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Isotopic tales of ancient continents[1]

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Abstract

One of the features which makes the planet Earth a strange anomaly in the Solar System is the presence of continental crust, unknown on the other terrestrial-type planets. In contrast to the basaltic crust underlying the ocean basins, which is nowhere older than 200 Ma (million years) because of its continuous recycling through the mantle by plate-tectonic processes, the continental crust (with the average composition of diorite) is composed of rocks with ages ranging from zero to ~3,900 Ma. It is thus on the continents that 95% of the Earth's geologic record is to be found, including the origin of the unique geosphere-biosphere system. The mainstream of present-day opinion is that these ancient entities were derived from the mantle through magmatic processes, but conflicting views prevail with regard to their growth rates. In this lecture the view is taken that the continents have grown through geologic time by irreversible differentiation from the mantle. The task of gaining insight into the evolution of the continental crust has fallen mainly to the isotope geochemists. Application of radiogenic isotopes (Pb, Hf, Nd, Sr) provides reliable constraints on the age and temporal evolution of rock units, on the time of primary separation of continental material from the mantle, and on the assessment of the relative contributions of mantle and ancient continental crust to magma genesis.

Recorded geologic history begins at ~3,900 Ma with the oldest preserved continental crust. There is evidence that (some) continental crust was already in existence by ~4,300 Ma ago. Most of the insight into the pre-recorded history of the Earth stems from studies of meteorites and other planets. One of the results of planetary exploration is the recognition of large-scale impact cratering by giant meteorites as a major, possibly even dominant geologic process in the early evolution of the Earth. The termination of the 'great bombardment' coincided approximately with the formation of the oldest preserved continental crust by ~3,900 Ma ago. Another major factor to be taken into account in the reconstruction of the early history of the Earth, is the much higher radiogenic heat production than today. Between 3,000 and 2,500 Ma ago, the Earth acquired its modern appearance with the formation of huge volumes of juvenile continental crust from the mantle, and 50–80% of the present-day continental mass was in existence by ~2,500 Ma ago. This fundamental change in the Earth's character marks the Archaean-Proterozoic transition. The growth of continental crust was not continuous through geologic time, but episodic, with five distinct periods of accelerated growth.

After two successful decades of concentrated ocean-floor exploration, the target of international earth science in the 1980s has become the continents.

The odd planet

If a group of scientists from another world explored the inner planets of our Solar System, they would be struck by a number of features at the surface of the Earth that makes this planet a strange anomaly. The first startling observaion of our imaginary visitors would be the highly exotic chemistry of the

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Earth's atmosphere, which at first sight would appear to be a curious and improbable mixture of gases that seems to violate the rules of chemical equilibrium. Further study should reveal, however, that the gaseous envelope of the Earth has to be understood as a complex entity involving not only the atmosphere, but also the oceans, the surface of the continents and – a feature unique to the Earth – the biosphere. It is this presence of life and its continuous interaction with its environment, sustained by the constant energy flux from the sun, that makes the Earth's atmosphere and the physico-chemical reactions at the Earth's surface an oddity in the Solar System.

Another feature unique to the Earth which would puzzle our imaginary extramundane explorers is the conspicuous topographic dichotomy of the surface, reflecting the fundamental difference between continents with a mean elevation of 126 m above mean sea level, and ocean basins with a mean depth of 3.8 km below that level. The crust underlying the ocean basins has a relatively uniform basaltic composition, which is also a widespread crustal component on the Moon, Mars, Venus and Mercury, the other so-called terrestrial planets. Terrestrial-type continental crust, on the other hand, does not exist on the other planets and is a feature unique to the surface of the Earth. Why and how did this uniqueness of our planet come about? The answer to these questions is also of direct interest to the understanding of the complex processes that make up the geosphere-biosphere system, as the operation of this system is strongly influenced by the continental masses. Moreover, the emerged continental crust provided a suitable environment for the later stages of the evolution of life. Without the existence of continents, the evolution of life would have been restricted to the oceans and volcanic islands, and would have taken another course. For example, there would then have been no platform for the appearance of the human race.

The continental crust

The total area of continental crust, including the submerged continental shelves, occupies some 40% of the Earth's surface. It has an average thickness of about 40 km and a highly variable chemical composition. For the assemblage of rocks that constitutes the upper 10 km or so, rather good estimates of the average composition are nowadays available based on large-scale sampling programmes of crustal rocks and analysis of clastic sediments, which monitor the exposed continental crust from which the sediments were derived ('natural composite samples'). On the basis of these data it is now generally agreed that the upper continental crust has the average composition of granodiorite (Fig. 1). For the lower continental crust, the only available samples are xenoliths in volcanic pipes and outcrops of granulitic terrains. The lower continental crust therefore remains an enigmatic region, but all evidence indicates that it is more basic in composition than the upper crust. Estimates of the average composition of the lower continental crust and, consequently, of the bulk (upper + lower) continental crust are model dependent. I consider the best available assessment at present to be that of Taylor & McLennan (1985), who estimate an average composition approaching that of high-alumina basalt for the lower and that of diorite for the bulk continental crust (Fig. 1). Continental rocks have ages ranging from zero to 3,900 Ma. In contrast, the basaltic crust underlying the ocean basins is nowhere older than 200 Ma. This young age indicates that oceanic crust exists for no more than 200 Ma before being returned to the mantle through subduction – the present-day 'surface roll-over time' of the ocean floor (Elder, 1976). Only those minor portions of oceanic crust which are incorporated into the continents by tectonic or magmatic processes, escape this recycling and are preserved. The continental masses, on the other hand, preserve a geologic record of some 3,900 Ma, meaning about 95% of the total history of the Earth. Insight into the evolution of the Earth through geologic time can thus exclusively be gained from the continental crust.

