configuration much resembling the African rift valleys.

(5) The records of the Hungarian earthquakes of 20th February, 1950 and 12th January, 1956 were also evaluated from the point of view of crustal thickness determination. In this work, the importance of the records of the Hungarian stations was rather slight, because of their small distance from the epicenter.

The evaluations using the data of Hungarian, Austrian, Czechoslovak, German, Polish, Rumanian and Soviet recordings give a crustal thickness of 33 to 35 km. This value is significantly greater than those yielded by reflection and refraction work. The deviation can be explained by the fact that most of the stations involved are situated beyond the Carpathian ranges.

(6) The Budapest recordings of two shocks in Burma were also utilized to determine crustal thickness. The evaluation based on the dispersion of the Love wave yielded a rather small crustal thickness, although most of the wave path lay in the high mountainous areas of Asia.

Figure 14 shows the crustal thickness data (in kilometres) obtained for Hungary by the methods listed under 1, 2 and 3.

The study of the Earth's crust in Hungary is interesting in a number of respects. The country occupies part of the Carpathian basin. This basin subsided considerably in the Pannonian stage and subsequently. The subsidence was not uniform; the basin can be divided into some seven smaller basins; nevertheless, most of the country is covered by early Tertiary and Quaternary deposits seldom thinner than 1000 m.

Consequently, Hungary is an area where the features of the crustal structure peculiar to basin areas can be excellently studied.

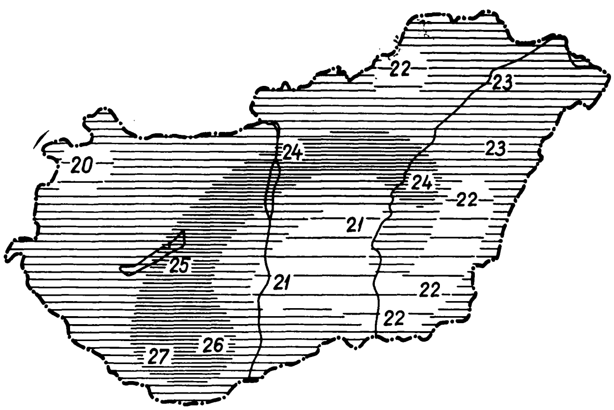

FIG. 14.

Thus far, the results permit the following conclusions to be drawn.

1. The twofold subdivision of the crust — Conrad and Mohorovičić interfaces — is also characteristic of Hungary.
2. The boundary velocities of the individual layers do not differ from values obtained elsewhere in Europe.
3. The thickness of the Hungarian part of the crust is remarkably small. This is due partly to the thinness of the gabbro layer.
4. It seems that, as regards areal distribution, the crust is especially thin along the borders of the basin and between the Danube and Tisza rivers. In the other parts of the basin, the crust is relatively thicker.

As regards further research, in addition to the necessity for a further development of the network of crustal thickness determinations, it would be necessary to obtain data from the neighbouring areas, so as to be able to compare the crust of the basin with that of the surrounding mountains. It is our intention to carry out such measurements in co-operation with the geophysicists of neighbouring countries.

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5.8. Etudes concernant la constitution de la croûte terrestre sous le territoire de la Roumanie

La structure de la croûte terrestre sous le territoire de la R.P.R. a fait l'objet de quelques études dont nous résumons ci-dessous les résultats.

1. DEMETRESCU, G. 1953. *Bull. Sc. Acad. R.P.R.* V, No. 2.

Epaisseur de la couche de granit, nord-est de Bucarest 29 km.

Epaisseur de la couche de basalte 25 km.

2. DEMETRESCU, G. et PETRESCU, G. 1954. *Revue de l'Université*, No. 4–5.

Epaisseur de la croûte terrestre dans la région de Vrancea 75 km, dans la région de Bucarest 54 km.

3. DEMETRESCU, G. 1958. *Studia Geophysica*, No. 2, Prague.

L'auteur signale certaines impulsions secondaires entre les ondes *P* et *S*, sur les enregistrements des séismes de Vrancea dans les stations roumaines et indique une méthode d'utilisation de ces impulsions pour la determination des épaisseurs des couches de la croûte terrestre.

4. ENESCU, D. et RADU, C. 1958. *Stud. Cerc. Astr. Seism.* No. 1.

Epaisseur de la couche de granit à Bucarest 35 km, de celle de basalte 28 km.

5. RADU, C. 1958. *Stud. Cerc. Astr. Seism.*, No. 2.

Epaisseur de la couche de granit à Campulung 33 km, de celle de basalte 24 km.

6. DEMETRESCU, G. et ENESCU, D. 1960. *Com. Acad. R.P.R.*, No. 10.

Epaisseur de la couche de granit, région de Bucarest 43 km, de celle de la couche de basalte 22 km.

7. DEMETRESCU, G. et ENESCU D. 1960. *Stud. Cerc. Astr. Seism.*, No. 2.

Epaisseur de la couche superficielle à Focsani 12 km.

5.9. The United States program of seismic crustal studies during the IGY

by G. P. WOOLLARD

5.9.1. *Introduction*

The United States program of seismic crustal studies during the International Geophysical Year can be divided into two parts, marine and continental measurements. Of the two, the marine program was the more extensive both in terms of number of observations and geographic areas studied. The following report is subdivided on the basis of areas investigated.

5.9.2. *Marine program*

Three scientific institutions collaborated in carrying out the marine program, the Scripps Institution of Oceanography of the University of California, the Lamont Geological Observatory of Columbia University and the Woods Hole Oceanographic Institution. The program was also materially aided by the direct participation of scientists from other nations and by ships of the Brazilian Navy, the Argentine Navy and the South African Navy. Although some of the measurements were studies of the surficial bottom sediments and the crustal material immediately beneath them, about one fourth were complete

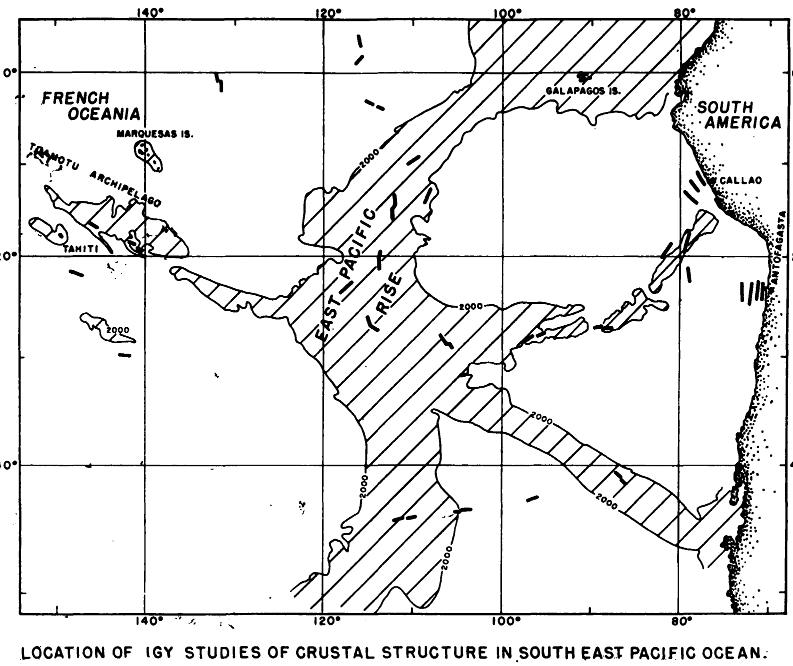

FIG. 15. Location of IGY studies of crustal structure in Southeast Pacific Ocean.

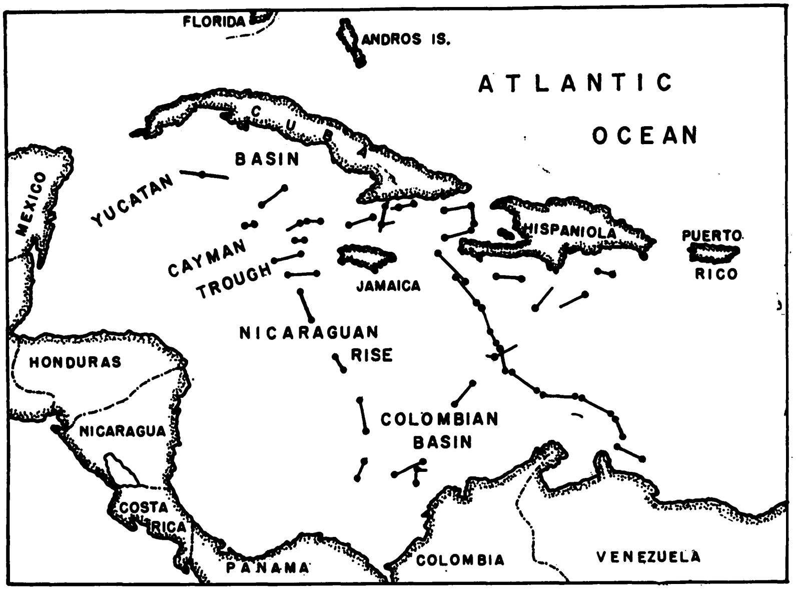

FIG. 16. Western Caribbean, showing locations of seismic observations.

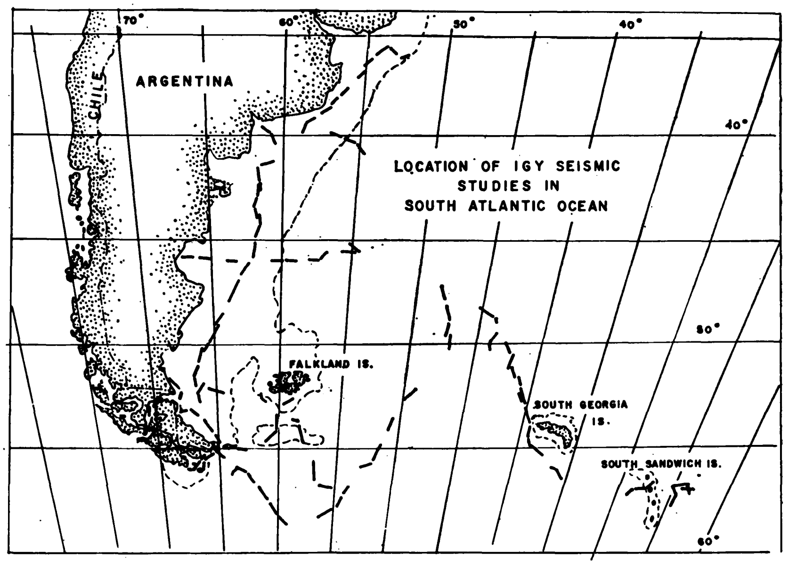

FIG. 17. Location of IGY seismic studies in South Atlantic Ocean.

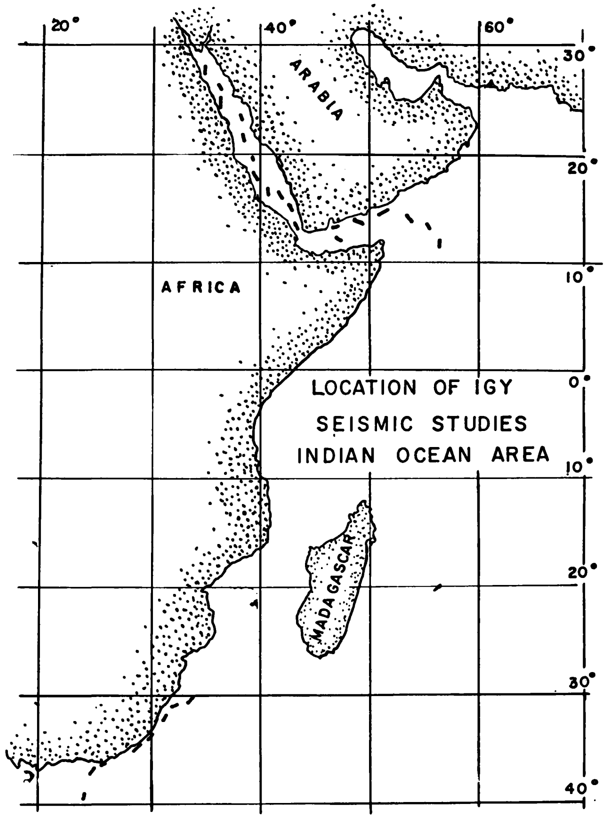

FIG. 18. Location of IGY seismic studies, Indian Ocean area.

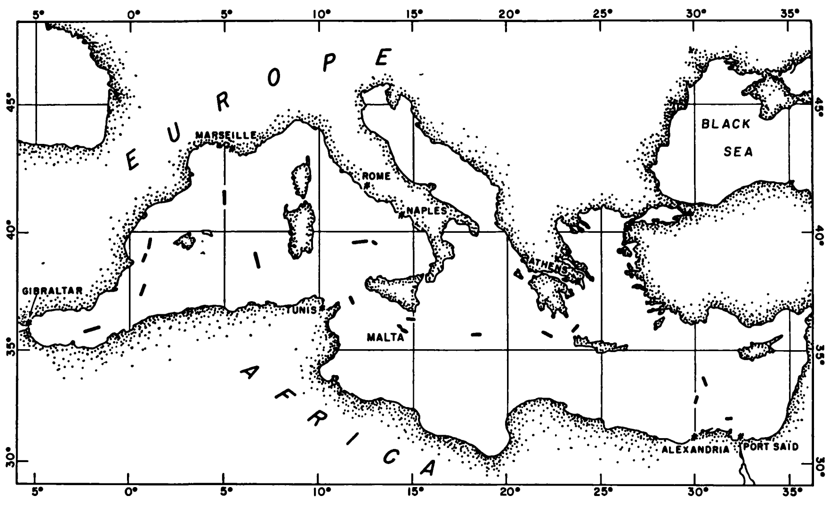

FIG. 19. Location of IGY seismic studies of crustal structure in Mediterranean Sea.

measurements of the crust above the Mohorovičić seismic discontinuity. The program of seismic measurements was integrated with other scientific studies of a geological nature, including continuous recording fathometer measurements, continuous magnetic measurements, heat-flow measurements, bottom photography, bottom cores and rock dredge hauls. Although most of the seismic studies were refraction measurements, reflection measurements were also made where only the bottom sedimentary column was being investigated. Although no seismic measurements of crustal thickness were attempted in the Arctic Ocean from the drifting ice floe stations, many of the other related investigations such as bottom cores, depth of water, reflection measurements, bottom photography, magnetic measurements and gravity observations were conducted. Gravity measurements are useful for studying the earth's crust and, although not a part of the regular oceanographic program, were observed with considerable success on one trans-Atlantic cruise by means of a shipboard gravimeter.

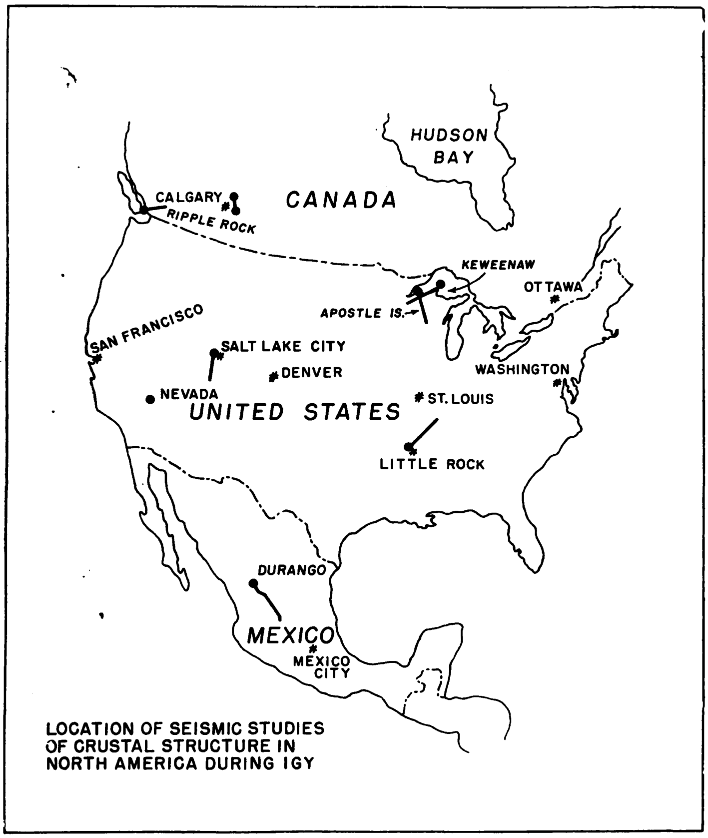

FIG. 20. Location of seismic studies of crustal structure in North America during IGY.

The operating areas covered by the respective groups were almost global in extent. The program of the Scripps Institution of Oceanography was concentrated in the South Pacific area between French Oceania and the coast of South America. Areas of particular interest studied were the Tuamotu Archipelago platform, the East Pacific Rise and the Peru–Chile Trench. In all, thirty nine seismic refraction stations were established. With the exception of five of these, which were shot for the study of the sedimentary column only, all were long range measurements for the determination of crustal thickness. The location of these measurements is indicated in Fig. 15. A narrative account of the Scripps expedition is given in *IGY General Report* No. 2, 26 June, 1958, published by the U.S. National Academy of Sciences for IGY World Data Center A.

The Lamont Geological Observatory, working in part with the collaboration of the Woods Hole Oceanographic Institution and in part with research vessels made available through the IGY National Committees for Brazil, Argentina and South Africa, participated in what was probably the most extensive series of seismic refraction measurements ever undertaken. The areas studied were the Caribbean Sea between Central America and the Greater Antilles of the West Indies, the eastern continental border of South America from Buenos Aires, Argentina to Cape Horn, the Scotia Sea area between the Falkland Islands and the Sandwich Islands, the Cape of Good Hope area and the Gulf of Aden, Red Sea and Mediterranean Sea. In all, 199 seismic refraction stations were established. Of these approximately one fifth yielded information on the thickness of the crust. Figs. 16, 17, 18 and 19 show station locations in the respective areas studied.

5.9.3. *Continental program*

The continental measurements of crustal structure, based on explosive blasts, were carried out in five areas during the International Geophysical Year as part of the U.S. seismic program: the central shield area of exposed crystalline basement rocks and pre-Cambrian sediments in the Great Lakes Region of the United States; the buried shield area near the head of the Mississippi embayment in the southern United States; the high plateau of central Mexico; the Andean high plateau of South America; and the adjacent Andean foothill region. In addition to the measurements made under the IGY program, the University of Utah Department of Geophysics made a seismic study of crustal structure in the Great Salt Lake area of central Utah. United States government agencies also made a study of crustal structure in Nevada. These measurements, plus two made under the Canadian IGY program, one on the Pacific coast and one in the Rocky Mountain foothill region of Alberta, give a total of eight measurements in North America during the IGY period. Fig. 20 shows the locations of the measurements.

Seismic measurements in South America were conducted by the Department

FIG. 21. Location of IGY studies of crustal structure in Andean region of South America.

of Terrestrial Magnetism of the Carnegie Institution of Washington in collaboration with scientists from the University of Chile and San Marcos University in Peru, and with the assistance and cooperation of the governments and IGY Committees of Chile and Peru. Seismic measurements were conducted by the University of Wisconsin in Mexico with the help and collaboration of scientists of the University of Mexico and the Mexican government and IGY Committee; seismic measurements in the Great Lakes region were conducted through the kind agreement of the Canadian government.

Although the basic procedures used by both groups were the same, the recording procedure varied somewhat. As shown in Fig. 21, the Department of Terrestrial Magnetism group recorded partial profiles along several azimuths around the shot sites. The University of Wisconsin recorded essentially in line along a single azimuth laid out so as to minimize possible changes in structure, as indicated by the gravity anomaly pattern and the surficial geology. Another difference was that the University of Wisconsin group took detailed measurements at close ranges in order to determine the effect of the surficial geology whereas this factor was disregarded in the Department of Terrestrial Magnetism measurements. Both procedures can be justified; that adopted will depend upon the type of area, accessibility, the frequency of shots, and the facilities and time available.

The Department of Terrestrial Magnetism utilized large blasts fired in connection with mining operations. The University of Wisconsin fired their own charges in the Great Lakes area and utilized quarry and mining operation blasts in the other areas studied. Measurements by both groups were carried out beyond 300 km in order to obtain first arrivals from the underlying mantle rock. Because of the great distances involved the instrumental requirements are much more critical than in marine crustal measurements, and the return in terms of number of measurements is much more limited for the amount of time and effort expended. Other factors, such as land use, property rights, background noise, seismic disturbance of isolated seismometer installations, the need for many recording sites and sets of equipment, etc., make continental crustal measurements a rather complicated undertaking requiring much planning as well as time.

5.9.4. *Results of marine crustal program*

5.9.4.1. *Pacific area measurements*

In general the oceanic measurements corroborated earlier seismic studies in the North and Central Pacific. As a rule the sedimentary layer is only a few hundred meters thick except in the equatorial carbonate area where the thickness is about two times the normal oceanic value. There is perhaps more variation in overall crustal thickness than observed previously, and although the thickness of the basic crustal layer having a velocity of about 6·7 km/sec is fairly constant, many stations showed an intervening intermediate velocity layer which averages about 1 km in thickness and has a velocity varying from 4 to 6 km/sec.

Unusual results were obtained in four areas. (1) In a deepwater embayment in the Tuamotu Island platform, normal oceanic structure is found, although the surrounding platform is a shallow water area. (2) Just east of the Easter Island Rise, the crustal layer appears to be only about 4 km thick. Conditions are also different in that the thickness of the upper crustal layer, about 2 km, equals that of the lower layer, which has a velocity of about 6·7 km/sec. (3) Under the Nasca Ridge, beneath 2900 m of water, the crust appears to be about 15 km thick. (4) At four locations on the Easter Island Rise, despite unusually long profiles, no velocity greater than 7·5 km/sec is observed. This is also an area where the heat flow is observed to be about five times the normal value. As other parts of the Easter Island Rise appear to be characterized by normal trustal conditions and normal heat flow, this correlation strongly suggests that the mantle velocity may be affected by abnormal thermal conditions.

The measurements in Peru–Chile Trench area are of particular interest. Off Antofagasta, Chile, the line of measurements, parallel to the axis of the trench and located about half way between the trench axis and the top of the continental slope, shows that the base of the crust is at about 23 km. Along the axis of the trench the base of the crust is at about 18 km and west of the axis of the trench the mantle is at a normal depth of about 11 km. This suggests a normal pattern of thickening as the continent is approached with no observable superimposed effect related to the trench. On the series of measurements conducted at similar locations off Callao, Peru, although the oceanic and trench axis measurements corroborate those to the south, it was not possible to obtain a measurement on the continental side of the trench. However, measurements were not carried beyond 160 km, and it is probable that this length of line was not sufficient to obtain first arrivals from the mantle for the thickness of crust present.

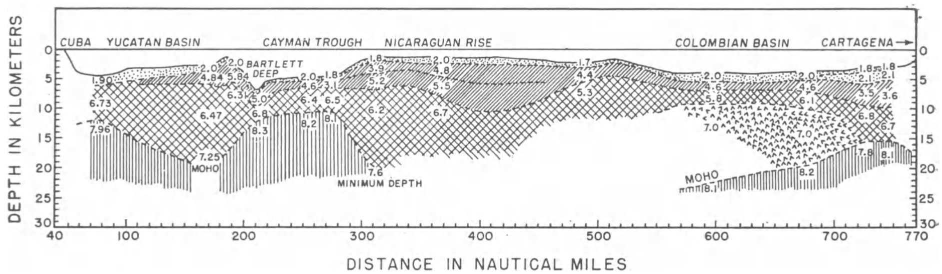

FIG. 22. Structure cross section of Western Caribbean from Western Cuba to the continental slope off Colombia.

5.9.4.2. *Caribbean Sea*

Although most of the measurements in the Caribbean Sea were made prior to the IGY, the results are of particular interest. Geographically and in terms of depth of water, it has not been clear whether such areas should be regarded as part of the continents or of the oceans, or whether they represent a transition phase between the two. The results obtained here are best summarized in Fig. 22, which shows a profile section across the area from Cuba to Venezuela. The outstanding points brought out by this profile are:

(1) The crustal thickness does not always conform to depth of water.
(2) The internal structure of the crust is complex both structurally and in terms of number of layers present.
(3) Both continental type structure and that associated with oceanic conditions are present.
(4) As was observed in the Peru–Chile Trench, the Bartlett Trough does not appear to involve the crust as a whole; however, the configuration of the upper crustal rocks and the overlying layers of unconsolidated and semi-consolidated sediments does conform to that of the trough.

The geological significance of the intermediate velocity values in the range 4 km/sec to 5·5 km/sec observed here is not too clear. Because of the geographic location and in view of velocities observed in surface limestones, these values could be assigned to carbonate sedimentary rocks. The values are also found associated with volcanic rocks and certainly anything over 5 km/sec can be associated with crystalline granitic rock at shallow depths.

5.9.4.3. *Continental border area of East South America*

Seventy odd measurements were made along the Argentine coast, mostly within the continental shelf area. Although none of the measurements give the total thickness of the crust, a considerable amount of geologic information was obtained. In general, the section on the continental shelf is characterized by a layer of bottom sediments having a velocity of 1·6 to 2·4 km/sec and a thickness which averages approximately 2 km. However, beneath the continental slope this layer thickens to over 5 km; beyond the slope area the thickness is only about 2 km. In places, particularly south of Bahia Grande and in deep water beyond the continental slope, the bottom sediments can be divided into an upper member with a velocity of 1·6 to 2·0 km/sec and a lower member with a velocity of 2·1 to 2·4 km/sec.

Beneath the bottom sediments a layer with a velocity of 2·5 to 3·8 km/sec is found in many places. It is not, however, a continuous feature and varies from zero to 2 km in thickness. In many areas this layer is replaced by one having a velocity in the range 4·2 to 4·9 km/sec and in a few areas both layers are present. As with the 2·5 to 3·8 km/sec layer, the higher speed layer is not continuous, and it varies in thickness from zero to 5 km. Both of these layers probably represent consolidated sediments.

Beneath the above sediments crystalline rock is encountered which has a velocity that varies from 5·0 to 6·1 km/sec. In some areas an underlying higher speed horizon with a velocity of 5·9 to 6·4 km/sec is also observed at depths of the order of 5 to 7 km. On a section perpendicular to the coast from Comodoro Rivadavia, the velocity of this intermediate velocity layer is found to increase to 7 km/sec beneath the continental slope where it reaches a maximum depth of 10 km, and then decreases in velocity again as it rises toward the surface beyond the slope. The oceanic section observed at the end of this traverse shows 2 km of bottom sediments represented by a surficial layer 0·3 km thick of 1·8 km/sec material, and 1·5 km of material of 2·3 km/sec, which lie directly on crystalline rock material with a velocity of 5·6 km/sec. Beneath this velocity horizon, which is only about 1·5 km thick, an intermediate velocity layer with a velocity of 6·6 km/sec is observed which extends down to a depth of about 13 km where the Mohorovičić discontinuity is encountered with a velocity of 8·0 km/sec.

5.9.4.4. *Scotia Sea — Falkland Islands area*

Two traverses were established in the Scotia Sea — Falkland Islands area. One extends westward to South America and the other directly south to about

57° South latitude. Within the South American Continental Shelf area, a 6 km section of sediments is found which is composed mostly of material having a velocity of 4·4 to 4·9 km/sec. This layer rapidly thins eastward and is replaced by sediments with a velocity of 2·9 km/sec in the Argentine Basin, but is again encountered with a thickness of 2 km as the Falkland Islands are approached.

South of the Falkland Islands the section crosses a 100 fathom bank about 100 km wide and 330 km long. The very marked change in geologic structure associated with this bank poses an interesting problem as to its origin. The structure is defined principally by the consolidated sediment layer with a velocity of 4·5 km/sec which thickens abruptly from about 1 km on the north, and zero thickness on the south, to 8 km beneath the bank. The lower velocity bottom sediments first thicken on the flanks from 0·5 km to about 3 km and then thin to about 0·5 km over the center of the bank. The underlying crustal rock surface, characterized by a velocity of 6·1 km/sec, although quite irregular in general, is depressed about 5 km beneath the bank to give an inverted picture of the surface topography. South of the bank a normal oceanic section is observed although the depth of water is about 3 km. The surficial sediments are less than 0·5 km in thickness and lie directly on crustal material with a velocity of 6·3 km/sec which persists to a depth of 11 km, where the Mohorovičić discontinuity is found with a velocity of 8·3 km sec. This pattern is similar to that observed in the Pacific Ocean rather than that found in the Atlantic Ocean.

5.9.4.5. *South Georgia area*

The measurements in this area result from a traverse extending from about 48° to 57° South latitude, which crosses the island arc and its associated fore-deep. Although the total thickness of the crust was not mapped, it is found that the sedimentary layers and the underlying crustal layer are conformable to the surface relief and that each of the sedimentary layers thickens as the depth of water increases. The pattern whereby the high velocity crustal rocks are at a higher elevation beneath the island arc than beneath the trench agrees with that observed in the West Indian Arc area. Velocity horizons observed in the Scotian trench are 2·0 km/sec, 4·0 km/sec and 5·6 km/sec.

South of South Georgia, the crust is found to have a layered structure with the lower member having a velocity of 7·6 km/sec. A similar high velocity basal crustal layer is observed in the Caribbean area, as shown in Fig. 22.

5.9.4.6. *South African area*

Twelve stations were run near the southern coast of Africa by the *Vema* of Lamont and the *Vrystaat* of the South African Navy. The locations of these stations are shown in Fig. 18. Analysis of the observations at these stations is incomplete at the time of this writing.

At the present stage of analysis, only tentative velocities have been obtained for these stations. A typical mantle velocity, however, appears to be present at station No. 161 (east of Durban). It is hoped that the final results from the stations near Durban can be related to the profile obtained inland from Durban by HALES of the Bernard Price Institute in Johannesburg.

5.9.4.7. *Gulf of Aden–Indian Ocean area*

In this area none of the measurements are believed to have been carried out far enough to give a measure of total crustal thickness although velocities on reverse profiles as high as 7·8 km/sec were recorded. At all sites two sedimentary layers having velocities of 1·85 km/sec and 3·9–4·6 km/sec are recognized. The aggregate thickness of the two varies from 2 to about 4 km. Beneath these sedimentary rock layers crustal material with a velocity of 6·4 to 6·8 km/sec is found. The very high velocity material 7·6 to 7·8 km/sec was observed at only two stations and the depth of this interface ranges from 9 to 11 km. In view of the shallow depth of water, 1310 to 2670 meters, these values cannot be reconciled with the thickness of crust to be expected or with the gravity anomalies in the area, which suggest that the base of the crust should be at about — 17 km. The occurrence of such high velocity material within the crustal layer is significant in view of the fact that in other areas such velocities have been identified with the mantle rock beneath the crust.

5.9.4.8. *Red Sea area*

In this area a local gravity and magnetic "high" is identified with a submarine trench extending over most of the length of the Red Sea. Although the total thickness of the crust was not measured, the seismic values do give an indication of the cause of the anomalies in the potential field. In the bordering shelf area the thickness of sediments varies between 2 and 6 km except in one area near St. John's Island where the total thickness is 10 km. The underlying crust is observed to have a velocity of 5·9 km/sec. Within the range of measurement, no layer of greater velocity was detected. In the central trench area, the sedimentary rock thickness is the same as in the shelf area (2–6 km) but the underlying crustal material is characterized by a high velocity of 7·1 km/sec. As the magnetic and gravity anomalies are missing in the graben area of the Gulf of Suez and Agaba, it appears that the 7·1 km/sec material found in the Red Sea trench is a basic rock intrusive that has invaded a crustal fracture system over this portion of the graben.

5.9.4.9. *Mediterranean area*

Seismic measurements in the Mediterranean Sea, Red Sea and Gulf of Aden were made by the *Vema* of Lamont and the *Atlantis* of Woods Hole. Fifteen seismic stations were run in the Mediterranean; the locations of these stations are shown in Fig. 19. At the time of this writing, velocities and thicknesses have been tentatively determined for these stations, but analysis is incomplete.

5.9.5. *Results of measurements in continental areas*

5.9.5.1. *United States — Great Lakes region*

In this area two profiles were made approximately at right angles to each other. The U.S. Coast Guard collaborated on this program by making a ship available so that explosives could be fired under water; this not only materially reduced the amount of explosive required but also gave much better energy coupling than could be obtained by firing on land. Charges of 750 lb of nitramon could be detected at distances greater than 275 km; no charges greater than 1500 lb were fired.

5.9.5.2. *Keeweenaw Peninsula profile*

This profile in northern Michigan extends from Keeweenaw Point into western Wisconsin and is aligned with the geologic strike and the pattern of the gravity anomalies. The structure found was approximately 1·5 km of material having a velocity of 4·7 km/sec overlying material of 6·3 km/sec down to —36 km where the mantle was found with a velocity of 8·05 km/sec. No intermediate layering was detected. The surface rocks are all crystalline: for the most part basalt with some gabbro and granitic schist and gneiss. The mean Bouguer gravity anomaly is +15 mgal, which might be interpreted as evidence of a thinner than normal crust. The source of this anomaly, however, appears to be related principally to the surficial rocks of basic composition. The crust may also make a contribution here as the velocity of 6·3 km/sec is somewhat higher than the 6·1 to 6·2 km/sec values usually associated with the crust.

5.9.5.3. *Apostle Island profile*

This reversed profile, oriented at right angles to the Keeweenaw profile, extends from the Apostle Islands in Lake Superior to south-central Wisconsin. Geologically this profile is oriented across a major structural dome composed almost entirely at the surface of granitic crystalline rocks, with pre-Cambrian sedimentary clastic rocks at the north (shot) end. The gravity anomaly pattern shows a pronounced minimum associated with the area and along the profile there is a negative gradient toward the south. The mean anomaly along the profile is —40 mgal. On the north end (Apostle Islands) the seismic section shows about 2 km of material having a velocity of 3·5 km/sec which is probably related to the pre-Cambrian sediments; 3 km of material having a velocity of 5·2 km/sec with the underlying crust down to the mantle at —37 km having a velocity of 6·15 km/sec. The mantle velocity was 8·15 km/sec. On the south end, in the vicinity of Wausau, Wisconsin, the section shows surface material (outcropping granite) with a velocity of 5·2 km/sec, which changes to a velocity of 6·15 km/sec at a depth of 1 km. No intermediate layering is evident and the mantle with a velocity of 8·15 km/sec is indicated at a depth of 36·4 km.

5.9.5.4. *Arkansas profile*

This profile is based on quarry blasts in the vicinity of Little Rock and extends up to the Cape Girardeau area of Missouri on the Mississippi River. The entire area is characterized by slightly positive Bouguer gravity anomalies and it has been noted for many years that earthquake wave arrivals in the area were earlier than would be predicted from normal continental travel-time tables. Although the shot point was located in an area of crystalline syenitic rocks, most of the profile extended over an area of increasing sedimentary section composed of Paleozoic, Cretaceous and later material. The seismic results were: 2 km of 4·65 km/sec material; 10 km of 5·5 km/sec material with the underlying crust having a velocity of 6·9 km/sec down to a depth of —43 km where the mantle was encountered with a velocity of 8·15 km/sec. On the basis of magnetic and well data, it is clear that the 5·15 km/sec material constitutes an upper layer in the crust rather than carbonate material such as is observed at the surface to the north of the profile. This is one of the deepest crustal sections reported so far in the United States and it is significant that it occurs in an area of low elevation rather than beneath a mountainous or plateau region where a thick crust would be expected. The implied high density of around 3·0 gm/cm^3 from the velocity of 6·9 km/sec explains the lack of an appreciable negative gravity anomaly. The high crustal velocity also explains the early earthquake arrival times in the area.

5.9.5.5. *Plateau of Mexico profile*

This profile is based on open pit mining blasts near Durango, Mexico. The profile extends southward from this shot point paralleling the Sierra Madre Range and the gravity anomaly pattern. Near-surface detail was obtained by firing a reverse spread over the first 20 km of the profile. As the mean anomaly value is about —200 mgal, a thick crustal section was anticipated. The surface geology is largely volcanic tuffs but there is reason to believe it may, in part, be underlain by Jurassic or Cretaceous limestones which outcrop farther south. The seismic results indicate a surface layer which has a velocity of 3·0 km/sec that is about 0·5 km thick which is underlain by a layer about 3·5 km thick having a velocity of 4·9 km/sec. This in turn is underlain by material having a velocity of 6·0 km/sec. Two interpretations appear to be possible for the thickness of this layer and that of underlying deeper layers. The preferred interpretation, on the basis of a theoretical analysis of the data as to the times for critical reflections from the base of the crust, is that the material with a velocity of 6·0 km/sec is 28·9 km thick. Beneath these is a basal layer having a velocity of 7·6 km/sec that is 11 km thick. The mantle rock velocity is 8·2 km/sec. The total thickness of the crust under this interpretation is 43·8 km. Under the alternate interpretation the 6·0 km/sec layer is 16·2 km thick and it is underlain by two basal layers. The velocity of the upper layer is 6·4 km/sec and its thickness is 15·3 km. The velocity of the lower layer is 7·6 km/sec and its thickness is 8·8 km. The total thickness of the crust under this interpretation is 44·2 km. As none of the arrivals from the deeper intermediate layering appears as a first arrival and the ambient noise level is high, the preliminary interpretation on this measurement given in IGY World Data Center A General Report No. 3 (10 July, 1958) did not show the deep intermediate layers.

A second measurement was attempted using quarry blasts at a limestone site about 40 miles north of Mexico City, but had to be abandoned because of an excessively high ambient noise level.

5.9.6. *South American measurements*

5.9.6.1. *Toquepala, Peru profiles*

These profiles were based on mining operation blasts at the Toquepala Copper Mine about 60 miles northwest of Tacna. Recording was carried out along 10 azimuths on profiles of varying lengths, as shown in Fig. 21. Although explosive shots up to 50 tons were used, it was not possible to obtain refracted first arrivals or reflections from the base of the crust in the altiplano proper. First arrivals from the base of the crust, however, were obtained in the flank area of the altiplano along the profiles paralleling the structural grain of the mountains. The apparent velocities are: 5·3 km/sec; 6·2 km/sec; a suggestion that there is a 6·7 km/sec layer; and 8·2 km/sec. On this basis the computed thickness of the crust is about 46 km. If second arrivals are used in the altiplano proper, a depth of the order of 65 km is suggested beneath the high plateau.

5.9.6.2. *Chuquicamata, Chile profiles*

This area, as the former one, lies on the high Andean altiplano, and the shot site, a mining operation, lies about 125 miles north-east of Antofagasta. As on the Toquepala profile, it was not possible to obtain first arrivals from the base of the crust in the altiplano although first arrivals were obtained to the south

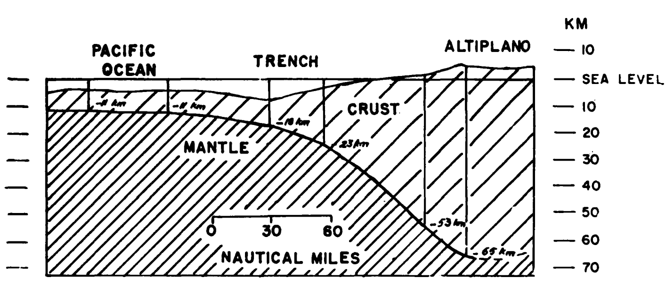

FIG. 23. Crustal section across continental border of Western South America.

over the flank area. The velocity structure observed was about 5·5 km/sec for the surface material, 6·0 km/sec to possibly 6·4 km/sec with a definite indication by first arrivals of a 7·0 km/sec basal layer overlying the mantle with a velocity of about 8·0 km/sec. On the basis of this structure, an overall thickness of about 56 km was computed for the crust. If second arrivals are used to compute the crustal thickness beneath the altiplano, a thickness of 70 km is found. Without first arrivals, however this value as well as that in the altiplano of Peru can not be said to be well substantiated. A more complete report on the above South American studies is given in Carnegie Institution of Washington Year Book 57, issued 19 Dec., 1958, p. 106–111.

