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CURRENT PROBLEMS IN RESEARCH

Geochronology

New techniques provide knowledge of the time and mode of formation and subsequent evolution of rock systems.

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Shortly after the discovery of radioactivity it was realized that the decay of a long-lived radioactive element might be used to determine the age of the mineral in which it is found. In the first part of this century Boltwood (*1*) showed that the ratio of lead to uranium tends to have a characteristic value for uranium-rich minerals from any particular district. Since this pioneer work appeared, the science of geochronology has undergone many refinements. With the discovery of isotopes came the application of the mass spectrometer to age determination (*2*). This made it possible to evaluate separately (i) the two uranium-lead ages resulting from the decay of the two isotopes of uranium to lead and (ii) the thorium-lead age. More recent advances include the development of the potassium-argon and rubidium-strontium methods and improvement of the analytical techniques so that common rock-forming minerals with low concentrations of the radioactive parent element may be dated.

It is now possible in many cases to determine the time of crystallization of common rocks with considerable accuracy. Equally important is the ability to recognize, independently of geological criteria, when postcrystallization alteration (metamorphism) has disturbed the age record in a rock. In favorable cases the time of metamorphism, together with the time of earlier crystallization, for a metamorphic rock can be determined. The basic importance of these developments is evident when it is realized that for about 90 percent of geologic time there is no other means of determining absolute ages, or even relative ages except in localized areas. The known fossil sequence is of great value in determining relative ages over the past half billion years but yields no information about still older (Precambrian) rocks. The radiometric methods are making it possible to prepare geochronological maps of the continents that include Precambrian rocks. Existing data indicate some interesting regularities, particularly for North America. When the North American ages are studied it is found that their distribution is not random but that there is grouping both areally and in time. Such data are pertinent to theories of the origin of continents.

There are additional areas in which geochronology is contributing information of fundamental importance. Careful studies of metamorphic rocks are providing clues concerning diffusion of the radiogenic daughter products—argon, strontium, and lead—from minerals. The relative abilities of various minerals to preserve their age record through periods of heating is beginning to be understood. Studies of the temperature dependence of apparent ages in the vicinity of intrusive rocks have yielded estimates of activation energies for diffusion of certain elements in their host minerals. The diffusion studies can be thought of as "natural tracer experiments." Many kinds of tracer experiments involving the decay systems are feasible and lead to information, obtained under natural conditions, that would be difficult or impossible to obtain for elements that are not involved in radioactive decay schemes. Since the ratio of radioactive parent to daughter element varies from mineral to mineral within a rock and from one rock to another, the isotopic composition of the daughter element varies between minerals in a rock and between rocks. Because of this it is possible to detect mixing of daughter elements between minerals or to reach conclusions concerning the source material for some types of rock.

This article deals primarily with data from North America and emphasizes advances in geochronology since 1958. Earlier reviews (*3, 4*) dealt with progress up to the that time. In particular, the earlier reviews summarized the results of studies aimed at comparison of the different methods and illustrated the agreements that can be found. Work since 1958 requires that several points be discussed anew. First, knowledge of the patterns of age distribution on continents has been substantially increased. Second, studies of the response of minerals to metamorphism have, for the most part, been made since the earlier reviews appeared. Likewise, results from many of the natural tracer experiments have been published only recently. These topics form the substance of this discussion. In the space available it is impossible to give the complete background; rather, our object is to present a general picture of the types of information that are being obtained. Admittedly, we have chosen the simplest and clearest examples. We do not consider the improvement in the relationship between the paleontologic and the absolute time scale, as this has been covered recently (*5*).

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Decay Systems and Constants

Mineral age measurements have for the most part been made by the five decay systems given in Table 1. The K^{40}-Ca^{40} system has not been used to any great extent. Calcium-40 is the major isotope of normal calcium, and the potassium-to-calcium ratio is too unfavorable in most minerals to permit accurate determination of the radiogenic component.

To determine an age it is necessary to measure the amount of radioactive parent and the amount of radiogenic daughter that have been produced. The time, t, required for the parent to have produced the daughter is given by the following equation:

$$t = \frac{l}{\lambda} \ln\left(1 + \frac{D}{P}\right)$$

where P is the present-day concentration of parent atoms, D is the concentration of radiogenic daughter atoms, and λ is the decay constant, equal to the natural logarithm of 2 divided by the half-life. This equation is valid for the uranium-lead, thorium-lead, and rubidium-strontium systems. The equation for potassium-argon ages must be modified to allow for the branched decay of K^{40}.

Several conditions must be fulfilled in order for t to represent a meaningful "age." At a distinct time in the past the mineral must have contained no radiogenic daughter element. This might be the time of crystallization or that of complete removal of pre-existing radiogenic daughter element as a result of metamorphism. Since this time the mineral must neither have gained nor have lost parent or radiogenic daughter by processes other than radioactive decay.

Evaluation of these factors has been a major objective of all work in geochronology. Since there are five decay systems, involving different elements in different minerals, it has been possible to make considerable progress in this evaluation. When agreement is obtained between the various decay systems there is little doubt that a significant age has been measured. This is usually, if not always, the time of crystallization of the rock. Conversely, when the ages do not agree, a complex history is indicated.

Table 1. Decay systems used in geochronology.

Parent	Daughter	Type of decay	Decay constant (yr^{-1})
U^{238}	Pb^{206}	$8\alpha + 6\beta$	1.54×10^{-10}
U^{235}	Pb^{207}	$7\alpha + 4\beta$	9.72×10^{-10}
Th^{232}	Pb^{208}	$6\alpha + 4\beta$	4.99×10^{-11}
Rb^{87}	Sr^{87}	β	1.39×10^{-11}*
K^{40}	Ar^{40}	Electron capture	5.85×10^{-11}
K^{40}	Ca^{40}	β	4.72×10^{-10}

* See text for a discussion of the uncertainty.

One other type of age is in common use in such evaluation—the so-called lead-lead age. Since two isotopes of uranium decay to two isotopes of lead, two equations of the type given above can be combined to give

$$\frac{Pb^{207}}{Pb^{206}} = \frac{U^{235}(e^{\lambda(235)t} - 1)}{U^{238}(e^{\lambda(238)t} - 1)}$$

Since the isotopic composition of uranium is essentially constant in nature, t can be calculated from knowledge of the ratio of radiogenic Pb^{207} to radiogenic Pb^{206} in a mineral. This age is less affected than the U^{235}-Pb^{207} or U^{238}-Pb^{206} ages by loss or gain of parent or daughter elements. For example, loss of half the radiogenic lead from a mineral in recent time would drastically lower the uranium-lead ages but would leave the lead-lead age unchanged, provided Pb^{206} and Pb^{207} were lost in the same proportions. It is the practice to give the two uranium-lead ages and the lead-lead age in reporting isotopic lead ages, but only two of the three values are independent ones, since only two decay systems are used.