After the intense exploration of the ocean floor during the 1960s and 1970s, international efforts in the earth sciences are now concentrating once again on the continents. However, compared to the ocean floor, the continents are extremely complex

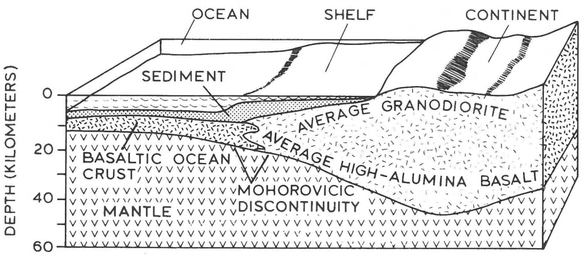

Fig. 1. Schematic representation of basaltic oceanic crust (density 3.0 and about 6 km thick) and continental crust (mean density 2.8 and average thickness about 40 km). The continental crust is divided into an upper crust (enriched in Large Ion Lithophile, LIL, elements, down to a depth of about 10 km), and a lower crust (LIL depleted). (Modified after Moorbath, 1977a.) Following Taylor & McLennan (1985), some key abundances and ratios for the continental crust are:

	Upper crust	*Lower crust*	*Bulk continental crust*
SiO_2 (%)	66.0	54.4	57.3
Al_2O_3 (%)	15.2	16.1	15.9
Na (%)	2.89	2.08	2.30
K (%)	2.8	0.28	0.91
U (ppm)	2.8	0.28	0.91
Th (ppm)	10.7	1.06	3.5
Rb/Sr	0.32	0.023	0.12
Sm/Nd	0.17	0.25	0.22

and the preserved continental record becomes ever more biased and distorted with increasing age. Ancient continental masses have gone through several periods of orogenesis, in contrast to today's oceanic crust which has been through only one event, the formation at its origin in the spreading ridges. This blurring of the older geologic record has resulted in profound disagreement on the older history of the Earth and the evolution of the continents. These subjects are today in the spotlight of the earth sciences, with many controversial opinions. The relevant data in this debate come from a wide variety of field and laboratory investigations in ancient terrains, from geophysical and experimental petrological studies, from studies of the Earth's thermal regime in present and past, from astronomical considerations, and from observations on other planets. Integration of all these different approaches is necessary to reach an understanding of the formation and evolution of the continental crust.

The continental controversy

If we leave aside a number of eccentric hypotheses such as that interpreting the continents as rim debris of huge meteoritic impact sites on the early Earth, now present as ocean basins (Harrison, 1960; Gilvarry, 1961), or the hypothesis that the continental crust represents meteoritic material that has accreted during the latest stages of the Earth's formation (Donn et al., 1965), the mainstream of present-day opinion is that both the oceanic crust and the continental crust have been generated from the mantle by processes of partial melting and magmatic differentiation. Many questions still remain, however. For example, does upper mantle undergo *in situ* partial melting to provide a source of continental magmas, or does its contribution to continental magmatism come via the more complex route of subduction of oceanic crust (altered by reaction with sea-water, metamorphosed during the subduction, and associated with pelagic sediments)? The task of answering these questions has mainly fallen to experimental petrology and trace element and isotope geochemistry, but whatever the actual mechanisms, all continental crustal material is thought to be ultimately mantle-derived. Since the entire continental crust comprises no more than about 0.5% of the total mass of the mantle, the extraction of this minor portion brought about only minimal changes to the major element composition of the residual mantle. This in contrast to the much higher concentrations of incompatible elements (e.g., K, Rb, U, Th, REE) in the continental crust relative to the mantle, which indicate that partial melting and crystal-liquid fractionation are the principal factors for the derivation of continental crust from the mantle. Various mass-balance considerations of incompatible elements indicate that about one-third to one-half of the mantle has been involved in the generation of the entire continental crust (e.g., Taylor & McLennan, 1985).