As the measurements in the Peru–Chile Trench offshore from Antofagasta, Chile, lie in an area immediately adjacent to the Chuquicamata profile, it is possible to draw a crustal section across the continental boundary as shown in Fig. 23. This profile, incidentally, is in excellent agreement with that which would be deduced from Bouguer gravity anomalies.

5.10. Crustal studies in the U.S.S.R. during the IGY

by I. P. KOSMINSKAYA

During recent years in the U.S.S.R., in connection with the solution of a number of geological and geophysical problems,[1] extensive efforts have been made to elaborate seismic methods for detailed study of the Earth's crust. An underlying feature of these methods is the use, when recording seismic waves, of systems of observation similar to seismic prospecting systems. They ensure the use of correlational principles of separating and tracking the waves.

The idea of detailed methods of studying the Earth's crust belong to G. A. GAMBURTSEV, under whose supervision the following two methods were developed: the method of deep seismic sounding (DSS), which utilizes special small blasts[2] as sources of waves, and the method of correlational studies based on records of weak local earthquakes by means of a special network of highly sensitive stations.[3]

During the IGY these methods were developed further.[4] In deep seismic sounding studies on land, continuous profiles with observation of both refracted and reflected waves came into wide use; the methodology of marine observations was also developed[5]. In seismology, observations of earthquake waves along a profile were accomplished and this made it possible to study the structure of the Earth's crust in seismo-active zones.[6]

In this summary report, attention will be concentrated on studies utilizing deep seismic sounding; such observations were carried out on land, on sea and on inland seas, and also in the transition zone from the Asiatic continent to the Pacific Ocean.

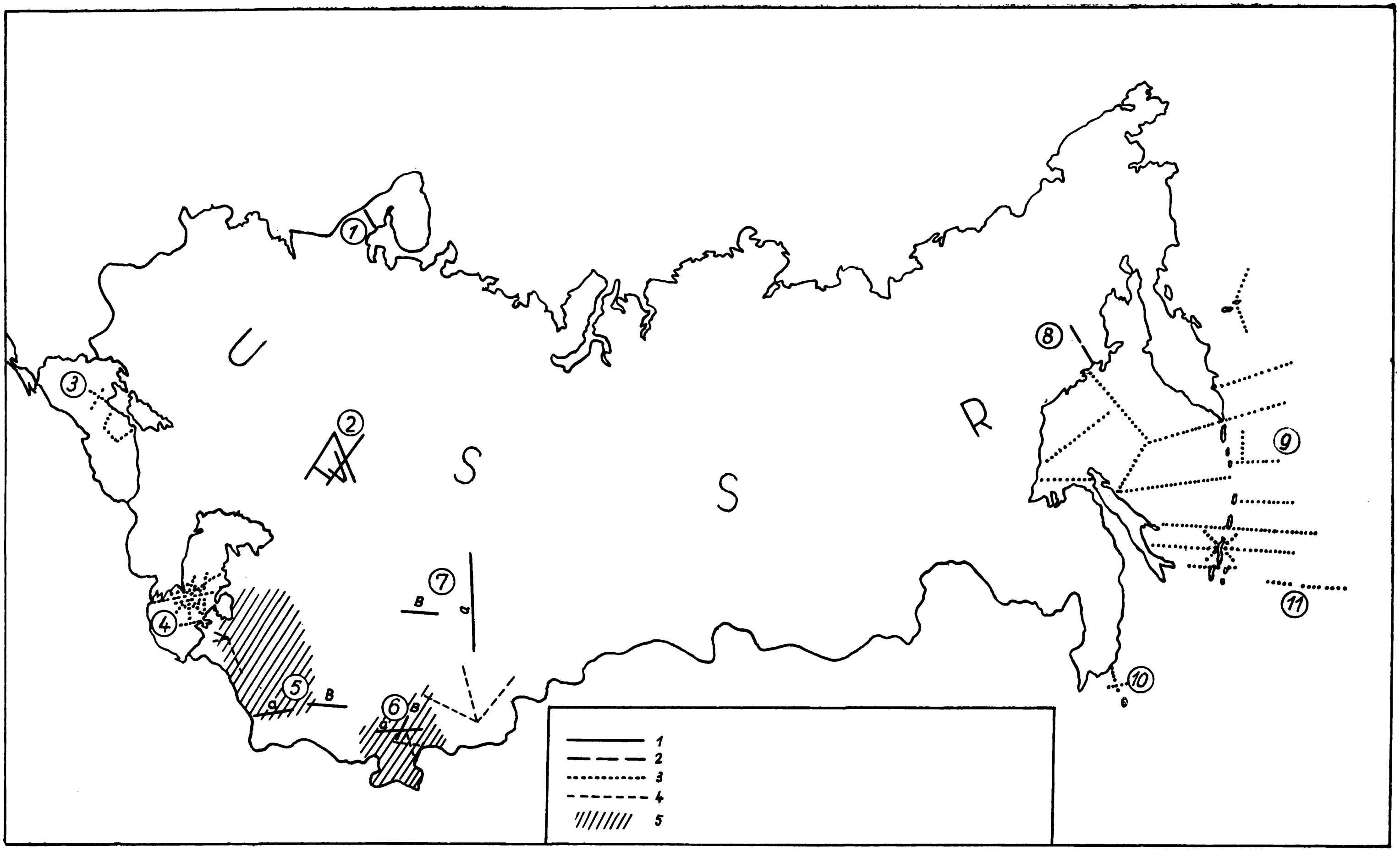

FIG. 24. Areas of detailed crustal studies on the territory of the U.S.S.R. Profiles of DSS (deep seismic sounding) carried out during the period 1956–59: 1 — continuous profiling, 2 — semi-continuous, 3 — point, 4 — DSS profiles carried out prior to 1956, 5 — regions of detailed seismological investigations.

Studies of crustal structure in the transition zone from continent to ocean were included by the U.S.S.R. in the IGY plan.[7–9] Studies in other areas were carried out in addition to and outside the program of the IGY, but the results are of general geophysical interest and, for that reason, are included in this summary.

Areas studied. Deep seismic sounding observations were carried out in areas with diverse geological structure. Figure 24 shows arrangement of DSS profiles carried out on land and at sea during 1956–59.

On land, the work was carried out on the territory of the ancient Baltic Shield,[10,11] in the southeast of the pre-Cambrian Russian Platform.[12–14] (Fig. 24, profiles, 1, 2), in a region of structures with a younger folded base: south-eastern Turkmenia[15] (profile 5a), Bukhara–Khivin region[16] (profile 5b), the folded zone of Kazakhstan[17,18] (profile 7), intermountain Fergana depression[19] (profiles 6).

At sea, DSS studies were initiated in complex tectonic zones, the geological structure of which had not been studied in detail; on the Black Sea[20–22] and the Caspian Sea,[23] where alpine and more ancient structures join (profile systems 3 and 4); in the transition zone from the Asiatic continent to the Pacific Ocean[24,25] embracing regions of the Sea of Okhotsk, the Kuril Island Arc and adjacent ocean areas (profile system 9), in the Sea of Japan (profiles 10).[26,27] The land profile Magadan–Kolyma[28] (Fig. 24, profile 8) is also in the profile system of the transition zone.

In land studies, DSS was part of a complex of regional seismic study and its principal aim was to study the surface of the basement and the deeper-lying boundaries in the Earth's crust[12].

In marine studies, DSS was usually carried out for the purpose of studying the common features of deep-seated crustal structure. The problem of studying the basement surface was posed only in places where this occurred at great depths. Deep seismic sounding studies, both during their execution and during analysis of the materials obtained, were broadly coordinated with other geophysical methods: gravimetry, magnetometry, seismological observations.

Seismological observations with the aim of studying the Earth's crust were carried out in Central Asia.[6,29]

5.10.1. *Land studies*

Crustal studies on land were carried out by means of deep seismic sounding and seismological observations.

(A) *Deep seismic sounding of the Earth's crust.*

Observational procedure. At the present time, two types of profile (continuous and semi-continuous) are used in the U.S.S.R., depending on the problems and the relief when operating on land. In marine studies, so-called point profiling is used.

Continuous profiling is the most detailed and ensures the greatest accuracy in crustal studies. It permits in principle, phase correlation of waves and makes it possible to obtain systems or travel-time curves for continuous correlational tracking of separate waves or groups of waves throughout the whole of the profile.

Continuous profiles are possible only in level country, but have proved particularly expedient in the case of regional seismic profiles for simultaneous studies of the upper and deeper parts of the Earth's crust.[12]

In mountainous regions or in areas with extremely broken ground, and also when studying only the general features of crustal structure, use is made of semi-continuous profiling in which continuous observations are made on separated sections of the profile. Here, the number of shot sites is chosen so as to ensure a system of reversed and overlapping travel-time curves for the principal wave groups. Continuous profiling was carried out in profiles 1, 2, 5–7 and semi-continuous, in profile 8 (Fig. 24).

The most detailed systems of observations have been obtained on the Russian Platform[12] and in the Bukhara.[14] The explosions were carried out in bore pits, shafts, and bodies of water. On the average, charges were between 500 and 1000 kg. In difficult seismogeological conditions they were increased to 3 tons, in which case they had a maximal range of 250 km, in rare cases up to 300–320 km.

The distance between blast sites was 40 to 60 km in most regions, with 100 to 200 m between seismographs.

Apparatus. For observations, use was made of multi-channel (26 and 60 channels) low-frequency seismic stations with a channel transmission band of 10–20 cps.

Observations were normally conducted by means of 2–5 stations, which for each blast ensured observations over a large section, up to 20 km, of the profile. Contact between the shot sites and the seismic stations was maintained by radio.

Principal deep-wave groups. Seismograms from deep seismic soundings reveal a large number, up to several tens, of regular cophase oscillations, which may be put into groups (Fig. 25) on the basis of the values of apparent velocities, arrival times and of peculiarities in the shape of the record.

To track separate waves along the profile, use is made, as in seismic prospecting, of phase and wave correlation. Group correlation is used to separate out and track groups of waves[30].

From the viewpoint of velocities, nearly all the observed deep waves belong to waves propagated in the Earth's crust, predominantly as longitudinal waves.

From the shape of the travel-time curves, the relation of overtaking travel-time curves, the sequence of arrival times and the regions of exit to first arrivals, the majority of observed wave groups refer to refracted waves and only some of the groups have properties characteristic of reflected waves (sub-critical and over-critical[31]).

Theoretical computations of the kinetic and dynamic characteristics of deep waves for the homogeneous-layer and inhomogeneous-layer models of the Earth's

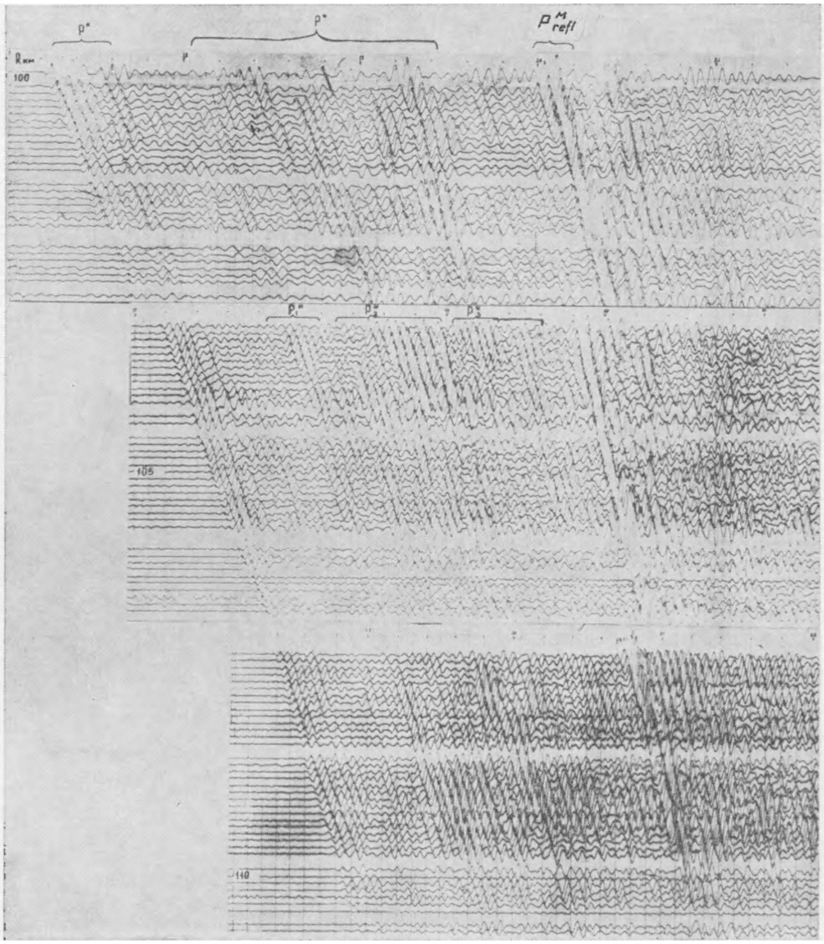

FIG. 25. DSS seismograms obtained in Kazakhstan[18] (Fig. 1, profile 7*b*). Distance between seismographs, 100 metres. The group of P^* waves may be subdivided by velocity into several subgroups. Phase and wave correlation is readily effected for P° and $P^M{}_{\mathrm{refl}}$ wave groups. Correlation of separate waves in the P^* group is disturbed.

crust have shown that refracted waves may in many cases be misinterpreted as weakly decaying, refracted waves.[32]

The wave pattern observed in various regions has much in common; however, significant differences are noted. The common feature lies in the similarity of the kinetics of the principal groups of waves, as is seen from the cases of the travel-time curves given in Fig. 26–31. In most regions, the groups of waves noted are those with velocities about 6 km/sec, 6·5–7 km/sec and 8 km/sec. The difference lies in the number of wave groups, regions of registration and relative intensity.

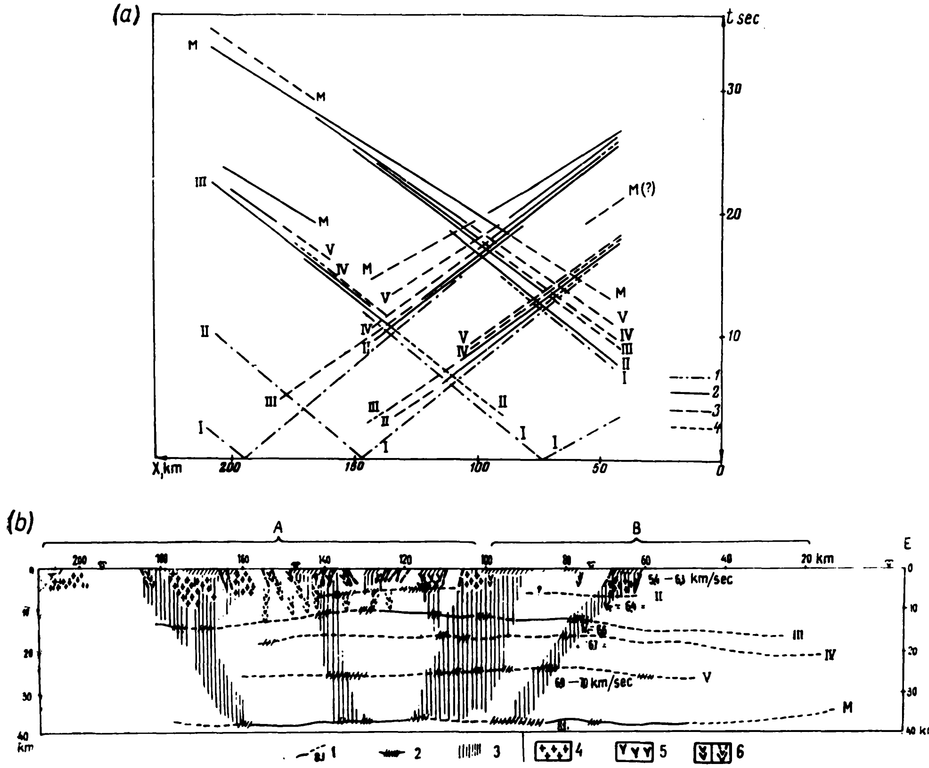

FIG. 26. System of travel-time curves (*a*) and section (*b*) of the Earth's crust from profile 1 (Baltic Shield), after I. V. LITVINENKO, and K. A. NEKRASOVA[10]. (*a*) 1 — refracted waves in the first layer, the velocity in which increases from 5·5 to 6·4 km/sec, 2 — refracted waves on the boundaries II—V, 3 — reflected waves from these same boundaries 4 — waves the nature of which is not clear, *M* — waves refracted on the Mohorovičić discontinuity, (*b*) 1 — averaged position of seismic boundaries constructed from refracted and reflected waves, 2 — sections of interface corresponding to correlation discontinuities, 3 — disruption zones based on DSS data, 4–6 — rock of Proterozoic age of different composition. (*A*) Karelian region, (*B*) White Sea region.

Wave groups which are dominant in one region may be less intense or not even recordable in another.

The following wave groups are distinguishable in most areas: P°, P^{*}, and P^{M}. In many regions, each of these groups exhibit two or three subgroups. Thus, seismograms obtained in the Russian Platform exhibit the following wave groups: P_0 and P°, P_1^{*}, P_2^{*} and P_{refr}^{M}, P_{refl}^{M} and P_2. The most stable are ordinarily the group of waves $P^{\circ}{}_1$ (velocity about 6 km/sec), which corresponds to the surface of a consolidated basement and the group of waves $P\ _{\mathrm{refr}}^{M}$ (velocity about 8 km/sec), which corresponds to the Mohorovičić discontinuity.

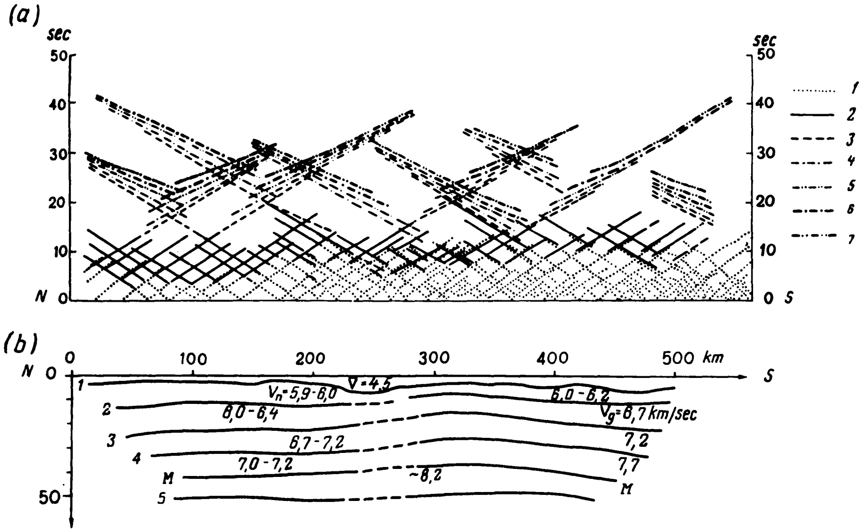

FIG. 27. System of travel-time curves (*a*) and section (*b*) of Earth's crust based on one of the profiles: 2 (Russian Platform), after YU. N. GODIN, R. A. BLOKHIN, N. P. IVANOVA, I. V. POMERANTSEVA, A. V. EGORKIN[12]: 1 — waves corresponding to boundaries in the sedimentary strata, 2 — refracted waves corresponding to boundary 1-surface of the pre-Cambrian basement. 3 — refracted waves corresponding to the intermediate boundary 2 in the earth's crust. 4,5 — refracted waves corresponding to intermediate boundaries 3,4; 6 — reflected waves reflected from *M* discontinuity; 7 — waves reflected from boundary 5 situated in subcrustal layer.

Of wave groups associated with intermediate boundaries in the earth's crust (P_2°, P_1^*, P_2^*), the following stand out: waves with velocity 6·5–6·9 km/sec, which have different designations according to the authors (P_2° — Russian Platform, P^* — Kazakhstan).

In all regions, except Kazakhstan, a wave group has been isolated with velocity about 7 km/sec (P_1^*). Waves P_2^* with velocity about 7·5 km/sec have been isolated only on the Russian Platform.

At considerable intervals, 50–70 km, the waves P_1° and P_2° are recorded in first arrivals. The waves P_1^* and P_2^* are recorded mainly in subsequent arrivals. The exit of the waves to the apparent first arrivals is usually due to damping of earlier waves. This is how the P_1° waves are supplanted by P_2° waves and the P_2° waves by P^* waves.

Waves P^M_{refr}, refracted at the Mohorovicić discontinuity are recorded in first arrivals from distances of 150–200 km. They are frequently very weak and are not reliably extracted from the noise background. In most areas we observe waves reflected from the Mohorovicić discontinuity, P^M_{refl}, that are recorded at distances greater than critical. These waves have a travel-time curve close to hyperbolic, nonparallel overlapping travel-time curves and, in the region close

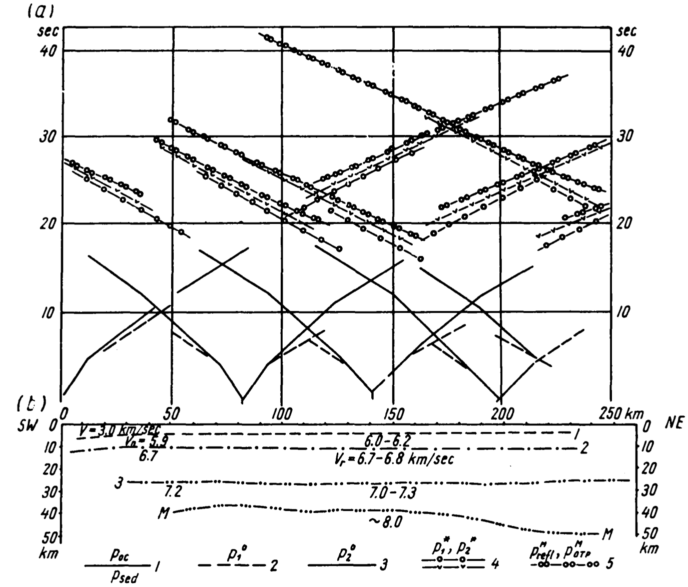

FIG. 28. System of travel-time curves (*a*) and section (*b*) of the Earth's crust from profile 5*a* (southeastern Turkmenia) after K. E. FOMENKO[15]. 1 — refracted waves P_{sed} corresponding to boundaries in the sedimentary strata, 2 — refracted waves $P^0{}_2$ corresponding to boundary 1 — surface of Paleozoic basement; 2 — refracted waves $P^0{}_2$ corresponding to boundary 3 — surface of lower Paleozoic basement, 4 — waves $P^0{}_1$ $P^0{}_1$ corresponding to boundaries in the lower part of the earth's crust. Boundary 3 is constructed from the first waves of this group; 5 — over-critical reflected and refracted waves from the *M* discontinuity.

to the initial point, dominant amplitudes. Their relative intensity diminishes with recession from the point of origin. In many regions, separation of refracted and reflected waves near the point of origin is not reliable.

At distances close to the region of exit of refracted waves P^{M}_{refr} to first arrivals, the reflected waves (at arrival times) are close to the waves corresponding to upper boundaries and their separation is difficult. For this reason, at large distances, over 250 km, from the shot point, the wave group designated usually as P^* is represented by complex oscillations. Methods for reliable separation of these oscillations into groups P^* and P^{M}_{refl} have not yet been worked out sufficiently.

It is ordinarily difficult to isolate reflection of a wave from deep-lying boundaries at large distances from the shot point. An analysis of DSS records by means of laboratory procedures of controlled and directed reception have shown[33] that the recording of reflections is hampered by numerous interfering waves propagated at high velocities and representing various kinds of multiple and exchange waves associated with the upper boundaries in the Earth's crust. This conclusion was corroborated by experiments in obtaining deep-seated reflec-

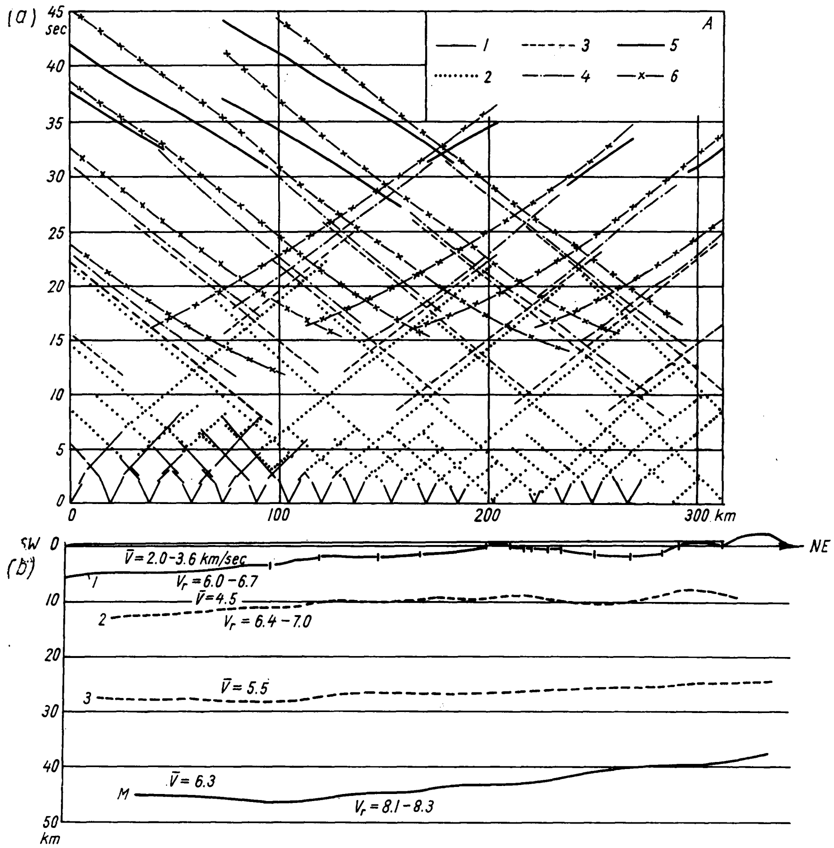

FIG. 29. System of travel-time curves (a) and section (b) of the Earth's crust from profile 5b (Bukhara area) after YU. N. GODIN, B. S. VOLVOVSKY, I. S. VOLVOVSKY[16]: 1 — waves corresponding to the boundaries in the sedimentary strata, V — mean velocities in km/sec, 2 — refracted waves on boundary 1 — surface of Paleozoic basement; 3 — refracted waves on intermediate boundary 2; 4 — waves reflected from intermediate boundary 3; 5 — waves refracted on M discontinuity; 6 — meta-critical waves reflected from M discontinuity.

tions,[34] in which it was possible to improve perceptibly the ability to correlate waves, by using linear grouping of the shots and seismographs to suppress the interference waves.

Velocities in the crust. Detailed systems of observations have made it possible to make sufficiently reliable determinations of boundary velocities by constructing time fields on the basis of encountering systems of travel-time curves. The mean velocities to the basic boundaries in the crust,[13,34] were determined from the travel-time curves of the first waves, the coordinates of the initial points, the points of intersection of the travel-time curves and the totality of

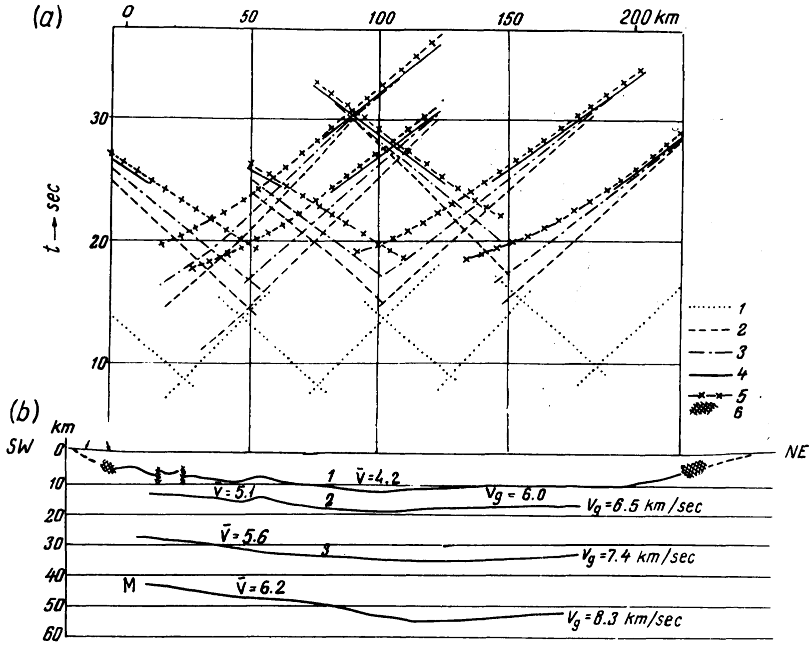

FIG. 30. System of travel-time curves (a) and section (b) of the Earth's crust from profile $6a$ (Fergana valley) after YU. N. GODIN, B. S. VOLVOVSKY, I. S. VOLVOVSKY[19]: 1 — refracted waves corresponding to boundary 1 — surface of Paleozoic basement; 2 — refracted waves corresponding to intermediate boundary 2; 3 — refracted waves corresponding to intermediate boundary 3; 4 — refracted waves on M discontinuity; 5 — waves reflected from M discontinuity; 6 — region of disturbances detected from seismic data.

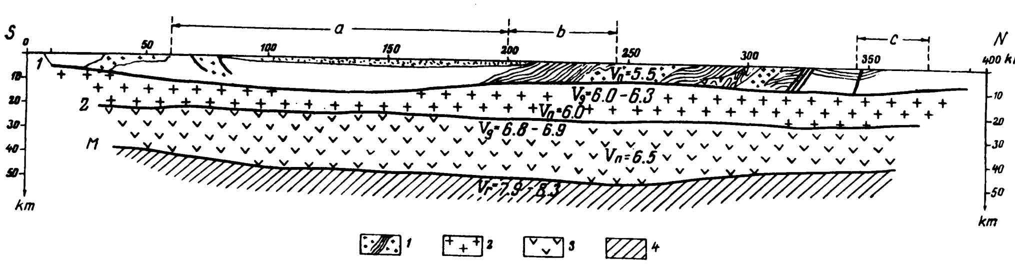

FIG. 31. Section of Earth's crust from profile $7a$ (Kazakhstan) after D .N. KAZANLI and A. A. POPOV[18]: 1 — sedimentary and metamorphic rocks of middle and upper Paleozoic; 2 — consolidated basement ("granitic" layer); 3 — intermediate ("basaltic") layer; 4 — subcrustal layer, (a) synclinorium, (b) zone of antilinorium, (c) region of carboniferous trough.

travel-time curves of reflected waves. On the basis of time-depth curves computed from mean velocities, graphs were constructed of stratum velocities as a function of depth.

The most accurate determinations were made of the boundary velocities along the deep-seated boundaries and the mean velocities to the Mohorovičić discontinuity (Table 1). From this table it will be seen that the least spread for boundary velocities is observed for the M discontinuity and the first deep-seated boundary (6·0–6·2 km/sec), despite the fact that in different areas they occur at different depths. The first of the intermediate interfaces, with velocities 6·4–6· 9 km/sec has stable limits of boundary velocities.

Let us now compare the boundary velocities with the stratum velocities under the interfaces (see Figs. 26, 27, 31). In some areas (Fig. 31) they are not definite, but merely taken or computed approximately. For the majority of represented sections the stratum velocities are usually less, by 0·2–0·5 km/sec, than the boundary velocities.

The reason for this discrepancy has not yet been established, but it may, obviously, be explained as follows: the observed refracted waves are associated with transitional layers[13] possessing several high velocities. These layers divide the regions, in which inhomogeneities exist, for which alternation of zones is possible, both with lower velocities and with velocities roughly the same as the boundary velocities in the transition layer. When there is a predominance of zones with lower velocities the mean stratum velocities in the layer under the boundary layer will be lower than the boundary velocities. On the whole, there may also be observed, in a thick layer, a certain very slight increase in velocity with depth.[31]

Crustal sections. Figures 26–31 show crustal sections constructed, for the most part, from systems of reversed travel-time curves by the method of time fields. The majority of the profiles (Figs. 26, 27, 28, 31) take into account refraction at intermediate boundaries. The medium and stratum velocities for deep-lying boundaries are taken as being constant.

The construction of the M discontinuity, and also the intermediate boundaries, from the travel-time curves of reflected waves is carried out taking account of variations in the mean radial velocity.

At all sections obtained on platforms in regions where the surface of the consolidated basement occurs regularly the deep-seated boundaries are approximately horizontal (Figs. 26, 27). In certain cases where the basement plunges, a subsidence of deep-lying boundaries (Fig. 28) is also observed. Occasionally, such agreement between the surface of the basement and the deep-lying boundaries is not observed (Fig. 31).

The total thickness of the Earth's crust within the Baltic Shield and the Russian Platform varies but slightly from 37 to 40 km. There are more noticeable variations in the depth of the M discontinuity in Turkmenia (40–50 km, Fig. 28), in the Bukhara region (40–45 km, Fig. 29), in the Fergana Valley (43–

TABLE 3
Boundary and Mean Velocities in Earth's Crust, depth of M Discontinuity and Thickness of Sedimentary Rock

No. Area	Baltic Shield	Russian Platform	South-eastern Turk-menia	Bukhara Region	Fergana Depress-ion	Kazakh-stan	Name of boundary
1. Boundary velocities for interface							
1	5·4–6·3	6·0–6·2	6·0–6·2	6·0–6·2	6·0	6·0–6·3	Surface of consolidated basement
2	6·6	6·6–6·7	6·7–6·8	6·4–6·8	6·5	6·8–6·9	
3	6·8(?)	7·1–7·2	7·0–7·3	—	7·4	—	Intermediate boundaries
4	6·9–7·0	7·6(?)	—	—	—	—	
M	8·1	8·1–8·2	8·3–8·6	8·1–8·3	8·3	7·9–8·3	Mohorovičić discontinuity
2. Mean velocity in Earth's crust	6·4	6·1–6·2	6·2	6·3	6·2	6·1	
3. Depth of M Discontinuity, km	37–40	38	40–50	40–45	43–53	33–48	
4. Thickness (in km) of sedimentary rock with mean velocity less than 4·5 km/sec	—	4	3–5	0–5	0·12	—	

63 km, Fig. 30), and in Kazakhstan (40–50 km, Fig. 31). The increase in crustal thickness in nearly all these areas (with the exception of Kazakhstan) is due to an increase in the thickness of sedimentary rock. The thickness of the crystalline crust varies only slightly and ranges between 35 and 40 km.

Intermediate crustal boundaries have been noted in all areas. The most reliably established boundary is that of 6·4–6·9 km/sec situated in most areas at a depth of about 10–15 km, while in Kazakhstan it goes down to 20 km. A boundary with velocity exceeding 7 km/sec has been noted in the same regions (except Kazakhstan) at depths of 25–30 km. Thus, in most areas, with the exception of the upper and lower boundary of the earth's crust, two more intermediate boundaries are observed.

How should we interpret the observed boundaries from the viewpoint of the classic concept of the existence of "granitic" and "basaltic" layers of the crust? In this case, apparently, such a division would be purely formal in character, although it is frequently necessary when generalizing data obtained in different regions.[35]

In velocity, the first boundary and the layer under it, ($V_{first} = 6$ km/sec), unambiguously belong to the "granitic" layer. As for the "basaltic" layer, judging by boundary velocities (following, for instance, B. GUTENBERG), one would say that the first boundary, with $v_g = 6{\cdot}4$–$6{\cdot}9$ km/sec, already belongs to it; however, the stratum velocity in the layer under this boundary is 6·0–6·4 km/sec from data on the Russian Platform,[13] which is nearly "granitic". On the Baltic Shield, the stratum velocity under this boundary is more in keeping with a "basaltic"[11] layer.

Judging by the boundary and stratum velocities, over 6·5–7 km/sec, the deeper boundaries may be unambiguously placed in the category of the "basaltic" layer. However, the relativity of this concept is now so clear that later, until more complete data are obtained, we should use the terms "granitic" and "basaltic", adding in all cases their seismic characteristics, namely, the velocities of longitudinal waves.

As regards the petrographic composition of the crust, one should, obviously, as proposed by certain geologists,[36] overstep the limits inherent in "granitic" and "basaltic" rocks and include in any consideration the entire complex of volcanic and metamorphic rocks, the density and velocities of which are close to those observed in the Earth's crust.

New views on the geologic nature of seismic boundaries may likewise rest on certain contradictions between deep-seated and surface structures that have been established in the studies described. The form of the relief of deep-lying boundaries may be most vividly correlated with the principal structures of the basement in the case of the Baltic Shield and in Kazakhstan. In these areas, the seismic boundaries are not in agreement with the borders of different structures (Figs. 27, 31), but, as it were, intersect them. In the Baltic Shield, one should, according to the authors of these studies,[10,11] consider only deep-seated fractures (which manifest themselves as a disrupted correlation in the DSS records) as the reflection of tectonic zones.

In other areas, the basement structures are not so well studied; however, on the basis of available data on the history of their formation and also of an analysis of geophysical fields we may suppose that both the young and old basements are complex joints of folded regions with deep-lying roots. But in nearly all regions, the seismic boundaries are almost horizontal.

The process of formation of seismic boundaries is what now remains obscure. Do these boundaries reflect definite historical stages in the development of the earth's crust or are they due to other physical causes? It seems to us, that the internal seismic boundaries of the earth's crust are of secondary origin. They are due to the combined action of temperature, pressure and permeation of fluids from the mantle. As to M discontinuity, it is probably due to changes both of the state and the composition of the matter.

(B) *Seismological observations*

Seismological profiling in the Garm–Dushanbe region[6]. A network of 12 seismic stations was set up in a string along the profile Gissar Valley — Alai Valley for over 200 km. To the west and east of this profile are the epicentral zones of the Southern Tian-Shan Mountains.

The small distances between stations, the comparatively large time sweep (4 mm/sec) and the good time service made possible a reliable isolation and determination of the absolute arrival times of the waves; in some cases it was possible to correlate them.

The distance between the extreme epicentral zone was about 700 km.

The situation of the epicentral zones made possible systems of encountering travel-time curves. For the compilation of overlapping travel-time curves, use was made of the records of more distant earthquakes.

For each epicentre, an individual travel-time curve for the longitudinal waves was constructed, taking account of the position of the stations with respect to the epicentre. Transverse waves were not used.

The seismograms usually exhibited two, less often three, arrivals attributable to direct longitudinal waves, waves refracted from the Mohorovičić discontinuity, and also waves reflected from this discontinuity. It was not possible on the seismograms to separate out waves associated with intermediate boundaries in the earth's crust. In some cases, there was a clear-cut record of a wave being propagated transversely along the M discontinuity.

The velocities of the direct waves were about 6·0–6·1 km/sec. The mean apparent velocity for all waves associated with the base of the crust is about 7·9 km/sec. For waves reflected from this boundary, the apparent velocity at a distance of 150–300 km diminishes from 6·8 to 6·3 km/sec.

The same procedure, with a sparser network of stations, was used to determine the depth of the M discontinuity along the profile Dushanbe–Frunze[6].

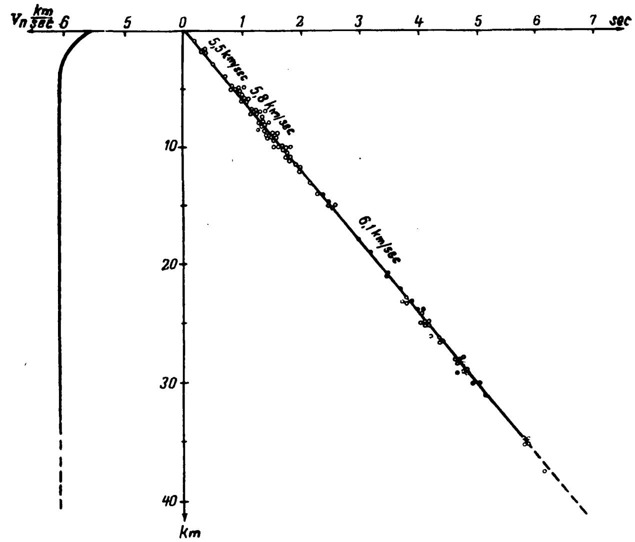

FIG. 32. Generalized time-depth curve and graph of stratum velocities for the Garmsky district after I. L. NERSESOV[6].