The decay constants given in Table 1 are not all known with equal accuracy. Decay constants are discussed in an article by Aldrich and Wetherill published in 1958 (*3*), and the situation is essentially unchanged since their work appeared. Best known are the constants for U^{238} and Th^{232}, which have uncertainties of ±1 percent. The greatest problem exists in the case of Rb^{87}, where the uncertainty is at least 5 percent. This is due to the unusual spectrum of the beta particles emitted by Rb^{87}, the high proportion of low-energy particles making it difficult to determine the specific activity. Values for the decay constant as determined by counting experiments have a large spread, and workers are currently divided over which of two values should be used—a geologically determined value of $1.39 \times 10^{-11}\ yr^{-1}$ (*6*) or a value of $1.47 \times 10^{-11}\ yr^{-1}$, determined by liquid scintillation counting (*7*). In this article we use the former.

Concordant Age Results

Table 2 illustrates the agreement that exists between various methods and minerals in favorable cases. In none of these instances is perfect agreement found, since the analytical uncertainty in the age values is less than ±3 percent for practically all values; however, there is no doubt that these age values provide reasonably good age estimates for these particular rocks. In three cases the rocks cited in Table 2 have clearly been involved in orogenic (or mountain-building) processes subsequent to the original crystallization of the rock. The Beartooth Mountain pegmatites were involved in the Laramide orogeny about 55 million years ago during the formation of the Rocky Mountains, and the Bear Mountain and Shenandoah National Park rocks were involved in Appalachian folding about 350 million years ago. It is fortunate that such good agreement can still be found in these cases. Discordant ages are often encountered in rocks that have been involved in orogenies after crystallization.

The data from Llano, Texas, deserve special mention, since the age values listed in Table 2 reflect only a small fraction of the work that has been done in the area. An extensive study of rubidium-strontium and potassium-argon ages has been made by Zartman (*8*), who has obtained approximately 75 results from the various granites and pegmatites. (Data from some of this work are shown in Fig. 4.) With few exceptions the rubidium-strontium and potassium-argon ages agree closely, at around 1050 to 1100 million years. The Llano rocks are probably the most thoroughly studied of any in North America from the standpoint of dating. The agreement of ages shows that these rocks crystallized about 1100 million years ago and that nothing has happened since to seriously alter the age record. Many additional ages of approximately 1100 million years from central and western Texas have been reported by Wasserburg, Wetherill, Silver, and Flawn (*9*).

In favorable cases, then, it is possible to find agreement to within about 10 percent between the rubidium-strontium, potassium-argon, and isotopic lead methods as applied to various minerals in a rock or assemblages of rocks in a particular area. Often these relatively concordant results are found for rocks that are thought to have had rather simple geological histories since their crystallization, although there is still no way of predicting with certainty from geological or petrological evidence when a rock will give concordant age results.

Metamorphic Effects on Mineral Ages

Geologists recognize many instances in which rocks have been altered after their formation, usually by effects of elevated temperature or pressure, or of both together. Hydrothermal alteration may also accompany such an event. Many petrological studies have dealt with the mineral assemblages that result from the various conditions of metamorphism. It is also important to evaluate the influence of metamorphic episodes on mineral ages. One of the first studies of this type dealt with the influence of the metamorphism that accompanied the Appalachian orogenies on the mineral ages of the basement rocks—that is, of rocks that pre-date these orogenies *(10)*. In the Baltimore area, the results clearly demonstrate the differences in response of several minerals to metamorphism. The Baltimore gneiss occurs as domes of basement rock in the Piedmont Province, where metamorphism was relatively intense during the Appalachian revolution. The zircon ages from the gneiss are nearly concordant, at 1100 million years. These results are closely comparable to those at Bear Mountain, New York, and Shenandoah National Park, Virginia, (Table 2) for rocks from the Blue Ridge Province, where Appalachian metamorphism was of relatively low grade. The rubidium-strontium age values of several potassium feldspars from the Baltimore gneiss and associated Hartley Augen-gneiss are also about 1100 million years. These observations are interpreted as showing that the Baltimore gneiss crystallized approximately 1100 million years ago. The biotite age values are only about 300 million years, according to both the rubidium-strontium and the potassium-argon methods; biotite from younger granites intruded during the Appalachian revolution also gives the same age in this area. The biotite age value is thus clearly related to the effects of Appalachian metamorphism. Additional measurements on feldspar and muscovite from the younger granites and pegmatites give ages of 350 to 450 million years, hence the biotite age value represents a late phase in a period of metamorphism that encompassed 150 million years or more.

Other examples of zircon ages greater than biotite ages have been reported in several areas, including eastern Finland *(11)*, southwest Africa *(12)*, Minnesota *(13)*, and New York *(14)*.

While the discordant age patterns in many areas can be attributed to post-crystallization regional metamorphisms, there is no adequate way of relating the degree of discordance to the intensity of the metamorphism. This is because discordant ages may be produced by heating which is too slight to produce any other effect in the rock, such as changes in mineralogy and texture. There are, however, some geologic phenomena which produce relative thermal effects that can be evaluated a priori. The simplest of these would be a sheet of molten rock intruding a homogeneous country rock. This will produce a thermal gradient in the country rock, with time-temperature characteristics that may be calculated from heat-conduction theory.

Given an ideal case of this kind, it should be possible to correlate the effect of heating on mineral ages with the relative intensity of the heating. This is obviously best done for a case where the intrusive is much younger than the country rocks.

A study of this type has been carried out near Eldora, Colorado *(15)*, where Precambrian crystalline rocks about 1300 million years old have been intruded by a large granitic stock of much younger age (54 million years). This stock is roughly brick-shaped and about 10,000 feet wide. Rocks in a zone 20,000 feet wide adjacent to the intrusive contact were studied for changes in geochronologic and mineralogic character. Figure 1 shows the age values for various minerals from this

Table 2. Some nearly concordant ages from North America.

Rock	Mineral *	Age (millions of years)						Reference
		U^{238}/Pb^{206}	U^{235}/Pb^{207}	Pb^{207}/Pb^{206}	Th^{232}/Pb^{208}	Rb^{87}/Sr^{87}	K^{40}/Ar^{40}	
Beartooth Mountains, Montana								
Pegmatite A	U	2600	2640	2700				*(51)*
Pegmatite B	M					2800	2470	*(51)*
Pegmatite B	F					2700		*(51)*
Rainy Lake, Ontario (Rice Bay)								
Gneiss A	Z	2450	2600	2730				*(52)*
Gneiss B	B					2630	2520	*(45)*
Gneiss B	M						2600	*(45)*
Viking Lake, Saskatchewan								
Pegmatite	U	1850	1880	1910	1670			*(53)*
	B					1970	1780	*(16)*
Keystone, S.D.								
Pegmatite	U	1580	1600	1630	1440			*(16)*
	L					1650	1380	*(16)*
	M					1730	1550	*(16)*
Wilberforce, Ontario								
Pegmatite	U	1020	1020	1020	1000			*(16)*
	Z	900	930	1000	990			*(16)*
	B					1030	960	*(16)*
Shenandoah National Park, Va.								
Gneiss	Z	1070	1100	1150	1110			*(54)*
	B					880	800	*(54)*
Pegmatite	H						900	*(20)*
	F					980		*(55)*
Bear Mountain, N.Y.								
Gneiss	Z	1140	1150	1170	1030			*(54)*
	B					880	780	*(54)*
Granite	Z	960	990	1060	850			*(54)*
	B					930	840	*(54)*
Llano, Texas								
Granite	Z	970	1020	1120				*(9)*
	Z	950	990	1070	890			*(16)*
	B					1100	1060	*(16)*
Spruce Pine, N.C.								
Pegmatite	U	385	390	400±50				*(16)*
	U	370	375	420±50				*(16)*
	M					375	335	*(16)*
	F					385		*(16)*

