

Measuring Geological Time

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The age of rocks and minerals can be determined by measuring the accumulation of the products of radioactive decay in them. The result is a time scale of fundamental significance to most branches of the Earth sciences.

IN RECENT YEARS enormous progress has been made in the study of the historical aspect of all branches of the Earth sciences by the development of precise and accurate methods for measuring the age of geological materials. These methods, which are based on the radioactive decay of certain nuclides with very long half-lives, are applicable over virtually the entire range of geological time (ca. 4600 million years) and to a wide variety of commonly occurring rocks and their constituent minerals.

The types of geological events which can be dated by these methods are (i) the crystallisation of igneous rocks from a magma, (ii) the recrystallisation of pre-existing rocks, of any type, to form metamorphic rocks, usually with a totally new mineral assemblage, (iii) uplift, cooling and erosion of fold mountain chains, and (iv) deposition of a sedimentary rock, provided that one or more new minerals are formed during, or very soon after, sedimentation.

Direct dating, expressed in millions of years measured backward from the present, can clearly yield fundamental information on the timing and duration of some of the major events and processes which have shaped the surface of the Earth in the geological past and given it the features which we observe in the geological present. This will be evident from many of the other contributions to this book.

The science of dating rocks in all its aspects is termed *geochronology*. In using the basic equations of radioactive decay* to calculate the age of a geological sample from the measured parent/daughter ratio and the appropriate decay constant or half-life, it is necessary to remember that the calculated age will yield the time of crystallisation

*The fundamental law of radioactive decay states that the number of atoms disintegrating per unit time ($-dN/dt$) is proportional to the total number of radioactive atoms, N , present. Hence we can write $-dN/dt = \lambda N$, where the proportionality factor, λ , is termed the *decay constant* and has a characteristic value for each radionuclide. It represents the probability that an atom will disintegrate in unit time (year for long-lived radionuclides). On integrating the above equation, we get $N = N_0 e^{-\lambda t}$, or $N_0 = Ne^{\lambda t}$, where N_0 indicates the number of radioactive atoms at some original time t , and N the number present to-day. This equation forms the basis for calculating a geological age from the analytical data; the number of radiogenic daughter atoms, N_D , formed during a period of time t is given by

$$N_D = N_0 - N = N_0 e^{-\lambda t} - N = N(e^{\lambda t} - 1).$$

Therefore, $N_D/N = (e^{\lambda t} - 1)$, and solving for t , we have

$$t = \frac{1}{\lambda} \log_e \left[1 + \frac{N_D}{N} \right].$$

Several varieties of this basic equation are widely used in geological age calculations. Note that the commonly used term *half-life* ($T_{1/2}$) i.e. the time required for the radioactivity of a given amount of radionuclide to decay to one-half of its initial value is related to the decay constant (λ) by the following equation: $T_{1/2} = 0.693/\lambda$.

or recrystallisation of the rock only if (i) there have been no losses or gains of either daughter or parent nuclide by processes other than radioactive decay of the parent, (ii) the half-life of the parent nuclide is accurately known, (iii) precise correction can be made for any daughter nuclide incorporated into the mineral or rock at time of crystallisation, (iv) the mineral or rock was formed in an interval of time that is short when compared with its actual age.

When more than one dating method is applied to a mineral, or to a cogenetic mineral assemblage, and the resulting dates agree within analytical error, the above conditions are clearly obeyed. Such dates are termed *concordant* and are found in many rocks which have had a relatively simple geological history. However, *discordant* dates are also frequently found in certain types of geological environment. They can result from the failure of one or more of the above conditions in regions with a more complex geological history. As will become evident later, such discordant age-patterns can be extremely useful when several geological events have been superimposed on one another in a single area.

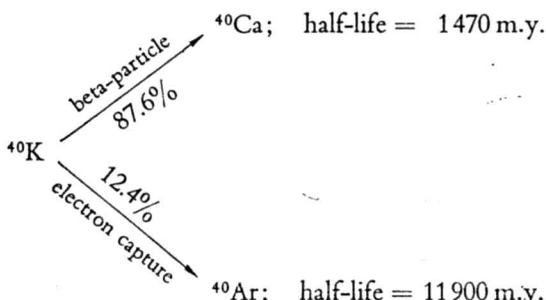
Direct measurements of long half-lives are obviously basic to geochronology and have been carried out in many physics and radiochemistry laboratories. Nevertheless, precise measurements of geological specimens are made difficult by the low-energy radiation and slow disintegration rates of long-lived radionuclides. In practice, radionuclides and their daughter nuclides can be respectively distinguished from stable and non-radiogenic isotopes of the same element by isotopic analysis with a mass spectrometer, which is by far the most important analytical tool of the geochronologist and which can also be used to measure precisely and accurately the frequently minute absolute quantities of parent and daughter nuclides in rocks and minerals. The actual laboratory techniques for determining so-called *radioisotope*, *isotopic* or *radiometric* ages of rocks (which used to be erroneously called *absolute* ages) are extremely complex. Further details of some of the principal techniques used by the fifty or more geochronology laboratories throughout the world may be found in the references at the end of this chapter.