A wide variety of opinions has been expressed as to the growth of the continental crust. One extreme, the *continental steady state* or *continental recycling model,* postulates that the differentiation of the entire continental mass was essentially completed very early in the Earth's history and that the continental mass has been recycled through the mantle ever since, without net change of the total amount (e.g., Hargraves, 1976; Fyfe, 1978; Armstrong, 1981). The fundamentally contrasted view, the *continental growth model,* is that the differentiation of the continental crust from the mantle is a process that proceeds through geologic time, starting with the development of small, localized patches early in the Earth's history (e.g., Moorbath, 1977a, 1977b, 1982; De Paolo, 1980; Moorbath & Taylor, 1981, 1984). In the recycling model continental material can be returned and efficiently remixed with the mantle in bulk, whereas in the growth model continental crust, once formed, is cratonized and, because of its lower density, cannot be returned and remixed with the mantle on a large scale. Obviously, some combination of these extremes, involving both continental growth and continental recycling, is also possible. These conflicting models with regard to the continental growth rate form one of the greatest current debates in the earth sciences. There is no doubt that at least some minor recycling of continental crustal material through the mantle takes place via the subduction of sediments of continental origin. For example, the occurrence of the short-lived radioactive ^{10}Be isotope (half-life 1.5 Ma, produced by cosmic rays in the upper atmosphere) in geologically young island arc volcanoes, proves that ocean floor sediments are subducted and incorporated, to some extent, in the upper-mantle source regions of subduction-related magmas. However, the overall weight of evidence indicates that recycling of continental crustal material in island-arc environments and calc-alkaline magmatism occurs only sporadically, and is of very restricted significance. For example, the current sedimentation rates appear to be far from sufficient to supply enough continental material to subduction zones, to sustain a recycling of continental crust through the mantle in bulk (Taylor & McLennan, 1985). Thus, the present-day regimes on Earth do not support a continental steady-state model, but can the present-day regimes be extrapolated back through time?

In the 'growth-versus-recycling' debate, data provided by geochemical, geophysical and sedimentological studies are all of key importance, but the leading part is being played by the study of radiogenic isotopes.

Radiogenic isotopes

Radiogenic isotopes have become the most powerful tool for elucidating the evolution of the continental crust. This is firstly because radiogenic isotopes put constraints on the time of formation of each sector of the crust (which provides, of course, the most unambiguous answer to the question if, and at what rate, the continental crust has grown through geologic time). Secondly, because radiogenic isotopes in continental magmas can be used as 'fingerprints' to distinguish between magmas derived from a mantle-like source or from ancient continental crust.

For isotopic dating the most direct way is to use the conventional methods (Table 1). These straightforwardly provide the formation ages of rocks, including, in many cases, the primary formation age of rocks that have been reworked by younger events. In this way an areal distribution of continental crust with regard to its age is obtained (Fig. 2). From this age distribution it is clear that rather more than half of the present continental crust was formed in Precambrian time. The areas which were cratonized more than 600 Ma ago, are designated as Precambrian shields. The actual pro-

Table 1. Long-lived radioactive decay systems applied in isotope geology.

			Half-life, 10^9 a
^{238}U	$6\beta^-$	^{206}Pb (+ 8^{4}He)	4.467
^{235}U	$4\beta^-$	^{207}Pb (+ 7^{4}He)	0.704
^{232}Th	$4\beta^-$	^{208}Pb (+ 6^{4}He)	14.01
^{176}Lu	β^-	^{176}Hf	35.7
^{147}Sm		^{143}Nd (+ ^{4}He)	106
^{87}Rb	β^-	^{87}Sr	48.8
^{40}K	el. capture	^{40}Ar (10.48%)	1.25
	β^-	^{40}Ca (89.52%)	

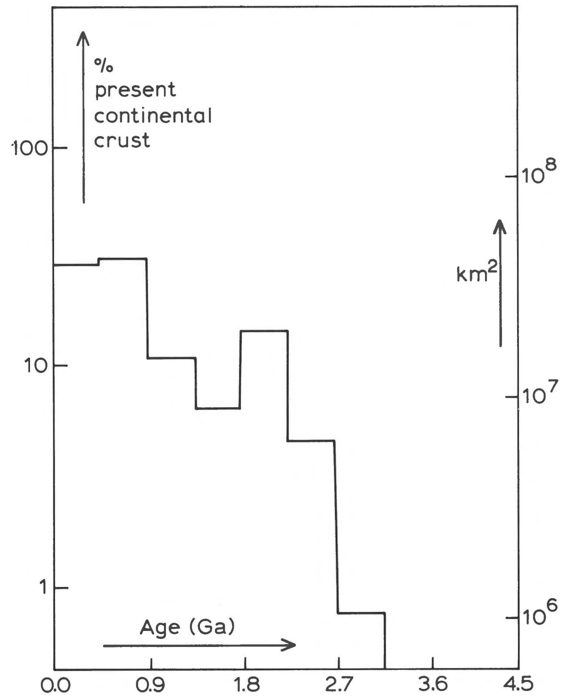

Fig. 2. Present-day areal distribution of the continental crust as a function of its geologic age (modified after Windley, 1986).

portion of Precambrian crust on the continents must be higher, however, as there are many small remnants of Precambrian rocks in Phanerozoic belts, while virtually all orogenic belts of any age contain older material that has been reworked beyond recognition.