* B, biotite; F, potassium feldspar; H, hornblende; L, lepidolite; M, muscovite; U, uraninite; Z, zircon.

zone as determined by both the potassium-argon and the rubidium-strontium methods. Also shown in Figure 1 is the approximate maximum temperature calculated for a magma, roughly brick-shaped, that was intruded at a temperature of 780°C, with latent heat of crystallization of 80 calories per gram. It is quite apparent that the heating by the intrusive has had a profound effect on most of the mineral ages and that there is a wide variation in the degree to which different minerals are affected.

When we consider first the presumably unaffected rocks at great distance from the contact, we see that even there the various ages are not the same. These age differences are probably due to the complex metamorphic history of this area during the Precambrian. Similar variations in Precambrian rocks have been reported in this Front Range area by other investigators (*16, 17*).

The potassium-argon age for feldspar at 20,000 feet is anomalously young—a phenomenon which is observed for potassium-argon ages for feldspar in nearly all areas (*18, 19*). Even in areas where all other ages are concordant, the feldspar age is almost invariably younger. The irregular behavior of the potassium-argon age for feldspar in the affected zone nearer the contact is probably related to this phenomenon. It is interesting to note that very near the contact this age pattern is reversed and the feldspar age is less affected than either of the biotite ages—a pattern observed in only one other locality in North America (*19*).

The potassium-argon ages for hornblende are the least affected of all the contact-mineral ages, changing significantly only within 100 feet of the contact. This remarkable retentivity prompted investigation of amphiboles as well as pyroxenes in other areas (*20*). The hornblendes in these other areas were not all as retentive as the contact hornblendes, sometimes being only moderately more retentive than biotites.

Both the rubidium-strontium and the potassium-argon ages for biotite in the contact zone show a very regular transition from apparently unaffected ages beyond 10,000 feet to almost completely affected ages near the contact. The rubidium-strontium age is greater than the potassium-argon age at all distances. This is typical of the pattern observed for discordant biotite ages in regional metamorphic areas, though the reverse pattern is occasionally found.

The curves for biotite age in Fig. 1 are for rather coarse-grained flakes (diameter, 5 mm) dominantly from pegmatitic phases of the country rock. A number of finer-grained biotites from country rock schists were also dated and consistently found to give much lower ages, suggesting that grain size has a strong influence in determining the losses of the daughter product. In principle, a low age can be due either to loss of daughter product or to gain of parent element. To demonstrate that the contact results do in fact indicate loss of daughter product and to investigate the influence of grain sizes, a study was made of separate radial zones from individual biotite flakes. A number of flakes of similar size were picked from the 248-foot sample, where age alteration was known to have occurred. The flakes were punched into concentric zones, and the potassium, rubidium, and strontium concentrations were determined for each zone. Figure 2 shows the way in which these elements varied with radial distance within the flakes. The potassium and rubidium are essentially constant throughout. The content of radiogenic strontium decreases from center to edge, showing that the low rubidium-strontium age is indeed due to loss of daughter product and not to gain of parent element. The radiogenic-strontium gradient also shows that diffusion loss is controlled by the actual grain size and not by a sub-grain-size mosaic. In Fig. 2 the rubidium-strontium ages are also compared with the theoretical age gradient which would result if diffusion were radial only (parallel to the cleavage) and if the external concentration of radiogenic strontium were zero. While there is qualitative agreement, the actual age gradient is somewhat flatter than the theoretical, suggesting that these two conditions are not completely satisfied in this case.

The control of diffusion by grain size in biotite shows that only samples of uniform grain size should be used

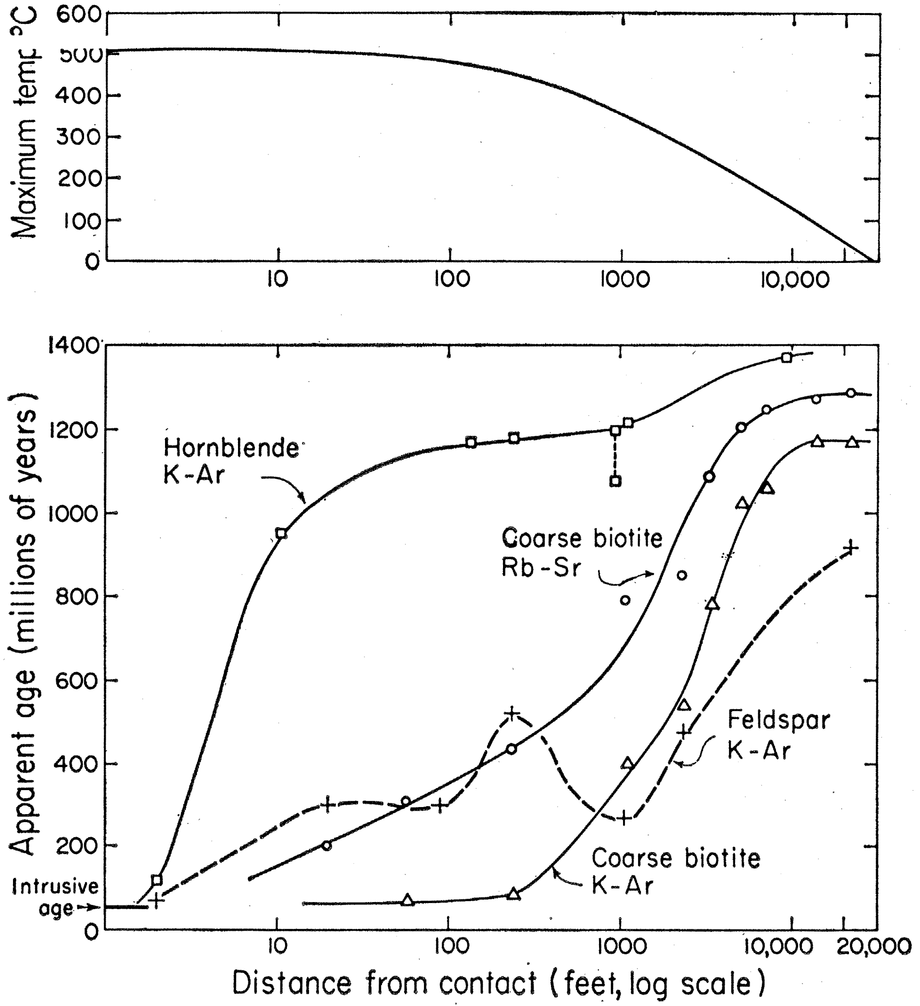
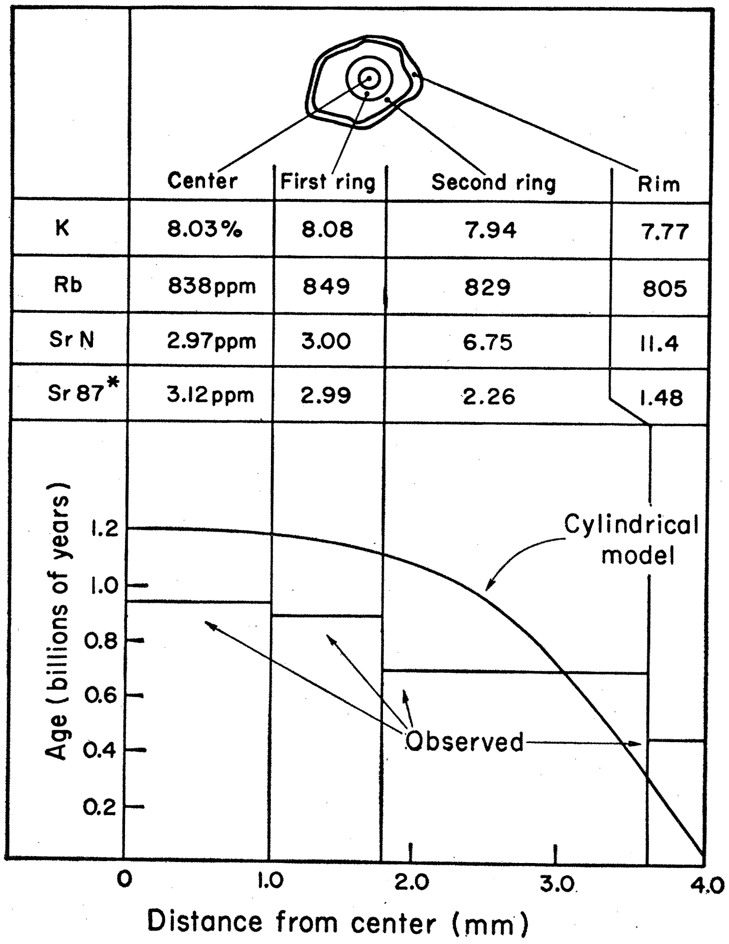