Published age determinations are usually quoted with an error which is most frequently somewhere between two to five per cent of the actual age, e.g. 1200 ± 40 million years, or 50 ± 2.5 million years. These errors refer only to uncertainties in the analytical measurements and represent a

probability—usually expressed as a 66 or 95 per cent confidence level—that a replicate measurement on the same sample would fall within the quoted limits. The errors do not take into account uncertainties in the half-life of the relevant radio-nuclide, or in the geological interpretation.

POTASSIUM-ARGON ($^{40}\text{K} - ^{40}\text{Ar}$ or K - Ar) METHOD

Potassium is a common and widespread element in many rocks and minerals (see Chapter 1). Of its three isotopes, namely ^{39}K , ^{40}K and ^{41}K , only ^{40}K is radioactive, forming a constant proportion of the total potassium of only 0.0119 atom per cent. ^{40}K decays by two entirely different processes, each with its own characteristic half-life, as follows:



The $^{40}\text{K} - ^{40}\text{Ca}$ method is only rarely used for dating, because most common rock-forming minerals already contain so much primary calcium, which is mainly ^{40}Ca , that the comparatively minute amounts of radiogenic ^{40}Ca cannot be determined with anything like sufficient precision for geological age work.

The K-Ar method can be used to date a wide variety of rock types over almost the entire geological age range, including meteorites (ca. 4 600 million years old) as well as terrestrial rocks older than 3 000 million years to younger than 1 million years. However, rocks younger than about 0.1 million years can be very difficult to date because of the minute amounts of radiogenic ^{40}Ar present.

Commonly occurring igneous or metamorphic minerals suitable for K-Ar dating include biotite, muscovite, hornblende and nepheline. High-temperature forms of the potassium-feldspars (e.g. sanidine) and of the sodium-calcium (plagioclase)

feldspars are generally suitable, particularly in Tertiary and Quaternary igneous rocks. In contrast, the most common potassium-feldspars, such as orthoclase and microcline, are unsuitable for K-Ar dating because they can lose argon readily even at atmospheric temperatures over long time intervals, due to the existence of gross structural defects in the crystal lattice and at boundaries between individual grains.

In many cases, whole-rock samples can be dated as they are found without mineral separation. (The term *whole rock* means just what it says: a piece of rock as it is found cropping out on the Earth's surface.) This is particularly useful for fine-grained volcanic rocks (lavas) such as basalts, andesites and rhyolites, and also for fine-grained, essentially micaceous, metamorphosed sediments, such as slates and phyllites. The marine sediments, limestone and sandstone, sometimes contain the potassium-bearing mineral glauconite, which is formed at time of sedimentation and can be dated by the K-Ar method.

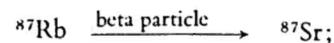
Measured K-Ar dates do not necessarily relate to the time of crystallisation, because diffusive loss of radiogenic ^{40}Ar can occur at temperatures well below those of igneous and medium-to-high grade metamorphic crystallisation (ca. 1200–1000°C and 700–400°C respectively). This is because the radiogenic argon atom does not easily fit into the potassium lattice site in which it was formed and can escape out of the crystal lattice by diffusion, eventually finding its way upwards to join the reservoir of atmospheric ^{40}Ar which makes up nearly one per cent of the Earth's atmosphere. The so-called 'blocking' or 'diffusion-threshold' temperatures below which different minerals become effectively closed systems to argon diffusion are not known precisely, but are probably about 150–250°C for biotite, 250–350°C for muscovite and sanidine, and 400–500°C for hornblende. Furthermore, the basic laws of diffusion show that the larger the grain size of a particular mineral, the higher will be the effective blocking temperature. It is now known that very large, deep-seated igneous intrusions and regionally metamorphosed parts of the Earth's crust which have gradually been uplifted from great depths and have cooled extremely slowly from the initial crystallisation temperatures may yield K-Ar dates younger than the true age of crystallisation by several tens, or even hundreds, of millions of years. Furthermore, different types of minerals may become closed systems to argon diffusion at different times, because of the respective differences in their block-

ing temperatures. For a given grain size the measured dates will clearly be in the order hornblende date > muscovite date > biotite date. These dates relate to various stages in the uplift and cooling history of the rocks. On the other hand, K-Ar dating of a lava-flow, volcanic ash or a small body of igneous rock emplaced high up in the Earth's crust may give the true age of crystallisation within analytical error, because such rock types cooled within a very short time interval compared to their actual age. Furthermore, K-Ar dates approaching the true age of metamorphism may be obtained from fine-grained, metamorphosed micaceous sediments, such as slates and phyllites, which recrystallised completely at only about 200–300°C, leading to complete expulsion of radiogenic argon inherited from the original sediment, but at temperatures only slightly above that at which subsequently formed radiogenic argon could begin to accumulate.