From the data of the Garm-station network, time-depth curves were constructed in addition to horizontal travel-time curves. To determine the vertical transit time, use was made of direct waves excited by foci situated at various depths, which in the region under investigation ranged from 5 to 30 km.

The averaged time-depth curve and the variation curve of stratum velocity of longitudinal waves with depth are shown in Fig. 32.

According to this graph, the velocity in the earth's crust is about 6 km/sec in the interval of depths 5–35 km. Data on velocity in the deeper parts of the crust have not been studied as yet. Here, the Mohorovičić discontinuity lies at a depth of about 50 km.

Crustal studies by means of recording exchange waves from distant earthquakes. Experience in recording earthquake waves in various areas has shown that the records exhibit prominently clearcut waves associated with exchange on the *M* discontinuity and the surface of the consolidated basement. The extensive systematic recording of these waves was carried out on the territory of the Turkmenian S.S.R.[29] The procedure was to record exchange waves from distant earthquakes (Pacific Ocean and Afghanistan) at points situated uniformly over the area under study. The distance between stations ranged from 5 to 20 km.

As in the case of seismological profiling, recording was done by means of highly sensitive seismological stations with a recording band from 1 to 5–8 cps. At each point, recording continued during 7 to 10 days. Several observing stations operated simultaneously.

The depths of the exchange boundaries were computed with allowance made

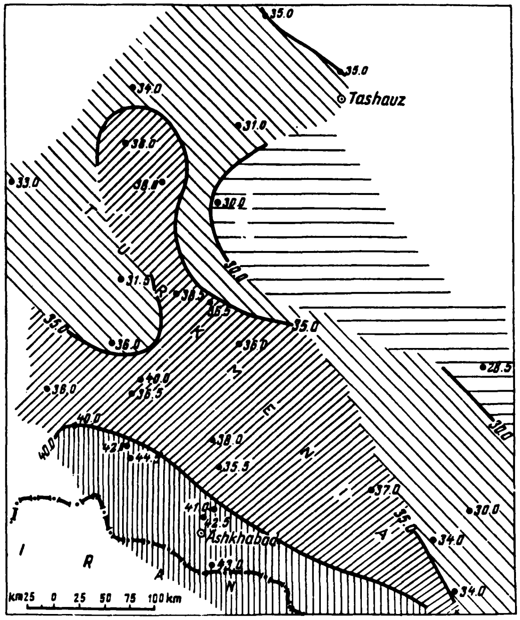

FIG. 33. Schematic map of depths of the M discontinuity after N. K. BULIN and YU. I. SYTIN[29]: dots — sites of observations; figures — depths of M discontinuity

for a velocity section of the crust determined in adjacent areas from DSS data and a seismic survey. Fig. 33 gives a schematic map of the depths of M discontinuity.

As yet, seismological investigations aimed at a detailed study of the crustal structure have been carried out on a small scale; however, their results point to great possibilities of these methods, which when coordinated with DSS, may enable the study of vast territories as seismo-active areas and as seismic zones.

5.10.2. *DSS studies on the Black and Caspian seas*

Observations on the Black and Caspian seas were conducted by means of point profiling with the use of fixed recording stations and a mobile shot point. [5] On the Caspian Sea use was also made of land recording stations. The average distance between shots was 5–10 km. The maximal recording range for charges of about 100 kilograms reached 200–250 km for the Caspian Sea and 120–170 km for the Black Sea.

The most detailed system of observations with simultaneous use of 3 or 4 recording stations was accomplished on the Caspian Sea. The distance between stations was 40 to 70 km, which permitted, on a number of profiles, to obtain sufficiently full systems of reversed and overlapping travel-time curves for certain deep waves.

Apparatus. Waves were recorded by means of low-frequency apparatus. Oscillations were recorded by receivers either deposited on the bottom (Caspian Sea) or suspended (Black Sea).

The amplifiers had a transmission band of 3 to 20 cps for deep waves and 200 to 600 cps for sonic waves. Recording was done simultaneously on several frequency characteristics.

Principal wave groups. Wave groups P°, P^* and P^M were observed which have kinematic and dynamic characteristics close to these same waves recorded on land (Table 24). Also isolated, in addition to deep waves, were P_{sed} waves that are propagated in sedimentary rock.

TABLE 4
Velocities for Principal Wave Groups

Wave groups	P_{sed}	P°	P^*	P^M
	Mean velocity, km/sec	Boundary velocity in km/sec		
Black Sea [20-22]	3	5·9–6·2	6·4–6·8	8·2–8·4
Caspian Sea [39]	2·5–3·5	5·9–6·4	6·5–6·8	8

The P_{sed} waves correspond to series of layers in sedimentary rock and characterize the mean velocity in them: the P° group is associated with the surface of the consolidated basement, which is corroborated by seismic-survey data on the eastern shore of the Caspian Sea.

The P^* group is referred to the surface of the so-called intermediate or "basaltic" layer, which, in regions where P° waves are encountered, obviously forms the surface of the crystalline crust. Judging by velocity values, the P waves can undoubtedly be placed in the category of the Mohorovičić discontinuity.

The difference in the relationship between the branches of the travel-time curves of the basic wave groups and the absolute arrival times of the waves is appreciable over different sections of the areas under investigation. This is particularly evident in the central and eastern part of the Caspian Sea and in the central and coastal parts of the Black Sea.

The absolute times of the travel-time curves of the first waves on the Caspian Sea differ on the platform and in the depression by a factor of nearly 1·5; this is due to variation in the thickness of sedimentary rock with low velocity.

Velocities in the earth's crust. In marine studies, as in land studies, the boundary velocities for deep-lying interfaces are most reliably determined. At sea, the mean velocities are usually determined with less reliability since there are large difficulties in introducing corrections for the influence of the low velocity upper layers. In interpreting the Caspian Sea materials, use was also made of

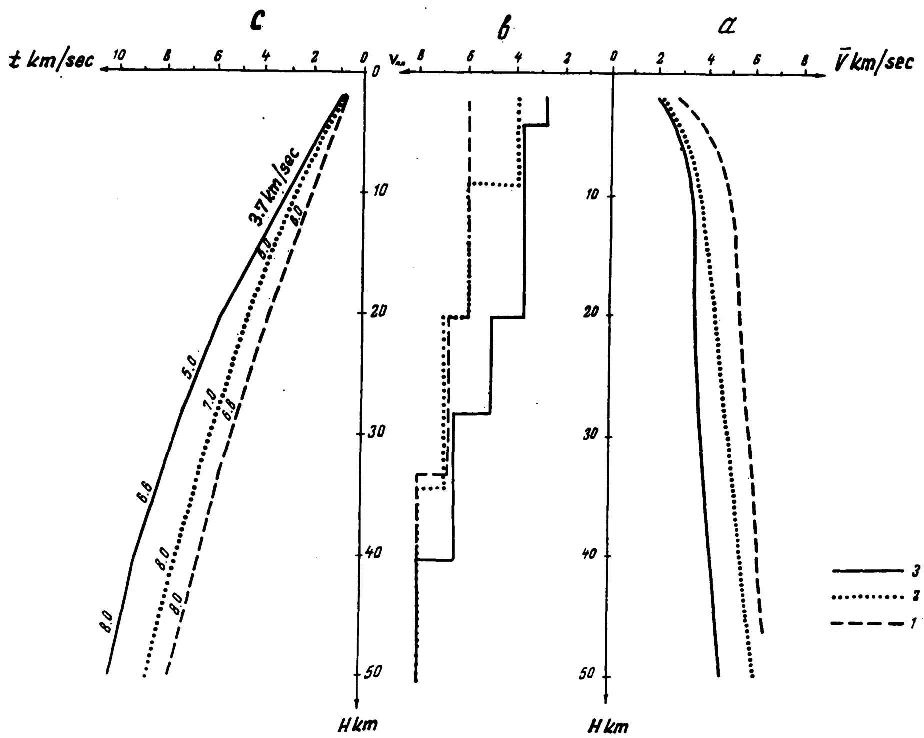

FIG. 34. Velocities in the Earth's crust in the central part of the Caspian Sea (after[38]): (*a*) averaged graphs of mean velocities V (H) for various zones: 1 — platform, 2 — region of depression, 3 — transition. (*b*) Graphs of stratum velocities computed from time-depth curves. (*c*) time-depth curves constructed from curves V (H).

determinations of mean velocities from travel-time curves of the first waves. For some sections of the Caspian Sea the reliability of mean-velocity determinations could be raised in certain cases through statistical averaging. The mean velocities in the Earth's crust vary respectively, just as much as the arrival times on the travel-time curves. Figure 34a, gives averaged curves of mean velocities as a function of depth computed for various sections of observations[38].

The first curve, with greatest velocities, belongs to the platform sections. The left one, to the depression region, the middle, to the transition zone, in which a considerable increase in sedimentary rock is observed. Characteristic of the central areas of the Caspian Sea are the anomalously low values of mean velocities in the crust, which do not exceed 5 km/sec. This is due to the special sedimentary — "basaltic" composition of the crust.

Figure 34 b, c shows graphs of the stratum velocities and time-depth curves constructed from mean-velocity curves. They indicate a difference in division of the earth's crust on different sections. Prominent in the platform sections are the sedimentary layer, the "granitic" layer, with velocity about 6 km/sec,

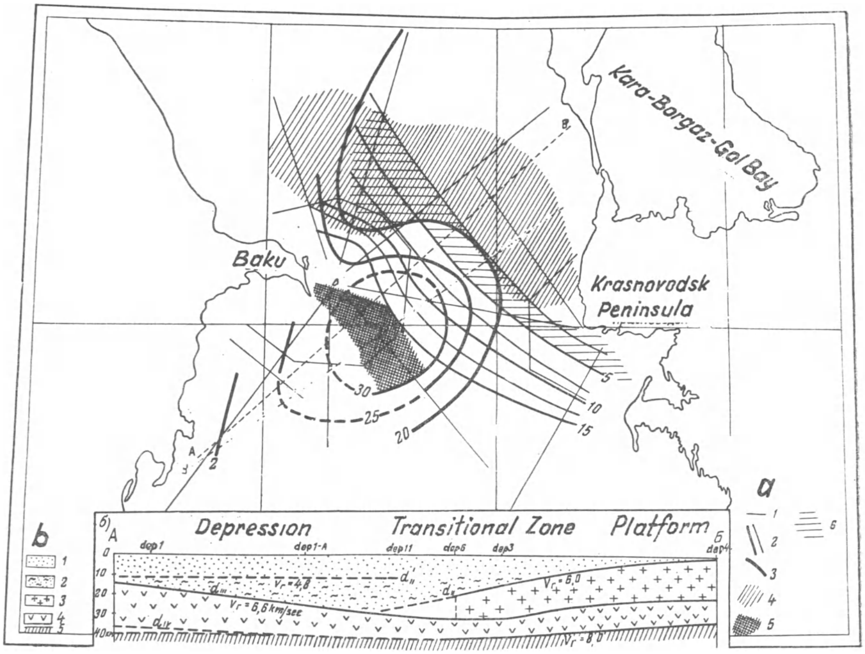

FIG. 35. Scheme of crustal structure of central part of Caspian Sea (after E. I. GALPERIN, I. P. KOSMINSKAYA, P. M. KRAKSHINA[38]). (*a*) 1 — DSS profiles, 2 — iso-depths of surface of crystalline basement; 3 — iso-depths of surface of "basaltic" layer; 4 — region where crustal thickness is less than 40 km; 5 — region where crustal thickness is greater than 45 km. 6 — zone dividing region of regular positions of depth boundaries from region of sharp subsidence. (*b*) Section of earth's crust along line AB: 1 — sedimentary rock with low (3 km/sec) velocity; 2 — compacted sedimentary rock; 3 — consolidated basement — "granite" layer; 4 — intermediate "basaltic" layer; 5 — subcrustal layer with velocity 8 km/sec.

and the "basaltic" layer, with velocity about 7·0 km/sec. In the depression area, only the "basaltic" layer is evident.

Section of the Earth's crust. Figures 35 and 36 give overall sections of the crust for the Caspian and Black seas, showing the principal peculiarities of the deep-seated structure of these widely dispersed regions. Figure 35a gives a scheme of the crustal structure for the Caspian Sea. According to the section and to this scheme, when passing from the epi-Hercynic Turkmenian platform to the Caspian depression the basement is observed to descend to depths exceeding 20 km and the so-called "granitic" layer seemingly disappears. The surface of the "basaltic" layer also experiences a considerable lowering from 20 to 35 km in the centre of the sea; it then rises towards the west to depths of about 15 km, forming a deep depression in the region of the deep-water trough. The M discontinuity likewise falls in this region, but it is much less marked. Its depth ranges from 37 to 45 km.

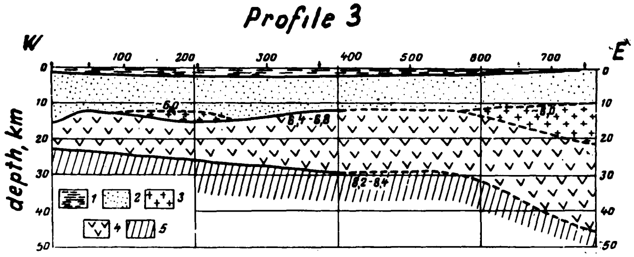

FIG. 36. Section of Earth's crust in eastern part of Black Sea (after Yu. P. NEPROCHNOV[22]): 1 — aqueous layer, 2 — sedimentary layer, 3 — "granitic" layer, 4 — "basaltic" layer, 5 — subcrustal layer. The eastern part of the profile is extrapolated with account taken of seismological data[39].

From seismic data on the behaviour of deep-lying boundaries of the crust and, similarly, from the data of other geophysical methods, we find, in the eastern part of the Caspian Sea, a clear-cut boundary area between the platform, which is characterized by small thickness (about 2–3 km) of sedimentary rock and conformable occurrence of all deep boundaries, and the depression, in transition to which all deep boundaries experience a sharp inclination.

On the Black Sea (Fig. 36), when passing from the continental slope to the deep-water part of the sea, we note an increase in the thickness of the sedimentary rock and, as it were, a tapering out of the "granite" layer. The thickness of sedimentary deposits with low mean velocity in the centre of the Black Sea reaches 12 km, and the depth of the M discontinuity, 22 km. As one approaches the Crimean peninsula and the Caucasus, the M discontinuity gets lower and seismic velocities appear that correspond to the "granitic" layer[39].

The fact that seismic studies in the Black and Caspian seas have elicited specific areas in which there is no typical "granitic" layer with a velocity of about 6 km/sec, or is very thin and not discernable by the given accuracy of observation, can, obviously, be of considerable interest in explaining the conditions of formation of these inner basins separated by the large Caucasian mountain system.[39]

5.10.3. *Studies in the transition zone from the Asian continent to the Pacific Ocean*

Crustal studies in the transition zone are of great interest for the solution of a number of problems in geology and geophysics and, in the first instance, for clarifying the question of the origin of the continents and the oceans. When selecting the area of study it was very important to cover regions with definitely different crustal structure. This requirement was completely satisfied by the far-eastern coast of the USSR, where an extraordinary diversity of geophysical and geological conditions occurs: enhanced seismic activity, modern volcanism, the prominent Kuril and Aleutian arcs, skirted by one of the deepest ocean trenches in the world, and, finally, the vast border seas of Okhotsk and Bering with their deep-water, nearly oceanic, depressions that separate the islands arcs from the continent.

Deep seismic sounding observations were included in the complex of geologico-geophysical investigations that involved gravimetric, aeromagnetic, and also seismological observations. The latter were carried out by a network of special high-sensitivity seismic stations situated on the southern islands of the Kuril arc (Iturup, Kunashir, Shikotan)[25, 48].

Observational procedure and apparatus[40]. Deep seismic sounding observations were conducted mainly at sea. One profile about 400 km long was traversed on land over the automobile road Magadan–Kolyma River. Observations on this profile were accomplished by means of semi-continuous profiling with the use of five shot sites which yielded a system of reversed and overlapping travel-time curves.

Recording was handled by standard multi-channel seismic stations with a transmission band of 10–18 cps. The apparatus and procedure of marine observations was similar to that described above for the Black and Caspian seas. The distinguishing feature of these studies was an expansion of the range of recorded frequencies towards low-frequency components (up to 1 cps).

For studying reflections from boundaries in sedimentary rock, use was made of a recording installation similar to that used in water-borne seismic prospecting[41-43].

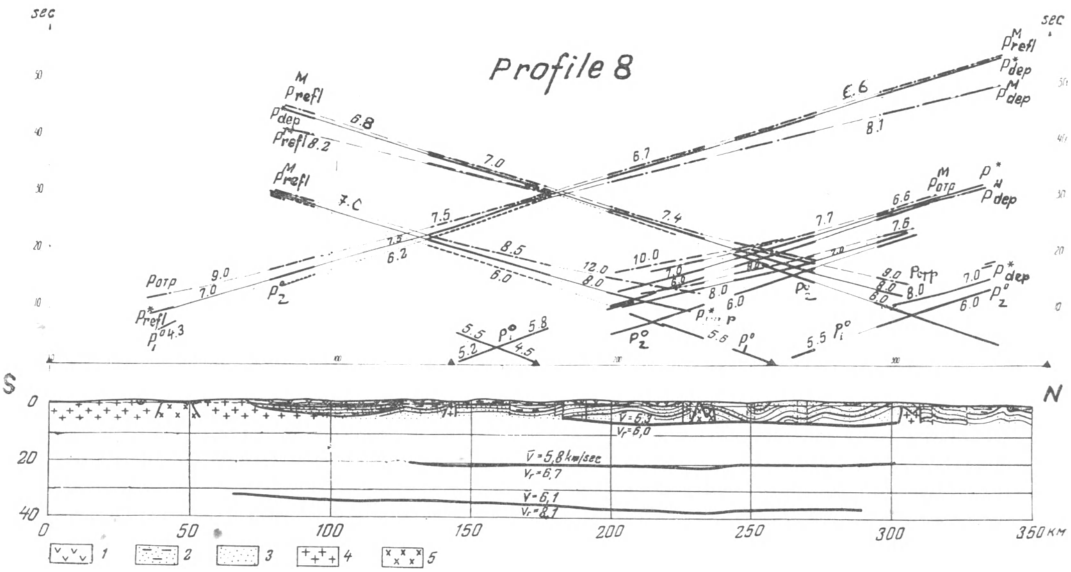

FIG. 37. Hodograph (*a*) and section (*b*) from profile 8 (Magadan-Kolyma) after B. Ya. SHVARTS, N.I. DAVYDOVA, and G. A. YAROSHEVSKAYA. The upper part of the section gives, schematically, occurrence of rocks of different age.

Peculiarities of the records obtained in various regions of the transition zone. The seismic material obtained on DSS profiles in the transition zone was extremely diverse. There are intricate records that are difficult to decipher and are characteristic of near-island and coastal regions, and comparatively simple ones typical of regions of the oceanic plateau and of certain parts of the Sea of Okhotsk.

Prominent on the seismograms are waves of different types associated with boundaries in the earth's crust and with bottom sediments, and also waves propagated in the water layer — direct and penetrating (of different multiplicity), reflected from the water surface of the sea bottom and from the interfaces within the sediments.

Waves associated with boundaries in the crust are ordinarily recorded first, or within the first five seconds of the record. All other waves are, as a rule, evident in the later arrivals.

As in other regions, kinematic data reveal the principal groups of deep waves P°, P^{*}, and P^{M}_{refr}, P^{M}_{refl}. All these groups have been reliably traced only on the land profile Magadan–Kolyma (Fig. 37).

In certain regions, only P^{M}_{refr} waves were recorded in all places. P° waves are found in near-island zones and in the northern and central parts of the Sea of Okhotsk.

P^{*} waves are reliably traced in the region of the oceanic plateau and in the deep-water parts of the Sea of Okhotsk and the Bering Sea.

Data on the boundary velocities for the principal wave groups are given in Table 5. Here also are stated the regions in which P^{M}_{refl} waves are recorded.

TABLE 5

Boundary Velocities for Principal Wave Groups, km/sec

Regions	P°	P^{*}	P^{M}_{refr}
Magadan–Kolyma profile	6·0–6·2	6·7	8·1
Deep-water part of Sea of Okhotsk, and Bering Sea		6·6–7·0	7·8-8·2
Northern and central part of Sea of Okhotsk	5·8–6·4	6·7	8·0

The P° waves are associated with the surface of the consolidated basement covered with sedimentary rock with sufficiently low (about 3 km/sec) velocity. In velocity, the P^{*} waves belong to the so-called oceanic or "basaltic" layer. The waves P^{M}_{refr} and P^{M}_{refl} are associated with the Mohorovičić discontinuity. P^{M}_{refr} are waves refracted on the M discontinuity. The P^{M}_{refl} waves are found

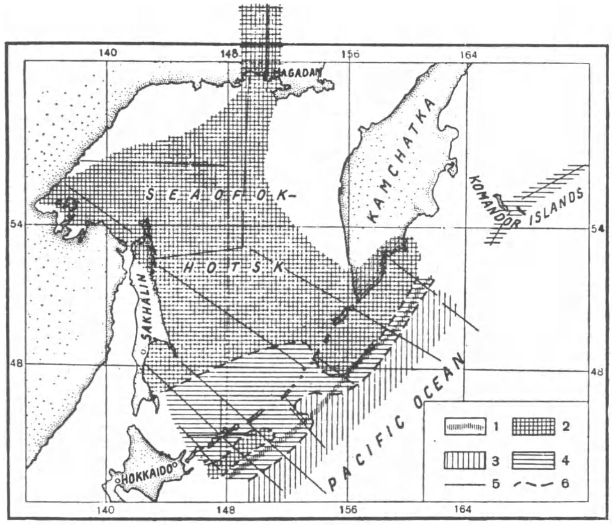

FIG. 38. Diagram of arrangement of zones with different types of crustal structure[4]; 1 — axis of deep-water trench; 2 — continental type; 3 — oceanic type; 4 — intermediate type; 5 — DSS profiles; 6 — arbitrary boundaries between zones.

only in subsequent arrivals; their apparent velocities diminish with distance within the limits from 8 km/sec to 6·5–6·0 km/sec.

From the form of the travel-time curves of P^M_{refl} waves, and also from their relationships with the travel-time curves of the P^M_{refr} waves, P^M_{refl} waves belong to over-critical reflections from the M discontinuity[44].

The P^M_{refl} waves have limited regions of existence and are found only on the profiles of the northern and central parts of the Sea of Okhotsk. They are distinguishable in the records only from distances of 40 to 70 km from the shot point. The P^M_{refl} waves ordinarily dominate the record and are registered up to large (150–250 km) distances from the shot point. In the area where both P^M_{refr} and P^M_{refl} waves exist jointly, the P^M_{refl} waves are far more intense than the P^M_{refl} waves.

Types of travel-time curves. The character of the seismic records, the number of branches of the travel-time curves and their temporal characteristics differ on different sections of the transition zone. This permits isolating types of travel-time curves that characterize specific geographical zones, which should correspond to different types of crustal structure.

The longest times, close to the times on the Jeffreys–Bullen travel-time curve, are characteristic of the continental type of travel-time curves. The shortest are typical of the oceanic type. The continental type travel-time curve has branches $P°$, $^*P^M_{refr}$, and P^M_{refl}. The ocean-type travel-time curve has only the branches P^* and P^M_{refr}.

Arrangement of zones with different types of crustal structure. In accordance with the types of travel-time curves, there are three prominent types of crust: continental, oceanic, and intermediate[41,46,47] (Fig. 38).

The distinguishing feature of the continental type is the presence of the "granite" layer with a velocity of about 6 km/sec and a sufficiently thick crust (20–40 km). Within the area studied, there are sections where a continental-type crust exhibits three clear-cut layers: sedimentary, "granite", and "basaltic". This subdivision is well established on the continent, under the northern islands of the Kuril arc in the region of Sakhalin Island.

The northern and central parts of the Sea of Okhotsk clearly exhibit a sedimentary and "granite" layer, whereas the "basaltic" layer is not detected at all by seismic data in the central part of the Sea of Okhotsk, or its presence reliably established, and there are not sufficient data to determine its position in the section.

The ocean-type crust consists of a thin sedimentary layer and "basaltic" layers with velocity about 6·6 km/sec. This type of crust has been established in the region of oceanic deeps of the order of 5 to 6 km. Unlike the crust of the oceanic plateau of the central parts of the Pacific Ocean, the thickness of the "basaltic" layer on the ocean adjacent to the Kuril depression is greater, reaching 6 to 10 km.

The intermediate type crust is relatively thin (of the order of 10–18 km) and consists either of sedimentary and "basaltic" layers of roughly the same thickness (such a crust is characteristic of deep-water depressions of the Sea of Okhotsk and the Bering Sea) or of crust like that of the southern islands of the Kuril arc, which consists of volcanic rock with velocity from 5 to 6·3 km/sec. It was not possible to make a clear distinction between layers with different velocities, which may be due primarily to inadequacy of the observational system.

Crustal sections in the transition zone. Figure 39 gives a summary of the crustal sections in the direction from the Asiatic continent (Kolyma River) to the Pacific Ocean, through the northern and southern parts of the Sea of Okhotsk.

A characteristic peculiarity of these sections is the increase (by a factor nearly of 4) of the crustal thickness when moving from the ocean to the continent: the velocity composition of the crust varies too. In the ocean, we observe a layer with velocity 6·6–6·8 km/sec and a thin sedimentary layer (about 0·2–1·0 km); in the intermediate zone the thickness of the sedimentary layer increases, approximately up to 3–6 km.

When moving to the central and northern parts of the Sea of Okhotsk, the velocity composition of the crust varies. Here, the predominant velocities are 6 km/sec. Velocities about 6·5–6·7 km/sec are evident in the lower parts of the crust and then largely conditionally. Yet, on the continent, this layer is reliably established.

In the foregoing summary there are no data on the crustal composition of the

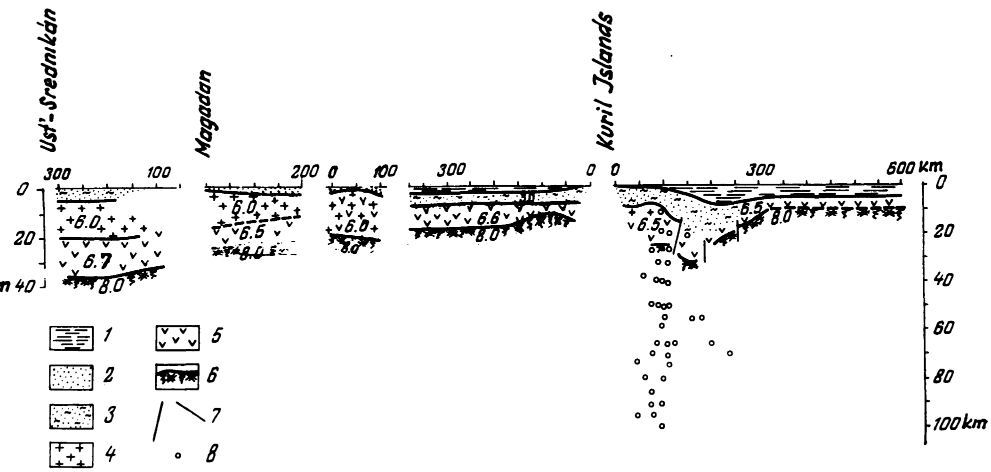

FIG. 39. Summary of seismic sections of Earth's crust in transition zone from Asiatic continent to Pacific Ocean in the direction of the Kolyma River (Ust-Srednikan) — Pacific Ocean (southeast of Iturup Island ([45])): 1 — water with velocity 1·5 km/sec; 2 — sedimentary layer with velocity about 2 km/sec. 3 — sedimentary rocks with mean velocity 3·5 km/sec.; 4 — "granite" layer with velocity about 6 km/sec.; 5 — "basaltic" layer with velocity 6·5–6·8 km/sec; 6 — subcrustal layer with velocity about 8 km/sec; 7 — regions of sharp variation of angle of incline of seismic boundaries; 8 — seats of earthquakes based on the data of special network of seismic stations on the Southern Kuril Islands after S. A. FEDOTOV[48].

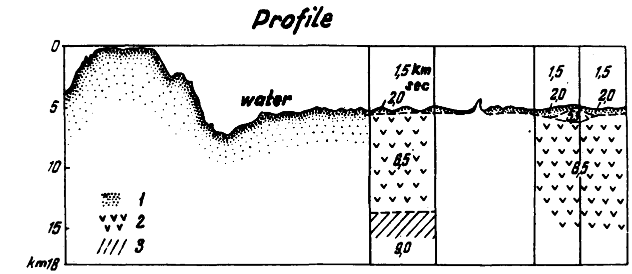

FIG. 40. Section of Earth's crust from profile 11 to the southeast of Hokkaido Island after G. B. UDINTSEV[49]: 1 — sedimentary deposits; 2 — "basaltic" layer; 3— subcrustal layer.

Kuril Islands, since the material obtained in this region is very complex and is not amenable to deciphering. The approximate depth of the M discontinuity under Iturup Island is 14–18 km. The crust of the islands is made up of dense rock with velocity in excess of 6 km/sec. To the southeast of Iturup Island there is a narrow band with continental-type crust. To this zone are confined the foci of subcrustal earthquakes determined from the data of a special network of seismic stations situated on the southern islands of the Kuril arc.[48] Here, also,

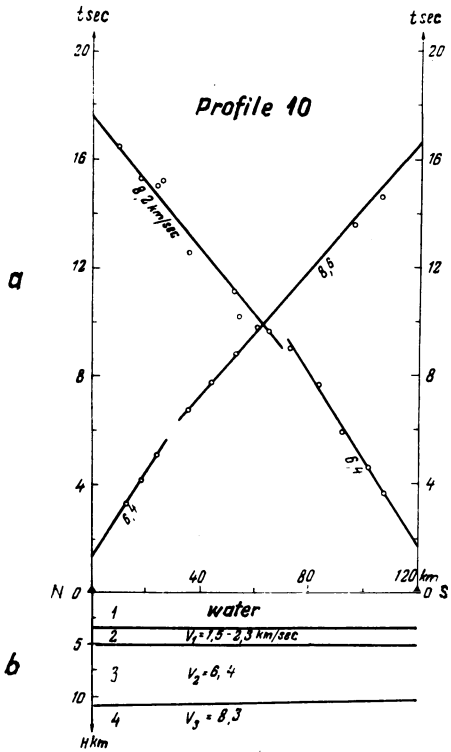

FIG. 41. Travel-time curve (*a*) and section of earth's crust from profile 10 in the deep-water part of the Sea of Japan to the southeast of Vladivostok after I. V. ANDREYEVA and G. B. UDINTSEV[27]: 1 — water, 2 — sedimentary layer, 3 — "basaltic" layer; 4 — subcrustal layer.

vertical displacements of the *M* discontinuity in the form of faults have been detected on the basis of the complicated nature of the seismic records.

In addition to the foregoing data on the transition zone, we should also note the crustal sections obtained in the ocean at roughly the latitude of Iturup Island[49] and in the Sea of Japan[27], not far from Vladivostok (Figs. 41, 42).

The first section (Fig. 40) is close to the crustal section obtained south of the Iturup Island. This section exhibits high velocities (8·6 km/sec) for the *M* discontinuity, thus indicating peculiarities in the composition of the subcrustal layer in this area.

The second section (Fig. 41) is similar to the section in the deep-water part of the Sea of Okhotsk. Here, the crustal thickness is 11 km and velocities in the lower layer underlying sediments are 6·1–6·4 km/sec. There are grounds for believing that these velocities are somewhat underestimated*.

* Our interpretation of the data utilizing only the first arrivals of the waves yields velocities of 6·5–6·7 km/sec

A similar crustal section with a sedimentary layer with low velocities and a "basaltic" layer close to oceanic thickness is also observed in the Bering Sea. We thus note similarity in type as concerns crustal structure of the deep-water depressions of these three seas adjoining the Asiatic continent, which, obviously, is indicative of similarity in type and conditions of formation.

Presently available, incomplete data of processed deep seismic sounding obtained in the transition zone show that the characteristic of this zone is the presence of areas with different crustal sections that differ both as regards the position of the M discontinuity and also the velocity characteristics of the crust itself. A correlation of this series of sections with the history of the geological development of the entire transition zone may be of great interest in studies of the nature of evolution of the earth's crust in the process of formation of the continents and the oceans[50,51].

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SEISMIC INVESTIGATIONS OF ANTARCTIC STRUCTURE

by K. E. BULLEN and M. J. GOODSPEED

6.1. Introduction

INTENSIFIED geophysical research on the Antarctic continent was a major feature of the 1957–58 IGY programme. This chapter is concerned with seismic measurements leading to the determination of thicknesses of the ice-cap in many places, of seismic velocities in the ice, and of some features of the structure below the ice.

The greater part will be concerned with the ice measurements, the results of which were among the more dramatic IGY discoveries and which led to an upward revision of 40 per cent in the estimated quantity of ice in Antarctica. The setting up of many new, well-equipped Antarctic bases during the IGY enabled field parties to operate over traverses exceeding 2000 km in some instances, and a principal task in many of these traverses was the recording and subsequent analysis of seismic waves from controlled explosions.

This work was supplemented by gravity and magnetic measurements. In the present chapter, relevant gravity results are included on certain of the diagrams. For further detail on the ice-thickness work, the reader is referred to the volume of the *IGY Annals* dealing with Glaciology.

In various cases, the procedures designed to obtain ice-measurements were extended to yield results on velocities and structures below the ice. In addition, observations of seismic surface waves from natural earthquakes led to broad inferences on crustal structure in Antarctica.

The ice-thickness and related measurements were carried out by parties from six countries which (in English alphabetical order) are: Australia, France, Japan, Union of Soviet Socialist Republics, United Kingdom of Great Britain and Northern Ireland, and the United States of America. (The British Commonwealth Trans-Antarctic Expedition results will be listed under the United Kingdom as the country responsible for this phase of the Expedition's work.)

Surface wave studies from natural earthquakes were carried out by seismologists in New Zealand and the United States.

Subsection 2 contains a brief review of ice-thickness measurements prior to the IGY. Subsection 3 gives details of the work and results of parties from the six countries which made ice-thickness measurements. Subsection 4 summarizes the seismic surface wave results.

The various results presented have been derived from data in papers referred to in the list of references on p. 247 and from data kindly sent to us before

publication. We wish to thank individuals and members of national committees who have kindly supplied us with needed materials.

The quantitative results on ice-thicknesses are presented at the end of the chapter in the form of both tables and diagrams. The first of two maps of the whole of Antarctica shows all the relevant traverses, and the second gives an overall picture of the ice-thickness results. Other diagrams show detailed cross-sections for individual traverses. Available gravity results are included on the latter.

The IGY station classification number (*IGY Annals*, VIII, 1959) and the geographical coordinates are given for each Antarctic station when the station is first referred to in the text.

In Figs 1 and 2, letters are used to denote key points on the traverses. The same letters are shown at the corresponding places on the appropriate detailed profiles, in order to facilitate cross reference between the overall maps and the individual profiles.

The drawings were prepared in the Geophysical Branch of the Bureau of Mineral Resources, Geology and Geophysics, of the Commonwealth Department of National Development, Australia. It is desired to thank J. M. RAYNER, Director of the Bureau of Mineral Resources and Member of the Australian National Committee for the IGY, and R. F. THYER, Chief Geophysicist, for their interest in this work and for permitting one of us (M. J. G.) to work on the collation of the material.

The work for this Chapter was completed in February 1961 and incorporated material which the authors had received by that date. Reference is made in an Appendix to new and revised material later received.

6.2. Summary of pre-IGY seismic investigations on ice

Seismic explosion methods were first applied on ice in the Austrian Alps[4]. Greenland ice-cap thicknesses were measured by the German Expedition to Greenland, 1930–31[5, 42]. During the second Byrd Antarctic Expedition[9], 1933–35, measurements were made by POULTER[32, 33] in the vicinity of the Bay of Whales, and by MORGAN in Marie Byrd Land.

Extensive measurements in Greenland were made by Expéditions Polaires Françaises[24, 25] in 1949–52, using the "weasel" tracked vehicle and improved electronic recording equipment. During this work, A. JOSET and J. JARLS lost their lives on 4th August, 1951, when their vehicle fell into a concealed crevasse.

The British North Greenland Expedition,[8] 1952–54 and a team from the Lamont Geological Observatory[3] made further seismic measurements in Greenland in the years just prior to the IGY.

The Norwegian–British–Swedish Antarctic Expedition of 1949–52 carried out noted ice-thickness measurements along a traverse of 615 km from the coastal base of Maudheim (71°03′ S, 10°56′ W). The experience gained in this project[37] was of immense value to the subsequent IGY projects.

6.3. IGY ice-thickness determinations and related work

6.3.1. *Australia*

A. Traverses

The base station was Mawson (A 980, 67°35′ S, 62°55′ E), some 400 km west of the western edge of the Amery Ice Shelf. The traverses fall under three headings:

(i) Regional traverse (1957–58): essentially a traverse (*VWX*, Fig. 1) along a single meridional line running south from Mawson for 642 km, with return along the outward route.

(ii) Regional traverse (1958–59): a loop 430 km long, based on (i).

(iii) Semi-detailed traverses near Mawson (1958): concerned primarily with a study of ice-flow in the vicinity of the Framnes Mountains, near Mawson.

B. Procedures

Seismic *P*-wave reflection and refraction methods were used.

The seismic recording equipment was fitted with 12 channels. The amplifiers and control system were Texas Instruments model 7000-B. The camera recorded 25 channels, including the separate outputs of the 12 amplifiers and mixed combinations of these outputs. The electronic equipment included a radio transceiver. A power-driven ice-drill was carried, capable of drilling a 9 cm diameter hole to about 60 m.

During the 1957–58 traverse, shot-holes averaged about 30 m depth. During the 1958–59 traverse, the ice-drill became unserviceable, and measurements were made almost exclusively using patterns of charges supported above the snow surface.[32] Charges used in shooting from bore-holes ranged from 0·1 kg to 2·3 kg; a total of 4 kg was used in each air-shot. Refraction measurements, supplementing reflection measurements, were made during these traverses, first to find corrections for low velocities near the ice-surface, and secondly to determine *P* velocities deep in the ice. Up to 10 kg of explosive were used for the refraction shots.

Both reflection and refraction methods were also used in the semi-detailed traverses. In all cases, the seismic measurements were supplemented by gravity measurements at closer intervals, made with a Worden gravimeter. Temperatures were measured at several depths in all shot-holes before each explosion. Altitudes were measured barometrically, with simultaneous readings at stations about 8 km apart, yielding altitude increments along the traverses.

C. *Results*

Tables 1(a), 1(b) and Figs. 3, 4 exhibit the ice-thickness and altitude results.