Fig. 1. Variation of apparent ages of minerals as a function of distance from an intrusive contact. Also shown are the theoretical maximum temperatures.

for comparing loss of daughter elements in the contact zone. When grain size is uniform, the age curves can be used to calculate relative diffusion coefficients and activation energies for strontium and argon diffusion. For samples of constant grain size the fractional loss by diffusion will depend on the diffusion coefficient and the amount of heating. The fractional loss is obtained directly from each age value, and the amount of heating is derived from theoretical heat-flow calculations. Since the parameter D/a^2, or diffusion coefficient divided by the square of the crystal radius, occurs throughout the calculations, it is not necessary to know these quantities independently, and it is customary to discuss diffusion in terms of this parameter. In Fig. 3 the reciprocal of maximum heating temperature for a particular sample (the temperature curve in Fig. 1) is plotted against log D/a^2. The points fit linear relationships rather well. According to simple diffusion theory the slopes of these lines are proportional to activation energies, and it is clear that the activation energies for strontium and argon diffusion in biotite are essentially equal. The diffusion coefficients differ by a factor of about 5. This difference alone accounts for the relationship between the curves for the potassium-argon and the rubidium-strontium ages for biotite in the zone of contact.

The role of temperature in producing discordant age patterns is clearly illustrated by these results for the contact zone. The regularities observed can be rather simply interpreted. The heating, however, did not produce large effects in the mineralogy and textures of the rocks beyond the first few feet from the contact. In fact, the only observable petrographic change produced by the heating was a change in the structural state of the potassium feldspar, from triclinic beyond 1000 feet to monoclinic nearer the contact. This lack of change emphasizes the fact that the mineralogy and textures of rocks in other areas may be of only limited help in determining a priori whether mineral ages from those areas will be concordant or discordant. A study of biotite ages from a contact zone in Minnesota shows a transition in ages similar to that at Eldora (*21*).

The contact metamorphism results are closely related to results of other studies that also show differences in response of mineral ages to metamorphic environments. Where a pegmatite cuts another rock, the pegmatite is obviously the younger of the two rock units. A number of workers have found, however, that the mineral age values from a pegmatite and its enclosing rock do not always agree with this sequence. Perhaps the most complete set of results are those obtained at Cutler, Ontario (*22*). Here the rubidium-strontium ages of muscovite and microcline from two pegmatites are each about 1750 million years, but five samples of biotite from surrounding granite and schist have rubidium-strontium and potassium-argon ages of about 1300 million years. The potas-



	Center	First ring	Second ring	Rim
K	8.03%	8.08	7.94	7.77
Rb	838ppm	849	829	805
Sr N	2.97ppm	3.00	6.75	11.4
Sr 87*	3.12ppm	2.99	2.26	1.48