During the last decade, a number of indirect dating methods has been developed which enable us to bring to light the hidden age record of continental rocks. Two methods have become particularly important:

- The U-Pb dating of detrital and xenocrystic zircons in sedimentary and igneous rocks. Zircons have proven to be extremely resistant with regard to their U-Pb systems to processes of weathering, sedimentation, metamorphism and magmatism. An inherited zircon in a granite can therefore have retained the precursor age, even when present as older cores within a younger generation. For example, our investigation in Amsterdam of suites of zircons from gneissified ~470 Ma old granites in NW Spain revealed discordia plots indicating that two zircon generations are present: a ~2,500 Ma old generation in cores, and a main ~470 Ma old generation (Kuyper et al., 1982; Priem & den Tex, 1984). This proves that a ~2,500 Ma old continental crustal component has been involved in the formation of the continental crust of NW Spain, although no rocks of this age are anymore present. The zircon-dating approach has a great potential in studies of the evolution of continental rocks, particularly with the recent developments in ion-microprobe techniques in Canberra which permit the dating of individual zircons and even different portions of a single zircon crystal.
- The Sm-Nd model-age dating of sedimentary, metamorphic and igneous rocks. This method is based on the close coherence of Sm and Nd (both REE), so that geochemical processes bring about virtually no fractionation between both elements. The most simple model is to assume that the isotopic evolution of Nd in the mantle took place in a uniform reservoir (CHUR) whose Sm/Nd ratio is equal to that of chondritic meteorites (Fig. 4). As a rule, the Sm/Nd ratio and Sm-Nd systematics are not substantially disturbed under conditions of erosion, sedimentation and metamorphism. This makes it possible to determine for any rock the 'mean crustal residence time', i.e. the time-span that the Sm-Nd system in the rock has resided in the crust since its separation from the mantle. The actual formation age of the rock, as it is now exposed to us, can be much younger. For example, in the granitic province of NW Spain and northern Portugal we obtained in Amsterdam Rb-Sr whole-rock data which indicate that the granites were emplaced during two distinct episodes of magmatism, one ~470 Ma ago and the other between 330 Ma and 280 Ma ago (Priem & den Tex, 1984; Priem et al., 1984b; Amsterdam, unpublished data), but the Sm-Nd systematics indicate model ages of about 1,400–1,600 Ma (Tung Chooi Liew, unpublished data Max-Planck-Institut für Chemie, Mainz). The granitic magmas have thus been generated from much older protoliths, in accordance with zircon U-Pb

data, and the crustal material forming this continental segment has been derived from the mantle at least some 1,000 Ma before the formation of the now exposed Paleozoic crust. Such 'times of addition to continents' of new, mantle-derived crustal material are not only traceable for (meta)igneous rocks, but also for (meta)sedimentary rocks. The 'mean crustal residence time' applies then to the exposed continental crustal area from which the sediments were derived (McCulloch & Wasserburg, 1978).

Isotopic fingerprinting of a magma is based upon the radiogenic isotope signature, the initial isotopic composition of Sr, Nd and Pb. The mantle has characteristic time-dependent arrays of $^{87}Sr/^{86}Sr$ (Fig. 3), $^{143}Nd/^{144}Nd$ (Fig. 4) and $^{208}Pb/^{207}Pb/^{206}Pb/^{204}Pb$ (Fig. 5). These arrays differ significantly from the isotopic compositions of Sr, Nd and Pb which have resided for a substantial time in a continental environment, because of the greatly different ratios of Rb/Sr, Sm/Nd, U/Pb and Th/Pb. Magmas inherit the radiogenic isotope signatures of their source and this signature remains unaffected by later igneous differentiation processes (e.g., Stosch et al., 1980; Allègre, 1982; Anderson, 1982; White & Hofmann, 1982). Granitic rocks and related extrusives which originate by partial melting of much older continental material, have therefore initial radiogenic isotope signatures which strongly contrast those of mantle-derived rocks. For example, the Rb-Sr investigations in Amsterdam of Hercynian granites in NW Spain and northern Portugal yield initial $^{87}Sr/^{86}Sr$ ratios between 0.709 and 0.731 (Priem & den Tex, 1984; Priem et al., 1984b; Amsterdam, unpublished data), versus mantle-type ratios of 0.703–0.704, which confirms the U-Pb zircon and Sm-Nd evidence that these granites were derived from ancient continental crust. In this way it is thus possible to distinguish between reworked ancient continental crust, and juvenile continental crust.