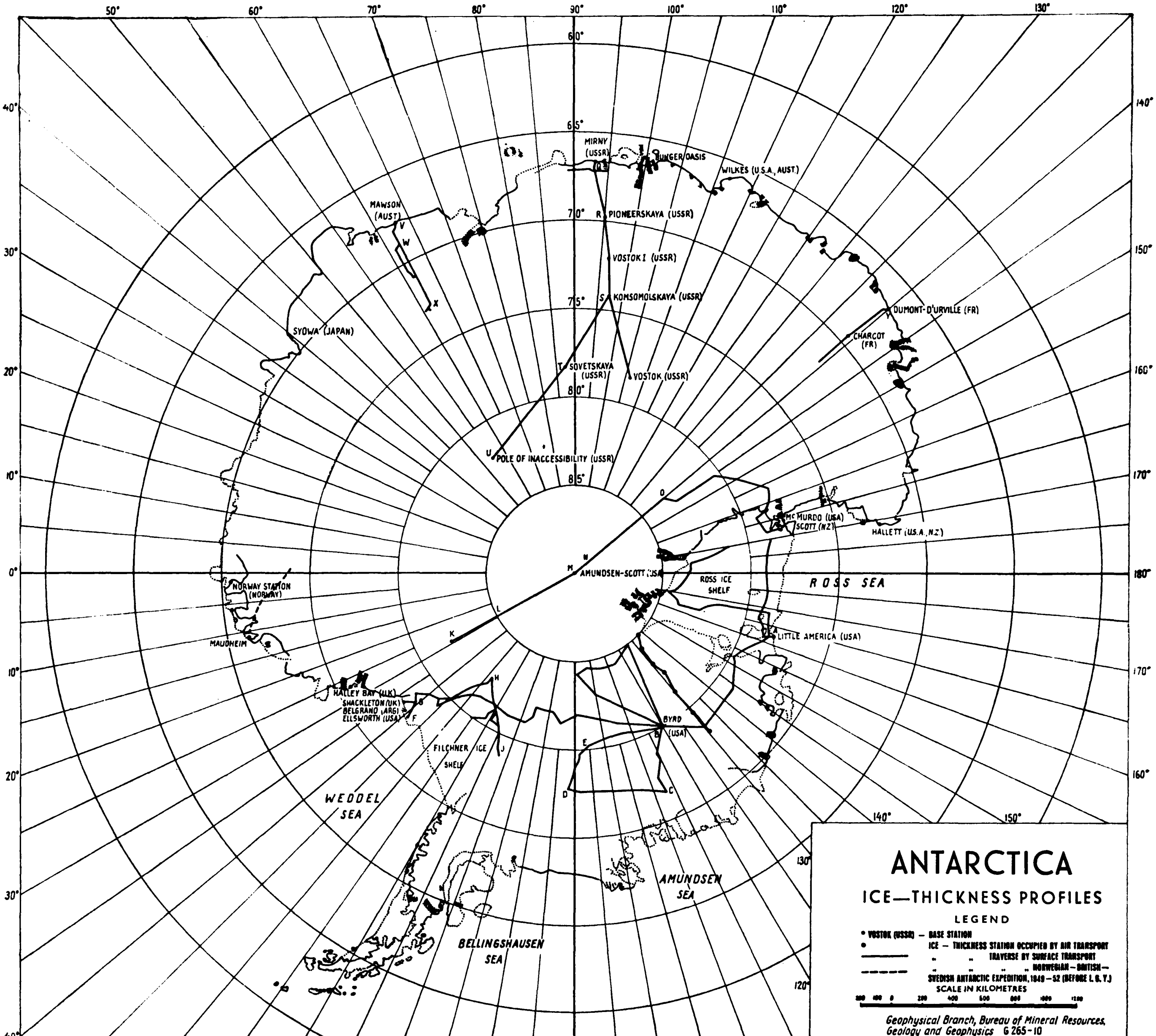


FIG. 1. Map of Antarctica showing IGY and certain earlier traverses.

Figure 5 shows the P velocity variation with depth in the ice at station SP 25, as deduced from refraction measurements made to a distance of 4300 m from the shot-point. This curve, with limiting velocity 3895 m/sec, was used in computing ice-thicknesses. A P velocity of 4850 m/sec in the subglacial rock

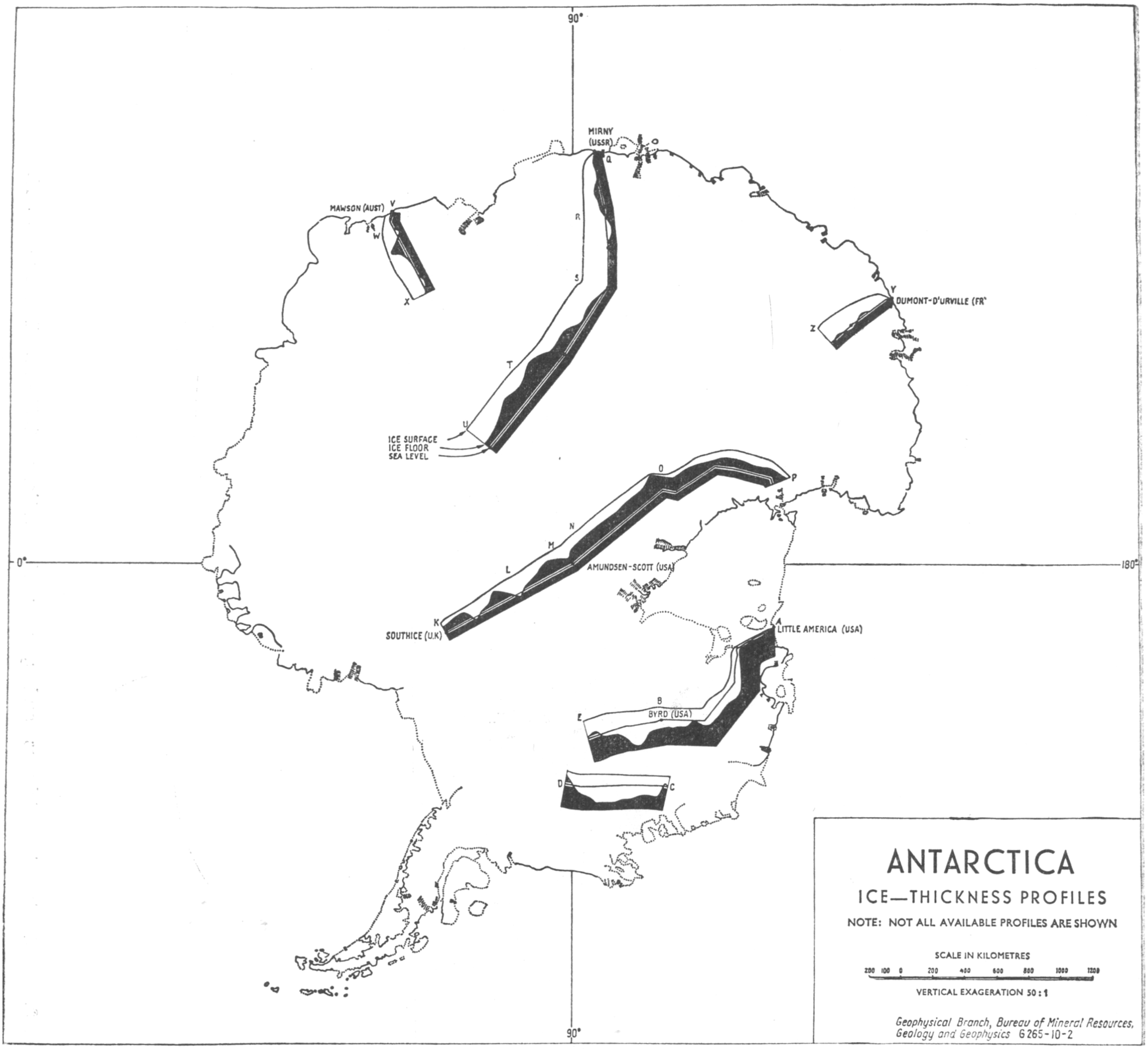


FIG. 2. Overall picture of selected ice-thickness results.

near the Framnes Mountains was derived from measurements in the semi-detailed traverses.

Attempts were made to measure S velocities in the upper ice-layers, but results are not available yet.

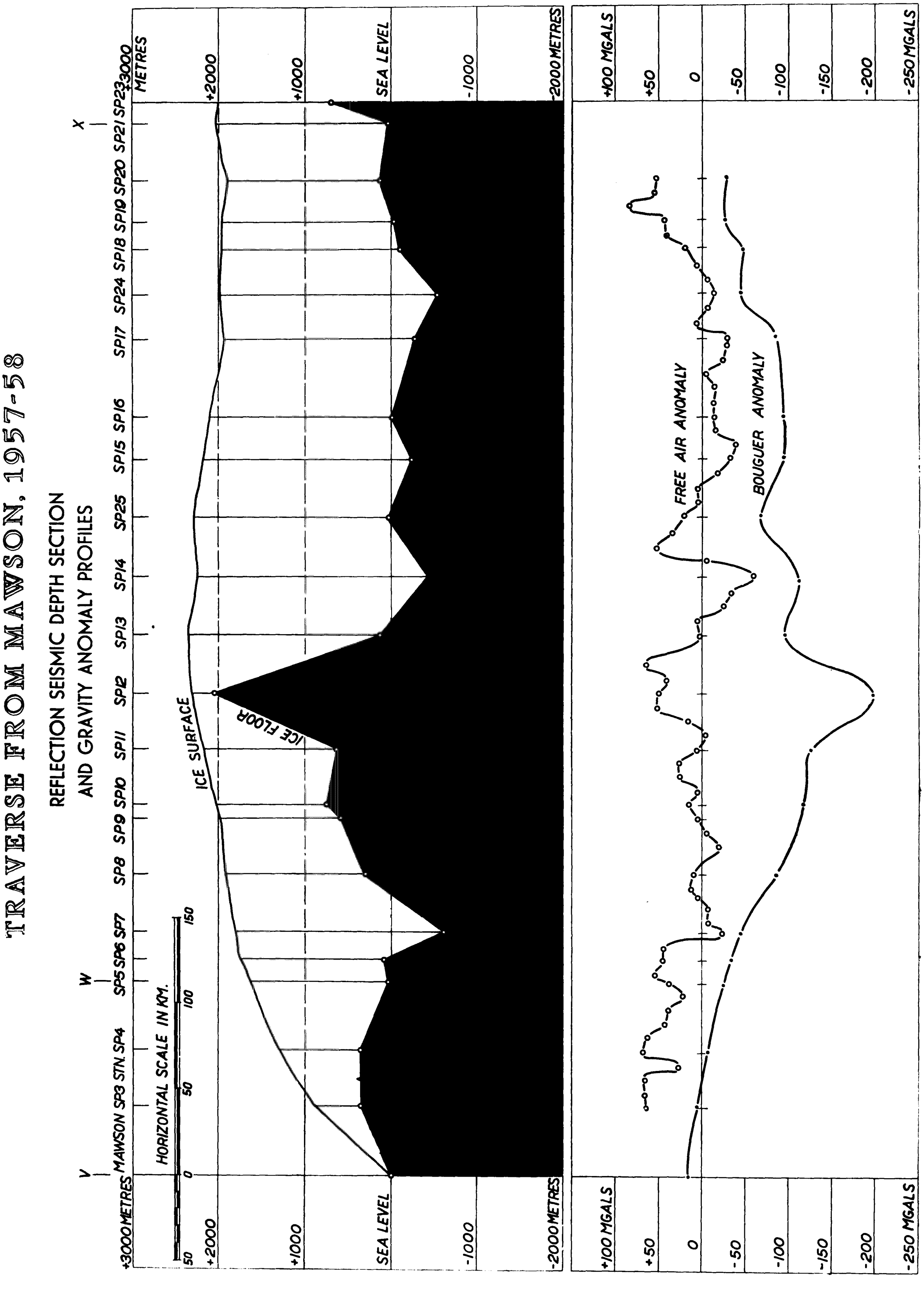


FIG. 3. Australian ice-thickness and gravity observations, 1957-58. Traverse VWX.

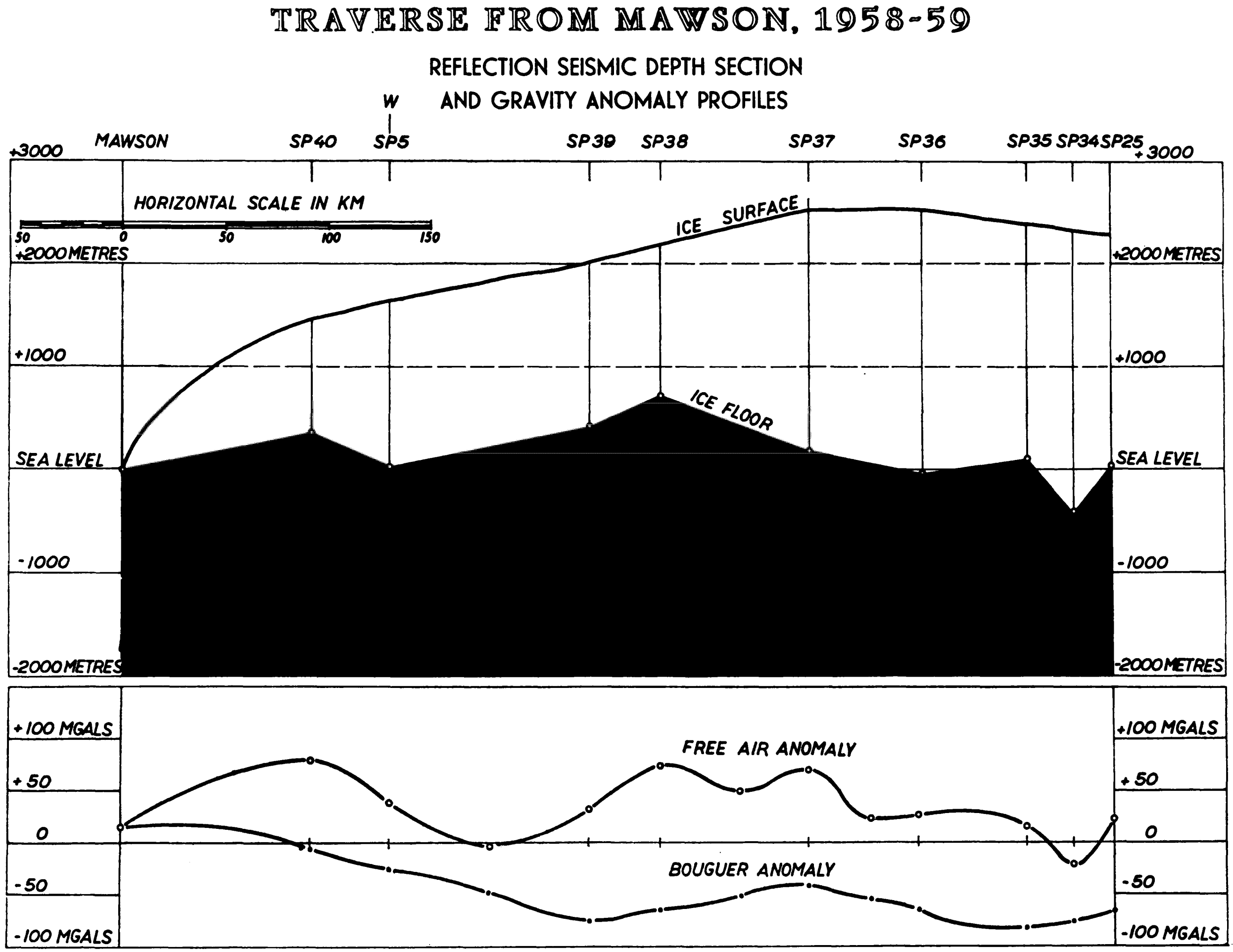


FIG. 4. Australian ice-thickness and gravity observations, 1958–59. Traverse through W.

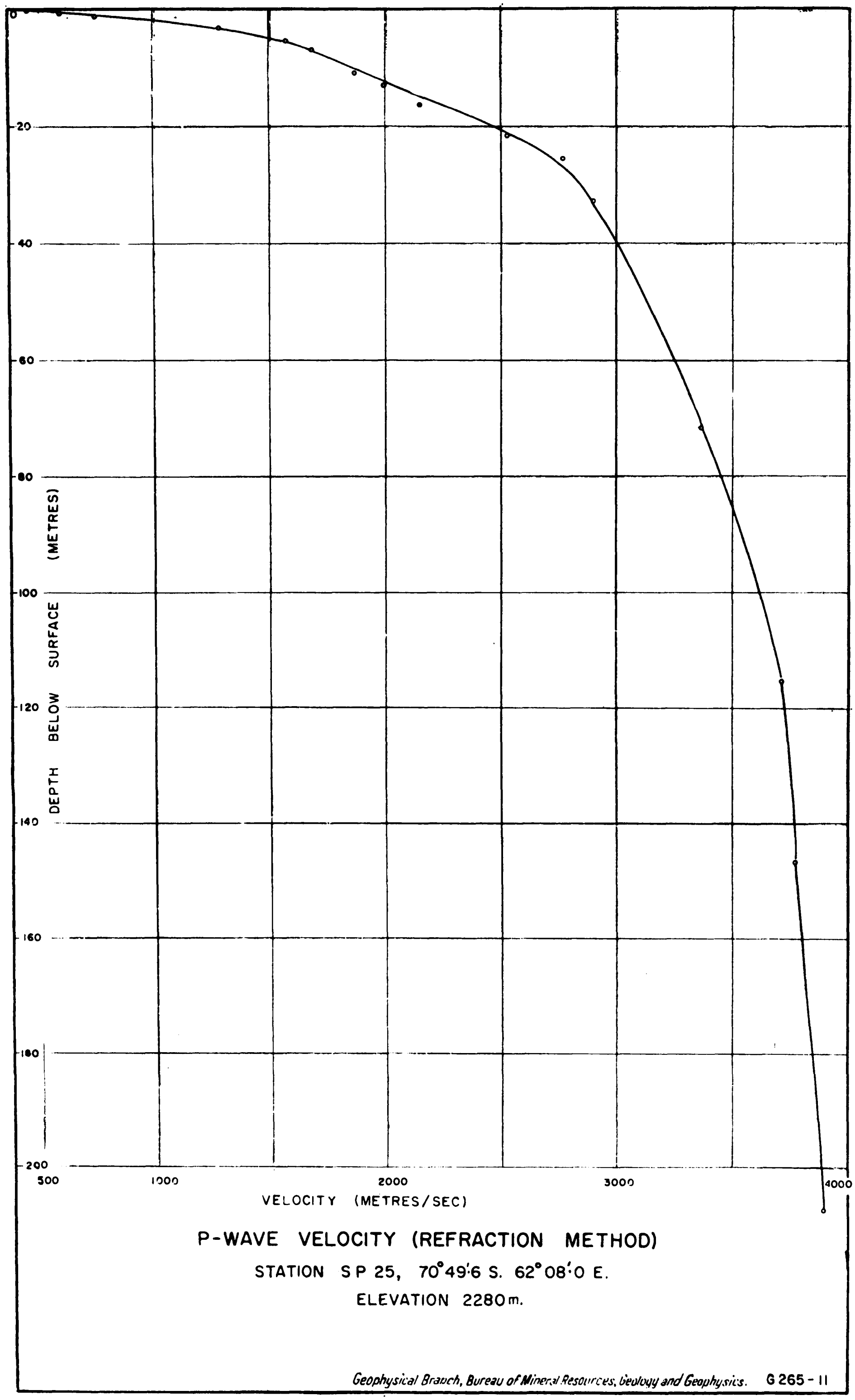


FIG. 5. *P* velocity against depth, Australian measurements.

The greatest ice-thickness found was 2730 m at station SP 34. The greatest ice-surface elevation was found to occur about 200 km inland from the coast over a submerged mountain range. This range is a westward extension of a range of exposed peaks 80 km away. The regional traverses showed that both ice-surface and rock-surface topography are related to the Lambert Glacier-Amery Ice Shelf depression[15, 29].

D. *Personnel*

The following personnel were engaged in the seismic field work either exclusively or for a considerable proportion of the time during the journeys.

Regional traverse (1957–58): K. B. MATHER*, M. J. GOODSPEED**, M. MELLOR, R. L. WILLING, N. COLLINS, B. SHAW;

Regional traverse (1958–59): I. ADAMS*, E. E. JESSON**, J. R. BLAKE, D. A. BROWN, F. A. SMITH;

Semi-detailed traverses (1958): F. A. SMITH*, E. E. JESSON**, G. A. KNUCKEY, I. R. McLEOD, P. W. KING, G. CHANNON, R. E. T. OLDFIELD.

* Traverse party leader
** In charge of seismology

6.3.2. *France*

A. *Traverse*

The base station was Dumont d'Urville (A 979, 66°40′ S, 140°01′ E), on the Antarctic coast.

A meridional traverse (*YZ*, Fig. 1) was made from Dumont d'Urville through Charcot (A 985, 69°22′ S, 139°01′ E) to a total distance of 480 km, with return along the outward route.

B. *Procedures*

Seismic *P*-wave reflection methods were used for ice-thickness measurements. Refraction measurements to a horizontal distance of 1733 m from Charcot were made to determine velocities in the upper snow levels. The seismic recording equipment, model G-33 of Southern Industrial Electronics, was fitted with 12 channels.

In reflection shooting, charges were at first fired on the snow-surface. But when interference from surface waves etc. was found to be too marked, shots were fired at the bottom of holes up to 18 m deep, drilled using an S.I.P.R.E. hand auger. Charges up to 1 kg were used. Temperatures in bore-holes were taken at some stations as part of another project[16].

Velocity at depth in the ice, for use in converting reflection times to ice-thicknesses, was derived from bore-hole temperatures, using empirical relations of Robin[37] between ice temperature, stress and *P* velocity.

Altitudes were measured barometrically and tied to a value for the altitude of Charcot station. This value was obtained from a study of atmospheric pressure and temperature records from Charcot and radiosonde results from Dumont D'Urville, Vostok I and Komsomolskaya, taking into account the effects

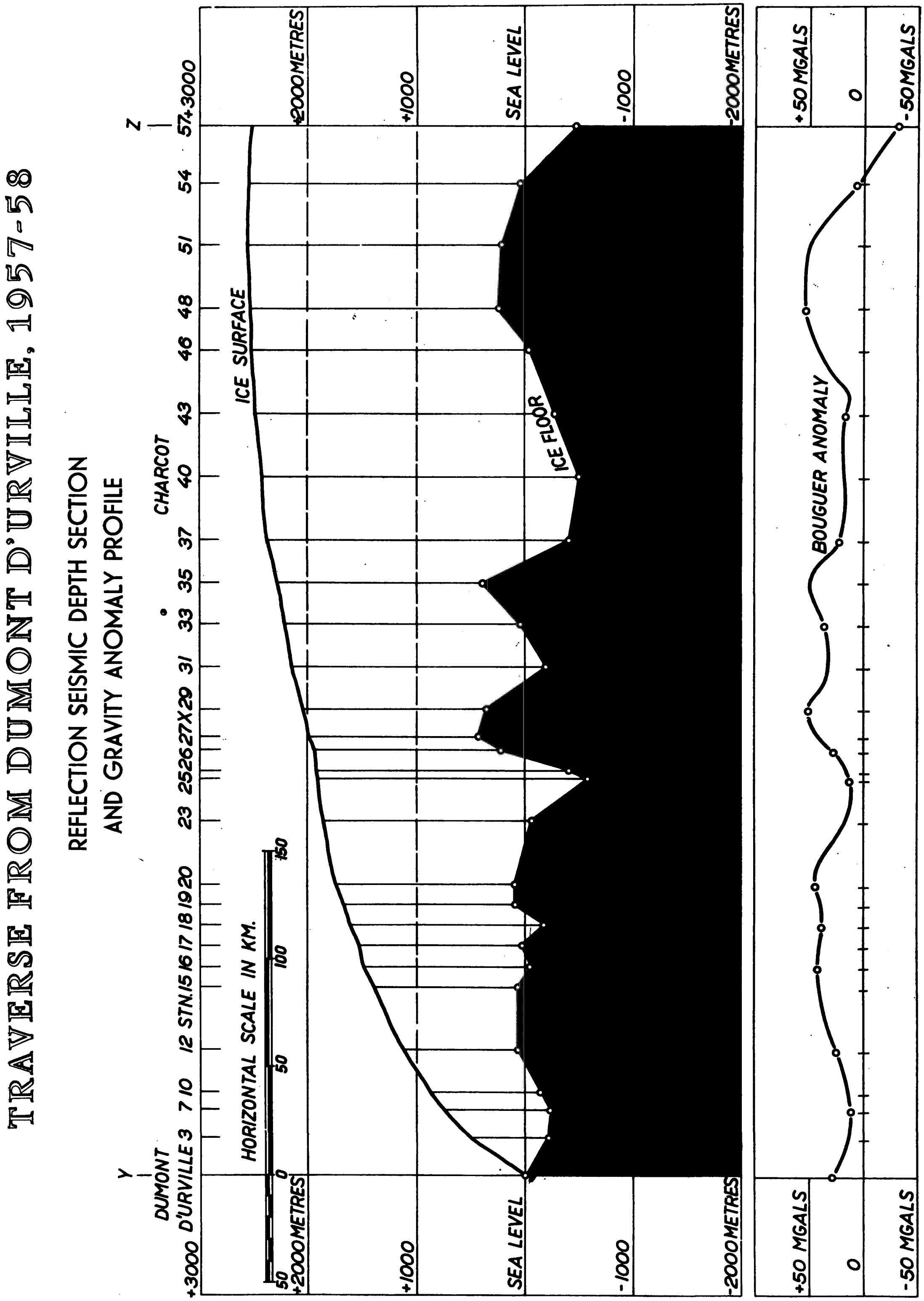


FIG. 6. French ice-thickness and gravity observations, 1957–58. Traverse YZ.

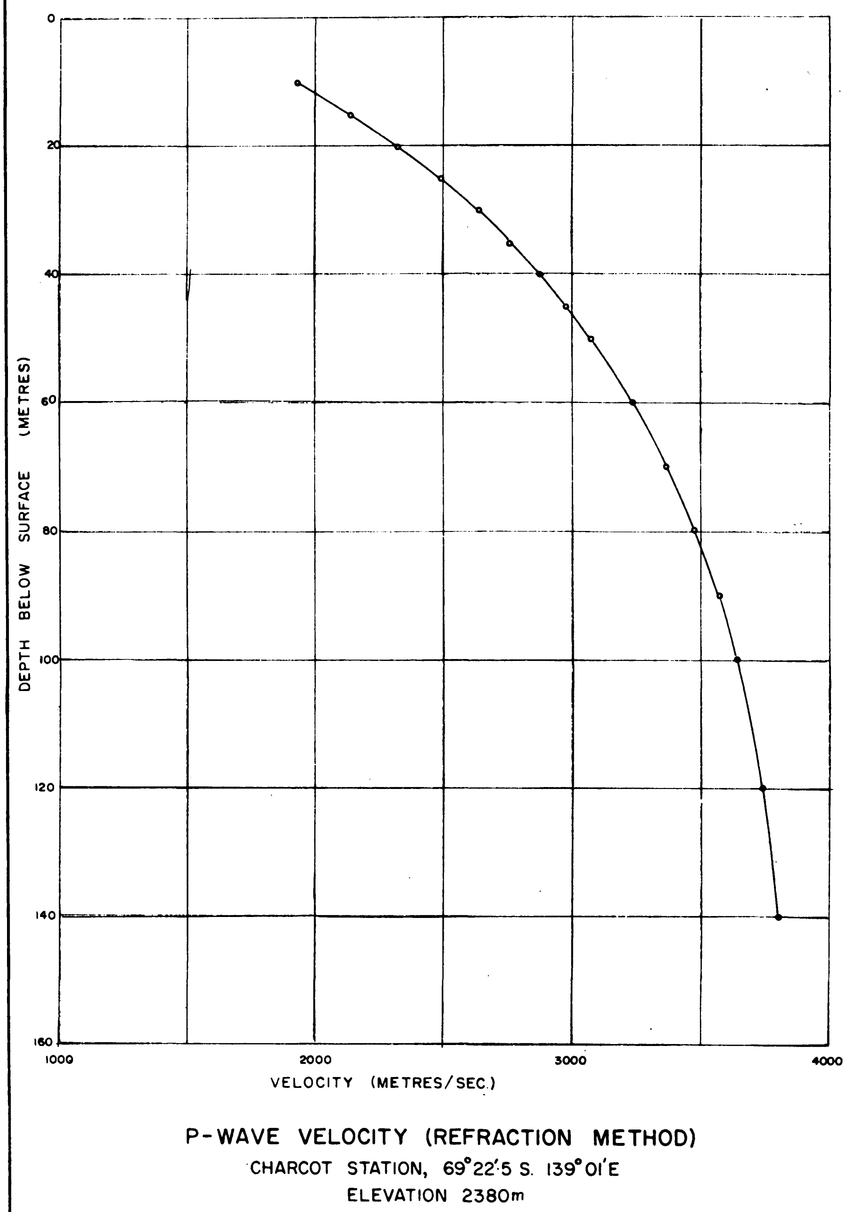


FIG. 7. *P* velocity against depth, French measurements.

Geophysical Branch Bureau of Mineral Resources, Geology and Geophysics. G 265-12 of the katabatic wind and the temperature inversion above the snow surface[13].

C. *Results*

Table 2 and Fig. 6 exhibit the ice-thickness and altitude results. Figure 7 shows the P velocity variation with depth at Charcot, with limiting velocity of 3800 m/sec at 140 m depth.

On this traverse, the ice-floor was found to be nearly horizontal, the depth averaging 40 m below sea-level. The ice-surface profile is well approximated by an arc of the ellipse

$$\left(\frac{h}{2{\cdot}68}\right)^2 + \left(\frac{550\text{-}x}{550}\right)^2 = 1,$$

where h km is the altitude and x km the distance from the coast[22].

D. *Personnel*

The following personnel carried out the ice-thickness field work during 1957–58: B. IMBERT*, R. SCHLICH.

6.3.3. *Japan*

A. *Programme*

Plans for ice-thickness measurements were based on Syowa (A 984, 69°00′ S, 39°36′ E). The programme was severely curtailed through the enforced abandonment of the base, due to natural causes, from February 1958 to January 1959. Preliminary studies were carried out on an iceberg and on fast ice near Syowa, in preparation for later work on the ice-cap.

B. *Procedures*

The seismic equipment used was the M-3 Model of Electro-Technical Laboratories. A hand auger and two mechanically-powered drills were available for boring holes. On the iceberg, shots were fired at the bottom of holes 4 m deep. From fast ice close by the iceberg, charges were exploded in the sea-water.

C. *Results*

Surface, refracted, reflected and other waves were identified, and compared with results in other areas[30].

D. *Personnel*

The following expedition members carried out the field work: S. MURAUCHI, T. TATEISHI, T. MATUMOTO.

6.3.4. *U. S. S. R.*

A. *Traverse*

Initially, traverses were based on Mirny (A 978, 66°33′ S, 93°01′ E), on the coast. During 1957–59 traverses were made:

(i) from Mirny through Komsomolskaya (A 960, 74°05′ S, 97°29′ E), to the Pole of Inaccessibility (82°06′ S, 54°58′ E), a journey of 2100 km (Q*RSTU*, Fig. 1);

* Traverse party leader.

(ii) between Komsomolskaya and Vostok* (A 996, 78°27′ S, 106°52′ E), some 550 km;

(iii) between Vostok and South Pole;

(iv) in some coastal regions.

The traverses under (i), (ii) and (iii) used seismic reflection methods. The traverses under (iv) used both reflection and refraction methods. Gravity measurements under (ii) and (iii) were made to interpolate thicknesses between seismic stations.

Efforts under (iv) were largely directed towards recording waves that had penetrated the subglacial rock. This was with a view to aiding in the identification of waves reflected near the ice-rock boundary.

Small-scale measurements were made on Drygalski Island.

B. *Procedures*

The seismic equipment included model SPM-16 geophones. Twenty-four channels were recorded, using portable PSS-24 equipment, and in some cases heavier SS-26-51-D equipment. Altitudes were measured by radio-altimeter from an aircraft. The aircraft flew above the near-surface temperature inversion and its altitude was measured barometrically[6]. Gravity was measured using three model SN-3 gravity meters.

In reflection shooting in areas of deep ice cover, shots up to 5 kg were fired at depths up to 60 m, where the velocity is not too far from the limiting value. Single geophones were used.

In shallow refraction shooting, shots were fired at the surface. Velocities in the upper ice-layers were further studied by "uphole shooting" in 60 m boreholes.

Independent measurements were made *in situ*, on a moraine exposure near Bunger Oasis, of elastic parameters of the heterogeneous medium formed of moraine material embedded in ice.

C. *Results*

Table 3 gives the results of all ice-thickness measurements by seismic and gravity methods. Figures 8 to 11 show a profile of the results of the main traverse from Mirny, through Komsomolskaya and Sovetskaya (A 998, 78°24′ S, 87°35′ E) to the Pole of Inaccessibility. It should be noted that the Figures were prepared from earlier data than that tabulated, and in cases where discrepancies exist the Table should be taken as authoritative. Some post-IGY results are included.

Figure 12 shows the P velocity variation with depth for a typical station in "central Antarctica" (specified as relating to all regions more than 250 km from the coast).

The limiting P velocity in the ice was found to increase from 3800 m/sec

* The distinction between the U. S. S. R. Antarctic stations "Vostok I" (72°08′ S, 96°35′ E) and "Vostok" (78°27′ S, 106°52′ E) should be noted.

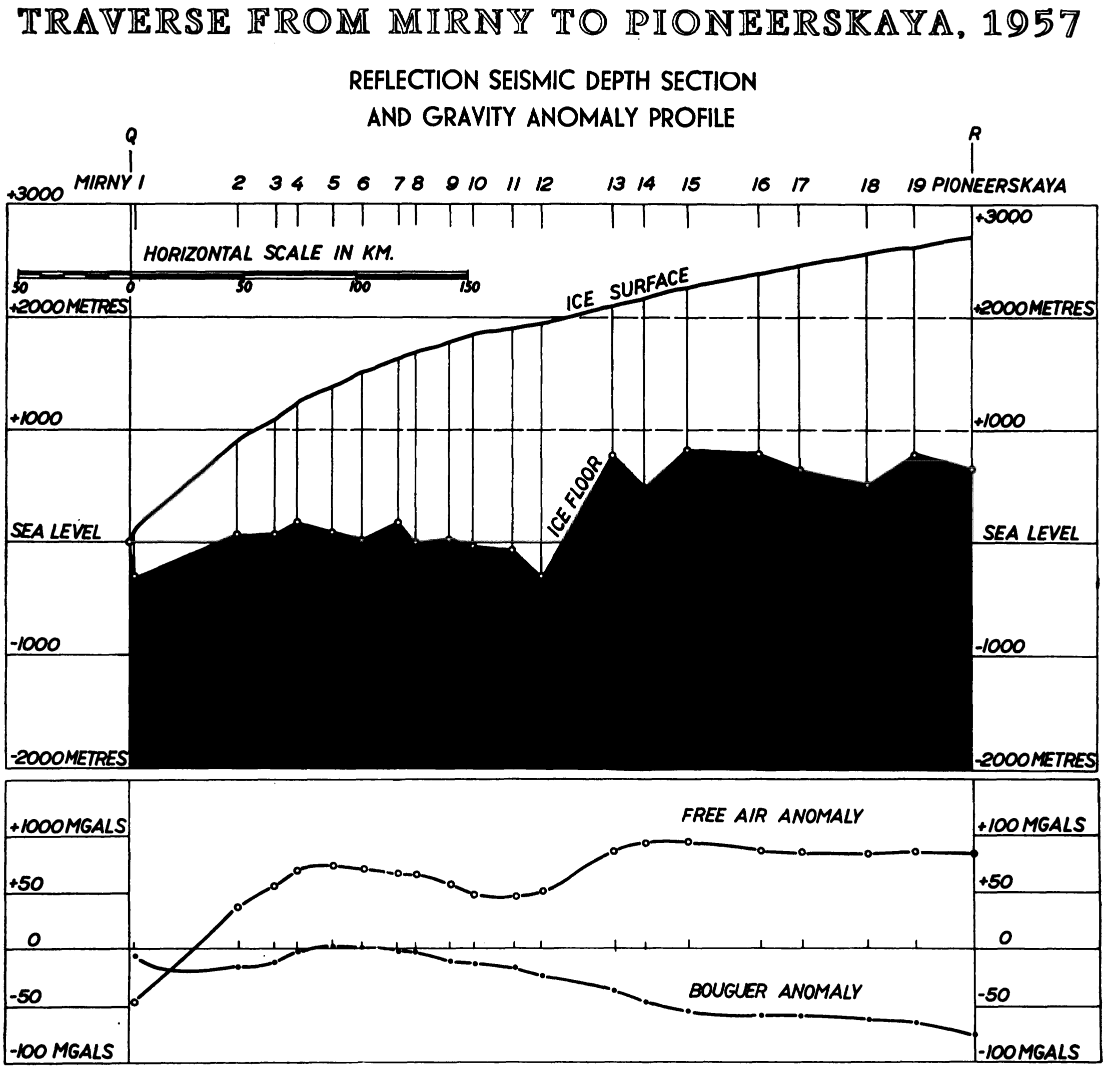


FIG. 8. U.S.S.R. ice-thickness and gravity observations. Traverse QR.

near the coast to 4000 m/sec near the Pole of Inaccessibility, presumably as a consequence of decrease in the ice temperature.

In coastal regions, some reflections were associated with moraine-bearing layers, not usually exceeding 50–100 m in thickness, at the bottom of the ice cover. Amplitude studies and measurements made *in situ* (referred to above) implied that these layers contained about 20 per cent of detrital material. Bouguer and free-air (Faye) gravity anomalies were inferred from gravity measurements in conjunction with seismically determined ice-thicknesses at control points.

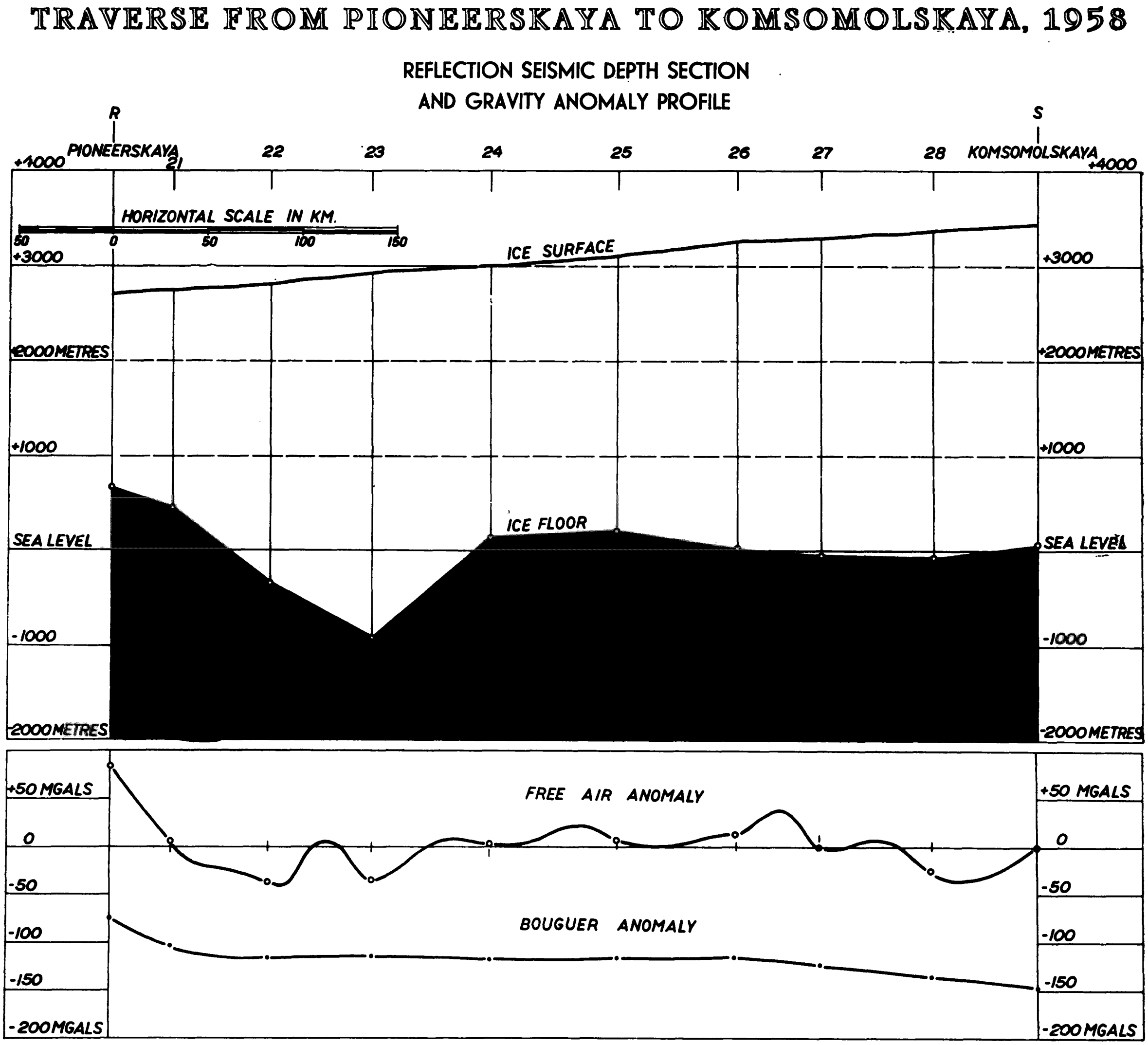


FIG. 9. U.S.S.R. ice-thickness and gravity observations. Traverse RS.