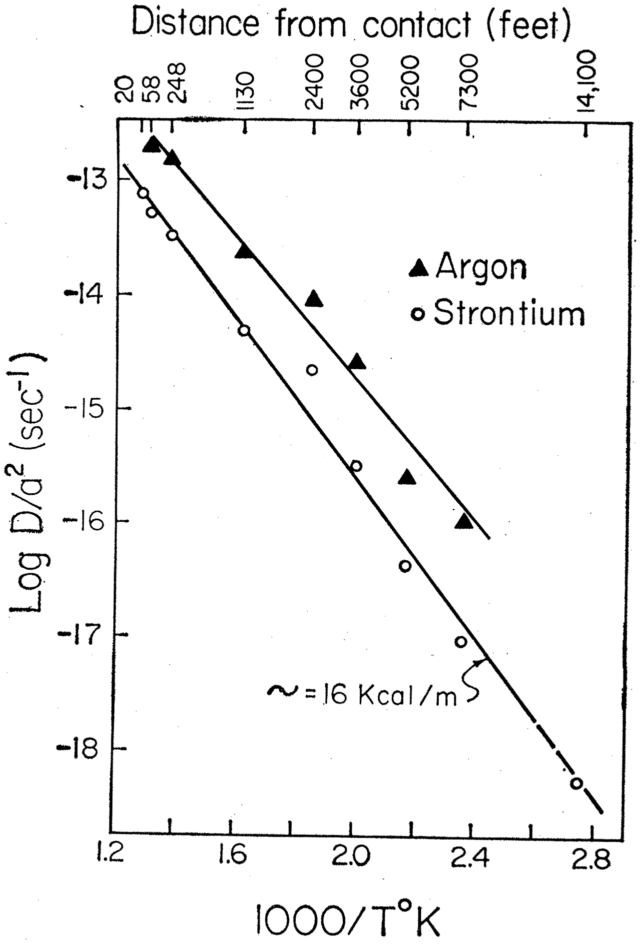


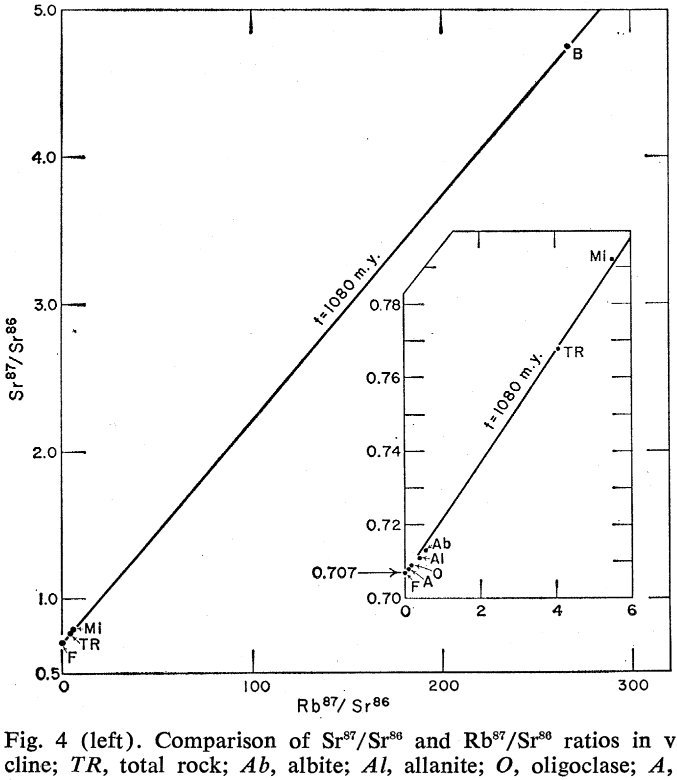
Fig. 2 (left). Variation of apparent rubidium-strontium age in single plates of biotite as a function of distance from the center of the plate. The theoretical curve for diffusion from a cylinder with impermeable ends is shown for comparison. Fig. 3 (right). Variation of the diffusion coefficients of strontium and argon in biotite as a function of the reciprocal temperature given by the theoretical heat-flow model. Minerals are assumed to be spherical crystals of radius *a*.

sium-argon ages for muscovite from the pegmatites are intermediate between 1300 and 1750 million years. Two factors may contribute to the older age values found for the minerals in the pegmatites: radiogenic strontium may be retained better by muscovite and microcline than by biotite under conditions of metamorphism, and the larger crystal size of the minerals in the pegmatites may result in more complete retention of strontium. Muscovite from two of the granites gave rubidium-strontium age values of approximately 1550 million years, suggesting the importance of grain size in the case of this mica. The results from the biotite at Eldora, Colorado, have already demonstrated the role of grain size for that particular sample.

If a mineral such as biotite loses its radiogenic strontium during an episode of metamorphism, the question arises, Where is the strontium relocated? In a large body the loss of strontium from the total rock could conceivably be quite small. Some interesting data illustrate this point. The rubidium-strontium ages mentioned so far have all been calculated on the basis of the assumption that, for the correction for nonradiogenic or common Sr^{87}, the isotopic composition of strontium found in strontium-rich minerals can be used. This composition is essentially constant and agrees well with that observed in young basalts from all over the world (*23, 24*) and in ocean sediments of various ages (*25*). It is customary to correct for nonradiogenic Sr^{87} by using a value of approximately 0.71 for the Sr^{87}/Sr^{86} ratio of common strontium.

Let us consider first a simple case in which all minerals in a rock were formed at a single time and all incorporated common strontium having a Sr^{87}-Sr^{86} ratio of 0.71. If all the minerals have been closed systems with respect to rubidium and strontium since that time, the ratio of radiogenic Sr^{87} to nonradiogenic strontium should be a linear function of the ratio of rubidium to nonradiogenic strontium in the rock or minerals. If, for example, one plots the atom ratios Sr^{87}/Sr^{86} against Rb^{87}/Sr^{86}, a straight line (called an isochron) will result whose slope is a function of the age of the suite of samples. The intercept of the isochron on the *y* axis will be 0.71. An excellent example of this is the work of Zartman (*8*), illustrated in Fig. 4. Zartman has determined the rubidium and strontium concentrations and the isotopic composition of strontium in seven minerals and in the total rock for a granite from the Petrick quarry near Llano, Texas. A plot of these values yields a line with a slope that gives an age of 1080 million years for the suite, and the intercept on the *y*-axis shows that the Sr^{87}-Sr^{86} ratio was initially 0.707 in all the minerals. A number of similar, but less complete, examples can be found in the literature. The work of Schreiner in the Bushveld complex (*26*) is of special significance, since this was the first published work that included an age measurement for the total rock.

The pattern of rubidium-strontium ages to be expected from rocks that have undergone metamorphism has been discussed by a number of authors (*27–29*). If the metamorphism is intense enough, the radiogenic Sr^{87} may be completely mixed with the nonradiogenic strontium in the rock—that is, the isotopic composition of strontium in all minerals is the same after metamorphism. A plot of the values for the rock and the constituent minerals yields a straight line whose slope gives the time of metamorphism, while the intercept on the *y*-axis will indicate the isotopic composition of strontium in all the minerals after metamorphism. The ratio Sr^{87}/Sr^{86} for this intercept will be greater than 0.71, since the strontium now includes radiogenic strontium made in the rock between



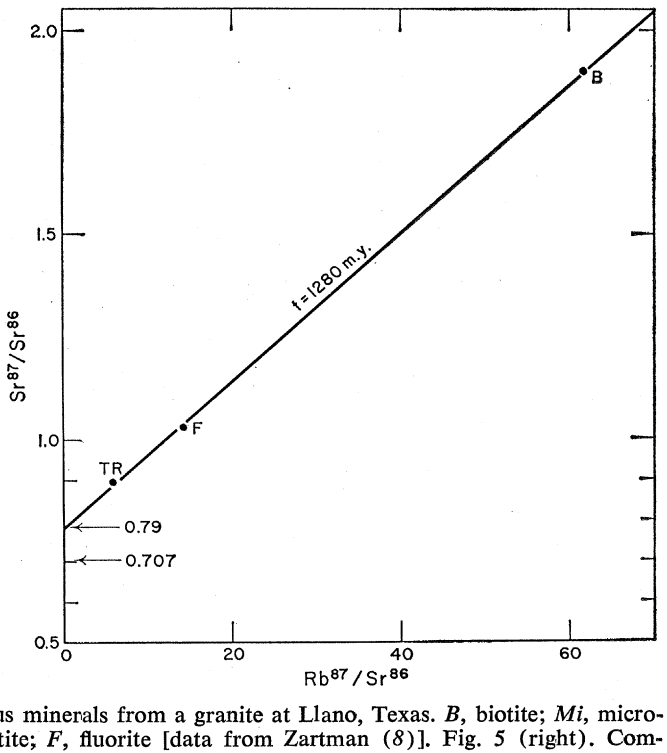
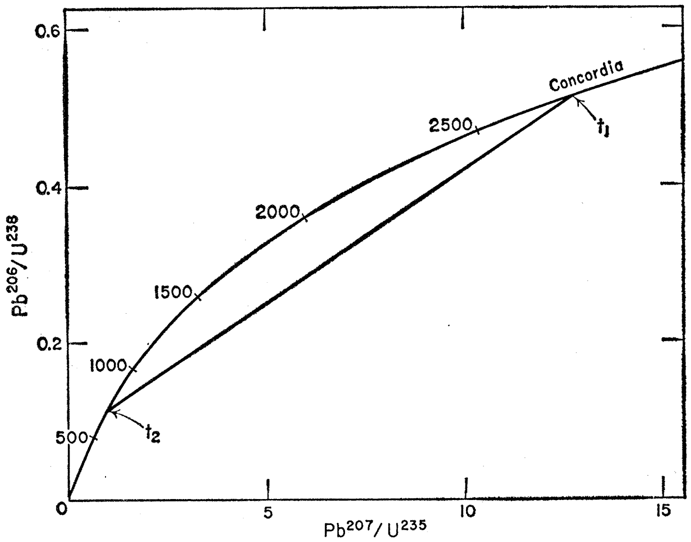