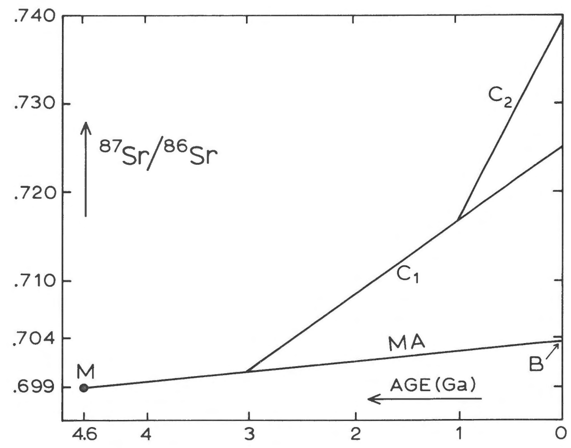

Fig. 3. Increase in $^{87}Sr/^{86}Sr$ through geologic time. M, initial $^{87}Sr/^{86}Sr$ in meteorites, 0.698990 ± 0.000047 (BABI, Basaltic Achondrite Best Initial, Papanastassiou & Wasserburg, 1969) and in lunar samples; B, average composition of modern ocean basalts, 0.704 ± 0.002, taken to represent the average composition of the upper mantle (but significant variations within this narrow interval signal small isotopic heterogeneities); MA, average upper-mantle growth line, assuming a constant Rb/Sr of 0.027. In the simplified example shown here juvenile continental crust with higher Rb/Sr was derived from mantle material 3.0 Ga ago and produced growth line C_1, beginning with $^{87}Sr/^{86}Sr = 0.701$. Granitic magma, again with higher Rb/Sr than its source, was generated from the juvenile continental crust 1.0 Ga ago and produced growth line C_2, beginning with $^{87}Sr/^{86}Sr = 0.717$. Effects of Rb depletion in the residual rock systems are neglected. In this example, the initial Sr isotopic composition of continental crust generated from ancient continental crust contrasts markedly with that generated from mantle-material or mantle-derived basalts.

However, at this juncture it should be noticed that a new segment of continental crust formed by differentiation of mantle material, begins its isotopic evolution with a radiogenic isotope signature identical to that of the mantle-source from which it was derived. If the time-interval between the generation of juvenile continental crust from the mantle and subsequent granite production by partial melting from this new continental material is relatively short (<100–200 Ma), insufficient time will have elapsed for the new continental segment to develop distinctive radiogenic isotope characteristics. In that case, reworked older continental crust is not distinguishable from primary mantle-derived crust exclusively on the basis of the radiogenic isotope signature. This limits the possibility of making unequivocal statements about magma provenance.

It should be mentioned that some students of the Earth (e.g., Collerson & Fryer, 1978; Hart et al., 1981) take the view that during intense reworking

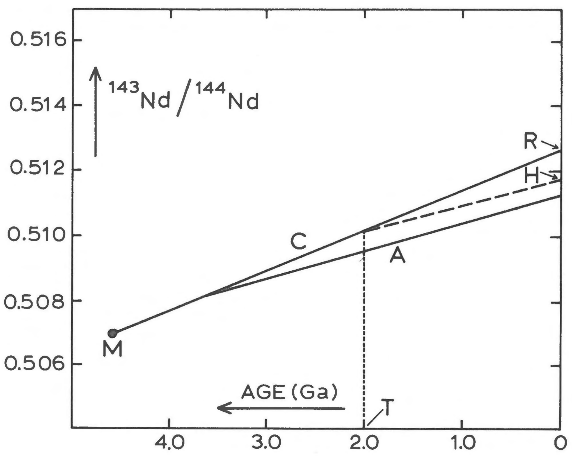

Fig. 4. The increase in $^{143}Nd/^{144}Nd$ in the mantle through geologic time is expressed in terms of the CHUR (CHondritic Uniform Reservoir) model (De Paolo & Wasserburg, 1976). The CHUR growth line (C) is based upon the present-day $^{143}Nd/^{144}Nd$ of 0.512638 (R) and Sm/Nd of 0.308. M, initial $^{143}Nd/^{144}Nd$ = 0.50684. (Both $^{143}Nd/^{144}Nd$ ratios are normalized to $^{146}Nd/^{144}Nd$ = 0.7219). Continental crust formed by partial melting of mantle material has a lower Sm/Nd than CHUR (leaving the residual, depleted mantle with a higher Sm/Nd), so that the growth lines of continental rocks are less steep than that of CHUR. As an example the average growth line A of the 3.8 Ga old Amîtsoq orthogneisses in West Greenland is shown (Sm/Nd ratio after O'Nions & Pankhurst, 1978). For a rock with present-day $^{143}Nd/^{144}Nd$ = H and Sm/Nd corresponding to the slope of the broken line, the intersection of this line with CHUR gives the 'Sm/Nd model age' or 'crustal residence age' T relative to CHUR (T^{Nd}_{CHUR}) of that rock.