SOROKHTIN *et al*[40] draw the following conclusions from the results obtained: Subglacial rock is generally close to sea-level in the first 200 km from Mirny. Between 200 and 400 km, there is a major plateau-like uplift averaging 600–700 m above sea-level (Sub-glacial Mt. Golicyn). Between 500 and 550 km, there is a pronounced trough. This is followed by a region, extending to the vicinity of Komsomolskaya (about 870 km from Mirny), where the rock-surface is again close to sea-level (Schmidt Plain). Next comes a very large mountain massif, culminating in a peak nearly 3000 m high at about the 1700 km point (Sub-glacial Mt. Gamburtsev). From here to the Pole of Inaccessibility, the height

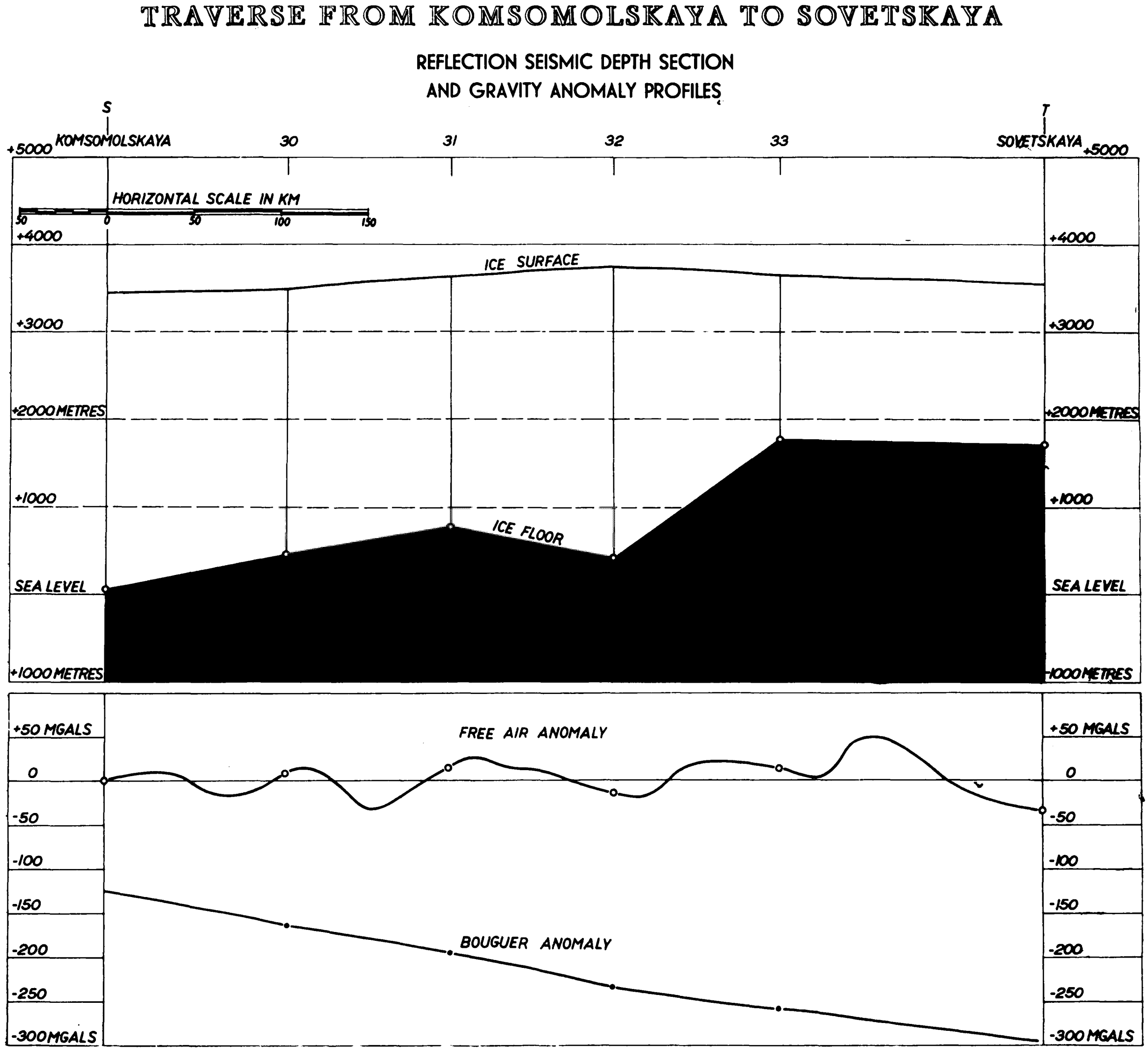


FIG. 10. U.S.S.R. ice-thickness and gravity observations. Traverse ST.

of the rock-surface steadily declines to 800 m above sea-level. Approximately 85 per cent of the whole profile shows rock-surface above sea-level.

The gravity anomalies support the inference that this part of Antarctica is continental in character. The free-air anomalies are generally small and positive and significantly correlated with the local rock relief. The Bouguer anomalies are negative and increase in absolute magnitude with distance from the coast until they reach a value of about—300 mgal in the area of the Gamburtsev mountains. The general smallness of the free-air anomaly is taken as evidence that the ice-burden is nearly completely isostatically compensated.

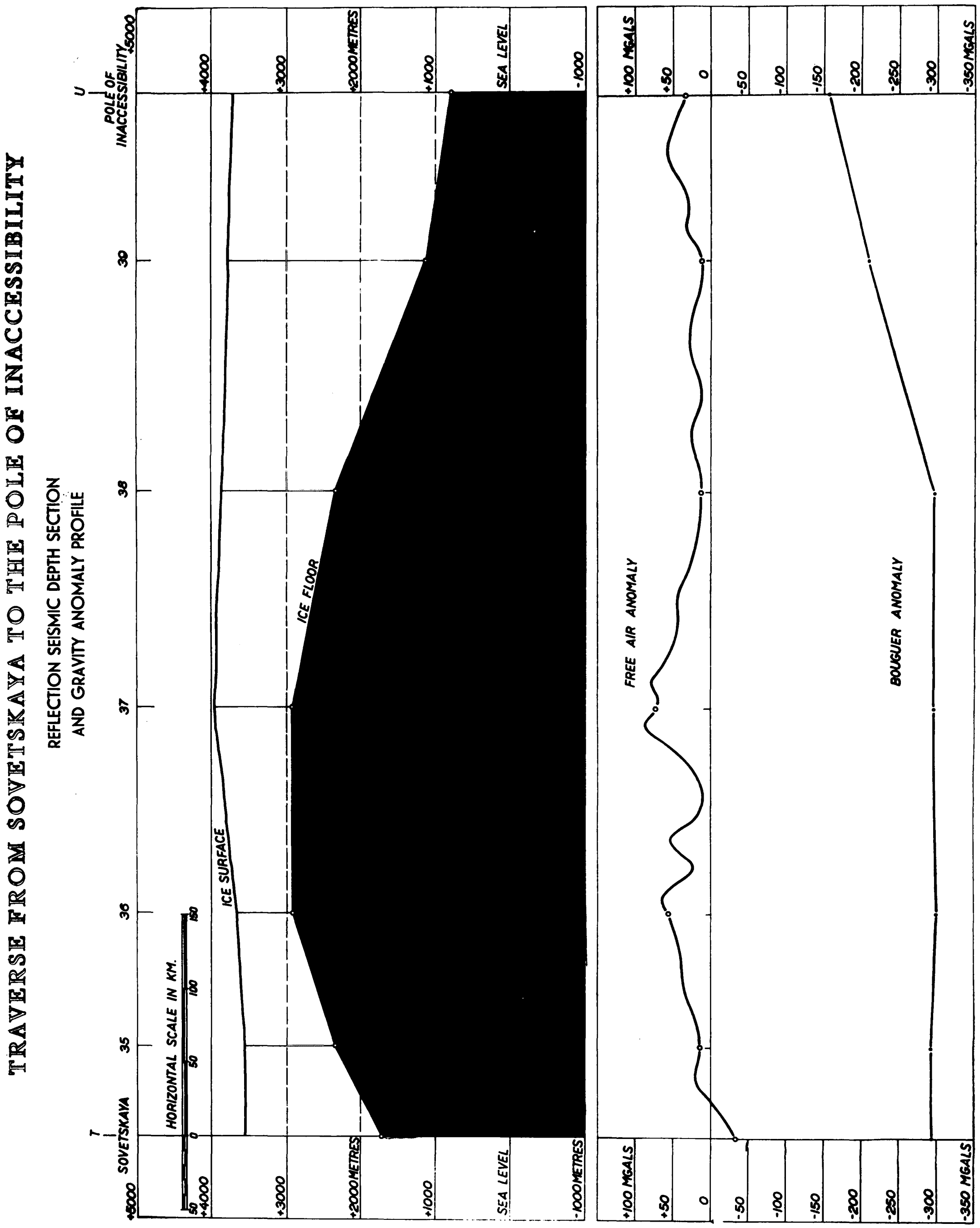
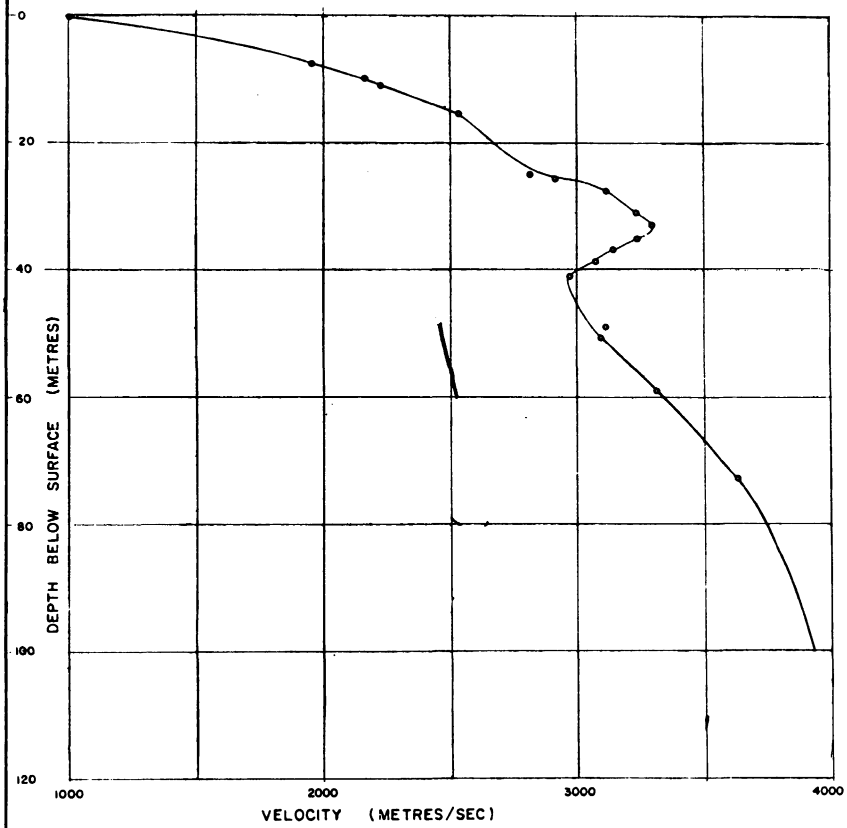


FIG. 11. U.S.S.R. ice-thickness and gravity observations. Traverse TU.



P-WAVE VELOCITY (VELOCITY LOGGING METHODS)
NEAR PIONEERSKAYA, 69° 44′ S. 95° 30′ E.
ELEVATION 2700 m.

FIG. 12. *P* velocity against depth, U.S.S.R. measurements.

Geophysical Branch, Bureau of Mineral Resources, Geology and Geophysics. G265-14

D. *Personnel*

The following personnel participated in the ice-thickness measurements in the field during the IGY:

Participants of the Second Complex Antarctic Expedition 1956–57: O. K. KONDRATYEV, S. S. LOPATIN, S. A. MANILOV.

Participants of the Third Complex Antarctic Expedition 1957–58: O. G. SOROKHTIN, V. I. KOPTEV, I. N. AVSIUK.

Participants of the Fourth Complex Antarctic Expedition 1958–59: A. P. KAPITSA, YU. F. DURYNIN, S. A. KHRUSHCHEV.

6.3.5. *United Kingdom*

A. *Traverse*

The British Commonwealth Trans-Antarctic Expedition started from Shackleton base (A 993, 77°59′ S, 37°10′ W). Ice-thickness measurements were carried out from Southice (A 958, 81°57′ S, 28°50′ W), through Amundsen–Scott (A 999, 90° S) to Plateau Depot (78°01′ S, 158°23′ E), near the terminal point of the journey, Scott Base.

B. *Procedures*

Seismic *P*-wave reflection methods were used for the ice-thickness measurements, supplemented by refraction measurements to a horizontal distance of 3525 m from Southice. The seismic recording equipment comprised twelve channels of Electro-Technical Laboratories M 4a equipment. The camera displayed the straight outputs of the twelve amplifiers and also the mixed outputs of these amplifiers. Multiple geophones were used. A group of four EVS-2 geophones, connected in series, provided the signal for each channel. The electronic equipment included radio transceivers for refraction shooting. Two augers, making holes of 4 and 7·5 cm diameter, were carried.

Gravity was measured at intervals between seismic stations, using a Worden gravity meter. Altitudes were measured by barometric methods, but the results are not as yet finally available. Where gravity measurements suggested that the ice was thin, shots were often fired in shallow pits less than one metre deep. Elsewhere, shots were fired at the bottoms of bore-holes 12 m deep.

C. *Results*

Table 4 gives the ice-thicknesses at the seismic stations.

Figures 13–16 show the resulting profile, but it should be noted that the *altitudes are only provisional and subject to revision.* (Meteorological records covering the period of the journey are to be further analysed.)

Figure 17 shows the *P* velocity variation with depth at Southice. A limiting velocity of 3936 m/sec was indicated 88 m below the surface.

Refracted *P* waves were recorded at Southice, but derivation of the rock-velocities requires reverse-shooting, not yet carried out.

D. *Personnel*

The following personnel were engaged in the seismic field work, either exclu-

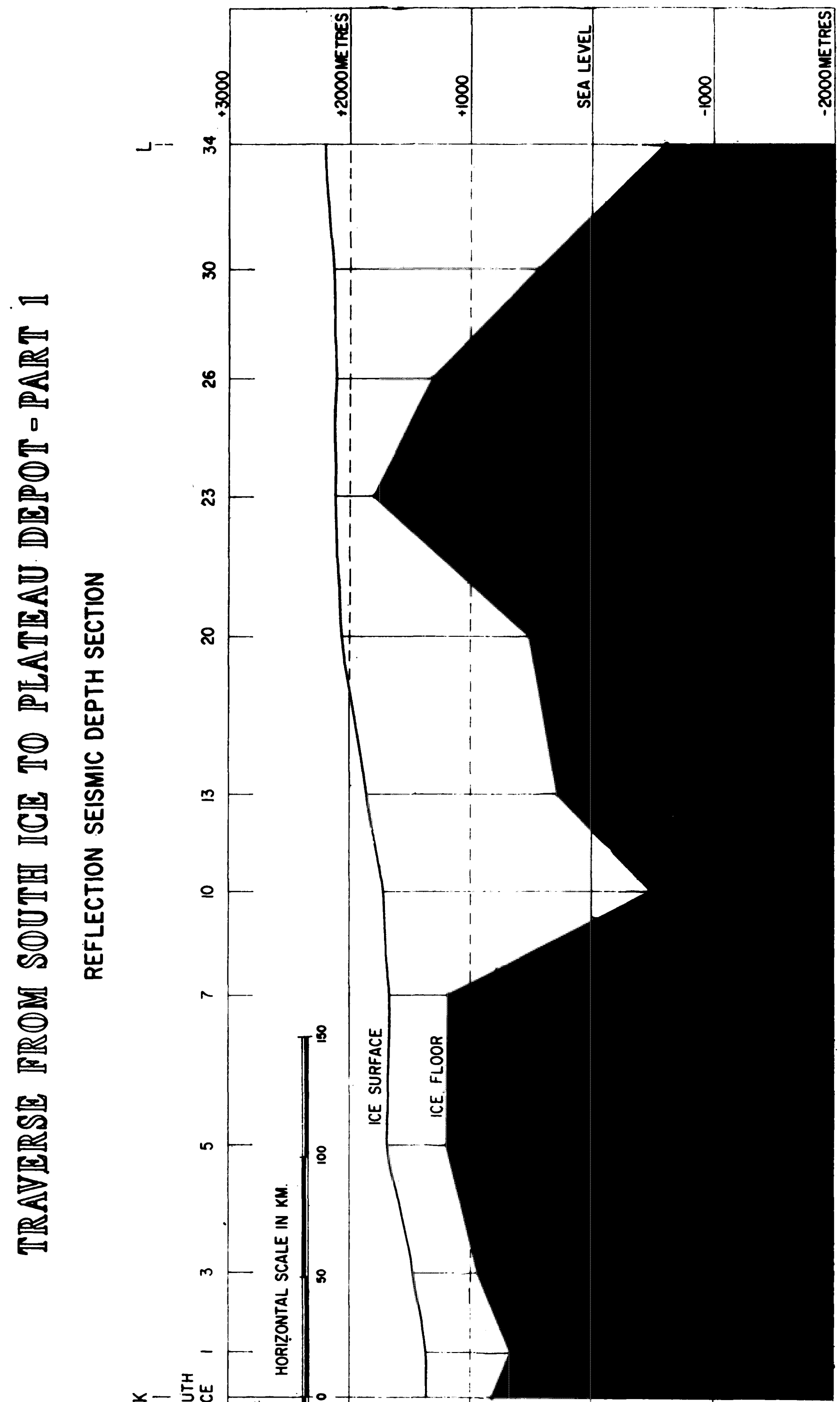


FIG. 13. U.K. ice-thickness observations. Traverse KL.

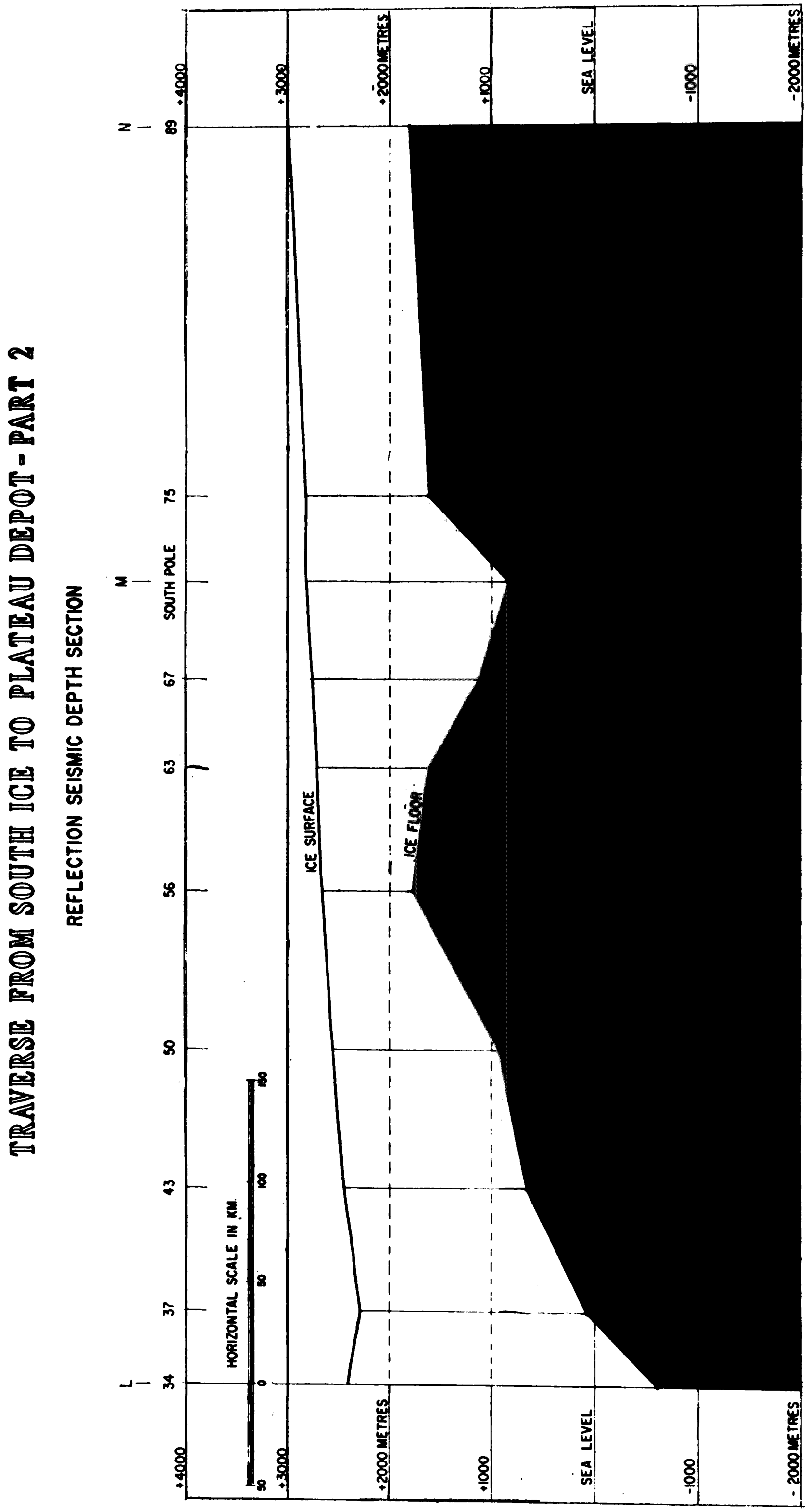


FIG. 14. U.K. ice-thickness observations. Traverse LN.

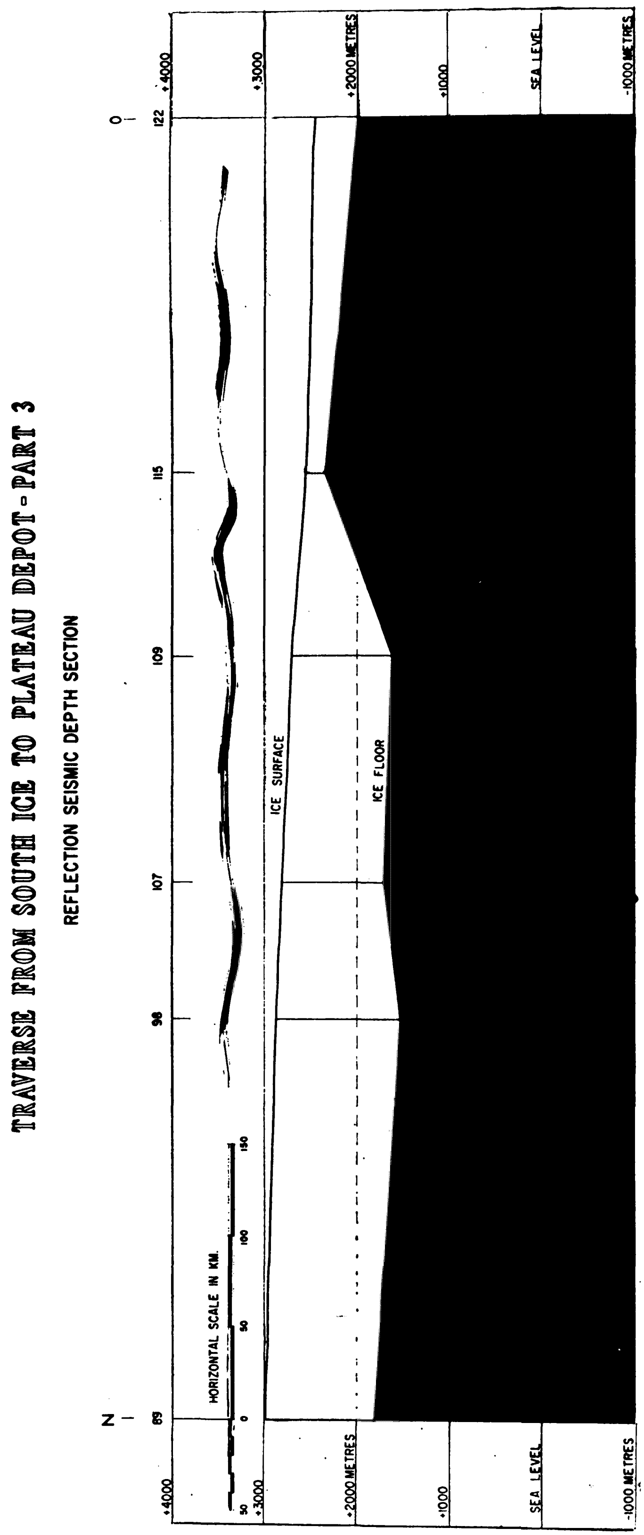


FIG. 15. U.K. ice-thickness observations. Traverse NO.

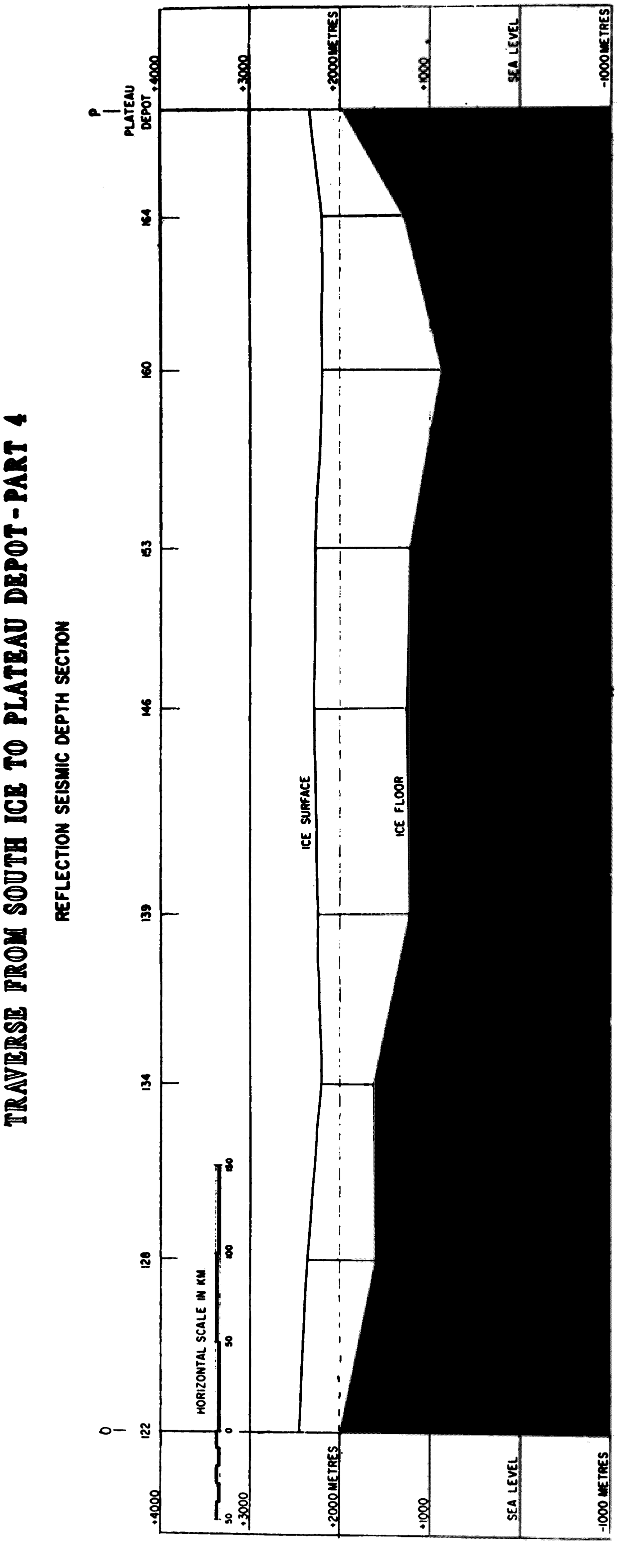


FIG. 16. U.K. ice-thickness observations. Traverse OP.

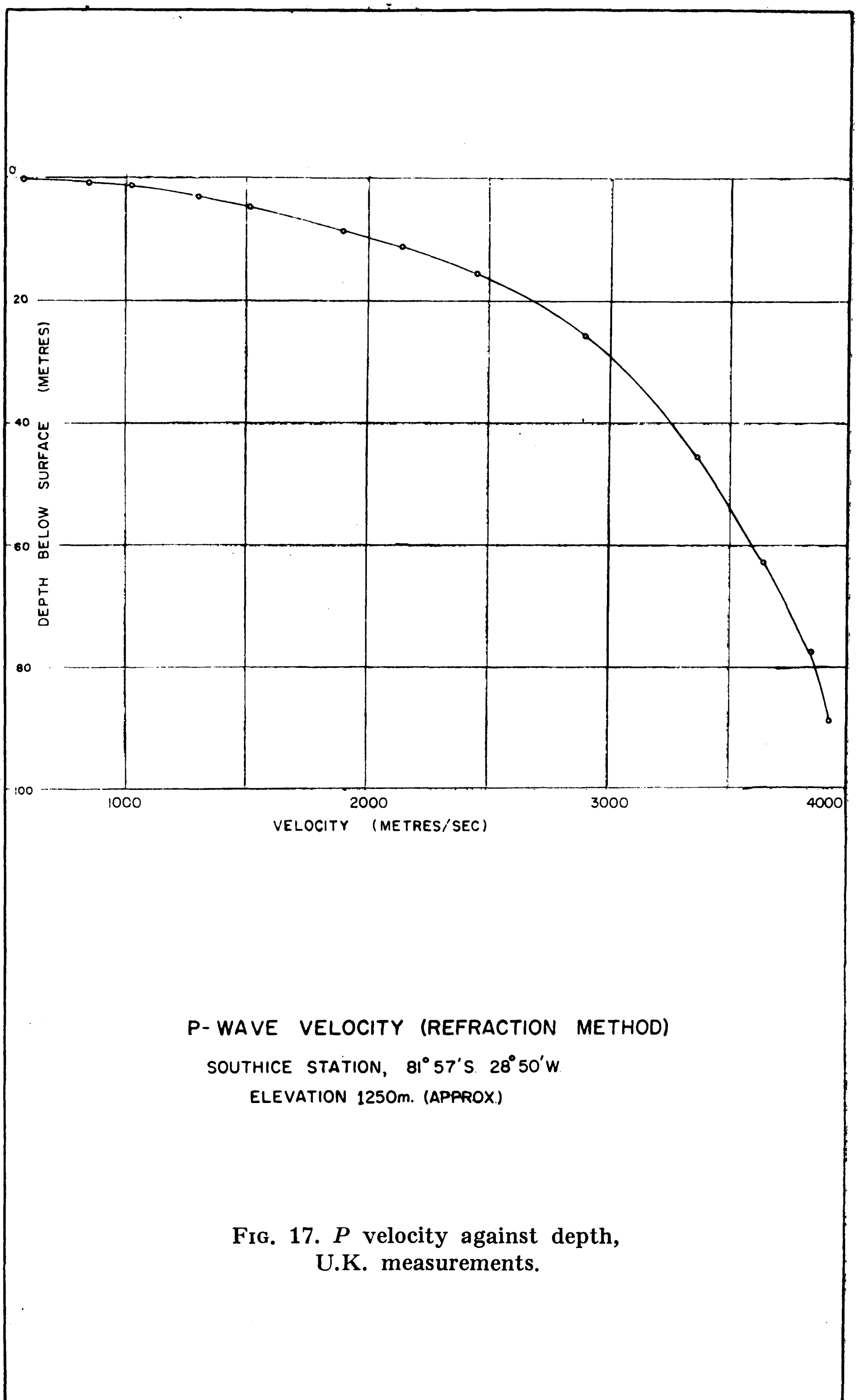


FIG. 17. *P* velocity against depth, U.K. measurements.

sively or for a considerable proportion of the time during the journey: V. E. FUCHS*, D. G. STRATTON, K. V. BLAIKLOCK, D. L. PRATT, D. E. L. HOWARD, R. A. LENTON, J. J. LA GRANGE (South Africa), J. G. D. PRATT**, A. F. ROGERS, H. LISTER, P. J. STEPHENSON (Australia), W. G. LOWE (New Zealand).

6.3.6. *United States of America*

A. *Traverses*

Traverses were made:

(i) From Little America V (A 995, 78°12′ S, 162°15′ W) to Byrd (A 997, 79°59′ S, 120°01′ W), a distance of 1038 km (*AB*, Fig. 1);

(ii) From Byrd, a closed loop extending to the east and north and covering 1900 km (*BCDEB*, Fig. 1);

(iii) From Little America V, a closed loop on the Ross Ice shelf covering 2320 km;

(iv) From Ellsworth (A 990, 77°43′ S, 41°08′ W), inland on the Filchner Ice Shelf, for a distance of 2010 km (*FGHJ*, Fig. 1);

(v) From Ellsworth to Byrd, covering 2010 km;

(vi) From Byrd, a further loop extending to the east and south and covering 1475 km;

(vii) In Victoria Land, based on Little America V and covering some 1300 km;

(viii) From Little America V, an airborne traverse with measurements at seven stations along the 130° W meridian.

B. *Procedures*

Seismic *P*-wave reflection was used in general for ice-thickness measurements. On floating ice shelves, several methods using multiple travel-paths or *S*-wave reflection times, or both, were used to determine the thickness of floating ice.

The seismic recording equipment was the Texas Instruments model 7000-B, recording 24 channels. A variety of geophones were used in recording vertical and horizontal motion. Long refraction profiles were shot at several locations. Charges up to 400 kg were fired and recorded at distances up to 22 km. Shots were fired on the surface and at depths up to 14 m in holes drilled with an S.I.P.R.E. hand auger. Short refraction profiles were shot at many stations.

Gravity values were obtained with Frost gravimeters on the oversnow traverses, and with a Worden gravity meter at the airborne stations. Altitudes were measured with Wallace and Tierman altimeters, using leapfrog methods whenever possible.

C. *Results*

Tables 5a–5c and Figs. 18–24 give the ice-thickness and altitude results for the traverses (i), (ii), (iv), respectively. Detailed results of other traverses are not yet available. The results for (ii) are shown in Figs. 19–22 in four separate legs, and for (iv) in Figs. 23, 24 in two legs.

* Traverse party leader.

** In charge of seismology.

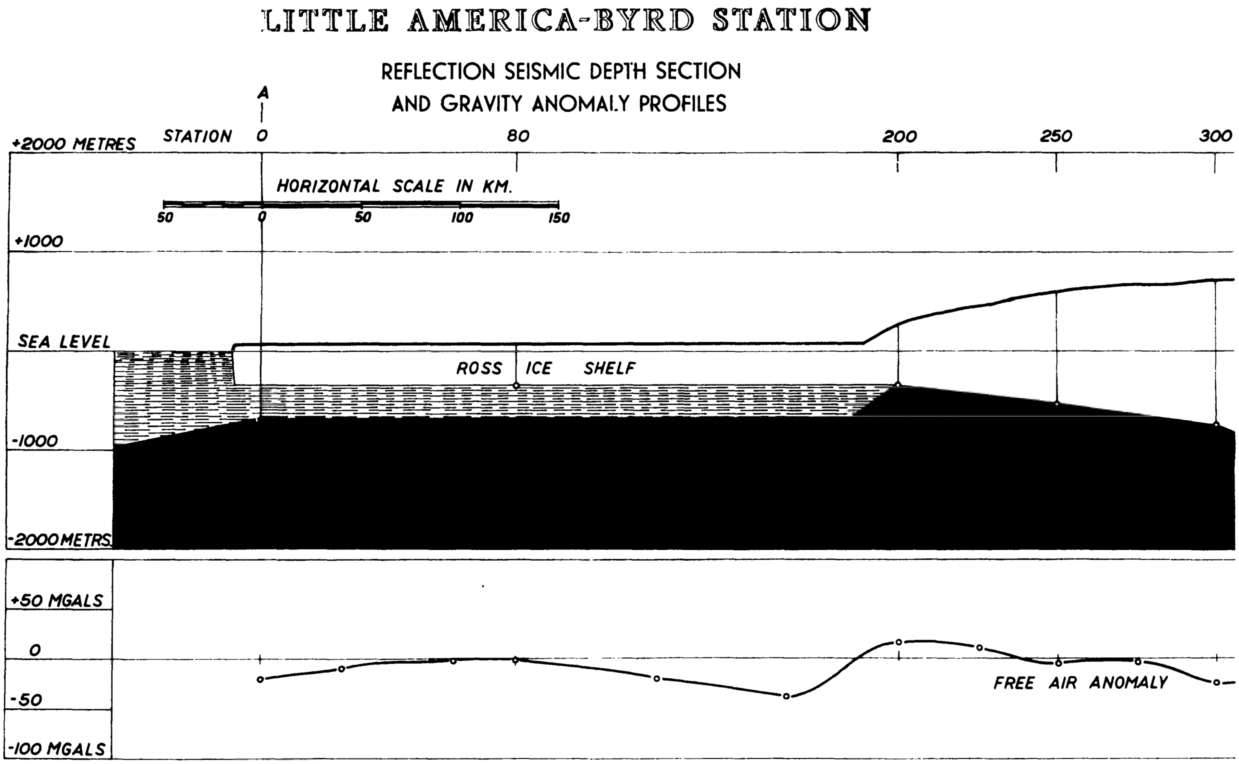


FIG. 18. U.S.A. ice-thickness and gravity observations. Traverse AB.

Fig. 25 shows the P velocity variation with depth at station 27 of the Filchner Ice Shelf traverse.

Bouguer and free-air gravity anomalies were inferred from gravity measurements and seismically determined ice-thicknesses.

BENTLEY *et al.*[2] report the existence of a low-velocity layer near the base of the ice-cap in Marie Byrd Land. They consider that doubts on the thickness and other features of the layer, and on the question of its presence in many places, diminish the accuracy with which ice-thickness values can be inferred from reflection times.

BENTLEY *et al.*[1] draw the following further conclusions from the results on hand:

A major channel exists between the Ross and Bellingshausen Seas, sufficiently deep to have been below sea-level when allowance is made for possible depression due to the present ice-burden. A deep basin within the channel, situated between Byrd Station and the Sentinel Mountains, reaches a maximum depth of 2500 m below sea-level.