Cases have been recognised where mineral samples yield anomalously old K-Ar dates—sometimes greater than the accepted age of the Earth—because they have apparently incorporated radiogenic ^{40}Ar into their crystal lattices during crystallisation. This comparatively uncommon phenomenon tends to show up in potassium-poor and/or geologically young samples which contain very little intrinsic radiogenic argon, although it has also been observed in old micas with high potassium contents. This so-called 'excess' argon is probably produced by heating and outgassing of ancient, potassium-bearing rocks at some considerable depth and pressure in the Earth's crust, where some of it became incorporated into crystallising igneous and metamorphic minerals.

RUBIDIUM-STRONTIUM (^{87}Rb - ^{87}Sr or Rb-Sr) METHOD

The two isotopes of natural rubidium are ^{85}Rb and ^{87}Rb , of which the atomic abundances are 72.8 and 27.2 per cent respectively. The decay scheme is as follows:



half-life = 47 000 or 50 000 million years.

Unfortunately, there is still no agreement on the exact value for the half-life of ^{87}Rb because this is so difficult to determine by direct radioactivity measurements. Most workers nowadays seem to prefer the higher value because better agreement is obtained with K-Ar and U-Pb ages on identical, or other, minerals from the same rock. However, both half-life values are still widely used and it should be noted that ages calculated with the 47000 million year half-life are 6 per cent lower than those calculated with the 50000 million year value. Although this half-life uncertainty is much greater than for the other radionuclides used in rock dating, the advantages and potentialities of the Rb-Sr method far outweigh it.

The rare element rubidium forms no minerals of its own, but is always found in potassium minerals where it can easily substitute for the chemically closely related potassium in the crystal lattice because of similarity in ionic radius and charge. Most of the suitable minerals for Rb-Sr dating contain between about 100 to 1000 parts per million of rubidium.

The Rb-Sr method can be used to date such common rock-forming minerals as muscovite, biotite, and (in contrast to the K-Ar method) all types of potassium-feldspars, including orthoclase and microcline. Hornblende is not suitable because it contains too little rubidium. The sedimentary mineral glauconite has also been successfully dated. Rb-Sr dating may also be carried out on whole-rock specimens without prior mineral separation, particularly those types of igneous and metamorphic rocks that are rich in potassium-feldspar and mica minerals, such as granites and granitic gneisses. Basic igneous and metamorphic rocks are usually not suitable. In whole-rock work, a sample of rock weighing several pounds is completely crushed to a fine powder, after which small aliquot samples are taken for rubidium, strontium and strontium isotope analyses. The Rb-Sr method may be applied to rocks of almost any geological age, although the comparatively slow decay rate of ^{87}Rb can create analytical difficulties for material younger than about 20–30 million years, unless it has an exceptionally high ^{87}Rb content.

In Rb-Sr dating it is usually necessary to make a correction for any so-called 'common' or 'normal' strontium which was incorporated into the mineral or rock when it crystallised. Although most common strontium follows the geochemically closely related calcium, a small amount is usually also incorporated into calcium-poor,

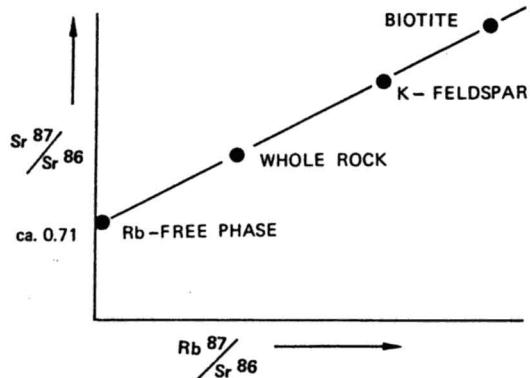
rubidium-bearing potassium minerals. For example, micas and potassium-feldspars commonly contain between about 5 and 300 parts per million of common strontium, which may actually exceed considerably the amount of radiogenic strontium formed since crystallisation. Common strontium consists of isotopes of mass 84, 86, 87 and 88, in which the relative abundance of ^{87}Sr to the others can be quite variable in different crystallisation environments. It is therefore essential to distinguish between common ^{87}Sr incorporated on crystallisation and subsequently formed radiogenic ^{87}Sr . This can sometimes be done by isotopic analysis of a coexisting, cogenetic calcium mineral such as plagioclase which characteristically has an extremely low Rb/Sr ratio and in which the strontium isotope composition has remained unchanged since crystallisation, regardless of the age of the rock. The non-radiogenic isotope ^{86}Sr is customarily used as a measure of the common ^{87}Sr component. The basic equation of radioactive decay can then be used in the form:

$$t = \frac{1}{\lambda} \log_e \left[\frac{(^{87}\text{Sr}/^{86}\text{Sr})_p - (^{87}\text{Sr}/^{86}\text{Sr})_0}{^{87}\text{Rb}/^{86}\text{Sr}} + 1 \right]$$

where t is the age of the rock, λ is the decay constant of ^{87}Rb , $(^{87}\text{Sr}/^{86}\text{Sr})_p$ is the measured isotope ratio in the dated mineral, $(^{87}\text{Sr}/^{86}\text{Sr})_0$ is the isotope ratio at time of crystallisation (i.e. in the plagioclase with extremely low Rb/Sr ratio). $^{87}\text{Rb}/^{86}\text{Sr}$ is computed from the analyses for total rubidium and strontium in the sample.

As stated above, the ratio $(^{87}\text{Sr}/^{86}\text{Sr})_0$ varies significantly in different types of rocks, depending upon the pre-crystallisation history of the rock, particularly as regards the age and Rb/Sr ratio of its source region. For most igneous rocks derived from deep-seated sources in the lower crust or upper mantle the $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratio varies from a minimum value of about 0.700 to about 0.715; for metamorphic rocks with a lengthy crustal pre-crystallisation history and with relatively high Rb/Sr ratios it can be much higher. The $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratio, usually referred to as the *initial* strontium isotope ratio, is an important parameter for studying the ultimate origin of igneous rocks.

In Rb-Sr dating it is common practice to analyse several minerals from the same rock, or several whole-rock specimens from a given body of granite or granite-gneiss, and to interpret them by a graphical method based on the above equation,



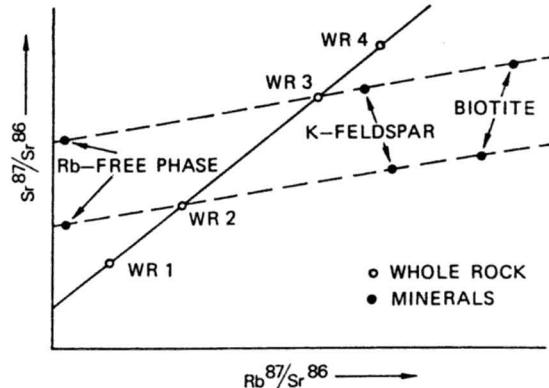
2.1 Rb-Sr isochron plot for a whole-rock specimen of granite and the minerals separated from it. The slope of the isochron is proportional to the age of the rock, whereas the intercept gives the isotopic ratio $^{87}\text{Sr}/^{86}\text{Sr}$ of common strontium when the rock crystallized.

which is termed an *isochron* plot. The equation can be rewritten as follows:

$$(^{87}\text{Sr}/^{86}\text{Sr})_p = (e^{\lambda t} - 1) ^{87}\text{Rb}/^{86}\text{Sr} + (^{87}\text{Sr}/^{86}\text{Sr})_0$$

This is clearly the equation of a straight line of the type $y = mx + b$, where y and x are the axes, m is the slope and b is the intercept. The age of a cogenetic mineral assemblage can be calculated from the slope of the isochron, whereas the intercept represents the initial strontium isotope at time of crystallisation. An isochron plot for a hypothetical granite sample is shown in Figure 2.1. Clearly, the principal requirement for successful application of the isochron method is the availability of mineral and/or whole-rock phases with different Rb/Sr ratios from within a given rock sample or rock unit respectively. It is frequently found that whole-rock samples from a given rock unit exhibit considerable variation in their Rb/Sr ratios, even where the rock unit appears to be petrologically quite uniform over large areas.

Because of the geochemical dissimilarity between rubidium and strontium, radiogenic ^{87}Sr can diffuse out of rubidium lattice sites at temperatures well below those of igneous and metamorphic crystallisation. Rather surprisingly, perhaps, the diffusion behaviour of radiogenic ^{87}Sr is very similar to that of radiogenic argon. Blocking temperatures are probably in the general range



2.2 Whole-rock and mineral Rb-Sr isochrons for a metamorphosed granite. The slope of the whole-rock isochron is proportional to the age of igneous intrusion. The slope of the mineral isochrons is proportional to the age of metamorphism.