(high-grade metamorphism, partial melting) of continental crust, all radiogenic isotope history can be completely wiped out by processes such as large-scale isotopic rehomogenization of daughter elements (Sr, Nd, Pb), purging of the (incompatible) parent elements (Rb, Sm, Th, U) and/or daughter elements, and percolation by mantle-derived fluids. In this way it would be possible, for example, that a granitic magma generated by partial melting from ancient continental crust, acquires a mantle-type radiogenic isotope signature. However, I adopt the view of those isotope geologists (e.g. Moorbath & Taylor, 1984) who consider that such claims of complete eradication of all radiogenic isotope evidence of the crustal residence time, have to be treated with extreme scepticism. Although

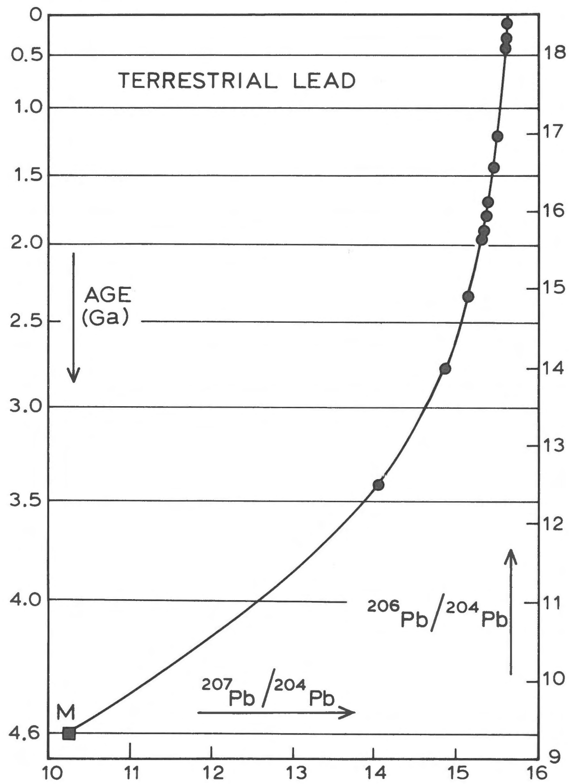

Fig. 5. Variation of the isotopic composition of mantle lead ($^{206}Pb/^{204}Pb$ versus $^{207}Pb/^{204}Pb$) through geologic time. M (square), troilitic lead from the Canyon Diablo meteorite. The other data-points (circles) represent the isotopic Pb compositions of a number of major conformable galena deposits of different ages, thought to have been derived from the upper mantle and emplaced without contamination of continental-crust lead. From the top downwards the following deposits are plotted: Hall's Peak, N.S.W., Australia; Bathurst, N.B., Canada; Cobar, N.S.W., Australia; Balmat, N.Y., U.S.A.; Sullivan, B.C., Canada; Mt. Isa, Queensland, Australia; Broken Hill, N.S.W., Australia; S.W. Finland, including Orijärvi; Sudbury, Ontario, Canada; Geneva Lake, Ontario, Canada; Barberton (Daylight Mine), South Africa (all data from Russell, 1972). These data-points lie close to the conventional Pb growth curve according to the single-stage Holmes-Houtermans model (μ value about 8.0). There are, however, to a greater or lesser extent discrepancies between the ages of the ore deposition obtained by other means and the corresponding calculated model ages. This and the fact that several other major conformable galena deposits do not fit to this curve, indicate that the isotopic evolution of conformable ore-leads is not (wholly) consistent with such a simple single-stage model. Nevertheless, the growth curve may be taken to sketch the broad features of the isotopic evolution of mantle lead.

the radiogenic isotope signature can be disturbed to a high degree, a rock will always maintain (part of) the radiogenic isotope memory of its source region. On the other hand, if the radiogenic isotopes do not betray any older crustal history, it can confidently be concluded that the rock is mantle-derived or has been derived from a continental source with a crustal residence time of at most some 100–200 Ma. For example, in the Arabo-Nubian shield of the Sinai peninsula, Rb-Sr whole-rock and mineral investigations in Jerusalem (Bielski, 1982) and U-Pb zircon investigations in Amsterdam (Priem et al., 1984a; Amsterdam, unpublished data) have demonstrated a Late Precambrian evolutionary history from about 800 to 580 Ma ago. Neither the initial Sr signatures (mantle-type ratios of 0.7030–0.7045 for the oldest, volcanic sequences), nor the zircon U-Pb systematics betray any older history. On the basis of this and other evidence it is concluded that the Sinai craton represents juvenile continental crust. Its formation started about 800 Ma ago with the deposition of volcano-sedimentary sequences, probably on oceanic crust, and it became cratonized about 580 Ma ago.

Date of birth

Radiogenic isotopes tell us thus about the time of formation of continental segments, and whether or not they were derived from ancient continental crust. How does the beginning of continental evolution relate to the early history of our planet? The Precambrian geologic record begins about 3,900 Ma ago with the oldest dated terrestrial rocks. No direct geologic records of the preceding period of the Earth's history have been preserved, but older rocks are available from extra-terrestrial bodies such as meteorites and the Moon. An impressive number of isotopic age determinations has shown that most meteorites were formed 4,550 ± 100 Ma ago. The same age has also been determined for the oldest rocks and soils from the Moon. It is not self-evident that ages of meteorites and lunar samples should define the age of the Solar System, but here again radiogenic isotopes provide the clue. The isotopic compositions of a number of elements indicate that in some meteorites, extinct radionuclides of short half-life were present during their formation (Table 2). This implies that the time-interval between the end of nucleosynthesis in our part of the galaxy and the formation of the parent bodies of the meteorites, including the differentiation into metallic and silicate phases and the subsequent consolidation and cooling, must have been relatively short – the very short half-life of ^{26}Al (0.76 Ma) indicates that at most a few million years have been involved. Meteorite ages can thus indeed be taken to approximate most nearly the age of the accretion of the Solar System from the ancestral solar nebula.