The Filchner Ice Shelf extends over a much greater area than was previously supposed. This result, together with the discovery of a major channel in Marie

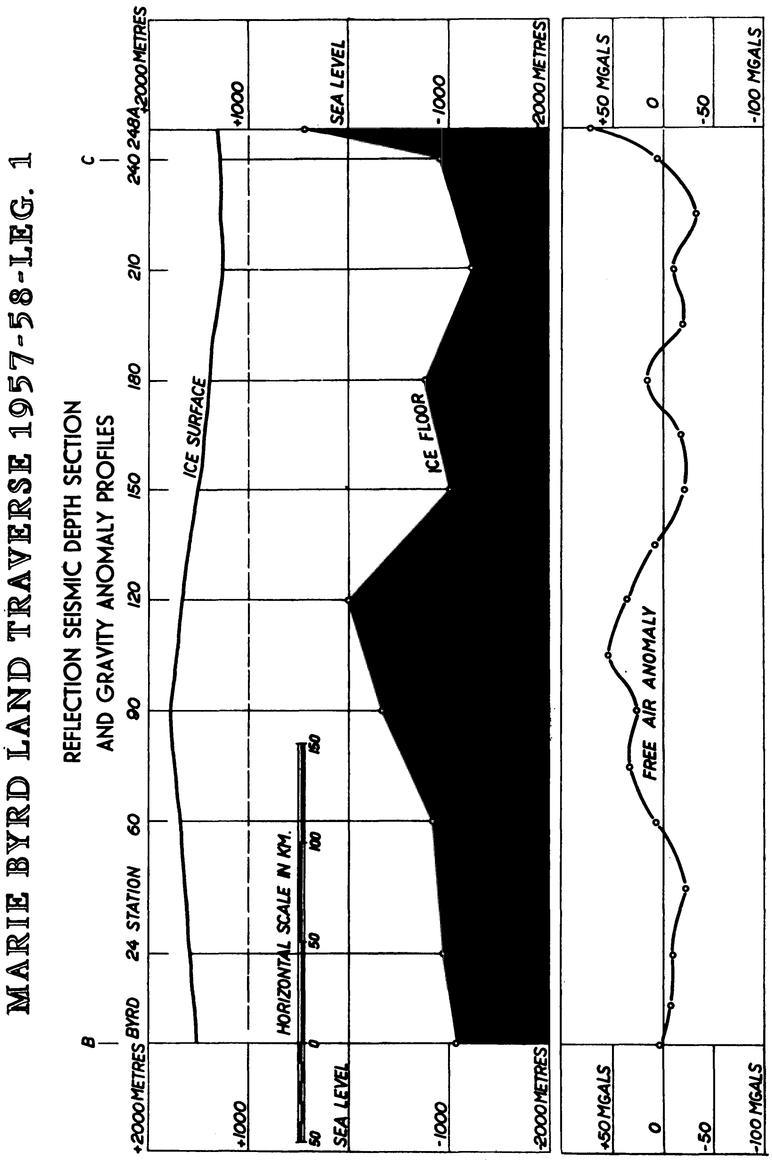


FIG. 19. U.S.A. ice-thickness and gravity observations. Traverse BC.

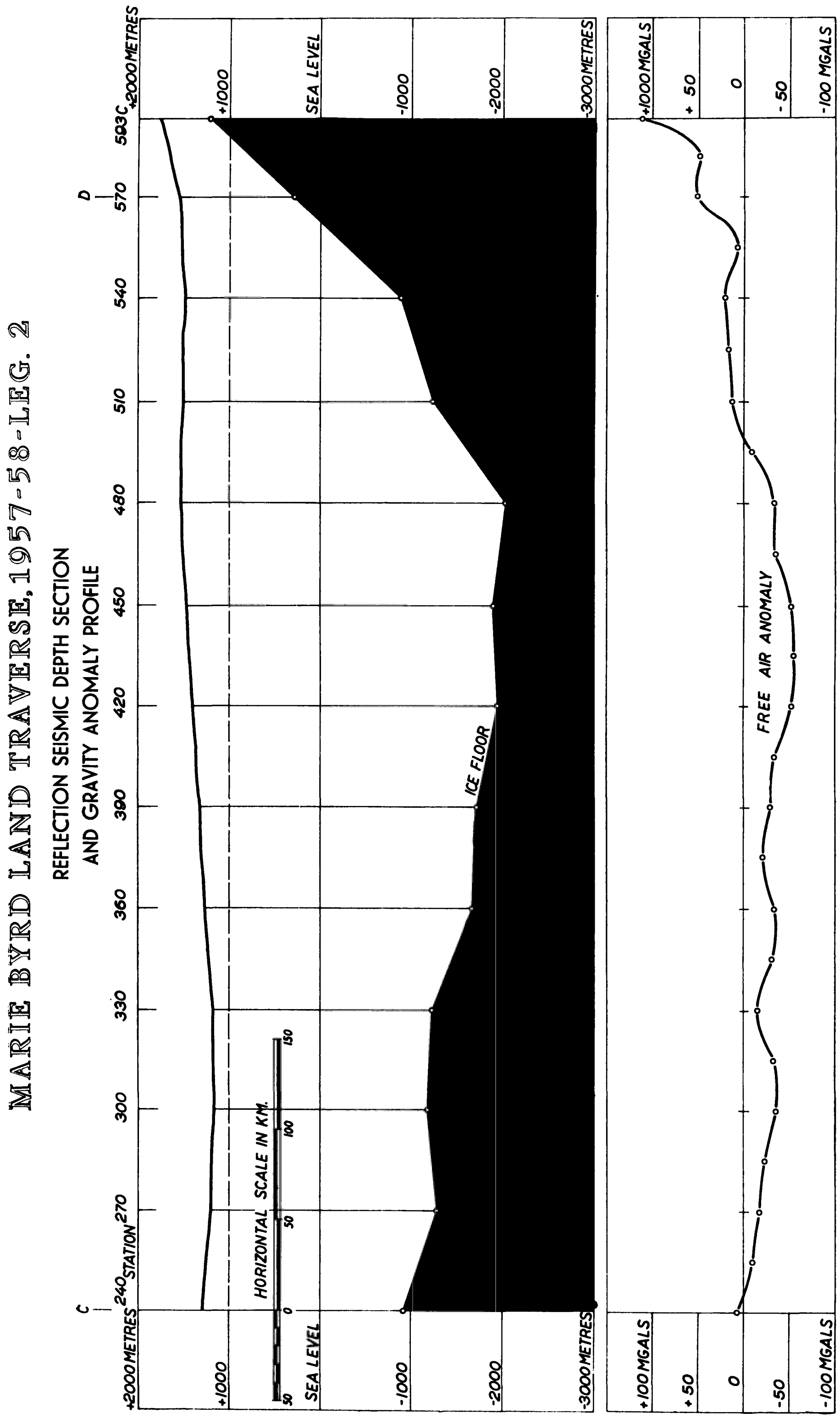


FIG. 20. U.S.A. ice-thickness and gravity observations. Traverse CD.

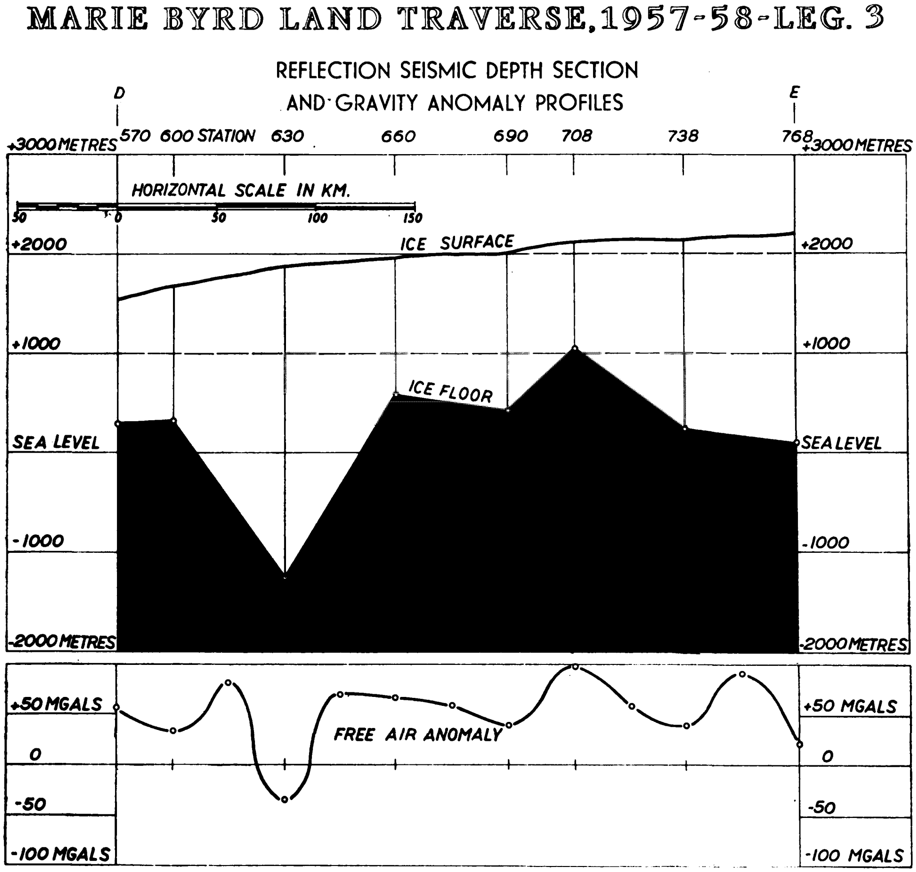


FIG. 21. U.S.A. ice-thickness and gravity observations. Traverse DE.

Byrd Land, is taken as evidence that the rock-surface of the major part of western Antarctica is below sea-level. A deep trough extends inland for several hundred km beneath the eastern area of the Filchner Ice Shelf. A trough is also found under the western and southern boundaries of the Ross Ice Shelf. The structure of the Palmer Peninsula (Graham Land) extends at least as far south as 83° S, and there is no broad connection below sea-level between the Ross and Weddell Seas.

The gravity anomalies are considered to indicate that western Antarctica is continental in character, but with the Mohorovičić discontinuity only about 30 km below sea level, except below mountain ranges. Western Antarctica appears to be in approximate isostatic equilibrium.

Bentley and Ostenso (1959), in a preliminary report, give the following detail on values of P and S velocities:

At Byrd Station (79°59′·2S, 120°01′ W), the maximum P and S velocities

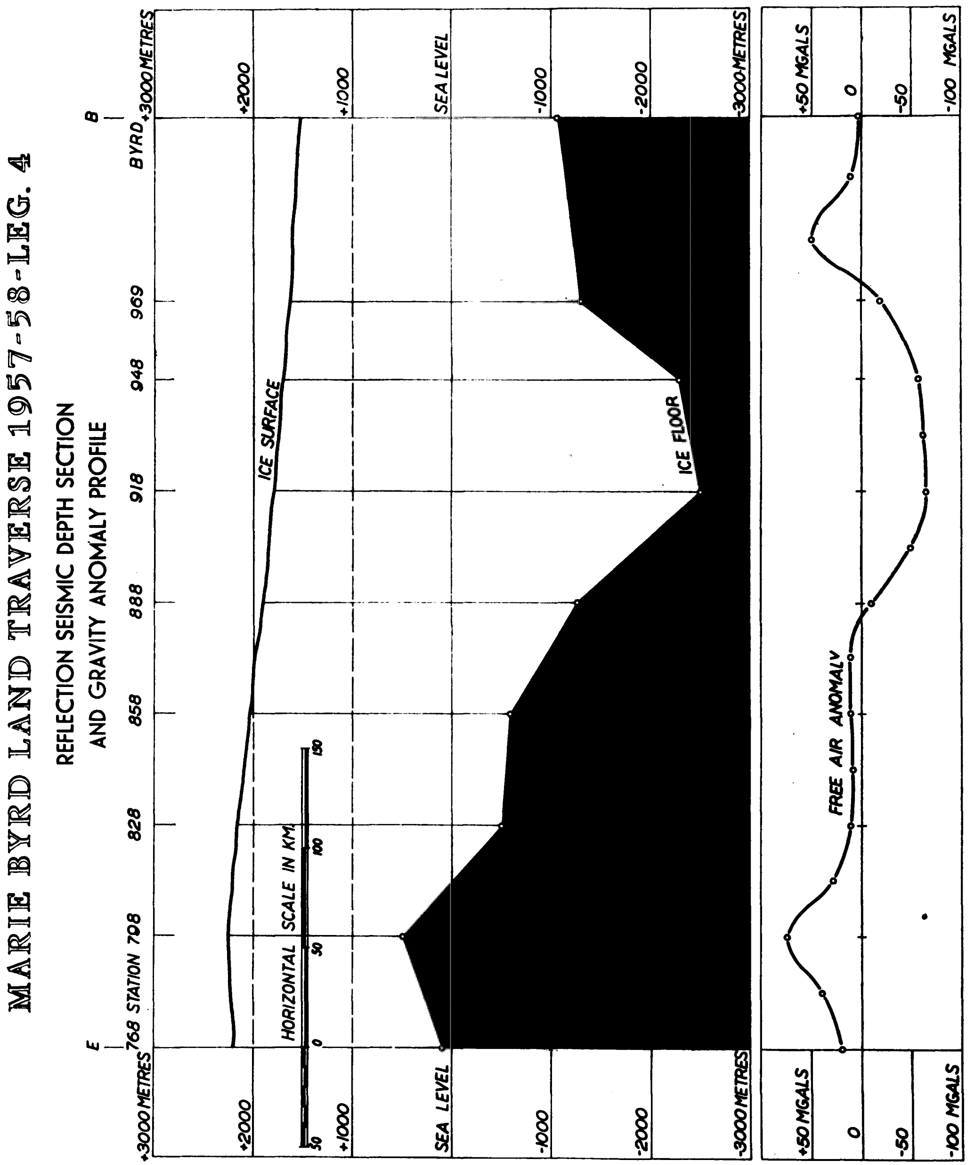


FIG. 22. U.S.A. ice-thickness and gravity observations. Traverse EB.

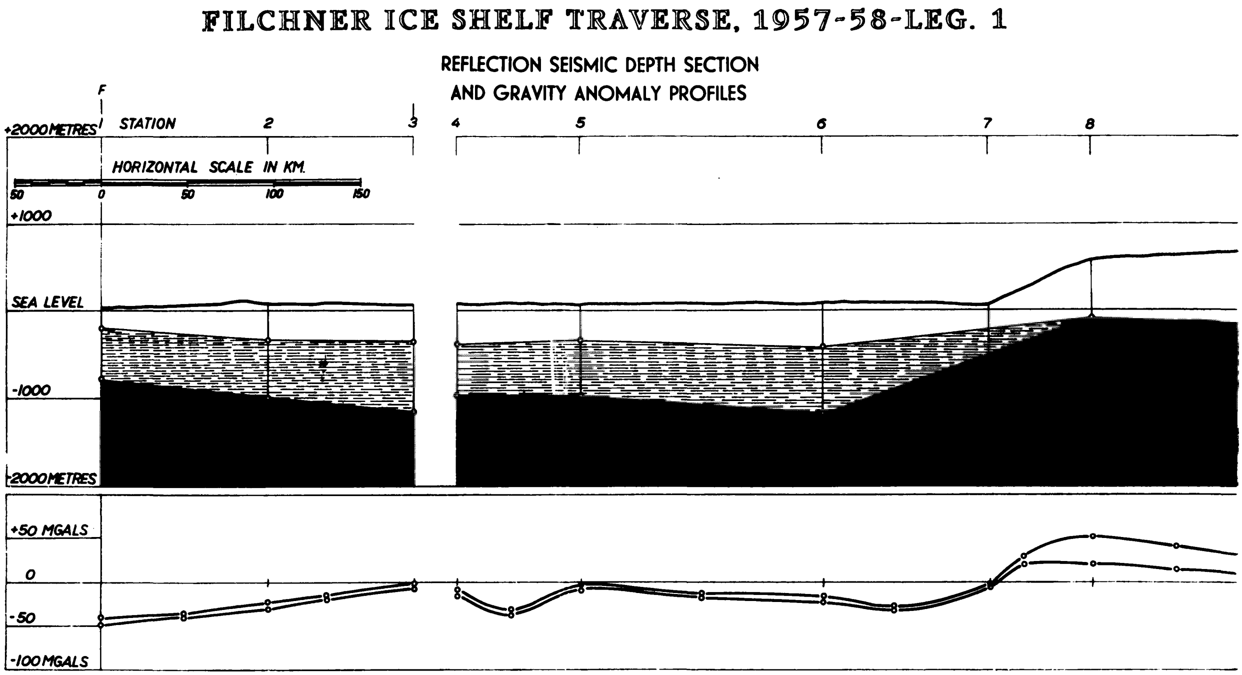


FIG. 23. U.S.A. ice-thickness and gravity observations. Traverse FGH.

in the ice are 3865 and 1945 m/sec, respectively, and the P velocity in the rock below is 4700 m/sec.

Five kilometres north of Byrd station, 2570 m of ice are underlain by 1400 m of material of P velocity 4300 m/sec (estimate), below which the P velocity is 5900 m/sec.

At 77°46′ S, 87°31′ W, 515 m of ice are underlain by 1250 m of material whose P velocity is 5200 m/sec, below which the P velocity is 6100 m/sec.

At 80°29′ S, 104°05′ W, 2630 m of ice are underlain by 1500 m of material whose P velocity is 5300 m/sec, below which the P velocity is 6300 m/sec.

THIEL *et al.*[43] find indications of S reflected waves in their studies on the Filchner Ice Shelf. From their studies of P and S velocities at the Ellsworth Snow Pit they have derived values for Young's modulus, incompressibility, rigidity and the Lamé parameters as functions of depth below the snow surface.

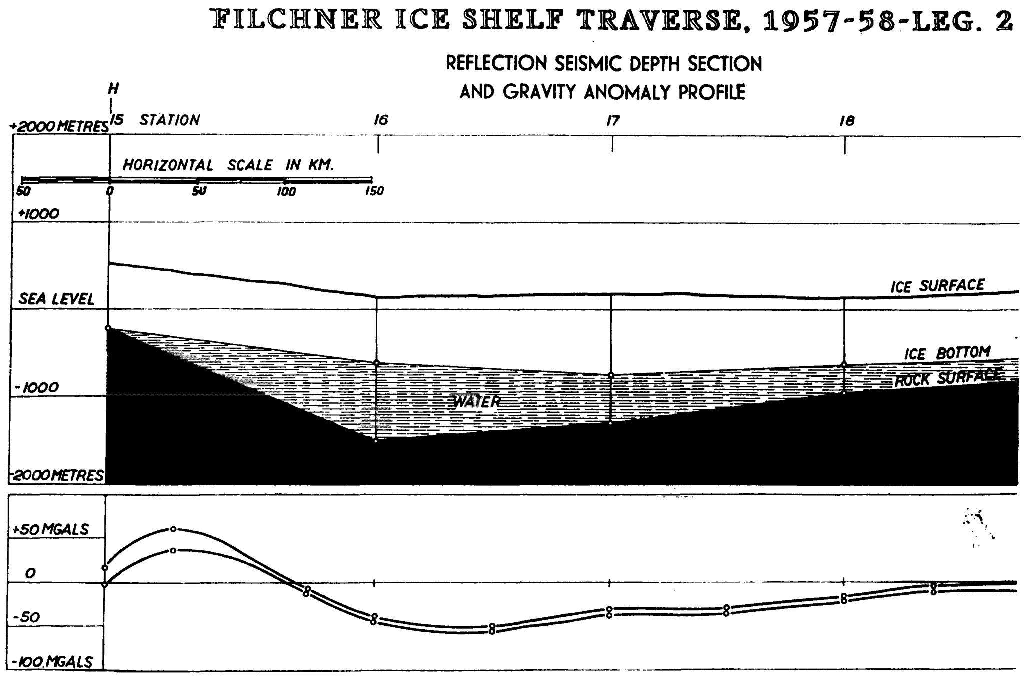


FIG. 24. U.S.A. ice-thickness and gravity observations. Traverse HJ.

D. *Personnel*

The following personnel participated in the ice-thickness field work during the periods listed. The list is further subdivided according to the bases from which the various parties operated.

February 1957

Byrd: V. H. ANDERSON, C. R. BENTLEY*, M. GIOVINETTO, A. MORENCY, N. A. OSTENSO.

1957–58

Little America V: H. F. BENNETT, W. W. BOYD, A. P. CRARY*, W. J. CROMIE, S. L. DEN HARTOG, F. LAYMAN, E. S. ROBINSON, P. SCHOECK.

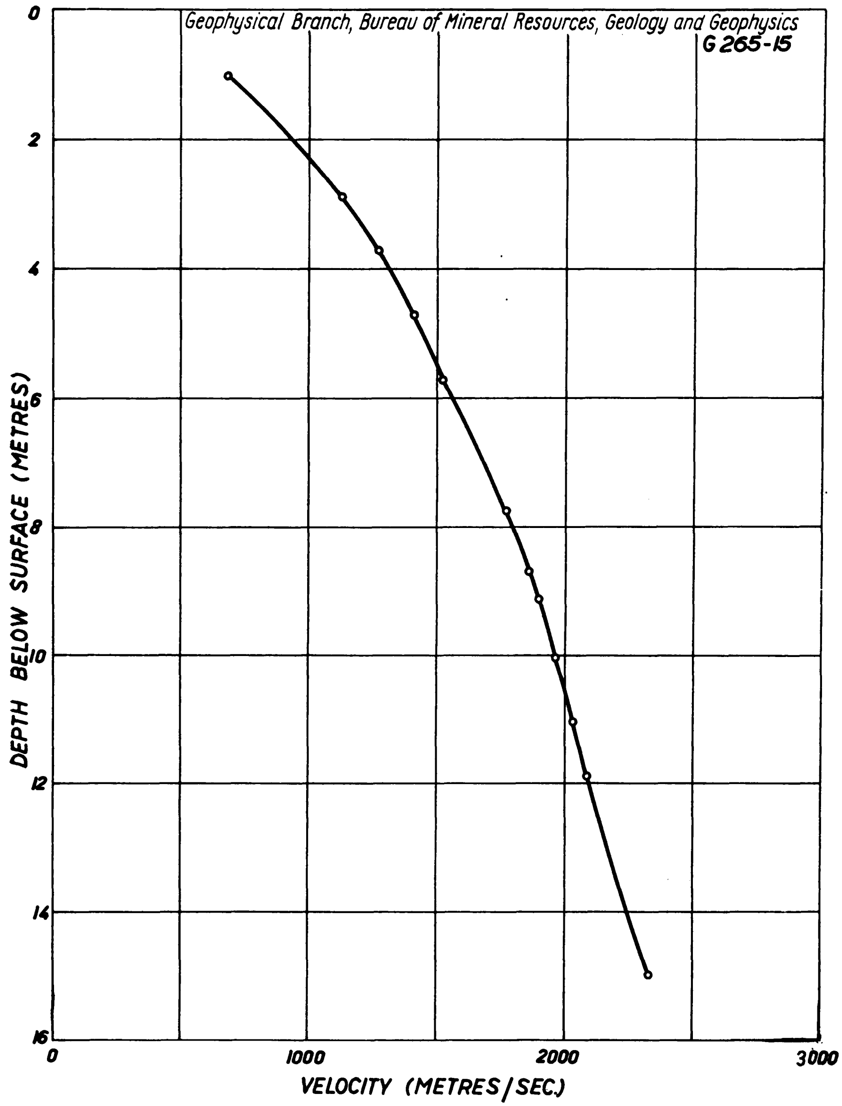
Byrd: V. H. ANDERSON, C. R. BENTLEY*, D. P. HALE, N. F. HELFERT, J. B. LONG, W. E. LONG, N. A. OSTENSO.

Ellsworth: N. B. AUGHENBAUGH, J. C. BEHRENDT, H. A. C. NEUBERG*, E. C. THIEL*, P. T. WALKER.

Airlifted traverses: J. C. COOK**, W. W. VICKERS.

1958–59

Little America V: A. P. CRARY*, S. L. DEN HARTOG, T. HATHERTON, F. LAYMAN, L. D. MCGINNIS, C. R. WILSON.



P—WAVE VELOCITY (REFRACTION METHOD)
STATION 27 (FILCHNER ICE SHELF) 80° 38'2 S. 58° 17' W.
ELEVATION 105 m.

FIG. 25. *P* velocity against depth, U.S.A. measurements.

Byrd: C. R. BENTLEY*, W. H. CHAPMAN, F. L. DARLING, L. A. LE SCHACK, J. B. LONG, W. E. LONG.

Ellsworth: E. A. BRADLEY, R. J. GOODWIN, D. HOFFMAN, J. PIRRIT*, F. T. TURCOTTE.

Airlifted traverses: H. A. C. NEUBERG, N. A. OSTENSO, E. C. THIEL*.

6.4. Evidence from seismic surface waves on Antarctic crustal structure

6.4.1. *New Zealand*

EVISON *et al.*[11,12] have studied Love and Rayleigh waves on records of eight earthquakes at Hallett Station, Scott Base and Mirny. The eight earthquakes were centred in Prince Edward Island, Bouvet Island, Drake Passage, Bolivia, the Indian Ocean, the Mid-Atlantic, and the South Pacific. In all cases, parts of the paths of the ensuing waves were across Antarctica, and allowance was made for parts of paths outside Antarctica.

The investigators infer an average crustal thickness in eastern Antarctica of about 35 km, as in typical continental regions, but only about 25 km in Marie Byrd Land. The data do not extend to the remainder of western Antarctica. It is also inferred from the Love wave dispersion that the thickness of the solid crust in the oceanic regions surrounding Antarctica ranges from 5 to 10 km.

6.4.2. *United States of America*

PRESS and DEWART[36] have carried out a similar investigation on Love and Rayleigh waves recorded at Wilkes station from eleven distant earthquakes in various azimuths.

They infer tentatively that at most only three-fourths of the Antarctic ice sheet is underlain by continent, the remaining area being oceanic in structure. Their results are stated to support the view that below-sea-level depths observed in measurements of ice-thickness are primary features and not the result of crustal sagging under an ice load.

6.5. Appendix — additional data

Additional U. S. S. R. data

A quantity of data obtained by U. S. S. R. traverses which commenced before the end of 1959 has been included by the Editor. The data relate to the traverse from Vostok to the South Pole.

Additional U. S. A. data

The following details of U. S. seismic traverses were received in October 1962, after the MS. had gone to press.

Traverse	*Route*	*Leader*
Byrd—Horlick Mountains 1958—59	Byrd—Horlick Mountains — along mountains — Byrd (1700 km)	C. Bentley
Marie Byrd Land 1959—60	Byrd—Amundsen Sea — Edsel Ford Range — Byrd (2250 km)	J. Pirrit

* Traverse party leader.
** In charge of seismology.

Traverse	*Route*	*Leader*
Victoria Land 1959—60	Scott Base — inland terminus of French traverse — terminal point (2460 km)	F. G. van der Hoeven
Airlift 1959—60	Eight stations along 88°W meridian and one at South Pole	E. C. Thiel

Additional data—other countries

Other countries have also obtained additional data after the end of the I. G. Y. (1957—58).

Reference may be had to an account by Robin for the Glaciological Volume of these Annals, written two years after this Chapter was prepared.

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TABLE 1(a). Antarctic ice-thickness measurements — Australian parties
Regional traverses from Mawson

Station	South Latitude	East Longitude	Surface Elevation metres	Rock Elevation metres	Ice-thickness metres
SP 3	67°53·0′	62°30·0′	876	+348	528
SP 4	68 06·7	62 07·0	1300	+357	943
SP 5	68 24·7	62 09·7	1638	+31	1607
SP 6	68 32·7	62 06·3	1752	+84	1668
SP 7	68 41·3	62 08·0	1796	−607	2403
SP 8	68 59·1	62 04·6	1916	+309	1607
SP 9	69 16·3	62 10·2	1980	+593	1387
SP 10	69 20·8	62 10·2	2029	+745	1284
SP 11	69 37·5	62 09·8	2170	+615	1555
SP 12	69 55·0	62 02·9	2299	+2043	256
SP 13	70 13·4	62 08·6	2352	+115	2237
SP 14	70 31·0	62 09·2	2234	−421	2655
SP 25	70 49·6	62 08·0	2281	+37	2244
SP 15	71 08·1	62 06·9	2169	−232	2401
SP 16	71 21·6	62 07·0	2100	+5	2095
SP 17	71 45·8	62 07·5	1930	−260	2190
SP 24	71 59·2	62 07·6	1978	−517	2495
SP 18	72 12·6	62 07·8	1965	−80	2045
SP 19	72 21·6	62 07·8	1963	−20	1983
SP 20	72 35·6	62 08·3	1895	+138	1757
SP 21	72 41·8	61 12·2	2031	+40	1991
SP 23	72 51·1	61 21·0	2007	+710	1297
SP 34	70 49·7	61 38·3	2320	−412	2732
SP 35	70 39·2	61 33·6	2379	+112	2267
SP 36	70 26·1	60 50·4	2532	−40	2572
SP 37	69 56·1	60 49·2	2518	+195	2323
SP 39	68 58 6	60 46·8	2007	+416	1591
SP 40	68 14·5	62 42·3	1462	+364	1098

TABLE 1(b). Antarctic ice-thickness measurements — Australian parties
Semi-detailed traverse near Mawson

Station	South Latitude	East Longitude	Surface Elevation metres	Rock Elevation metres	Ice-thickness metres
SP 26	67°44·3′	63°02·8′	538	+255	283
SP 27	67 44·2	63 02·5	529	+178	351
SP 29	67 44·2	62 57·9	501	+13	488
SP 31	67 44·3	62 46·5	447	+33	414
SP 32	67 44·4	62 37·4	454	−8	462

TABLE 2. Antarctic ice-thickness measurements — French parties
Traverse south from Dumont D'Urville

Station	South Latitude	East Longitude	Surface Elevation metres	Rock Elevation metres	Ice-thickness metres
3*	66°45′	139°40′	505	−209	714
7*	66 48	139 25	754	−233	987
10	66 50	139 15	880	−135	1015
12*	67 00	139 16	1119	+69	1050
15*	67 16	139 17	1392	+72	1320
16*	67 21	139 17	1488	−47	1535
17*	67 27	139 17	1523	+26	1497
18*	67 32	139 18	1610	−182	1792
19*	67 37	139 18	1670	+101	1569
20*	67 42	139 18	1744	+111	1633
23*	67 58	139 19	1855	−60	1915
25*	68 08	139 20	1912	−578	2490
26	68 10	139 20	1932	−401	2333
27*	68 15	139 19	1933	+227	1706
x*	68 19	139 18	1988	+452	1536
29*	68 25	139 16	2032	+365	1667
31*	68 36	139 13	2141	−175	2316
33*	68 46	139 11	2197	+52	2145
35*	68 56	139 08	2270	+400	1870
37*	69 07	139 05	2355	−399	2754
Charcot	69 22	139 01	2400	−490	2890
43*	69 37	139 03	2458	−265	2723
46*	69 52	139 05	2500	−38	2538
48*	70 03	139 06	2524	+250	2274
51*	70 18	139 08	2547	+222	2325
54*	70 33	139 10	2527	+40	2487
57	70 47	139 12	2509	−483	2992

* Positions given for these Stations are approximate only. The estimated accuracy of location is ± 1 nautical mile in latitude, ± 2 nautical miles in longitude.

TABLE 3. Antarctic ice-thicknesses measurements — Soviet Union parties 1957–59

Station	South Latitude	East Longitude	Surface Elevation metres	Ice-thickness metres	Rock Elevation metres
1	2	3	4	5	6
1 (Mirny)	66°33′	93°00′	60	170	−110
2	66°34′	93 00	120	458	−338
3	66 36	93 00	135	506	−371
4	66 37	93 00	156	458	−302
5	66 37	93 00	175	502	−327
6	66 37	93 00	190	518	−328
7	66 38	93 00	195	478	−283
8	66 38	93 00	285	390	−105
9	66 39	93 01	345	444	−99
10	66 39	93 02	391	521	−130
11	66 40	93 02	414	588	−174
12	66 41	93 03	425	628	−203
13	66 42	93 05	470	608	−138
14	66 43	93 05	535	656	−121
15	66 43	93 07	585	709	−124
16	66 46	93 09	593	801	−208
17	66 47	93 09	606	998	−392
18	66 48	93 10	645	1116	−471
19	66 50	93 10	645	1026	−381
20	66 51	93 11	689	958	−269
21	66 52	93 12	725	932	−207
22	66 53	93 13	705	862	−157
23	66 55	93 13	700	866	−166
24	66 56	93 14	790	825	−35
25	66 56	93 15	820	902	−82
26	66 56	93 15	820	915	−95
27	66 56	93 16	824	900	−76
28	66 57	93 17	880	851	+29
29	66 59	93 19	885	906	−21
30	67 01	93 21	940	784	+156
31	67 09	93 29	1047	1001	+46
32	67 13	93 35	1206	1022	+184
33	67 22	93 38	1337	1310	+27
34	67 26	93 46	1456	1466	−10
35	67 37	93 52	1570	1470	+100
36	67 42	93 58	1605	1755	−150
37	67 49	94 02	1734	1751	−17
38	67 50	94 03	1750	1706	+44
39	67 55	94 07	1803	1869	−66
40	68 02	94 15	1894	1920	−26
41	68 15	94 20	2008	2254	−246
42	68 28	94 29	2093	1300	+793

Station	South Latitude	East Longitude	Surface Elevation metres	Ice-thickness metres	Rock Elevation metres
1	2	3	4	5	6
43	68 50	94 35	2380	1590	+790
44	69 06	94 48	2460	1800	+660
45	69 21	95 05	2560	2040	+520
46	69 30	95 15	2620	1830	+790
47 (Pioneerskaya)	69°44′	95°30′	2740	2050	+690
48	70 00	95 32	2760	2300	+460
49	70 12	95 38	2790	3460	−670
50	70 21	95 41	800	3530	−730
51	70 26	95 48	2830	3160	−330
52	70 45	95 56	2840	3230	−390
53	70 51	95 51	2900	3600	−700
54	70 57	95 54	2930	4030	−1100
55	70 59	95 52	2930	3680	−750
56	71 26	96 42	3070	2890	+180
57	71 33	96 29	3080	2870	+210
58	71 37	96 33	3110	2870	+240
59 (Vostok I)	72 08	96 36	3210	2900	+310
60	72 13	96 45	3230	3340	−110
61	72 25	96 46	3280	3210	+70
62	72 35	96 49	3320	3150	+170
63	72 41	96 50	3350	3270	+80
64	72 56	97 00	3360	2930	+430
65	73 02	96 52	3380	3270	+110
66	73 07	96 53	3390	3350	+40
67	73 17	96 57	3400	3390	+10
68	73 31	97 20	3430	3200	+230
69	73 38	97 22	3440	3460	−20
70	73 47	97 24	3450	3530	−80
71 (Komsomolskaya)	74 05	97 29	3490	3360	+130
72	74 20	97 09	3520	3110	+410
73	74 33	96 44	3550	3250	+300
74	74 39	96 23	3570	3290	+280
75	74 53	96 03	3560	3240	+320
76	74 56	96 00	3580	3030	+550
77	75 19	95 16	3670	3430	+240
78	75 25	95 04	3690	3170	+520
79	75 32	95 51	3700	3050	+650
80	75 43	94 35	3710	2620	+1090
81	75 57	94 04	3740	2320	+1420
82	76 07	93 41	3790	2440	+1350

Station	South Latitude	East Longitude	Surface Elevation metres	Ice-thickness metres	Rock Elevation metres
1	2	3	4	5	6
83	76° 21′	93° 20′	3800	2460	+1340
84	76 34	93 03	3790	2970	+820
85	76 56	91 46	3770	1890	+1880
86	77 17	90 58	3710	1870	+1840
87	77 42	90 07	3670	1910	+1760
88	77 57	89 43	3680	1150	+2530
89 (Sovetskaya)	78 24	87 35	3650	1830	+1820
90	78 36	86 44	3640	1280	+2360
91	78 42	86 13	3650	1130	+2520
92	78 47	85 32	3650	1240	+2410
93	78 58	84 34	3660	1130	+2530
94	79 06	83 50	3680	910	+2770
95	79 16	82 55	3670	850	+2820
96	79 20	82 30	3740	750	+2990
97	79 31	81 21	3830	1670	+2160
98	79 38	80 37	3870	1360	+2510
99	79 42	80 17	3900	1490	+2410
100	79 48	79 38	3920	1710	+2210
101	79 56	78 53	3960	1370	+2590
102	80 02	78 14	3980	1210	+2770
103	80 08	77 27	4020	640	+3380
104	80 14	76 57	4030	1040	+2990
105	80 23	75 46	4010	790	+3220
106	80 30	74 23	3990	1110	+2880
107	80 45	72 15	3940	1180	+2760
108	80 50	70 39	3950	1850	+2100
109	80 54	69 44	3940	1540	+2400
110	81 01	68 41	3920	1870	+2050
111	81 11	67 04	3910	1930	+1980
112	81 19	65 28	3920	2160	+1760
113	81 27	64 22	3900	2260	+1640
114	81 35	62 56	3880	2290	+1590
115	81 42	62 00	3860	2680	+1180
116	81 50	60 22	3840	2320	+1520
117 (Pole of Inaccessibility)	81 55	58 29	2830	2890	+940
118	82 01	56 40	3810	2750	+1060
119	82 06	54 58	3790	2950	+840
120	74 48	98 20	3500	3450	+50
121	74 52	98 46	3510	3430	+80
122	75 56	100 36	3530	3300	+230
123	76 36	101 36	3500	3410	+90

Station	South Latitude	East Longitude	Surface Elevation metres	Ice-thickness metres	Rock Elevation metres
1	2	3	4	5	6
124	77° 03′	102° 45′	3500	3480	+20
125	77 18	103 23	3510	3220	+290
126	77 49	104 47	3500	3610	−110
127 (Vostok)	78 27	106 52	3490	3700	−210
128	78 27	106 52	3480	3700	−220
129	78 39	106 52	3500	3250	+250
130	78 54	106 52	3430	3370	+60
131	79 08	106 52	3430	3300	+130
132	79 28	106 52	3400	3350	+50
133	79 49	106 52	3400	3450	−50
134	80 13	106 52	3380	3440	−60
135	80 28	106 52	3390	3480	−90
136	80 54	106 52	3420	3450	−30
137	81 16	106 52	3440	3340	+100
138	81 58	106 52	3460	3410	+50
139	82 10	106 52	3390	3580	−190
140	82 34	106 52	3320	3750	−430
141	83 01	106 52	3300	3660	−360
142	83 24	106 52	3280	3290	−10
143	83 50	106 52	3300	3340	−40
144	84 35	106 52	3280	3130	+150
145	85 02	106 52	3310	3310	−100
146	85 36	106 52	3150	3200	−50
147	86 13	106 52	3120	3310	−190
148	86 54	106 52	3110	3380	−270
149	87 18	106 52	3100	3210	−110
150	87 44	106 52	3090	3150	−60
151	87 58	106 52	3080	2930	+150
152	88 36	106 52	3060	2950	+110
153	89 06	106 52	3020	2840	+180
154	89 44	106 52	2900	2850	+50
155 (South Pole)	90 00	—	2860	2810	+50

Seismic measurements were made: Stations 1–42 by KONDRATIEV, LOPATIN and MANILOV (1957), Stations 43–119 by SOROKHTIN and KOPTEV (1958), Stations 120–155 by KAPITSA (1959), gravity measurements were made: Station 49–126 by AVSYUK (1958), stations 128–154 by KHRUSHCHEV (1959).

TABLE 4. Antarctic ice-thickness measurements — United Kingdom party
Trans-Antarctic traverse

Station	South Latitude	Longitude	Ice-thickness metres
Southice	81°57′	28°50′ W	+540
1	82 04	28 52	+710
3	82 25	28 22	+530
5	82 51	28 53	+500
7	83 18	29 01	+490
10	83 45	29 01	+2200
13	84 09	29 00	+1580
20	84 43	29 00	+1560
23	85 14	29 06	+320
26	85 40	29 12	+800
30	86 05	29 18	+1700
34	86 30	29 30	+2820
37	86 48	29 36	+2220
43	87 25	29 36	+1790
50	87 57	29 36	+1640
56	88 31	29 36	+800
63	89 11	30 00	+1110
67	89 37	30 00	+1640
South Pole	90 00	—	+1990
75	89 39	151 00 E	+1220
89	88 02	139 42	+1190
98	86 02	139 06	+1350
101	85 23	138 54	+1100
109	84 18	139 42	+1090
115	83 27	141 18	+200
122	83 09	144 24	+460
128	82 15	146 07	+760
134	81 30	146 09	+580
139	80 48	145 28	+1030
146	79 51	148 00	+1020
153	79 12	150 54	+1060
160	78 31	153 44	+1340
164	77 56	155 58	+920
168	78 01	158 23	+380

Final altitude values are not yet available for this traverse.