200–500°C, depending upon type of mineral and grain size. The blocking temperature order for a given grain size is potassium-feldspar > muscovite > biotite, and this is frequently the observed order of measured dates in large bodies of granite or gneiss in orogenic belts which underwent slow uplift and cooling after crystallisation. In such a situation whole-rock specimens of granite or granite-gneiss and of some types of feldspathic and micaceous schists are more retentive than individual minerals. This is because radiogenic ^{87}Sr that diffuses out of rubidium-bearing minerals down to relatively low temperatures can easily enter the lattice of the almost invariably present surrounding calcium minerals such as plagioclase or apatite and homogenise and equilibrate with the common strontium there, so that a hand-specimen of whole-rock can remain a closed system with regard to radiogenic ^{87}Sr .

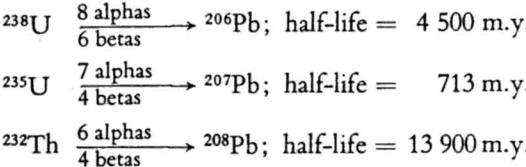
To carry this one stage further, it is often found that a granite intrusion may remain a closed system with regard to radiogenic ^{87}Sr (as well as common strontium and rubidium) even during a subsequent metamorphism of sufficient intensity to convert it into a gneiss. Exactly as in the simple cooling situation described above, radiogenic ^{87}Sr diffuses out of micas and potassium-feldspars during the metamorphism but then mixes and homogenises with common strontium

in neighbouring grains of calcium minerals, so that individual hand-specimens of the whole rock remain closed systems. Whole-rock Rb-Sr measurements on a body of granite-gneiss, even one occupying an area of tens or hundreds of square kilometres, can then yield a close approximation to the time of original crystallisation and intrusion of granite magma, whilst Rb-Sr and K-Ar mineral dates will record either the time of subsequent metamorphism, or the time of cooling and radiogenic closure after metamorphism, in the order of the mineral blocking temperatures for radiogenic ^{87}Sr and ^{40}Ar . A hypothetical example is presented in Figure 2. 2. The whole-rock isochron (points WR) yields the age of crystallisation and emplacement of the original granite, whilst the dashed lines show the isochrons for different minerals (all assumed to have become closed systems at the same time for the sake of simplicity) separated from whole-rock specimens WR-2 and WR-3, in which the slope is proportional to the time of metamorphism or radiogenic closure and the intercept yields the initial strontium isotope ratio at that time. The absolute differences between the initial $(^{87}\text{Sr}/^{86}\text{Sr})_0$ ratio obtained from a whole-rock isochron and the corresponding mineral isochrons depend, of course, on the length of the time interval between igneous intrusion and metamorphism (or radiogenic closure), as well as the Rb/Sr ratio in the individual systems.

The practical application of Rb-Sr whole-rock isochrons has been of the greatest importance to geochronology and has contributed an immense amount of fundamental information on the age of igneous and metamorphic provinces throughout the world.

URANIUM-LEAD (^{238}U - ^{206}Pb , ^{235}U - ^{207}Pb), LEAD-LEAD (^{207}Pb / ^{206}Pb) AND THORIUM-LEAD (^{232}Th / ^{208}Pb) METHODS

The decay schemes, omitting details of the complex chains of intermediate nuclides, are:



Uranium and thorium frequently occur in the same mineral and it is therefore possible, in principle, to make three independent age determina-

tions on one mineral sample. The basic equation for calculating an age, t , is:

$$t = \frac{1}{\lambda} \log_e \left[1 + \frac{\text{radiogenic } {}^{206}\text{Pb}}{^{238}\text{U}} \right] \text{ etc.}$$

Correction for common lead incorporated into the mineral during crystallisation is usually necessary, although the correction is mostly quite small. The isotopic composition of lead in nature is very variable. Common lead consists of the isotopes ^{204}Pb , ^{206}Pb , ^{207}Pb and ^{208}Pb . Of these, only ^{204}Pb is entirely non-radiogenic and is used to correct for the common lead component as already described for the analogous case of strontium, i.e. by means of a coexisting uranium- and thorium-free lead mineral such as galena (PbS), or by isochron methods.