Excesses in ^{129}Xe and fissiogenic xenon, derived from the extinct radionuclides ^{129}I and ^{244}Pu, respectively, have also been reported from mid-ocean ridge basalts (Staudacher & Allègre, 1982) and are taken to indicate that the formation, differentiation and consolidation of the Earth was contemporaneous within very narrow limits with that of the parental bodies of meteorites. The same conclusion is reached on the basis of the evolution of the isotopic composition of lead in the Earth's mantle.The measured growth through geologic time in the Pb isotope ratios ^{206}Pb/^{204}Pb and ^{207}Pb/^{204}Pb due to the radioactive decay of the U isotopes ^{238}U and ^{235}U (Table 1) approximates a simple curve, indicating that about 4,550 Ma ago the isotopic abundance ratios of mantle lead were identical to those in the parental bodies of the meteorites (Fig. 5). Students of the Earth therefore now agree that the Earth, along with the entire Solar System, was produced in something like its present shape about 4,550 Ma ago.

Table 2. Radiogenic or fissiogenic isotopes in meteorites produced by extinct radionuclides of short half-life.

Radiogenic/ fissiogenic isotope	extinct parent	Half-life 10^6 a	Decay or fission
^{129}Xe	^{129}I	15.7	Decay
$^{131-136}$Xe	^{244}Pu	82	Fission
^{107}Ag	^{107}Pd	6.5	Decay
^{26}Mg	^{26}Al	0.76	Decay

The turbulent beginning

It is now a generally accepted scenario that the story of the Earth began with gravitational accretion from the ancestral solar nebula. This process released enough heat (estimated at 38.10^3 J/g, which would have amounted for the present mass of the Earth to about 2.10^{32} J) for the Earth to melt and undergo chemical differentiation into a core composed of iron, nickel and presumably sulfur, and an ultramafic primitive mantle mainly made up of silicates and oxides. The process of core separation liberated another huge quantity of heat (estimated at 10^{31} J). All this heat must have caused in the very beginning a large degree of fractional melting in the primitive mantle. During convective and radiative cooling, this molten primitive mantle material must have progressively crystallized, leading to the generation of a globe-encircling protocrust of primarily basaltic and komatiitic composition. More complicated models imply that the melt residues differentiated further to produce other rock types, including residual magmas of dioritic and perhaps even granitic composition. As outlined earlier, isotopic evidence in meteorites, lunar samples and terrestral materials indicates that this whole sequence of events: accretion – core separation – formation of the protocrust, did not take more than a few million years.

The oldest known terrestrial rocks have an age of about 3,900 Ma. The formation of the Earth and the beginning of recorded geologic history are therefore separated by some 650 Ma, of which we have no preserved direct evidence. Of the next 400 Ma the preserved record is also scanty, because of the scarcity of terrestrial rocks older than about 3,500 Ma. All this makes the first billion years of the history of the Earth a period subject to much speculation, especially when models are developed which are exclusively based upon terrestrial data. Fortunately, however, since the beginning of the space age we are no longer confined to our own planet for finding the clues to the evolution of the early Earth. Planetary exploration missions in the inner Solar System since the 1960s have shown that, although each planet is different, there are broad similarities between the inner planets with regard to their early history. In terms of planetary evolution, the Moon, Mercury, Mars, Venus and the Earth constitute a series with the Earth being the most evolved planet. Apparently, evolutionary activity and complexity increase with the size of the planet, because of the greater thermal reservoir in the larger bodies. Useful information regarding the earliest Earth is therefore preserved on the other, more primitive planets – information which is not provided by the Earth itself because of the destruction of the early geologic record due to our planet's highly dynamic nature.

Analogies with crustal evidence from the other terrestrial planets, with due allowance for differences in size, mass, internal heat production and location in the Solar System, therefore play an important role in the reconstruction of the early history of the Earth. One of the results of the recent planetary exploration is the recognition of impact cratering as a major, possibly even dominant process in the crustal evolution of the early Earth. From observations on the other planets in the inner Solar System, we know that their surfaces were intensely bombarded by meteorites and asteroid-sized bodies. Isotopic dating of samples from the Moon revealed that this bombardment lasted for some 700 Ma after the accretion of the Moon, possibly with a culmination in a relatively short-lived bombardment by giant bodies at the end of this period, about 3,900 ± 100 Ma ago (Tera et al., 1974; Grieve, 1980). Then the meteoritic flux decreased steeply, to about the present level.