Fig. 4 (left). Comparison of Sr^{87}/Sr^{86} and Rb^{87}/Sr^{86} ratios in various minerals from a granite at Llano, Texas. *B*, biotite; *Mi*, microcline; *TR*, total rock; *Ab*, albite; *Al*, allanite; *O*, oligoclase; *A*, apatite; *F*, fluorite [data from Zartman (*8*)]. Fig. 5 (right). Comparison of Sr^{87}/Sr^{86} and Rb^{87}/Sr^{86} ratios from Creighton granite Sudbury, Ontario, showing redistribution of radiogenic strontium 1280 million years ago [data from Fairbairn, Hurley, and Pinson (*29*)].



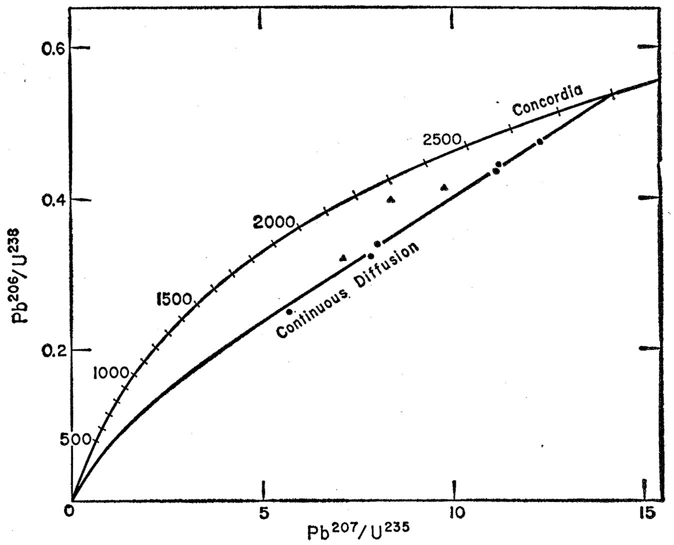


Fig. 6 (left). Concordia diagram showing the effect of a single episode of loss of lead. The curve labeled "Concordia" is the locus of points for which the U^{238}-Pb^{206} age is equal to the U^{235}-Pb^{207} age. Points for samples of age t_1 that lost lead at t_2 will fall on the chord connecting t_1 and t_2. Fig. 7 (right). Parent-daughter ratios for zircons from Karelian basement rocks of eastern Finland compared with the calculated pattern for loss of lead by continuous diffusion, with a constant value for the diffusion coefficient. The diffusion curve is drawn for samples that are 2800 million years old. The triangles represent zircons from mantled gneiss domes that lie to the west of the complex containing the remaining zircons.

the time of first crystallization and metamorphism. Figure 5 shows a set of observations taken from the work of Fairbairn, Hurley, and Pinson (*29*) that fit such a model. These results are from the Creighton granite near Sudbury, Ontario. When the values for the granite and the contained biotite and potassium feldspar are plotted in an isochron diagram, they give an age of 1280 million years and a *y*-axis intercept of 0.79. Biotite ages of 1300 million years have been found by several workers in the Sudbury district (*22, 29*), so this does in fact appear to have been a time of metamorphism in that region.

If the rock as a whole behaved as a closed system for rubidium and strontium while the radiogenic strontium was redistributed, then the rubidium-strontium age for the total rock, calculated on the basis of a value of 0.71 for the ratio Sr^{87}/Sr^{86}, would be the time of crystallization of the rock. This age for the Creighton granite is 2240 ± 100 million years. Fairbairn, Hurley, and Pinson report 11 additional whole-rock ages from the area. If their whole-rock data are plotted in an isochron diagram in the manner of Figs. 4 and 5, data for ten of the 11 samples fit a 2200-million-year isochron quite closely. This is good evidence that rocks did crystallize in the area at this time.

Several other examples of redistribution of strontium during metamorphism are reported in the literature (*27, 30, 31*). The paper of Compston and Jeffery (*27*) was the first convincing demonstration that such redistribution was a possibility. As might have been expected, the more complete studies show that radiogenic strontium is not always completely mixed with nonradiogenic strontium in all minerals during metamorphism. In other instances rocks have not constituted closed systems for rubidium and strontium during redistribution.

Discordant Lead Ages

Nearly all isotopic uranium-lead ages recorded in the literature are discordant—that is, the various isotopic age values differ by amounts that exceed the experimental errors. Except for uraninite it is unusual for the U^{238}-Pb^{206} and the Pb^{207}-Pb^{206} age values to agree to within less than 10 percent. As noted earlier, the ages determined from zircon may agree as well in rocks, such as the Baltimore gneiss, that experienced rather intense metamorphism during the Appalachian revolution as in the rocks from Bear Mountain and Shenandoah National Park where Appalachian metamorphism was of low intensity. This suggests that in some cases, at least, metamorphic intensity—or presumably temperature—is not the principal factor responsible for age discordances. These and related observations have led to a theory in explanation of discordant lead ages that does not rely on an episode of loss of lead during an interval of high temperature. The alternative theory assumes continuous diffusion of lead at a constant rate over the entire time the mineral has existed.

It is customary to examine discordant ages by graphical methods. The one developed by Wetherill (*32*), based on an earlier paper by Ahrens (*33*), is particularly useful; it is illustrated in Fig. 6. The curve labeled "Concordia" is the locus of points for which the age obtained by the U^{235}-Pb^{207} method equals that obtained by the U^{238}-Pb^{206} method—that is, the isotopic ages are concordant. Let us consider first the loss of lead during an episode of metamorphism. In the diagram, points for minerals of age t_1 that lost lead at time t_2 will fall on a chord connecting t_1 and t_2. Their position along the chord will be determined by the fraction of the lead that was lost. Wetherill showed that a suite of five monazites from the Rhodesian Shield in South Africa have age discordances that would result if the minerals are all 2700 million years old and lost different fractions of lead 500 million years ago. This is a logical time for the loss, since other minerals having ages of 400 to 600 million years are found in the area (*34*).