A semi-independent age can be obtained from the isotopic analysis of lead alone—without accompanying analysis for total amounts of uranium, thorium and lead in the mineral—namely from the ^{207}Pb / ^{206}Pb ratio. This ratio is itself time-dependent, because of the different decay rates of ^{238}U and ^{235}U . After correction for any common lead the age, t , may be calculated from the equation:

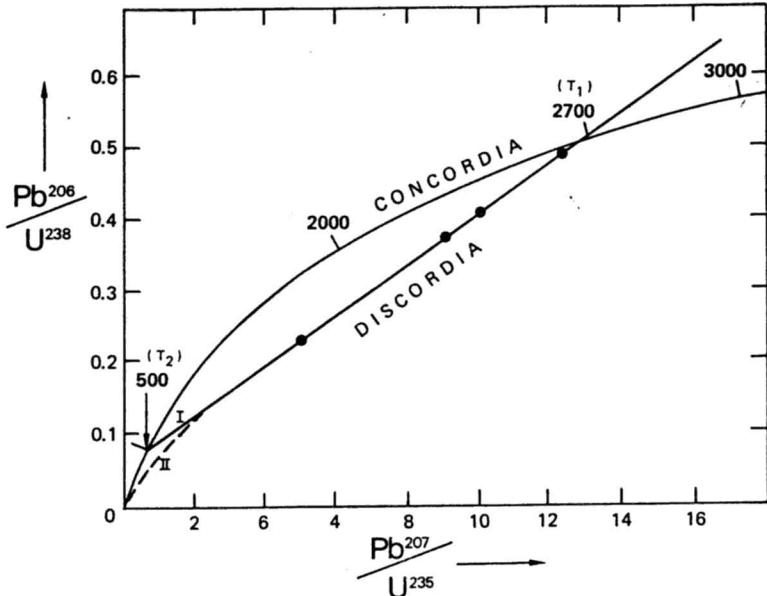
$${}^{207}\text{Pb}/{}^{206}\text{Pb} = \frac{1}{137.8} \frac{(e^{\lambda_1 t} - 1)}{(e^{\lambda_2 t} - 1)},$$

where λ_1 and λ_2 are the decay constants of ^{238}U and ^{235}U , and $1/137.8$ is the present-day atomic ratio $^{235}\text{U}/{}^{238}\text{U}$. This equation is solved for t by a trial and error method, or from specially prepared tables in which ${}^{207}\text{Pb}/{}^{206}\text{Pb}$ is given as a function of t .

It is of passing interest to note from the half-lives that since the formation of the Earth approximately 4600 million years ago, about 50 per cent of the original ^{238}U has decayed to ^{206}Pb , 99 per cent of the original ^{235}U to ^{207}Pb , and 20 per cent of the original ^{232}Th to ^{208}Pb .

These methods can be applied over the greater part of the geological age range, although their usefulness decreases sharply for rocks younger than about 100–200 million years, partly because the rate of production of ^{207}Pb has become so low. The methods have a much more restricted range of application than the K-Ar and Rb-Sr methods but they may be used to date rare minerals in

2.3 Concordia-Discordia
 Diagram showing (I) the effect of a single period of lead loss, so that samples of age T_1 (2700 million years) that lost variable amounts of lead at time T_2 (500 million years) fall on a chord connecting T_1 and T_2 , (II) the hypothetical effect of continuous lead loss by diffusion from samples of age T_1 .



which uranium and thorium are major constituents. Most frequently, however, the methods are applied to minerals such as zircon (zirconium silicate) and sphene (calcium titanium silicate) which occur in accessory amounts of up to about one per cent in many igneous and metamorphic rocks, and which have uranium and thorium contents ranging from a few hundred to a few thousand parts per million.

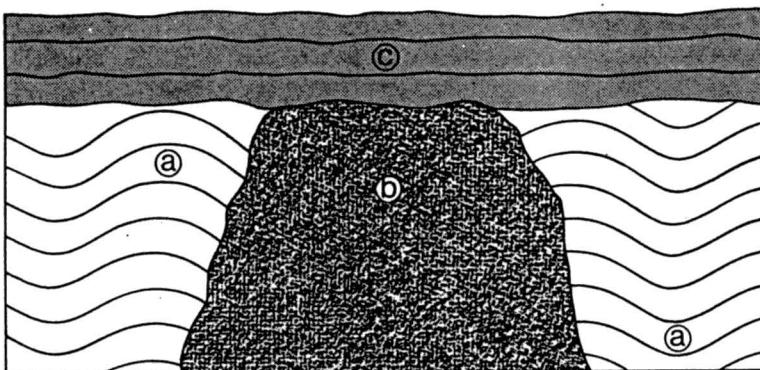
It should be noted that the U-Pb methods are of much greater practical value than the Th-Pb method. The latter is nowadays only rarely used because it often yields anomalous ages which can be difficult to interpret.