TABLE 5(a). Antarctic ice-thickness measurements — United States parties
Traverse from Little America V to Byrd station

Station	South Latitude	West Longitude	Surface Elevation metres	Rock Elevation metres	Ice-thickness metres
200	79°26·1′	150°25·9′	288	−342	630
250	79 10·7	146 48·5	603	−521	1124
300	78 44·8	143 49·0	719	−741	1460
350	78 47·2	139 58·4	649	−1376	2025
400	78 45·4	136 17·0	909	−896	1805
450	78 40·5	132 35·2	1233	−982	2215
500	78 47·3	129 13·0	1404	−266	1670
520	78 58·0	128 05·0	1466	−52	1518
540	79 07·8	126 57·0	1454	−496	1950
560	79 19·4	125 34·2	1435	−750	2185
580	79 28·7	124 24·2	1423	−762	2185
600	79 38·5	123 00·0	1422	−968	2390
620	79 47·2	121 46·0	1460	−875	2335
Byrd	79 59·2	120 01·0	1515	−1055	2570

Stations are identified by their distance along the traverse in statute miles (1 statute mile equals 1.61 km).

TABLE 5(b). Antarctic ice-thickness measurements — United States parties
First traverse in Marie Byrd Land

Station	South Latitude	West Longitude	Surface Elevation metres	Rock Elevation metres	Ice-thickness metres
Byrd	79°59·2′	120°01·0′	1515	−1055	2570
24	79 42·0	118 33·0	1579	−926	2505
60	79 11·0	116 36·0	1688	−817	2505
90	78 47·0	114 59·0	1787	−330	2117
120	78 18·5	114 31·0	1670	+10	1660
150	77 54·0	113 16·0	1510	−998	2508
180	77 28·0	112 22·0	1392	−748	2140
210	76 59·5	112 47·0	1249	−1226	2475
240	76 30·2	113 01·0	1292	−908	2200
248A	76 23·0	112 55·0	1313	+397	916
270	76 44·0	111 22·0	1196	−1284	2480
300	76 58·0	109 07·0	1164	−1181	2345
330	77 07·0	107 01·0	1171	−1229	2400
360	77 14·0	104 50·0	1256	−1674	2930
390	77 21·3	102 41·0	1318	−1707	3025
420	77 27·6	100 26·0	1388	−1937	3325
450	77 31·6	98 11·0	1456	−1894	3350
480	77 34·3	95 56·0	1519	−2021	3540
510	77 36·2	93 44·0	1501	−1234	2735
540	77 41·3	91 28·0	1484	−871	2355
570	77 44·0	89 08·0	1537	+286	1251
593C	77 46·3	87 31·0	1737	+1222	515
600	77 57·0	89 15·0	1668	+323	1345
630	78 25·0	89 17·0	1870	−1245	3115
660	78 52·6	90 11·0	1968	+583	1385
690	79 19·2	91 14·0	2017	+432	1585
708	79 36·9	91 11·0	2120	+1050	1070
738	79 52·7	93 27·0	2142	+242	1900
768	80 16·6	95 19·0	2196	+96	2100
798	80 25·5	98 10·0	2236	+491	1745
828	80 29·3	101 04·0	2158	−502	2660
858	80 29·5	104 05·0	2043	−587	2630
888	80 27·2	107 01·0	1899	−1261	3160
918	80 23·2	109 51·0	1780	−2490	4270
948	80 19·2	112 41·0	1686	−2284	3970
969	80 13·8	114 40·0	1631	−1289	2920

Stations are identified by their distance along the traverse in nautical miles. (1 nautical mile equals 1·86 km.)

TABLE 5(c). Antarctic ice-thickness measurements — United States parties
Traverse from Ellsworth on the Filchner ice shelf

Station	South Latitude	West Longitude	Elevation			Water depth metres	Ice-thickness metres
			Upper ice surface metres	Lower ice surface metres	Rock surface metres		
1	77°42·6′	41°08′	41	−191	−792	601	232
2	78 05·2	39 58	69	−325	−995	670	394
3	78 26·8	39 13	72	−339	−1145	806	411
4	78 33·9	37 58	80	−377	−976	599	457
5	78 53·0	38 16	73	−344	−986	642	417
6	79 21·2	40 25	86	−405	−1188	783	491
8	79 27·7	44 26	599	−77	−77	0	676
9	79 52·0	45 06	694	−159	−159	0	853
11	80 39·7	47 09	81	−382	−927	545	463
12	80 55·6	47 59	96	−452	−1195	743	548
13	81 27·5	49 50	90	−424	−1215	791	514
15	82 28·9	52 17	523	+119	+119	0	404
16	81 58·5	54 45	129	−608	−1495	887	737
17	81 30·0	57 19	158	−744	−1287	543	902
18	81 02·1	59 51	131	−617	−948	331	748
21	79 54·3	64 27	82	−386	−550	164	468
22	79 28·8	65 30	77	−363	−519	156	440
23	78 58·7	66 38	64	−301	−512	211	365
24	78 49·2	68 37	400	−172	−172	0	572
26	80 17·0	60 28	162	−602	−602	0	764
27	80 38·2	58 17	105	−495	−941	446	600

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7. LONG PERIOD WAVES AND THE Lg PHASE

by JACK OLIVER

7.1. Introduction

OF the many topics in the field of seismology, two were selected for special study during the IGY: long period waves (LP) and the Lg phase. Inasmuch as subsequent study revealed that there is no clear-cut physical distinction between these two subjects, they will usually be discussed together in this report.

Advantage was taken of existing seismological observatories, and arrangements were made by the Lamont Geological Observatory with various scientific institutions and national committees to install long period seismographs in observatories at: Agra, India; Buenos Aires, Argentina; Hallett, Antarctica; Hong Kong; Honolulu, Hawaii; Huancayo, Peru; Lwiro, Congo (Leopoldville); Mount Tsukuba, Japan; Resolute Bay, Canada; Rio de Janeiro, Brazil; Santiago, Chile; Suva, Fiji; and Uppsala, Sweden. Lg phase instruments were supplied to the observatories at Huancayo, Peru, and Rio de Janeiro. In addition, a long period installation was made at Wilkes Station, Antarctica, under the supervision of the California Institute of Technology.

The IGY LP and Lg programs for which the Lamont Geological Observatory served as center were carried out through cooperative arrangements between the Lamont Observatory and the following institutions:

Station	*Institution*
Agra, India	Government Meteorological Department
Buenos Aires, Argentina	National Meteorological Service
Hallett, Antarctica	Department of Scientific and Industrial Research, Wellington, New Zealand
Hong Kong	Royal Observatory
Honolulu, Hawaii	Honolulu Magnetic Observatory
Huancayo, Peru	Geophysical Institute of Huancayo
Lwiro, Congo	Institut pour la recherche scientifique en Afrique centrale
Mt. Tsukuba, Japan	Earthquake Research Institute
Resolute Bay, Canada	Dominion Observatory, Department of Mines and Technical Surveys
Rio de Janeiro, Brazil	National Observatory
Santiago, Chile	University of Chile
Suva, Fiji	Geological Survey
Uppsala, Sweden	Meteorological Institute

The program was also coordinated with stations at:

Bermuda	Columbia University Geophysical Field Station
Ottawa, Canada	Dominion Observatory, Department of Mines and Technical Surveys
Perth, Australia	Perth Observatory
Pietermaritzburg, South Africa	Bernard Price Institute of Geophysical Research
Waynesburg, Pennsylvania	Waynesburg College

all of which had previously operated cooperative programs with the Lamont Geological Observatory.

The following personnel participated in the operation and maintenance of the network at their respective stations: Sa. BASU, A. N. TANDON, L. S. MATHUR (Agra); Dionisio VALENZUELA, Francisco Lucio FERNANDEZ (Buenos Aires); R. C. HAYES (Hallett); J. E. PEACOCK, P. PETERSON (Hong Kong); Gerard HARADEN, Merril L. CLEVEN (Honolulu); A. Rodriguez BEGAZO, Albert A. GIESECKE, Jr. (Huancayo); Eduard BERG, J. Cl. DE BREMAECKER, L. VAN DEN BERGHE (Lwiro); T. HAGIWARA, Y. SATO (Mt. Tsukuba); P. L. WILLMORE, John H. HODGSON (Resolute Bay); Lelio I. GAMA (Rio de Janeiro); Cinna LOMNITZ, F. GREVE (Santiago); R. E. HOUTZ, N. J. GUEST, Karl R. FLEISCHMAN (Suva); Markus BÅTH (Uppsala); Gilbert DEWART, Henry F. BIRKENHAUER (Wilkes); N. PIMENTEL, G. HAMILTON (Bermuda); P. L. WILLMORE, John H. HODGSON (Ottawa); H. S. SPIGL, B. J. HARRIS (Perth); P. G. GANE, G. L. HALES, W. L. MOUTON (Pietermaritzburg); James B. SCHROYER (Waynesburg).

The following personnel at the Lamont Geological Observatory participated in the operation and coordination of the LP and Lg networks: Maurice EWING, Jack OLIVER, George SUTTON, Ruth SIMON, Paul POMEROY. Operation of the seismological equipment at Wilkes Station, Antarctica, was conducted under the direction of Frank PRESS, Director of the Seismological Laboratory, California Institute of Technology.

This report is confined, almost entirely, to the results from these programs. Some work on these subjects was done during the International Geophysical Year based on data from other seismograph stations, and this work is only touched on briefly here. The lack of emphasis on this work, however, is not meant to imply that significant progress was not made outside of this particular International Geophysical Year program.

The term, "long period waves" as used here may refer to seismic waves with periods greater than one or a few seconds and thus may include almost all seismic waves except certain compressional wave phases of teleseisms, as well as all short period phases of near earthquakes. Frequently, the term is restricted to waves with periods greater than about 10 sec, i. e., greater than the period of the strongest microseisms, but it is not possible to retain this dividing line when studying certain types of propagation. Although the term "long period waves" does not imply an upper limit on the period, almost all such waves studied during the IGY have periods less than about 600 sec. Since the long period wave components are dominant in surface wave trains recorded at some distance from the source, most studies of long period waves are concerned with surface waves, but long period wave components appear to be present in all or nearly all body phases as well.

The term "Lg phase" was defined originally by PRESS and EWING in their discovery paper, but in the light of subsequent papers it is clear that the waves described in the original definition correspond to more than one type of propagation. Thus it is important that each study of such waves state precisely the nature of the waves in question. Generally speaking, however, the term "Lg phase" refers to surface waves with periods less than about 7 sec and group velocity of about $3\frac{1}{2}$ km/sec or less, traveling over paths which are entirely continental in nature. Although some short-period, near-earthquake phases are now referred to, on occasion, as "Lg", the International Geophysical Year Lg phase program was devoted largely to the longer-period waves recorded at some distance from the epicenter, and it is here that the Lg studies overlap with the long period wave studies.

These two topics were selected for intensive studies during the International Geophysical Year primarily because relatively recent investigations had opened the field and it was clear that much information could be obtained on the nature of seismic disturbances and on the structure of the earth's interior by an accelerated effort in these fields.

The existence of long period components in seismic disturbances has been known since near the beginning of instrumental seismology. Many early mechanical seismographs, chiefly those designed to measure the horizontal components of ground motion, had relatively long free periods and hence responded to waves of long periods, providing the source of the disturbance was strong enough to excite such waves with amplitudes sufficiently large to overcome the insensitivity of these early instruments.

These waves were identified generally, i. e., surface and body waves were properly distinguished, and two major classes of surface waves were recognized and named after their discoverers, LOVE and RAYLEIGH. Prior to about the 1940's, however, results of experimental studies of surface waves were clearly secondary in importance to those of body waves. The important contributions made during this period by the few scientists working in the long period field should not be overlooked, however, and indeed many outstanding advances, particularly in theory of surface wave propagation, were made prior to 1940 by scientists in many parts of the world. The fact remains, however, that the more important results on the nature of the interior of the earth during this interval came largely from studies of body waves, and largely from studies of travel times of body waves. This is not surprising, for it was to be expected that seismologists would concentrate their efforts initially on the exploration of the deep interior of the earth, for it is there that the field of seismology has its greatest advantage over other geophysical methods of exploration and it is there that seismic body waves are clearly the most important source of information.

Most of the advances in our knowledge of the interior of the earth based on seismological data have come from measurement of time, either as travel time of seismic waves, or wave periods, or both. This is equally true in the case of both body and surface waves. Initially the most important data on body waves were the "travel time curves", i.e., the graph of travel time of the onset of a seismic phase as a function of distance of the station from the epicenter. The early studies of surface waves were strongly influenced by the precedent established in the body wave case and were concerned primarily with the travel time of the onset of the maximum of the seismic surface wave train. When the marked dispersion, i.e., the dependence of velocity on wave period, within the surface wave train became apparent, crude quantitative measurements of the dispersion were made by plotting the group velocity of the onset, or of the maximum of the train vs. the approximate period of the wave train at the corresponding time. Attempts were made to build up curves by assembling data from a number of shocks. Thus the essential measurements were, as they remain today, ones of time rather than of amplitude of seismic waves.

Beginning in about the 1940's with the work of WILSON, and EWING and his co-workers, attempts were made to determine dispersion, i. e., group velocity as a function of period for the entire surface wave train recorded for a given path from a single shock. This experimental approach coupled with a normal mode theory previously developed by SEZAWA and KENAI, and STONELEY and others, and later by PEKERIS, EWING and PRESS, and others, ultimately resulted in the explanation of the dispersive character of a large portion of the long period seismic wave train. A large portion of the seismic wave train which previously had frequently been labeled "coda" and summarily dismissed began to yield information about the interior of the earth.

For a review of the progress in the field prior to the inception of planning for the International Geophysical Year program, the reader is referred to EWING and PRESS, *Surface Waves and Guided Waves* Encyclopedia of Physics, XLVII, Geophysics I, 1956. A brief summary follows:

For the mantle waves, i. e., those waves with periods greater than about 75 sec and with lengths so great that they derive their properties primarily from the earth's mantle and not from the crust, both the Love (G) and Rayleigh types were identified in a limited number of cases.

For waves of periods less than 75 sec in the case of oceanic paths, both Love and Rayleigh types were recognized and at least parts of the dispersion explained in terms of crust-mantle structure. In particular, the dispersive nature of the long, sinusoidal train of Rayleigh waves which makes up the large part of many seismograms, was explained as the result of the effect of the water layer on the Rayleigh wave train. Wave trains of periods less than 10 sec on all three components of ground motion were known to come from small shocks for some oceanic paths, but could not be explained unambiguously.

For waves of periods less than 75 sec in the case of continental paths, long period waves corresponding to the fundamental Love and Rayleigh modes were known, and in addition, two phases called Lg and Rg were identified. The first, a surface or near surface shear wave, the second, a Rayleigh wave. Dispersive properties of most of these waves had been measured in terms of group velocity as a function of period, but only the simplest of theoretical models were available for comparison with data, so it was not possible to derive the detailed structure of the earth. This was the situation for dispersion studies of waves of all periods.

During this period prior to the International Geophysical Year, long period instrumentation was sparse, extremely heterogeneous; in particular, rarely were three matched components operated at one station. The instruments used in both the LP and Lg programs of the International Geophysical Year are based on prototypes developed at Lamont Geological Observatory. The basic seismometer consists of a hinged mass supported by the LaCoste or zero length spring suspension. Damping and coupling to the recording galvanometer are electromagnetic. The free period of the pendulum is adjustable from several seconds to more than 30 sec. In order to obtain uniformity and assure stability over a network that included a variety of operating conditions, these pendulums, both vertical and horizontal, were operated at a free period of 15 sec during the International Geophysical Year. The vertical component pendulums were built by the Sprengnether Instrument Company, which also built the recording drums, and the horizontal component instruments were built by the Lehner and Griffith Company, which also built the long period galvanometers. The short period galvanometers were built by Leeds and Northrup.

For the long period program the seismometers were coupled to galvanometers with free periods of about 90 sec and for the Lg program, to galvanometers with a free period of about 8 sec. Response curves for these instruments may be found in SUTTON and OLIVER. Recording in all cases was photographic with paper speed of 15 mm/min. All International Geophysical Year stations operated three component seismographs and once initial operating difficulties were overcome, the instruments proved quite reliable. The percentage of good results received was well beyond expectations.

Stations were selected on the basis of geographical location, particularly in location to earthquake epicenter belts, and on the basis of interest of local personnel. Operation of the special International Geophysical Year instruments at established stations, where qualified, interested scientists were in residence, did much to insure the quality of the data, and at the same time, gave these scientists the opportunity to work with what, in almost all cases, was different and unusual seismic data. This network of 20 stations distributed relatively uniformly about the world, and operating similar three component instruments, is in many respects unsurpassed in the history of seismology. Prior to the International Geophysical Year, comparable long period instruments had been in operation in a very few places throughout the world, so the International Geophysical Year program provided a manifold increase in data of this type. In most cases, data from these instruments were analyzed in a routine manner for the standard seismic phases and for microseisms, and the data were placed on file at the appropriate centers.

Of much greater importance, however, are the research results based on these data. At this writing, many such projects have been completed, many more are in the process of preparation, and certainly many more will be carried out. The following is a brief summary of the progress in this field during the planning and operational periods of the International Geophysical Year.

Detailed study in specific cases of the surface wave trains for continental paths revealed that a large portion of the seismogram corresponded to waves of higher modes of the Love and Rayleigh types. At the same time it became apparent that the Lg phase and, incidentally, also the Rg phase (at least such phases recorded at some distance from the source), could be interpreted in terms of normal mode propagation within a relatively simple model of the crust-mantle system. Once these higher mode waves were understood, they could be used for determination of crustal structure and they provide very effective limitations on the variety of hypothetical structures which may be used to explain the surface wave dispersion in a given case.

At about the same time, measurement of Rayleigh wave phase velocities were used for determination of crustal thickness for the first time, initially in limited regions, and later for the entire United States.

The application of high speed computers to the problem of normal mode propagation in multi-layered media, both homogeneous and with velocity gradients, revolutionized the use of surface wave data for determination of earth structure. Data which previously were fit approximately, through the use of models of one or two layers, can now be fit as precisely as the data warrant, and the resolving power of the method can thus be markedly improved. Such procedures led to the confirmation, from surface wave data, of the existence of the low velocity channel within the upper mantle and to its delineation in some detail beneath the continents and oceans. Differences in the mantle beneath the continents and oceans were discovered.

Other new means of data analysis were devised. A sound spectrograph was applied to the analysis of earthquake signals, with striking results.

Nuclear explosions generated seismic waves in many cases, and these sources were located much more precisely in time and space than is possible in the case of natural earthquakes. In some cases, waves were generated which are not commonly found when the source is a natural earthquake.

Methods of analysis of the dispersed surface wave train were developed which relate the motion at the source to phase velocity and the dispersion in such a way that if two of these quantities are known, the third may be obtained. Phase velocities of mantle Rayleigh waves were determined in this way for the first time, and several studies relating the nature of surface wave trains to the source were completed, clearly indicating that this method will become, at the very least, as important as the body wave method of such studies.

Long-period waves following P at moderate epicentral distances were clearly related to the near-surface wave guide and appear to correspond to leaking modes of this guide. New instrumentation of much greater sensitivity for the recording of long period waves was developed.

Readings from the special IGY installations in the LP and Lg program are reported in the regular station bulletins of the various stations, or in special supplements. Crustal structure of a variety of areas has been explored by means of data from the International Geophysical Year network; certainly many more such papers in this field will be forthcoming. Most of the papers relating to the above are described in the following author's abstracts.

7.2. Abstracts of papers based directly on data from the IGY LP and Lg programs

The Crust and Mantle of the Earth
by M. EWING

Three major problems about the crust-mantle system are the foci of much present research; these are the origin of the crust, mechanics of crustal deformation, and possibility of continued interchange of matter between mantle and crust. Lines of investigation of these problems are outlined. At the present time it is very difficult to judge which of the new results must be accommodated into our structure of geophysical knowledge, and which must be re-examined, reinterpreted and perhaps rejected. (Abstract taken from *Geophysical Abstracts* 176, January–March 1959.)

Seismology and the IGY
by J. OLIVER

The purpose of this paper is to give a brief survey of the methods, achievements, and potentialities of the field of seismology to serve as a background for the seismological program of the International Geophysical Year. Hence the following discussion will be limited to the topics of seismology which relate especially to that program. These are (1) seismicity, essentially the geography of earthquakes; (2) earthquake mechanisms; (3) seismic-wave propagation and exploration of the Earth's interior; and (4) microseisms, the minute tremors of the Earth present at all times at all places. Other topics of less pertinence to the IGY which are not discussed include (1) macroseismology, the study of the intensity of earthquake effects near the epicenter; (2) engineering seismology, the study of problems of an engineering nature, primarily construction, in seismic areas; and (3) seismic prospecting, the application of the methods of seismology to the search for minerals.

Surface Wave Dispersion for an Asio-African and a Eurasian Path
by R. L. KOVACH

Love and Rayleigh waves in the period range of 20 to 83 sec were well recorded at Lwiro, in the Congo, from an Aleutian shock on 20 March, 1958. The epicentral distance is 14,240 km (128°) and the direct path is almost entirely continental. The path through the antipodes is almost entirely oceanic, and the corresponding Rayleigh waves were recorded at Lwiro and at Pietermaritzburg, South Africa. Dispersion data indicate an average thickness of 0·8 km for the oceanic path.

Uppsala seismograms of the Sinkiang shock of 24 June, 1958 exhibited fundamental and higher mode Love waves. Rayleigh waves affected by the near-surface sediments, and higher mode Rayleigh or M_2 waves. An average crustal thickness of about 45 km is indicated for the path which was studied. This result agrees well with available Russian refraction measurements.

A Long-Period Seismograph System
by F. PRESS, M. EWING and F. LEHNER

A matched three-component seismograph system operating with pendulum period of 30 sec and galvanometer period of 90 sec is described. As a result of increased sensitivity for periods greater than 20 sec, seismograms from these instruments contain information not previously available.

Seismic Surface-wave Dispersion: A World-Wide Survey
by M. BÅTH

By means of the long-period records at Uppsala and Kiruna of thirteen specially selected earthquakes, the Love and Rayleigh wave dispersion is determined for all oceans and all continents, except for South America. The observations have given information on crustal structure along several paths not investigated earlier; in other cases, they have confirmed earlier results. The methods are discussed.

The observations demonstrate the oceanic structure of the central Arctic area as well as the complete similarity of the Atlantic and Pacific bottom structure. Continental dispersion curves are determined for very long paths over Euro-Asia and for the hitherto longest path over North America, as well as for a path crossing Africa. These observations indicate an average crust about 10–15 km thicker along the Euro-Asiatic path than along the North American path, whereas there is perfect agreement between the North American and the African continent. A shear-wave velocity of 4·3–4·4 km/sec is obtained for the upper part of the mantle under the continents. Further conclusions must await the computation of a greater number of theoretical dispersion curves.

Ultra-long-period motions from the Alaska earthquake of 10 July, 1958
by M. BÅTH

Ultra-long-period motions from the Alaska earthquake of 10 July, 1958 have been studied on records at Uppsala. Mantle Rayleigh waves with up to six passages around the earth were recorded. The dispersion curve for the period range 100–400 sec was deduced, showing a minimum group velocity of 3·56 km/sec at a period of 225 sec, thus confirming earlier results by Ewing and Press. A value of the internal friction in the mantle of 480×10^{-5} for the period range 120–260 sec was obtained. The particle motion of the mantle Rayleigh waves is retrograde elliptical in the plane of propagation, the vertical axis being 0·68 of the horizontal axis. A transverse horizontal wave motion is observed on the E-component. Both its period (12 min 10 sec) and the delay of its onset (8 hr after the earthquake) confirm the hypothesis that it represents a free torsional vibration of the whole earth, apparently observed for the first time.

Extent of the Antarctic Continent
by F. PRESS and G. DEWART

Group velocities of earthquake-generated Love and Rayleigh waves for certain trans-antarctic paths are abnormally high when compared with data from other continents. For these paths, the data indicate that at most only three-fourths of the antarctic ice sheet is underlain by continent, the remaining area being oceanic in structure.

Thickness of the earth's crust in Antarctica
by F. F. EVISON, C. E. INGHAM, and R. H. ORR

The existence of a true Antarctic Continent is confirmed by dispersion of seismic surface waves, from which a crustal thickness of about 35 km is calculated. Waves from the earthquake of 9 September 1957, were recorded at Scott Base and at Hallett station after traversing about 2000 km of Antarctica. Records at Mirny station indicated a crustal thickness of 9 km beneath the Indian Ocean. Preliminary analysis of records of another Indian Ocean shock on 4 August 1957, also indicates a thickness of 35 km. (Abstract taken from *Geophysical Abstracts* 176, January–March 1959.)

Seismographs of High Magnification at Long Periods
by G. SUTTON and J. OLIVER

Several types of high magnification seismographs with peak response at periods between about 7 and 80 sec are operated at the Palisades station of the Lamont Geological Observatory. Comparison of the seismograms from these instruments, and of the appropriate theoretical and experimental frequency-response characteristics, illustrates their value for the detection and resolution of the long-period components of both body and surface waves. New vertical-component seismographs detect surface waves from small shocks at very great distances and, apparently for the first time, record microseisms in the 10–30 second period range almost continuously.

Two other types of seismographs are operated throughout the world in a cooperative program with other institutions during the International Geophysical Year.

7.3. Abstracts of papers in preparation based directly on data from the IGY LP and Lg programs

The effect of oceanic sediments on surface wave propagation
by J. DORMAN, J. OLIVER and M. EWING

When the path between the epicenter and the station traverses deep oceanic basins primarily, the surface-wave trains from shallow earthquakes recorded by ordinary long-period seismographs may be separated into two classes on the basis of wave periods. The first class consists of Love and Rayleigh waves with periods greater than about 15 sec. The second class includes waves with periods generally between about 5 and 15 sec, but occasionally as great as 20 sec or more.

Numerical computations of surface-wave dispersion for theoretical models of structures which take account of the small but significant rigidity of the sedimentary layer can account for some previously unexplained features of these wave trains. For example, the low rigidity in the sedimentary layer affects the dispersion curves in such a way as to account for the long duration of the short-period Love-wave train. It does not appear likely at present that the short period train on the vertical and longitudinal components can be explained as shear-mode propagation, however. With such models, the T-phase can be explained by propagation in higher shear modes, although not by propagation in the fundamental, or Rayleigh, mode.

Leaking modes and the PL phase

by J. OLIVER and M. EWING

The *PL*-phase is a normally-dispersed train of waves of periods greater than about ten seconds, beginning at or near the time of the initial *P*-wave and sometimes continuing at least to the time of the beginning of the Rayleigh-wave train. With adequate instrumentation, the *PL*-phase is commonly observed at distances less than about 25° from shallow shocks. In general, surface particle motion is elliptical and progressive, and amplitudes are not greater than about one-quarter those of Rayleigh waves of the same period. Comparison of *PL*- and Rayleigh-wave dispersion shows that both waves propagate in roughly the same near-surface wave guide. Whereas Rayleigh waves correspond to normal- (non-leaking-) mode propagation, *PL*-waves appear to correspond to leaking-mode propagation within this wave guide.

Radiation Pattern of Rayleigh Waves from the S. E. Alaska Earthquake of 10 July 1958

by J. N. BRUNE, J. E. NAFE and J. E. OLIVER

The network of long period seismographs operated during the International Geophysical Year made it possible to study the radiation pattern of Rayleigh waves from the South East Alaska earthquake of 10 July 1958. A simple method for determining initial phases of surface waves, developed recently at Lamont Geological Observatory was used to determine the azimuthal pattern of initial phases. When this pattern is compared with observed fault motion and the fault plane solution from body waves, it appears that the surface wave pattern for this earthquake bears a simple relation to the fault motion and can be understood in terms of the elastic rebound theory of faulting. The azimuthal pattern of amplitudes was also determined. The use of surface waves for determination of fault motion should prove to be especially valuable in the case of small earthquakes, for the radiation pattern may be found using only stations near the epicenter. The method may be combined with the fault plane solution of body waves to give a more powerful method for the study of faulting mechanisms.

A simplified method for the analysis and synthesis of dispersed wave trains

by J. N. BRUNE, J. E. NAFE and J. E. OLIVER

A disturbance at one point of a dispersive medium resulting from an impulse applied at another point may be represented as a superposition of travelling plane waves. The phase and period of the disturbance at any instant are related by the principle of stationary phase to the phase and period of a travelling wave component. For the instantaneous phase of that travelling wave component the following equation may be written:

$$Ct\text{-}x = (N - 2\frac{\varphi_0}{2\pi}CT)$$

where C is the phase velocity, x the distance, T the period, t the travel time, N an intege and φ_0 the initial phase of the travelling wave component. Since T and t may be measured from a record of the disturbance and x may be determined, the equation may be used to compute the phase velocity as a function of period if the initial phases are known. If distance and the dispersion are known, initial phases may be determined. From distance, initial phases and phase velocities, the disturbance at any point may be constructed. The practical use of the method is demonstrated by application to antisymmetric waves in a cylindrical rod, Rayleigh waves from U.S. and Russian nuclear explosions, Rayleigh waves from the Hudson Bay earthquake of 30 January 1959, and Love waves from the Fairview Peak and Fallon, Nevada earthquakes of 1954.

Oceanic Rayleigh-wave dispersion in the period range 15–140 sec: the group velocity maximum and associated Airy phase
by G. H. SUTTON, M. MAJOR and M. EWING

Rayleigh waves recorded at Suva, Fiji, on Columbia long-period seismographs from four circum-Pacific belt earthquakes exhibit periods greater than the period of the group-velocity maximum, U_0, near 35 seconds. Resulting dispersion curves between 15 and 140 sec period have U_0 equal to 4·02, 4·02, 3·98 and 3·98 km/sec et periods T_0, to 42, 42, 34, and 32½ sec for waves from Andreanof Islands, Fox Islands, Gulf of California, and southern Peru shocks, respectively. Energy from the Airy phase associated with U_0 is evident on the seismograms more than 1 min preceding the arrival time of U_0. The observed Airy phase agrees well with the theoretically predicted disturbance before and for some time after the arrival time of U_0. The first few minutes of the Rayleigh wave trains were reconstructed using the Airy-phase contribution for times earlier than the time of U_0 and constant amplitude trains whose frequencies were determined from the observed dispersion curves for later times. The fit is good although neither amplitude variations nor instrument-phase shift were considered. Because of the Airy phase, the apparent velocity of the beginning of the Rayleigh-wave train depends upon epicentral distance and seismogram-trace amplitudes. Errors of as much as 0·14 km/sec would arise from the imprope interpretation of the apparently impulsive beginnings of the observed wave trains as the arrival time of U_0. The southern Peru shock (U. S. C. G. S. focal depth 100 km) was notable for lack of energy at periods less than T_0.

7.4. Abstracts of certain papers based indirectly on the IGY LP and Lg programs

Crustal Structure in the California-Nevada Region, by F. PRESS

A case is made for the combined use of three methods for the study of the earth's crust in a given region: seismic refractions, surface wave phase velocity, and gravity. Only by this approach can the fine details of crustal structure be revealed. The standard phase velocity curves are revised to take into account recent refraction results in South Africa and the Gutenberg low-velocity zone of the upper mantle. Restrictions on the use of the phase velocity method alone are discussed.

In the California-Nevada region seismic refractions reveal the following structure below the sediments: 23 km of granitic rock with velocities of 6·11 km/sec and 3·49 km/sec for compressional and shear waves: 26 km of gabbroic-ultramafic rock with compressional velocity 7·66 km/sec underlain by a zone of ultramafic rock with compression velocity 8·11 km/sec. When this structure is used to compute theoretical Rayleigh wave phase velocities and gravity anomaly, discrepancies are found with the observed values which can be resolved by reducing the mean shear velocity and density in the crust. It is probable that this reduction is limited to the intermediate crustal layer, and several modifications consistent with all three exploration methods are discussed.

Study of Earthquake Mechanism by a Method of Phase Equalization Applied to Rayleigh and Love Waves
by K. AKI

Rayleigh waves and Love waves are used for the study of the earthquake mechanism by the use of a method of phase equalization. In this method, an impulse response is computed from known phase velocity data and instrument characteristics, and is cross-correlated with an actual record. A comparative study of Love waves from Kern County aftershocks of 1952 and those from Nevada shocks of 1954, strongly supports the hypothesis of a pair of couples for the earthquake source rather than a single couple. Source functions for five Kern County aftershocks are derived from the Rayleigh waves recorded at Weston and Palisades. It was found that the sense of principal motion in the source function is in agreement with the fault plane solution obtained from the *P*-wave data. Mantle Rayleigh waves are found to be useful for this purpose also.

The Use of Love Waves for the Study of Earthquake Mechanism
by K. AKI

Long period Love waves of continental path were successfully used for the study of earthquake mechanism. The method used is simply the comparison between wave forms recorded by the same instrument at the same station due to different earthquakes of similar size which occurred within a limited area. The aftershocks of Kern County earthquake of 1952 and the series of Nevada shocks during 1954 were studied and it was found that the direction of lateral fault motion derived from Love waves perfectly agrees with that derived from the bodily wave data. We also found a very definite fact, which cannot be explained by postulating a single couple for an earthquake source, strongly supporting Honda's hypothesis of a pair of couples. This additional information from the Love wave data confirmed that the fault plane solutions have great geotectonic significances at least for the area studied.

Seismic surface waves at Palisades from explosions in Nevada and the Marshall Islands
by J. OLIVER and M. EWING

Surface waves were detected at Palisades, New York, for nuclear explosions on or above the surface in the Marshall Islands (at a distance of 105°) and in Nevada (at a distance of 33°). Signals from both sites consist entirely of Rayleigh wave trains. The dispersive pattern may be explained by using dispersion curves developed in studies of earthquake-generated surface waves. The small underground nuclear explosion was not detected at Palisades. (Abstract from *Geophysical Abstracts* 175, October–December 1958.)

Study of Shear-velocity distribution in the upper mantle by mantle Rayleigh waves
by J. DORMAN, M. EWING and J. OLIVER

Comparison of Rayleigh-wave dispersion computations on 11 models of shear-velocity structure of the continental and oceanic crust-mantle systems permits a detailed explanation of observed mantle Rayleigh-wave dispersion for periods less than 250 sec. Velocity distributions for the continental crust-mantle obtained by Gutenberg and by Lehmann from body-wave data, both of which include a region of low velocity in the upper mantle, are consistent with the dispersion data. Furthermore, it is shown that a structure which agrees with results from travel times of body waves from explosions and from distant earthquakes and with Rayleigh-wave dispersion must necessarily contain a low-velocity region of this type.

Shear-wave data for oceanic areas are meager, but from Rayleigh-wave dispersion there is firm evidence that the shear velocity immediately below the *M* discontinuity of deep ocean basins is about 4·6 to 4·7 km/sec, and the same as under the continental *M* discontinuity. It is also clear that the velocity distribution below the depth of the continental *M* discontinuity cannot be the same under continents and oceans. Instead, an oceanic model obtained by successive approximation to oceanic Rayleigh-wave dispersion data shows that the region of low shear velocity extends to much shallower depths under the oceans, thus being a much more prominent feature under oceans than under continents. This is similar to results obtained by LANDISMAN *et al.* using oceanic Love-wave dispersion, and provides a more detailed knowledge of the structure.

The use of galvanometers as band-rejection filters in electromagnetic seismographs
by P. W. POMEROY and G. H. SUTTON

The large-amplitude intermediate-period, 4-9 sec, microseisms are effectively removed from high-magnification long-period seismograms with maximum magnification near 50 sec period. This has been accomplished by the use of a galvanometer of natural period in the microseism range as a band-rejection filter. Addition of the filter galvanometer greatly increases the usefulness of these instruments for studies of long-period microseisms and for clear detection and resolution of the long-period components of both body and surface waves from small shocks at very great distances. Magnification equations and theoretical magnification curves show that extremely high sensitivities for desired periods and low sensitivities for undesired periods can be obtained by combining one or more filter galvanometers in seismograph systems, varying the damping constants and natural periods of the components, and varying the coupling between the components. Also, by proper choice of instrument parameters, undesired periods can be rejected without appreciable modification of the seismograph magnification for periods outside the rejection band.

The seismic noise of the earth's surface
by J. N. BRUNE and J. OLIVER

Maximum, average, and minimum values of surface particle displacement, velocity and acceleration of earth noise as a function of period are illustrated in graphical form. For periods less than about 5 sec the amplitude curves rise rapidly with increasing period. The most prominent feature of the illustration is the sharp peak in the 5- to 8-second period range. There are virtually no data on noise in the range of periods between 20 sec and the earth tide periods. With the exception of the 10- to 40-second period range, the data used are taken from the existing literature.

The second shear mode of continental Rayleigh waves
by J. OLIVER, J. DORMAN and G. SUTTON

Waves of the Rayleigh type corresponding to the fundamental and first shear modes for the continental crust-mantle system have been identified on seismograms previously. In this paper, waves corresponding to the second shear mode are identified for two paths, one from the Belgian Congo to Pietermaritzburg, South Africa; the other from Oklahoma to Palisades, New York. Comparison of the dispersion of these waves with theoretical dispersion for several crust-mantle models demonstrates the increase in resolving power of this method of obtaining crustal structure when data for several modes are available. There are small but measurable differences in the velocity structures averaged along these two paths.

The distortion of pulse-like earthquake signals by seismographs
by M. LANDISMAN, Y. SATO and M. EWING

A numerical method for calculating the seismogram of any seismograph for an arbitrary continuous ground disturbance is used to investigate the effect of instruments with various physical constants on both theoretical and actual impulsive signals. This calculation solves the equations of motion of the pendulum-galvanometer system by a fourth-order Runge-Kutta process. Variation of instrumental constants produces conspicuously different seismograms, in several cases of practical interest.

In order to resolve the apparent discrepancy between the observations of SATO, and those of BENIOFF and PRESS for the group velocity of long period Love waves, certain computed seismograms of the long period Love wave arrival, *Gl*, from the New Guinea shock of 1 February 1958 were compared with the linear strain recording of this signal. The results of this study confirm the observations of SATO, that the group velocity of Love waves is essentially flat in the long period range to several hundred seconds. The distortion introduced by instruments whose magnification changes rapidly with period is great enough to mask any small dispersion that might be associated with these pulse-like signals.

Numerical solutions for Love wave dispersion on a half-space with double surface layer
by J. DORMAN

The IBM 650 computer of the Watson Scientific Computing Laboratory, Columbia University, was programmed to obtain numerical solutions for the period equation for Love waves on a half-space with a double surface layer. Solutions including higher modes for seven models of the continental crust-mantle system are presented. This group of related cases shows that certain properties of the solutions are diagnostic of crustal structure. These relationships are illustrated graphically.

Note on the variational and homogeneous layer approximations for the computation of Rayleigh-wave dispersion
by F. PRESS and H. TAKEUCHI

An earlier study of upper mantle structure using a variational method is repeated using the homogeneous layer approximation programmed for an electronic digital computer to obtain dispersion curves. The dispersion curves computed by the two methods differ significantly but systematically so as to yield the same conclusion about the presence of the Gutenberg low-velocity zone in the upper mantle.

Determination of crustal structure from phase velocity of Rayleigh-waves Part III: The United States
by M. EWING and F. PRESS

Variations in phase velocity of Rayleigh waves from the Samoa earthquake of 14 April 1957 are reported for the United States. These variations are correlated with topography and Bouguer gravity anomaly on a continental scale, demonstrating regional isostatic compensation. The correlation of phase-velocity variations with crustal-thickness changes is justified, and permits specification of the mechanism of compensation as the regional Airy system.

Regional average crustal thicknesses are: Peninsular Ranges and Southwestern Desert, 40 km; Basin and Range Province, 48 km; Rocky Mountains, 47 km; Interior Plains, 35-41 km; Appalachian Mountains, 40 km.

Numerical integration of the equation of motion for surface waves in a medium with arbitrary variation of material constants
by Y. SATO

The calculation of surface-wave dispersion is difficult when the waves propagate in media whose physical properties change with depth, and only a few solutions are available for fairly simple cases.

These computations may now be performed with the aid of highspeed computers, even for media whose material constants change arbitrarily with depth. The dispersion of both Love waves and Rayleigh waves has been obtained for such cases by the numerical specification of surface displacement followed by numerical solution of the equation of motion.