Since Wetherill's papers were written, minerals with apparent ages of

2700 million years have been studied in other regions, notably Finland, Michigan, and Ontario. A remarkable coincidence was found: the discordant ages could be explained if it could be shown that the minerals on all three continents lost varying fractions of lead 500 to 600 million years ago. These results present a problem since there is no evidence of metamorphism 500 million years ago in the areas studied. Biotite ages were 1800 million years in the Finnish and 1300 to 2500 million years in the North American localities.

An attempt has been made to resolve this problem by assuming that loss of lead by diffusion has occurred continuously rather than at a single time during an episode of metamorphism. The continuous-diffusion calculations are based on the following model: (i) the minerals consist of spherical crystals of radius *a*; (ii) uranium is distributed uniformly within the spheres; (iii) diffusion of uranium and intermediate daughter products is negligible as compared with that of lead; (iv) the diffusion coefficient *D* has been constant over all time for a particular mineral; and (v) the diffusion of lead follows Fick's law—that is, it is controlled by concentration gradients. It is then possible to calculate the Pb^{206} and Pb^{207} contents of spheres of age t as a function of the parameter D/a^2, with the further assumptions that no initial radiogenic lead is incorporated in the spheres and that the concentration of lead at the surfaces of the spheres is zero at all times (*35*).

The discordant age patterns of the 2700-million-year-old minerals from Africa, Europe, and North America can all be explained on the basis of this model (*36*). The type of information obtained is illustrated in Fig. 7, which gives all the presently available data for zircons from basement rocks of the Karelian orogenic zone in Finland (*11, 37*). Since the calculated curve for diffusion loss is nearly linear over much of its length, it is apparent that the linear relationship of the points need not imply an episode of lead loss 600 million years ago, and in fact the evidence indicates that metamorphism did not take place in the region at that time. It is known that regional metamorphism did occur 1800 to 1900 million years ago (*11, 38*). The three zircons designated by triangles occur in mantled gneiss domes (*39*) and are the farthest to the west of all samples in the suite. These may indeed have experienced some episodic loss of lead during the regional metamorphism. Alternatively, they could be of somewhat younger age than the remaining zircons. Several additional examples of data that fit the continuous-diffusion hypothesis can be found in an article by Tilton (*36*).

It must be emphasized that loss of lead by continuous diffusion does not appear to be the sole explanation for all discordant ages. There are results from Precambrian rocks in the western United States that do not fit the theory well (*40*). Here it seems necessary in many cases to postulate influence of the Laramide orogeny on the discordances. Relatively few results from these studies have been published, and an attempt at interpretation would be premature. Although continuous diffusion can explain discordant lead ages very well in some instances, other factors related to episodes of metamorphism have affected at least some rocks. The problem is now to decide when, how, and to what extent the various factors have been of importance.

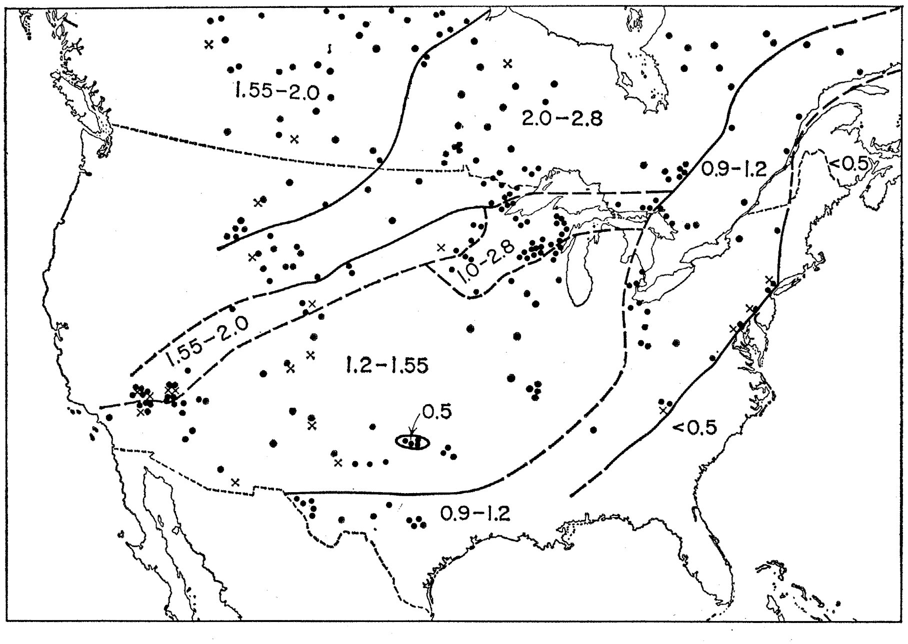


Fig. 8. Distribution of ages in crystalline rocks from the central part of North America. (Circles) Ages that are within the limits specified for a given zone on the map; (crosses) ages that are outside the limits. Age limits are given in billions of years.

Some General Applications

Of major interest is the oldest age value that can be found for the continents, since this is a minimum value for the age of the earth. It has been relatively easy to establish ages of 2500 to 2700 million years for all continents except Antarctica. The indications of ages of 2700 million years are indeed very pronounced on some continents—for example, nearly all of the points in the 2.0- to 2.8-billion-year age zone in Fig. 8 represent ages of 2500 to 2700 million years. It has been much more difficult to extend the record back beyond 2700 million years, but this has been accomplished recently. The oldest ages from Africa are 3000 to 3300 million years (*31, 41*); from North America, 3100 to 3500 million years (*13*); from Europe, 3500 million years (*42*); and from Australia, 3000 million years (*43*). Only a few areas having rocks with 2700-million-year ages have been systematically studied in an effort to find older ages, but it seems unlikely that the age record can be extended much beyond 3500 million years.

The age of the earth is believed to be about 4500 million years (*44*). This indicates a hiatus of 1000 million years or more in the age record. Two explanations seem likely: (i) granitic rocks were not made during the first 1000 to 1500 million years of the earth's history, or (ii) rocks older than 3000 to 3500 million years fail to preserve their age record, presumably because of subsequent metamorphisms. The discussions given here seem to favor the first alternative. Certainly the data are too incomplete for us to attempt a detailed evaluation of the problem, but it is possible that granites and other silicic rocks that have commonly occurred in orogenic belts during the past 2700 million years formed in negligible quantities, if at all, during the first third of the earth's history.

Also of interest is the pattern of ages found for the central portion of the continent of North America. This is illustrated in Fig. 8, on the basis of a survey of values in the literature. In the favorable instances, ages from mica, zircon, and feldspar are in agreement; in the least favorable ones, only potassium-argon and rubidium-strontium ages have been measured on a single mineral. No data from Precambrian rocks were rejected in preparing the figure, but the many Paleozoic ages (200 to 450 million years) have been omitted in the Appalachian chain, and the post-Precambrian ages have been omitted in the Rocky Mountain area in the western United States. For some areas more data exist than can be shown on the map—for example, Minnesota (*45*).