In some cases $^{238}\text{U}/^{206}\text{Pb}$, $^{235}\text{U}/^{207}\text{Pb}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ dates for a given mineral are found to be concordant within analytical error and thus yield a valuable geochronological reference point. More frequently, however, the three measured dates are grossly discordant outside any plausible limits of analytical error. A commonly observed pattern is:

$$^{207}\text{Pb}/^{206}\text{Pb} > ^{235}\text{U}/^{207}\text{Pb} > ^{238}\text{U}/^{206}\text{Pb}.$$

Such discordant patterns result from lack of closed-system behaviour for the radioactive and radiogenic nuclides, including the intermediate members of different decay chains. The degree of discordance itself usually varies in different samples of the same mineral from a given rock unit.

Close study of discordant U-Pb age patterns shows that, in many cases, they result either from lead loss from a mineral during a metamorphic event long after crystallisation, or from continuous diffusion of lead out of the mineral over its entire geological history. Extremely useful geological information may be obtained from such discordant age patterns, particularly from the mineral zircon. This is illustrated by means of the so-called *Concordia-Discordia* diagram (Fig. 2.3), in which the measured $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{235}\text{U}$ values (after correction for common lead) are plotted against each other. The Concordia curve represents the locus of all points for which the $^{238}\text{U}/^{206}\text{Pb}$ age equals the $^{235}\text{U}/^{207}\text{Pb}$ age. When, as usually happens, different zircon samples from a rock unit define a straight Discordia line, the upper intercept with Concordia gives the time of crystallisation, t_1 , of the zircon. The lower intercept with Concordia indicates the time, t_2 , at which a metamorphic heating event removed variable amounts of lead from zircons from different parts of the rock unit, whilst the actual position of a sample on Discordia is determined by the amount of lead lost by that sample at time t_2 . This situation is referred to as *episodic lead loss*. In other cases, however, there is simply no evidence from independent K-Ar and Rb-Sr measurements on associated rocks for a metamorphic event at the time given by the lower, extrapolated intercept of Discordia on Concordia. An alternative mechanism of continuous diffusion of lead out of the



2.4 Cross-section of a 'bracketed' igneous intrusion. Provided that sediment (a) is not much older than sediment (c), the radio-isotopic date of the intrusion (b) can be used as a calibrated point in the time scale.

mineral has been postulated in such cases in which the extrapolated Concordia bends down towards the origin, as shown in Figure 2.3. (No samples actually falling on the curved region have yet been reported.) Whichever mechanism is assumed for lead loss, extrapolation back to time of crystallisation, t_1 , remains valid, but the assumption that lead loss was episodic must be verified by independent K-Ar and Rb-Sr measurements on the same rock unit.

GENERAL DISCUSSION OF DIFFERENT TYPES OF AGE PATTERNS

Concordant Age Patterns

When all measured mineral dates from a given igneous or metamorphic rock unit agree within analytical error, they either represent the time of crystallisation, or the time at which the entire rock unit cooled rapidly enough for all minerals to become closed systems to radiogenic nuclide diffusion at the same time. In those cases where whole-rock Rb-Sr ages are in agreement with the mineral ages it can be concluded that the time of crystallisation is being dated. Mineral and whole-rock dates from small igneous intrusions, volcanic lavas and ash-beds (*tuffs*), from the sedimentary mineral glauconite, as well as from low-grade metamorphosed sediments, are mostly concordant and usually give the age of crystallisation.

Such dates on rocks of known stratigraphic-palaeontological assignment are essential for fixing the post-Precambrian *Phanerozoic* time-scale. The time of deposition of glauconite-bearing sediments can be dated directly. This is best for Mesozoic and Tertiary sediments, since older glauconites are rather susceptible to loss of radiogenic nuclides due

to the temperature effect of deep burial. A somewhat more widely applicable, indirect, method of dating sediments is by means of volcanic rocks such as lava flows and tuffs which were deposited rapidly and which are often interstratified with fossiliferous sediments of known stratigraphic age. Suitable minerals from lavas and tuffs include feldspar, biotite and zircon, whilst whole-rock specimens of acid type can be dated by the Rb-Sr method. Another related, indirect method is to date a small igneous intrusion such as a dyke, sill or stock ((b) in Fig. 2.4) which intruded a fossiliferous sediment (a) and which has been unconformably overlaid by a fossiliferous sediment (c). For example, if (a) is Upper Silurian and (c) is Lower Devonian, then dating the intrusion (b) will obviously give an age value for the boundary between the two systems. This is called the method of *bracketed intrusives*.