For example, with respect to the problem of Love waves, besides the ordinary boundary condition that the stress vanishes at the free surface, an extra condition is stated which requires that the displacement amplitude be unity at the surface. The equation of motion is then solved numerically for tentative values of frequency and wave number, and this solution produces the distribution of displacement amplitude in the half space. For all combinations of frequency and wave number which are not solutions, the values of computed displacement do not converge and tend to become positively or negatively infinite for increasing depth below the free surface. To obtain a solution, one of the parameters—for instance, wave number—is fixed, and frequency is varied in small steps until the computed displacement converges to zero at great depths. This combination of parameters fulfills all the standard boundary conditions and is the required solution. The problem of sound waves in an elastic liquid can also be solved with only a minor change in the physical properties.

The dispersion of Rayleigh waves propagating in a heterogeneous substance can also be obtained by a similar method. In this case, another parameter is needed, namely, the ratio of the amplitude of horizontal and vertical components of displacement at the free surface. Denote this quantity by a, and the phase velocity by c. Wave number is fixed, and a two-dimensional search in the a-c plane is used to locate the point that produces a convergent solution. For limited media a solution is required which satisfies the boundary conditions at the other surface.

After testing the method by applying it to cases which had been solved analytically, a few problems were solved. These include:

(1) Love waves in a medium with constant density and linearly increasing rigidity.
(2) Sound waves in a medium whose density and velocity are given by experimental curves.
(3) Rayleigh waves in a medium having constant density and equal rates of increase for λ and μ. Using an IBM "650" it takes a few minutes to get a point for cases 1 and 2, and from thirty minutes to two hours per point for Rayleigh-wave dispersion.

Normal modes of continental surface waves

by J. OLIVER and M. EWING

When the path between epicenter and station traverses only continental structure, the dispersion of the entire train of directly arriving seismic surface waves can be explained as the result of normal mode propagation in a crust-mantle system in which the velocity increases in some manner with depth within the crust. At least four modes, the Rayleigh mode, Sezawa's M_2 mode, and the first two Love waves, may appear prominently on the seismogram. The characteristics of the higher-mode dispersion curves permit the explanation of the L_g phase of Press and Ewing, Bath's L_{g1} and L_{g2}, and, in some cases, Caloi's S_a without recourse to a low-velocity layer in the crust or mantle. Speculation on changes in these curves for less simplified models indicates that the remaining cases of S_a as well as Leet's C or coupled wave may be explained by classical theory.

The occurrence of the higher-mode waves is widespread; they are found on the four continents for which data are available.

Higher-mode data, particularly when combined with information from the fundamental modes, make surface-wave dispersion, previously a useful tool, a much more potent method for the study of crustal structure.

The effect of surficial sedimentary layers on continental surface waves

by J. OLIVER and M. EWING

Surface waves in the $\frac{1}{2}$-second to 12-second period range, recorded at several stations in eastern North America from the eastern Tennessee shock of 23 June 1957, are the bases for several deductions concerning the effect of sedimentary layers on continental surface wave propagation. These are: (1) the velocities of surface waves of the fundamental Love and Rayleigh modes having periods less than about 10 sec may be strongly affected by sedimentary layers of average thickness. The decrease in velocity accounts, at least in part, for the prolongation of surface-wave trains in this period range when sedimentary layers of appreciable thickness have been traversed. (2) Higher-mode propagation for both types of surface waves is a possible explanation for the velocities, frequencies, and amplitudes of the phase S_g at moderate epicentral distances, and of its long-distance counterpart the high-frequency component of L_g. The lower-frequency components of L_g have been explained previously by other aspects of normal-mode propagation in the crust. (3) Study of dispersion of short-period surface waves can result in fairly detailed knowledge of velocity-depth relation within the sedimentary column and may also reveal information on anisotropy. (4) The results of this study must bear heavily on studies of microseism propagation. As an example, the increase of microseismic activity along the entire east coast of the United States when a storm moves onto the continental shelf may be attributed to channeling of the waves in the deep sedimentary trough beneath the shelf.

Attenuation, dispersion, and the wave guide of the G wave
by Y. SATÔ

Using the strain seismograms of the New Guinea earthquake of 1938 and the Kamchatka earthquake of 1952, the decrement of the G wave in the mantle of the earth was determined from the comparison of the amplitude of Fourier components, which are obtained by analyzing the G phases at different epicentral distances. The value of $1/Q$ thus obtained is a little larger than that given by M. EWING and F. PRESS using mantle Rayleigh waves, but is not much different. The phase velocity was also calculated using the argument of the Fourier transform. The dispersion curves obtained from (G_1 and G_3), (G_2 and G_4) of the New Guinea earthquake and (G_1 and G_3) of the Kamchatka earthquake agree quite well, giving a nearly constant group velocity of 4·4 km/sec as was anticipated. Theoretical consideration of the distribution of shear velocity that serves as the wave channel for the guidance of the G wave was given, and the shear velocity was calculated applying the method of T. TAKAHASHI to the dispersion curve derived from the condition of constant group velocity, which is a direct consequence of the fact that the G wave shows almost no dispersion. The $V_s(z)/V_0$ curve which was derived theoretically agrees well with the curve given by the distribution of shear velocity of Jeffreys-Bullen in the range between one and several hundred kilometers.

Higher modes of continental Rayleigh waves
by J. OLIVER and M. EWING

A long dispersive train of waves corresponding to higher modes of the Rayleigh-wave equation (including Sezawa's M_2 wave) for the continental crust-mantle system is positively identified, apparently for the first time. Observed particle motion is elliptical and retrograde, in agreement with theory. Although several theoretical studies have been published in which progressive elliptical particle motion was found, all of these involved values of the elastic constants unsuitable for the present problem.

The beginnings of the short-period branches of the higher modes can account for the high-frequency longitudinal and vertical components of the continental surface-wave phase Lg. The large amplitudes and the peculiar appearance of Rg appear to depend on the broad flat minimum of the group velocity curve of the lowest or Rayleigh mode.

Transient Analysis of Earthquake and Explosion Arrivals
by M. EWING, S. MUELLER, M. LANDISMAN and Y. SATÔ

An electronic sound spectrograph has been used to analyze the transients in complicated earthquake and explosion signals. Transient analysis with a sound spectrograph gives a means for directly obtaining group velocity at each of the spectral frequencies for each of the separate arrival branches, for any type of transient signal. Compared to frequency analysis, it presents the fine spectral structure of the signal as it changes with time, not an average over a time that includes many parts of the seismic signal.

Dispersion as well as the body wave spectrum of the P-wave has been observed by transient analysis of seismic signals. Dispersion has also possibly been observed in the S-wave, and in the various multiply reflected S-waves, which have been found to increase in period from one arrival to the next. The dispersed pattern of arrival of fundamental and higher mode surface waves has been observed for oceanic, continental, and mixed paths. Among these signals is a clear indication of the continental second shear mode. The separation of surface waves for the direct and complementary paths has also been accomplished. Our results compare well with those obtained by standard techniques.

The dispersion of the fundamental and higher mode signals from explosive sources in shallow water may be easily studied by making sound spectrograms and amplitude sections of these seismic signals. These shallow water shots show bubble pulses and Airy phases, which are clearly defined on the spectrograms and sections.

Long-Period seismographs
by H. BENIOFF

Descriptions and theories of a number of different seismographs developed particularly for recording of very long-period seismic waves are presented. These include (1) electromagnetic strain seismograph with galvanometer of 8 min period and photographic recording; (2) displacement transducer strain seismometer with resistance-capacitance network and short-period galvanometer photographic recorder or with ink-writing recorder; (3) electromagnetic pendulum seismometer with RC network having transfer characteristic of a long-period galvanometer recorder or a heated shunt capacitance; and (5) electromagnetic pendulum with condenser-lengthened period and triple RC integrating network recording with either heated stylus visible writer, ink writer, or short-period galvanometer photographic recorder.

Progress report on Long Period Seismographs
by H. BENIOFF and F. PRESS

Long period seismograph systems in operation in Pasadena are described. Extension of the group velocity curves for mantle Rayleigh waves and G-waves, the detection of these waves from earthquakes in the magnitude range $5\frac{3}{4}$–7, and the recording of unusual body waves with unsuspected long period components are among the results which have been achieved.

Rayleigh-wave evidence for the low-velocity zone in the mantle
by H. TAKEUCHI, F. PRESS and N. KOBAYASHI

Variational calculus methods are applied to the problem of dispersion of mantle Rayleigh-waves. In the present paper we have worked two models. One is Gutenberg's model with a low-velocity layer around 150 km depth. The other is a Jeffreys–Bullen model modified above 200 km depth so as to join smoothly to the explosion-determined velocities just under the Mohorovicic discontinuity. No low-velocity layer is assumed in this model. Both models give almost identical theoretical dispersion curves which agree well with the Ewing–Press observations of mantle Rayleigh waves for periods longer than 250 sec. This result means that the minimum group velocity at about 250 sec is mainly due to a sharp increase of shear velocity at about 400 km depth, which is a common feature for the two models. For periods shorter than 250 sec. Gutenberg's model gives results concordant with the observations. The modified Jeffreys–Bullen model disagrees significantly with the observations. This demonstrates the existence of a low-velocity layer in the upper mantle.

Some implications on mantle and crustal structure from G waves and Love waves
by F. PRESS

G-wave velocities for continental and oceanic paths do not differ by more than about 2 per cent. Since the *G*-wave velocity is controlled by the low velocity zone in the mantle, this zone is present beneath continents and oceans. This suggests that the composition and distribution of temperature are the same for depths greater than about 50 km under continents and oceans. The low velocity zone may be the source of the primary basaltic magma and could account for the long-period nature of *S* waves. It may also represent a zone of decoupling for relative movements between crust and mantle.

Love waves with long propagation paths recorded with long period seismographs are used to infer that the mean value of continental crustal thickness lies in the range of 32 to 37 km.

Seismic Waves from High Altitude Nuclear Explosions
by P. W. POMEROY

Seismic waves of long-period were well recorded at epicentral distances up to 9300 kilometers from the high altitude nuclear explosions, TEAK and ORANGE, which were fired in the Johnston Island area on 1 and 12 August 1958 respectively. At Honolulu, at a distance of 1300 kms, the recorded seismic waves may be divided into three types; (1) a normally-dispersed oceanic Rayleigh-wave train in which the wave periods decrease from about 35 to 14 sec as the corresponding velocities decrease from about 4·1 to 1·6, km/sec, (2) an inversely-dispersed oceanic Rayleigh wave train in which the wave periods increase from about 6 to 10 sec as the corresponding velocities decrease from about 1·3 to 1·0 km/sec, (3) a *T*-phase consisting of waves with periods less than about 0·5 seconds and corresponding to a velocity of 1·47 km/sec.

From the normally dispersed train, phase and group velocities of waves in the 35 to 14 second period range were computed for the paths which both traverse primarily a typical deep oceanic basin. The inversely dispersed train is highly unusual in that, although predicted by classical theory, these waves are not recorded when the source is a natural earthquake.

Surface waves of long periods were recorded at Palisades from both high altitude nuclear explosions and these waves have amplitudes comparable to those generated by the larger near-surface explosions in the Marshall Islands. A special instrument at Palisades, not operated for TEAK, indicates a similar relationship for body waves of long periods generated by ORANGE and by the Marshall Island shots. In contrast, seismic body waves of short periods are apparently generated much more efficiently by near-surface explosions than by high altitude explosions.

7.5. Abstracts of certain papers in preparation based indirectly on the IGY LP and Lg programs

Period equation and particle motion for the P/SV normal modes of a flat layered liquid-solid half space
by J. DORMAN

A period equation, that is, a relation between wave length and phase velocity, for plane surface waves travelling parallel to the interfaces of an elastic half space of homogeneous solid and liquid layers is obtained by means of a matrix formulation. The procedure is an extension of Haskell's derivation of the period equation for an *n*-layered solid half space. The present theory places no restrictions on the number of layers nor on the number of liquid-solid interfaces which may be considered. The generality of this structure makes it possible to approximate any hypothetical flat layered structure for purposes of guided wave dispersion computation. The form of the equation lends itself to rapid solution on an automatic digital computer. Methods for computing the solutions are discussed.

A simple method of calculating the particle motion and stress distributions as a function of depth is outlined in terms of the equations derived.

The same methods can also be applied to find the period equation and the particle motion and stress distribution in the region of an arbitrarily layered transition zone between two semi-infinite half spaces.

Particle amplitude profiles for Rayleigh waves on a heterogeneous earth
by J. DORMAN and D. PRENTISS

The relationship between vertical and horizontal particle amplitude and depth for Rayleigh waves was obtained for several models of a heterogeneous, solid half-space using a new computing program for the IBM 650. The data show how the well-known characteristics of Rayleigh wave motion on a homogeneous half-space are modified in the common case where compressional and shear velocity increase downward in the earth and also in the case where a low-velocity region such as the mantel or asthenospheric low-velocity channel exists.

Theoretical particle amplitude profiles and dispersion curves, computed on the basis of bore hole measurements of compressional and shear velocities are compared with the data of DOBRIN, *et al*, on explosion-generated Rayleigh waves of 4 to 8 cps recorded in the bore hole. Observed particle trajectories and computed amplitude profiles are in good agreement except in the upper 10 ft where the very large horizontal amplitudes predicted by theory, particularly for short periods, are not shown in the field observations.

Particle amplitude profiles based on Gutenberg's model of the mantle are given for a broad spectrum of mantle Rayleigh wave and long period crustal Rayleigh wave frequencies. These data show that the heterogeneous character of the mantle cannot be neglected in the problem of crustal Rayleigh wave dispersion. They also show that no "captured waves" or "channelled waves" of unusual character exist in the Rayleigh mode. Instead, particle motion profiles for the heterogeneous earth differ only slightly from the profile for Rayleigh waves on a homogeneous half-space throughout the spectrum of this mode.

Dispersion of Love waves in a heterogeneous spherical earth
by Y. SATÔ, M. LANDISMAN and M. EWING

Numerical solutions for the dispersion of Love waves in a heterogeneous spherical earth have been solved by programming the modified Adams method for an electronic digital computer. The method is similar to that previously employed by SATÔ for dispersion in heterogeneous media having no curvature.

The fundamental and several higher radial modes of vibration have been calculated for the Jeffreys–Bullen model and for one proposed by Miss. I. LEHMANN.

Surface wave dispersion in elastic media having gradients in their physical properties
by M. LANDISMAN, Y. SATÔ and M. EWING

Machine calculations have been applied to various cases of Love wave dispersion in elastic media having gradients in their physical properties. Among the cases considered are several which are directly related to the structure of the earth.

Love wave dispersion has been computed by two different methods. In the first method, the period equation is solved by the explicit series appropriate for the model under study. This has been done for MEISSNER's case, a half-space with a linear gradient in shear modulus, and for SATÔ's case, which is the same structure beneath two homogeneous layers. MEISSNER's case has also been solved by the second method, which is a numerical solution of the equation of motion, and the two methods are in excellent agreement. This agreement indicates that the numerical methods may be used to investigate Love wave dispersion in the multigradient structures which result from travel time studies.

Explicit series solutions of the period equation for models with two homogeneous layers over a gradient half-space were used to study the structure of the upper mantle under continents and oceans. These solutions have confirmed the conclusions, reported a year ago on the basis of Love wave dispersion studies, and since confirmed in a report by DORMAN and others on Rayleigh wave dispersion:

1. Under continents a zone of low shear velocities exists between depths of roughly 100 and 200 km.
2. The upper mantle beneath the oceans is different from that under the continents. Under oceans the region of low shear velocities rises to depths of about 50 km. The low velocity zone is thicker, and the velocities are lower, than under continents.

Observations of phase velocity for Rayleigh waves in the period range 100–400 sec
by J. E. NAFE and J. N. BRUNE

Phase velocity as a function of period has been determined for Rayleigh waves in the period range 100–400 sec. The results were derived from a study of seismograms from the Southeastern Alaska earthquake of 10 July, 1958 and from published data on the Assam earthquake of 15 August, 1950. The method depends on measurement of the travel time of wave crests along an arc of known length with proper correction for change of period with distance. For observations of a single Rayleigh wave train at a single pair of observing stations crest identification is uncertain and so too is the resulting phase velocity period curve. A set of phase velocity curves may be computed each one corresponding to a different choice of crest identification. Only one of these is consistent with the data from several earthquakes and several pairs of observing stations. In the present work, high precision in phase velocity measurement is achieved by using the observations of the Rayleigh waves R_3 and R_5 at Pasadena of the Assam earthquake. Data from the Southeastern Alaska earthquake are used to resolve the ambiguity resulting from uncertainty in crest identification. The final phase velocity curve is estimated to be accurate to better than 1 per cent in the range of periods 100–400 sec.

7.6. Conclusions

It is clear from an inspection of those papers already published and those in preparation that the Long Period and Lg programs of the International Geophysical Year produced scientific results of such quality and quantity that pre-IGY expectations have already been achieved or surpassed. There remains still a great quantity of scientific data which has not yet been treated, and which is likely to produce a volume of scientific results of at least equal importance to those already obtained. The author would like to take this opportunity to thank personally all of those who participated, and who, in doing so, not only contributed to the success of the program, but demonstrated again, the feasibility and the great value of international cooperation in science.

8. MEASUREMENTS OF STRAIN IN THE EARTH'S CRUST

by Hugo Benioff

8.1. Introduction

As part of the United States program in seismology for the International Geophysical Year, two fused quartz extensometers were installed at Ñaña (near Lima), Peru, and at Santiago, Chile. This program was undertaken under the direction of the California Institute of Technology Seismological Laboratory, operating jointly with the Peruvian Committee for the IGY for the Ñaña station, and the University of Chile for the Santiago station. J. A. Broggi, Chairman of the Peruvian Committee for the IGY, served as scientific advisor for the

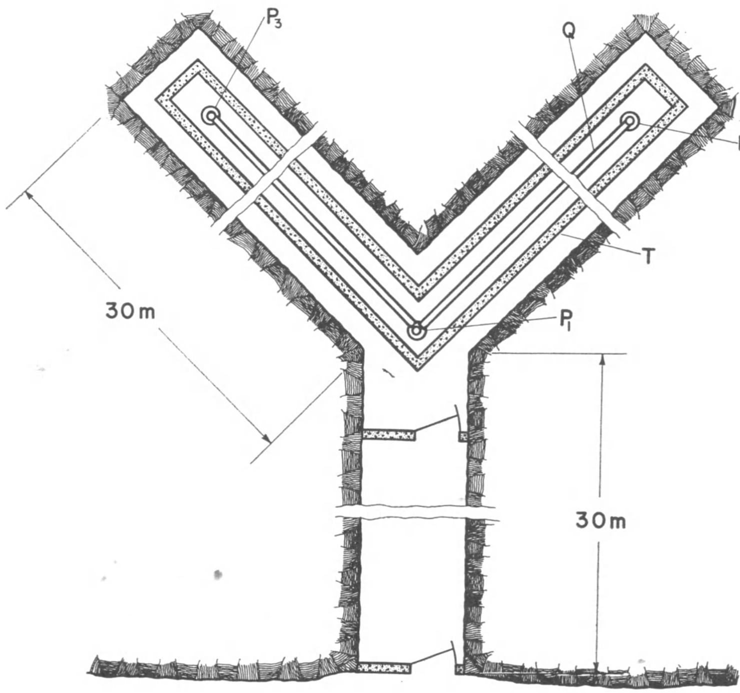

Fig. 1. Extensometer installation, Ñaña, Peru.

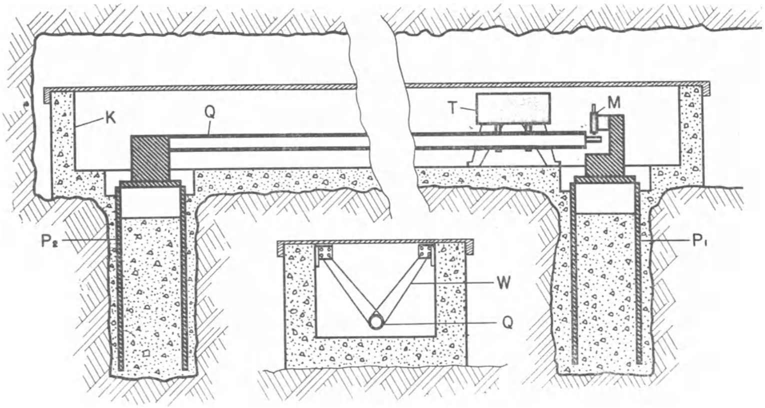

FIG. 2. Schematic longitudinal section of one extensometer component.

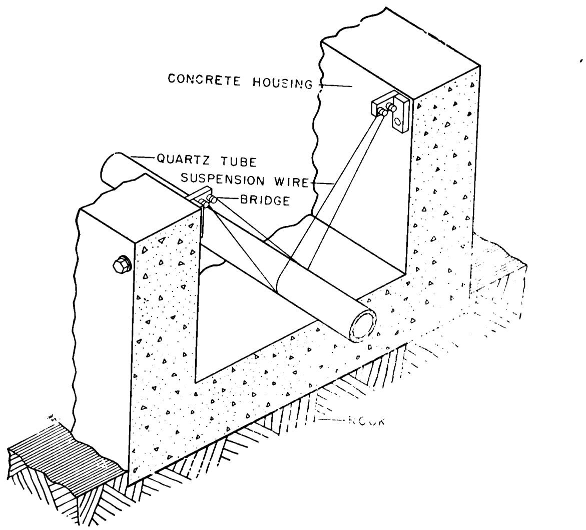

FIG. 3. Trough and suspension detail for extensometer.

Ñaña station with the active cooperation of A. GIESECKE, Jr., Technical Director of the Instituto Geofisico de Huancayo, and P. LEDIG, also of the Instituto Geofisico de Huancayo. C. LOMNITZ of the staff of the University of Chile was advisor for the Santiago installation.

As their contribution to the joint effort the local agencies excavated the tunnels in which the instruments are located, and in addition, they provided electric power service at the sites, maintenance and operating personnel.

The tunnels are constructed in the form of a Y with the two arms mutually perpendicular as shown in Fig. 1. The arms are each approximately 30 m long. In the Ñaña tunnel, the entrance portion is also 30 m, whereas at Santiago it is about half this length. The fixed ends of the quartz standards are mounted to the piers P_2 and P_3 (Fig. 1) which are constructed of 10-in (25·4 cm) galvanized steel pipe set 2 m into the rock of the floor and cemented in with concrete, as shown in Fig. 2. The quartz standards are approximately 25 m in length and are made up of 3-m sections of quartz tubing 6·4 cm in outside diameter with 0·6 cm wall thickness. The sections are cemented end to end with Sauereisen low-expansion ceramic cement with the addition of a wrapping of glass tape saturated with the same cement. The standards are housed within concrete troughs (Figs. 2 and 3) formed in 1·5-m sections with soft fiber spacers between sections to reduce differential expansion effects between the trough and the rock. The troughs are covered with resin-bonded waterproof plywood planks. The quartz tubes are supported at 1·5-m intervals by double slings of stainless steel wires, 0·036 cm in diameter, attached to machine bolts screwed into lugs anchored to the concrete troughs (Fig. 3). Accurate adjustment of the wires is, thus, effected by rotation of the bolts.

8.2. Measuring microscope

For measurements of secular-strain increments, the position of an accurately ruled glass scale attached to the end of each quartz tube is read in a measuring microscope mounted on a heavy stainless steel pedestal anchored to the pier. The glass scale is illuminated from below by a small incandescent lamp. The scale is ruled with lines spaced at 0·01 mm intervals. A crosshair in the microscope eyepiece is used as fiducial line and the position of the numbered lines of the scale can be estimated to tenths of their interval or 0·001 mm. With a length standard of 25 meters the least detectable strain increment as measured with the microscope is 4×10^{-8}.

8.3. Tidal rainst recorder

During the early stages of the tunnel construction it developed that suitable electric power in the form of uninterrupted service at constant voltage and frequency would perhaps not be available at the sites. Accordingly, the recording assemblies for the tidal strains were modified to operate free of power sources other than dry batteries. The transducer is a modified form of the 5 Mc/s resonant capacitance bridge assembly described earlier for pendulum instruments.[1] The capacitance bridge is formed by two fixed insulated plates, P, (Figs. 4 and 5) anchored to the adjacent pier, and a third grounded plate, F, mounted between them and attached to the quartz tube, Q, near its end. The capacitances formed between the grounded plate and the two fixed plates are shunted by two equal inductances, L, (Fig. 5) to produce two resonant circuits. These are loosely

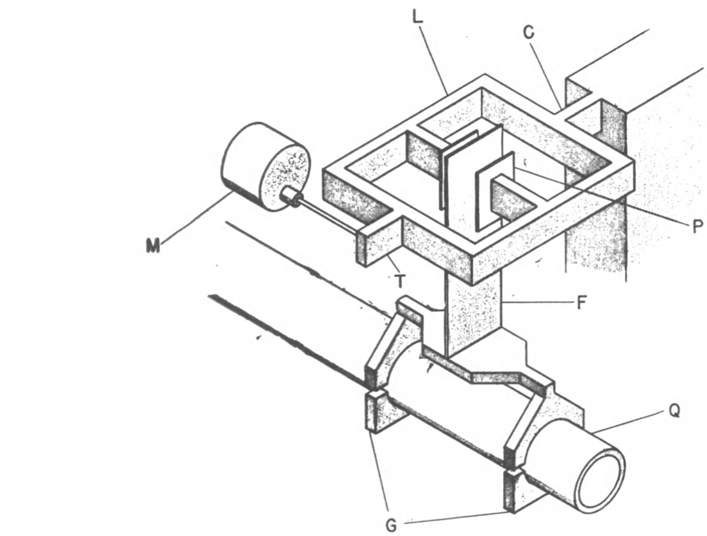

FIG. 4. Schematic representation of transducer assembly.

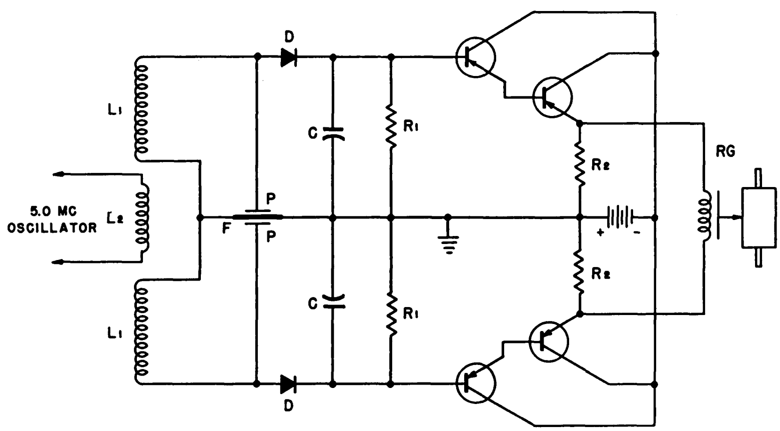

FIG. 5. Transducer circuit diagram.

coupled to a 5 mc crystal controlled transistor oscillator as shown. When the grounded plate is positioned midway between the plates, the two circuits are resonant at a common frequency which differs from the oscillator frequency sufficiently to reduce the currents to approximately 0·8 of their resonant value and in this condition they are also equal. Movement of the grounded plate relative to the fixed plates produced by strain increments in the ground between the piers varies the two capacitances oppositely, making one circuit more nearly resonant with the oscillator and the other less so. The resulting unbalance in the magnitude of the two currents is proportional to the displacement of the end of the quartz standard relative to the adjacent pier, which in turn is proportional to the component of ground strain parallel to the standard. The resonant circuits are connected to two germanium diodes, D, D, so that the rectified output voltages across the two resistors R_1, R_1 are proportional to the currents flowing in the two circuits. The two output voltages serve to drive a pair of transistor push-pull double cathode followers which supply sufficient power to drive an Estertine-Angus 0–1 mA ink-writing recorder. The recorder is spring driven and manually wound. Dry batteries are used for the power source for the transistors. One set lasts approximately three months in continuous service.

The insulated plates of the transducer are individually adjustable for fixing the operating capacitance air gaps and are arranged for common movement by means of a reversible motor and gear train, M, (Fig. 4). The transducer can thus be balanced remotely by operation of push buttons situated near the recorder. The transducer assembly is housed within a hermetic enclosure sealed to the moving member by a ring of silicone grease which also serves to damp transverse vibrations of the quartz tube.

The available magnification of the recording system is in excess of 100,000. In routine operation the magnification is limited to about 80,000 to prevent the maximum tidal strain deflection from going off scale. With a standard having a length, L, in millimeters, and a recorder magnification M, and trace amplitude A, in millimeters, the ground strain is $\sigma = A/LM$.

8.4. Santiago operation

Shortly after completion of the Santiago station, water began to leak into the tunnel in quantities sufficient to prevent satisfactory operation of the instruments. The inside of the tunnel was then treated with gunnite, a special concrete material and although the leakage was reduced, it is still large enough to affect adversely the stability of the instruments. The noise level is, thus, abnormally high, and it is difficult to keep the instruments in satisfactory condition, free of fungus and corrosion. If the leakage continues at its present level, it will be necessary to move the installation to another, drier site.

8.5. Ñaña operation

The Ñaña tunnel has proved to be an extremely dry one, so much so that operation of the fresh air blower was discontinued entirely. The noise level at this site is the lowest which the writer has encountered. Fig. 6 is a reproduction of the tidal strain recording of 6 November, 1958, showing waves of the Kurile Islands earthquake well recorded. The recorder was fitted with an electromag-

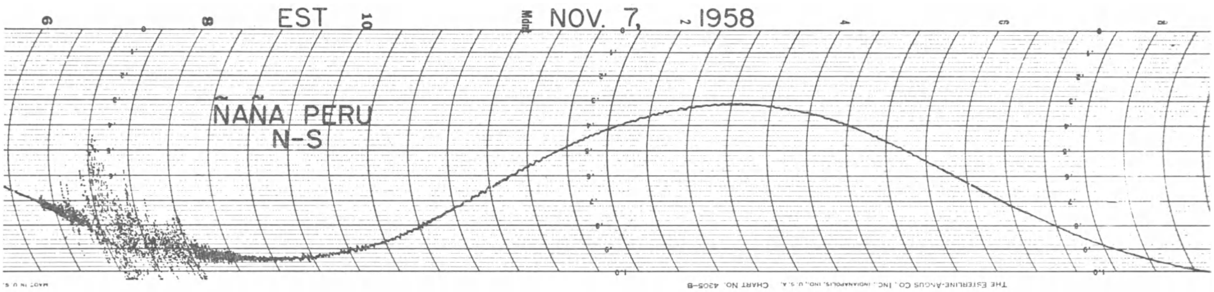

FIG. 6. Ñaña recording of the Kurile Islands earthquake of 6 November, 1958 (N33·75° W component).

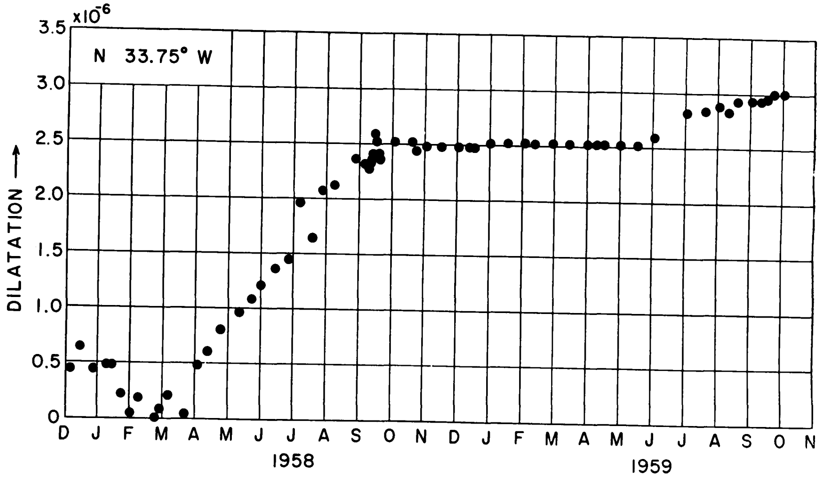

FIG. 7. Observed secular strain increments, N33·75° W component, Ñaña.

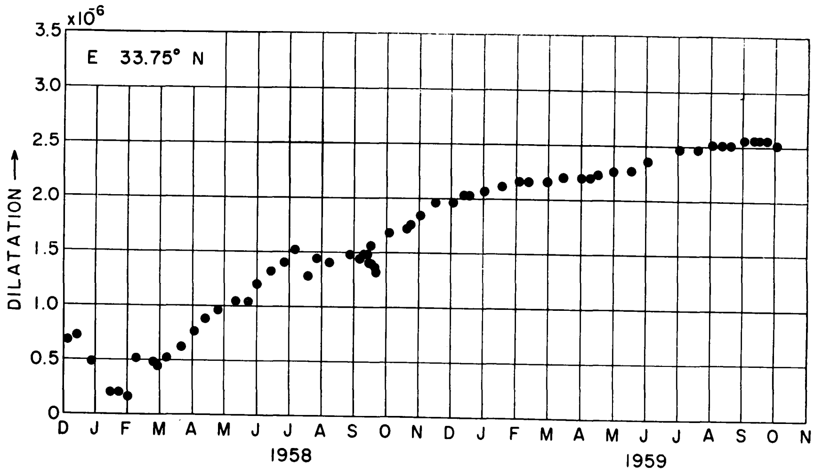

FIG. 8. Observed secular strain increments, E33·75° N component, Ñaña.

netic device for lifting the pen once every two seconds for reducing the effects of pen friction, and this has produced the dotted portions of the recording during the fast wave motions. This device has since been removed since it produced too many difficulties with ink flow in the pen. Owing to the low noise level, the recorded line is very steady when waves are absent and consequently, in this shock, wave motion is discernable for some 12 hours following the initial arrival.

8.5.1. Ñaña secular strain measurements

Measurements of the Ñaña secular strain increments are shown in Figs. 7 and 8. Figure 7 refers to the N33·75° W arm and Fig. 8 to the E33·75° N arm. Experience with the Isabella, California extensometer has shown that following completion of the extensometer installation, several years are required for the various elements to come to equilibrium. Among the sources of these movements are creep of the rock surrounding the tunnel in response to excavation, curing of the concrete structures, including the concrete bonding of the piers, curing of the cement joining the quartz tube sections, and creep of the quartz standards in response to strains set up during manufacture.

8.5.2. Naña tidal strain recordings

Continuous tidal strain recordings have been collected covering most of the interval from the beginning to the time of this writing.

8.6. New recorders

For progressive plane seismic waves, the strain is proportional to the time derivative of the wave displacement, i.e. the ground particle velocity. Thus, if the wave displacement amplitude is $\xi = a \sin w(t - \frac{x}{c})$, the strain is

$$\frac{\partial \xi}{\partial x} = \frac{1}{c}\frac{\partial \xi}{\partial t} = -\frac{\alpha\omega}{c}\cos\omega\left(t - \frac{x}{c}\right)$$

Since the strain amplitude is proportional to $\omega = 2\pi f$ where f is the frequency for waves of constant displacement amplitude, the direct strain recorder emphasizes the higher seismic frequencies in relation to the lower. In order to improve the recording of the very long period waves, an integrating circuit has been designed as shown in Fig. 9 for use with a Leeds and Northrup type *G* high impedance recorder having a tape speed of 6 in (15 cm)/hr. In addition to the integration portion of the circuit, a series condenser is also added to reduce the response to the tidal periods. With this circuit the recording trace amplitude is substantially flat for constant ground displacement for periods extending to several hundred seconds and falls to half amplitude at 1000 sec. The measured response curve is shown in Fig. 10.

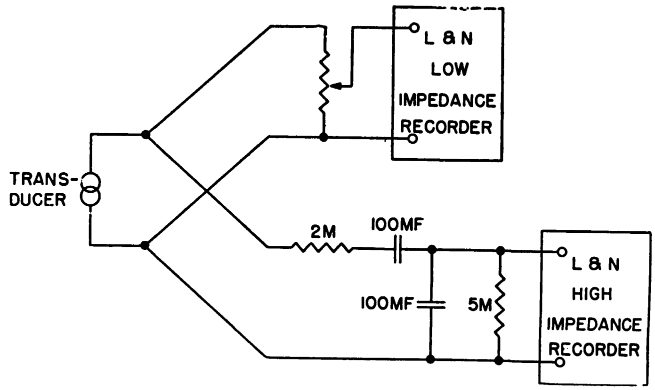

FIG. 9. Recording circuits for tidal strains and long-period seismic waves.

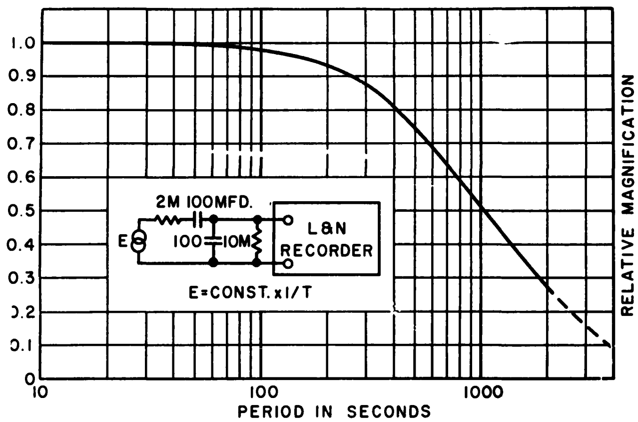

FIG. 10. Measured response of RC network on strain seismograph.

As indicated in the drawing, this recorder is operated in parallel with the tidal strain recorder from a single transducer. This network has been undergoing tests at our Isabella, California strain installation, prior to shipment to Ñaña. On a recording of the Yellowstone, Montana earthquake of 18 August 1959 made with this combination, the seismic wave pattern predicted by LAMB in 1904 has been observed for the first time. Although the pattern has been observed in model experiments[2], it has eluded detection in actual earthquakes. The reason is now obvious. The pattern exhibits P-wave and Rayleigh pulses with durations of from 2 to 5 min and existing seismographs were entirely too low in sensitivity to waves with periods in this range to be effective.

The recording of the Montana shock is shown in Fig. 11. Although the form of the later portions of the theoretical pulse shows a D. C. component, it is absent on the recording, since for the D. C. component and associated long pe-

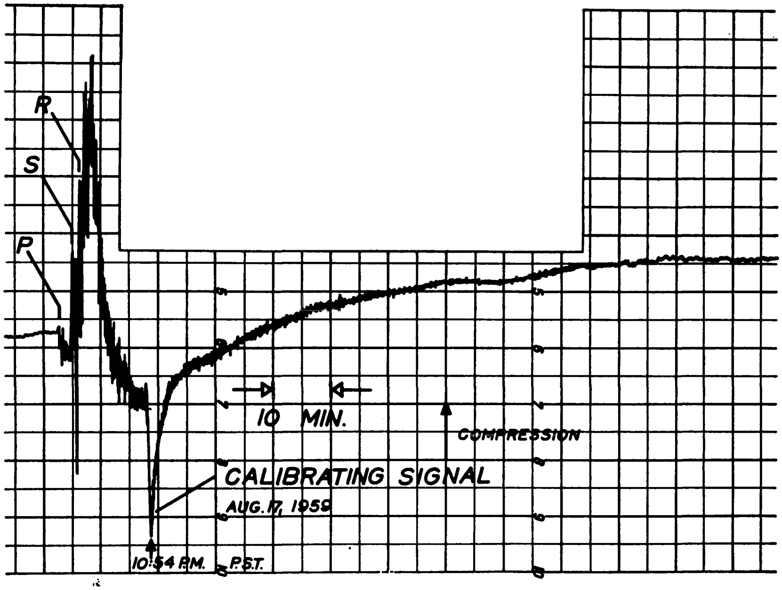

FIG. 11. Pulse pattern of Yellowstone, Montana, earthquake of 18 August, 1959, written with recorder combination of Fig. 10 at Isabella, California.

riods involved in this portion of the seismogram, the waves are no longer plane progressive waves and the network integration is inadequate to derive ground displacement from ground strain. Although the seismogram generally follows the predicted form it differs in detail owing to the presence of shorter period oscillations.

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2. TATEL, H., 1955. Note on the Nature of a Seismogram, 11, *Jour. Geoph. Res.*, 2, No. 2, July.