As Fig. 8 shows, the measurements can be grouped by age and geographic location in a manner such that few exceptions are found. In some cases the explanations of the exceptions are obvious. The 300-million-year biotite ages from the Precambrian Baltimore gneiss represent the influence of Appalachian metamorphism. On the other hand, the 500-million-year ages from the Wichita Mountains in Oklahoma (*46*) seem to represent an isolated area of younger rocks in an area containing generally older rocks. The area comprising much of Minnesota, Wisconsin, northern Michigan, and part of Ontario is a "mixed age zone" in which ages similar to those in each of the surrounding zones can be found. The reason for this is not known.

In Fig. 8 the belt of 2000- to 2800-million-year rocks is bounded on both sides by rocks of younger age. In a general way the belts to the southeast become progressively younger toward the Atlantic Ocean. This regularity suggests that the age zones may represent "growth rings" for the continent. The theory that continents have grown throughout geological time has been considered by many authors, most recently by Wilson (*47*). The age zones in Fig. 8 do not in themselves constitute proof of the growth of the continent. Orogenic belts contain recognizable remnants of older rocks, and it is necessary to determine the extent to which an orogenic belt may represent reworked older materials. The two possible extremes are as follows: (i) the younger age zones represent completely new additions to the continent of material from great depth, or (ii) they are completely reworked older material for which the earlier age record has been lost. In the latter case, no addition to the continent has occurred.

Table 3. Frequency of ages in various age groups and relative areal extent of rocks of the same age.

Age class (10^6 yr)	Area (%)	Age (%)
0–500	12	26
900–1200	20	16
1200–1500	33	20
1500–2000	12	22
2000–2800	18	15

A detailed discussion of this problem is beyond the scope of this article, but it should be noted that rubidium and strontium can be used as tracers in obtaining information pertinent to this problem in a manner analogous to their use in dating metamorphic rocks. The general aspects of the method have been discussed by Gast (*23*) and by Hurley, Hughes, Faure, Fairbairn, and Pinson (*48*). These authors give evidence from rubidium and strontium measurements that many granitic rocks in orogenic belts represent additions of new material to continental masses rather than a reworking of older granitic materials. It appears that both new and reworked materials are to be found in orogenic belts. Hurley and co-workers show how, in principle, the proportions of each may be determined. The analysis depends critically on small differences between large numbers, so it will be necessary to have very accurate data before meaningful results can be obtained. It should also be possible to apply uranium and lead data in an analogous way in studying this problem.

The data for Europe, Africa, and Australia indicate more complex patterns of age distribution than those for North America do. It is also apparent that the age distribution for the whole of North America is somewhat more complicated than that for the area covered in Fig. 8. Only the United States and southern Canada are represented in Fig. 8 because these areas have been more thoroughly studied than other parts of the continent; however, many potassium-argon ages have recently been measured in northern Canada by workers at the Geological Survey of Canada (*49*). Their data show a group of 2000- to 2500-million-year ages in the vicinity of Great Slave Lake. This is the only break in the pattern resulting from their work so far.

Since there appears to be a rather regular areal grouping of ages on the North American continent, it is interesting also to consider the distribu-

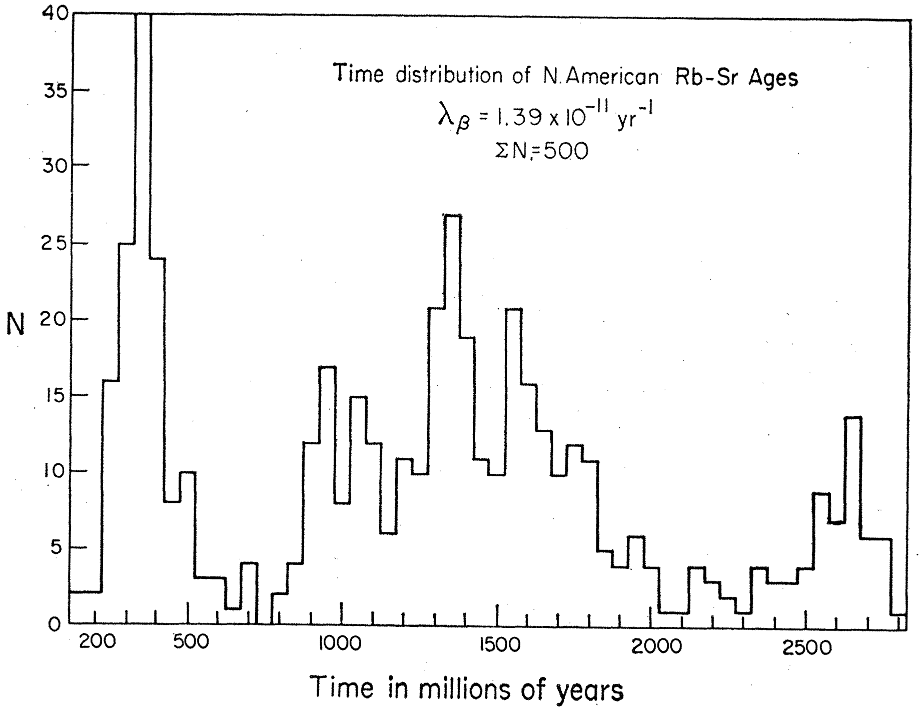


Fig. 9. Time distribution of all published rubidium-strontium ages from the continent of North America.

tion of ages in time, to see if any grouping exists there. Figure 9 is a frequency histogram of all the published rubidium-strontium ages from the North American continent. The published ages are not randomly distributed on an areal basis, and some biasing of the frequency histogram doubtless arises from the high concentration of effort in small areas by some investigators. An estimate of the magnitude of bias can be obtained from Table 3, in which the age frequency within various intervals is compared to the areal extent of rocks of the same age. The bias does not appear to exceed a factor of 2. There appear to be definite age groupings, and some of these occur more frequently than others. The peaks at 350, 1000 to 1800, and 2600 million years are especially distinctive. In addition there is a distinct valley between 500 and 900 million years, where there are relatively few age values. Most of the values that are found in this interval appear to be ages that have been lowered by metamorphism and are not true times of crystallization or metamorphism. Age values are related to igneous and metamorphic events which accompany orogeny, and the data of Fig. 9 suggest a somewhat episodic nature for orogeny on the North American continent. It is of course possible that additional data will alter this picture, indicating a more continuous spectrum of orogeny, but it seems unlikely that the quiescent periods from 500 to 900 and from 2000 to 2300 million years will be filled.

Conclusion

In summary, it now seems possible to obtain accurate ages for many of the common rock types. When rocks have been affected by metamorphic episodes after the time of crystallization, it is possible to recognize this from the age values and, in some cases, to date both the time of crystallization and the time of metamorphism. Work is under way that will make it possible to draw accurate geochronological maps of the continents. An impressive regularity is already evident in the map for North America. Knowledge of the temporal and areal distribution of ages should have an important bearing on theories of the origin of continents and of mountain-building processes (*50*).

Note added in proof: A recent paper by A. E. J. Engel (*56*) has discussed the bearing of geological and geochronological data on the evolution of the North American continent.

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