Using the above principles, first pioneered by Arthur Holmes over fifty years ago, a Phanerozoic time-scale calibrated by radioisotope dates has been produced. More and more relevant dates from all over the world are being published each year, so that the time-scale is constantly increasing in refinement and sophistication. Several hundred points covering the entire Phanerozoic are now available, but the Mesozoic and Tertiary parts of the time-scale are more satisfactorily calibrated than the Palaeozoic.

Discordant Ages Produced by Cooling

According to the principles outlined earlier, in a deep-seated igneous or metamorphic terrain from the heart of a major orogenic belt one can usually expect a significant interval between the time of crystallisation and the time when different minerals

reached their respective blocking temperatures for the radiogenic nuclides. In geological terms this corresponds to gradual uplift and cooling of the orogenic belt through a vertical distance of at least 15–30 km and also to gradual reduction of the thermal gradient at depth (see Chapter 5). The following sequence of measured dates would be quite typical from such an environment, although inversions and irregularities may well occur because of the effect of grain size, so that coarse-grained, pegmatitic minerals become effectively closed before fine-grained minerals of the same type:

Rb-Sr (whole-rock granite, or granite-gneiss)

- > $^{207}\text{Pb}/^{206}\text{Pb}$ (zircon)
- > Rb-Sr (potassium-feldspar)
- > K-Ar (hornblende)
- > Rb-Sr (muscovite)
- > K-Ar (muscovite)
- > Rb-Sr (biotite)
- > K-Ar (biotite).

Of course, not all these minerals are necessarily, or even usually, present in a single area, and one must also remember that in the more ancient rocks an intrinsically discordant age pattern may be masked by the inherently greater absolute error of individual age measurements. However, the oldest dates from such a sequence are clearly the ones approximating most closely to the true age of crystallisation, whilst the youngest dates relate to various stages in the uplift and cooling of the orogenic belt and also to the time of erosion and deposition of the eroded detritus in bordering sedimentary basins and geosynclines.

Detailed study of age patterns in the Precambrian Grenville orogenic belt of Canada, the Lower Palaeozoic Caledonian belt of the British Isles, and of the Tertiary Alpine belt of Switzerland (see Chapter 20) indicates time intervals between crystallisation and attainment of the blocking temperature in fine-grained micas of approximately 150, 70 and 20 million years respectively. The actual differences are probably a consequence of the different depths of erosion exhibited by these three orogenic belts. With continuing uplift and erosion of the Western Alps, a geochronologist 30 million years from now, sampling surface rocks at that time, might well find a difference of 50 million years between time of crystallisation and cessation of diffusion of radiogenic nuclides.

At any given structural level, consistent differences of the type muscovite date > biotite date of about 12 million years for K-Ar, and about 8 million years for Rb-Sr, have been reported from the Caledonian belt of Scotland and from the Swiss Alps respectively. The spatial distribution of 'cooling' dates within orogenic belts is becoming an important tool for tectonic and structural investigations, because both the overall pattern and the rate of uplift can be assessed.

Discordant Age Patterns produced by Overprinting

These are commonly observed where igneous or metamorphic rocks which crystallised at time t_1 were reheated at a much later time t_2 but did not undergo complete recrystallisation at this time. The surviving minerals lose variable amounts of the accumulated radiogenic nuclides depending on mineral type and grain size, as well as on the intensity of the metamorphic heating event at time t_2 . A whole range of so-called *overprinted* or *mixed* dates can result, falling between times t_1 and t_2 , or more strictly between the times at which the minerals became closed systems after the earlier and later events. By using the various dating methods on different types of minerals and whole rocks close limits can frequently be set for times t_1 and t_2 . The most retentive and coarsest-grained minerals formed at time t_1 yield the oldest dates, often very close to time t_1 , whilst the least retentive and finest-grained minerals give the youngest dates, often very close to time t_2 . Some of the mineral dates will fall between times t_1 and t_2 and give individually meaningless or so-called *apparent* dates. The actual order of radiogenic nuclide retentivities is, of course, the same as that mentioned previously with reference to discordant age patterns produced by cooling. It must not be forgotten that certain new minerals may be formed at time t_2 , but it is often possible for the field geologist to decide which minerals have survived from the first event and which are the new minerals formed during the second event. A hypothetical overprinted age pattern is shown in Table 1, such as would be expected in the above situation. It corresponds closely with reality, although it is unlikely that so many suitable minerals and rocks would be available from a single geological terrain. In each case the geochronologist has to do the best with the available material. Note in Table 1 how whole-rock Rb-Sr measurements typically give the oldest dates, simply because individual hand-specimens may