On Earth, no evidence for any impact craters dating from the early Archaean has been preserved because of later vigorous geologic and meteorologic activity. On the basis of the highland record on the Moon, making due allowance for the differences in size and gravitation, it is estimated that at least some 3,000 impact basins with a diameter greater than 100 km (of them, some 25 greater than 1,000 km), along with many more smaller craters, formed on the surface of the Earth during the episode of heavy bombardment (Grieve, 1980). The cumulative kinetic energy deposited on the early Earth during this period by the impacting bodies is estimated at 10^{28}–10^{29} J. The intense bombardment must have had profound effects on the early Archaean crust and upper mantle. The protocrust must have been continuously brecciated and modified by the impactions, while it seems probable that the heat released by the major impacts generated huge volumes of basaltic magma, as in the lunar maria. Also, it is postulated that the anomalous abundances of siderophile elements (such as Ni, Co, Cu, Au and Re) in the outer few hundred kilometers of the Earth, which are largely in excess of values expected in an equilibrium partition between a metallic core and a silicate mantle, testify to the large-scale addition of meteoritic material after the core separation.

The giant impacting interfered with the endogenic activity of the Earth, which must have been much greater in these early times because of the vastly greater internal heat production (see below) and associated more active mantle convection regimes. It is difficult to evaluate what the net effects of the combination of giant impact events and vigorous endogenic activity have been, but feasible scenarios (Grieve, 1980) are that the impacts have promoted localized reprocessing in the upper mantle of basaltic or komatiitic protocrust, which may have led to the production of sialic partial melts. In this way, the giant impacts could have triggered the production of continental nulei in the turbulent first 700 Ma of the Earth.

The cooling Earth

Another factor which must be taken into account when considering crustal evolution in the early history of the Earth, is the much greater internal heat production than today. The principal source of the Earth's internal energy is the radioactive decay of uranium, thorium and potassium. Because of the decreasing abundances of these heat-producing elements through geologic time, the thermal output of the Earth must also have steadily diminished (Fig. 6). If we assume an average chondritic composition for the bulk Earth, the total annual production of radiogenic heat at present amounts to 9.10^{20} J ($1.5.10^{-7}$ J/g). It is this radiogenic energy which is the driving force of plate-tectonic processes, of magmatism, and of all other endogenic

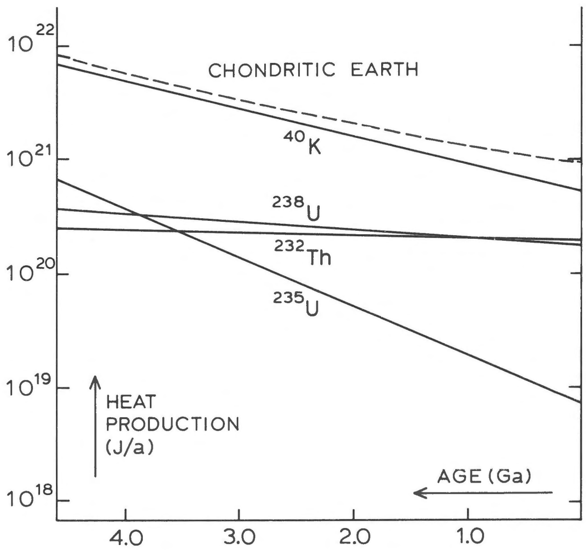

Fig. 6. Decrease in annual radiogenic heat production of the bulk Earth through geologic time, assuming an average chondritic composition (0.01 ppm ^{238}U, 7.10^{-5} ppm ^{235}U, 0.04 ppm ^{232}Th and 0.1 ppm ^{40}K, Durrance, 1987). The straight lines represent the heat-producing isotopes, the broken line the bulk Earth (6.10^{27}g). The present-day observed annual heat loss from the Earth is of the order of 10^{21} J, close to the calculated radiogenic heat production by a chondritic Earth of 9.10^{20} J. Equation of present-day heat production with present-day heat loss is invalid, however, because the low thermal conductivity values of rocks make that the present-day heat loss represents heat production some time in the geological past (when the abundances of the heat-producing isotopes were greater), and the contribution of primordial heat sources (accretion, core separation) to present-day heat loss is very uncertain (estimates range as high as 50% of the total heat flux; Durrance, 1987). By way of comparison: the annual energy input from the sun in the Earth's system (atmosphere, hydrosphere, land surface and biosphere) amounts to about $3.4.10^{24}$ J, and the present-day annual energy consumption of mankind to about 3.10^{20} J.

processes. In the early Earth the radiogenic heat production must have been much higher (for example, about $5.8.10^{21}$ J/year at the beginning of the recorded geologic history ~4.0 Ga ago, Fig. 6), which must have had a profound influence on all endogenic processes. As the Earth tends to thermostat itself by convective motions in the mantle and by magma production, the intensity of both these processes must have been several times that at present. In terms of modern plate tectonics, it has been calculated that the 'surface roll-over time' of the ocean floor 4.0 Ga ago might have been as short as 20 Ma (Elder, 1976), instead of the present-day time of 200 Ma. However, notwithstanding this tendency towards thermostasy, the Archaean Earth must have been hotter than today. For example, higher mantle temperatures ar signalled by the extrusion, unique to the Archaean, of ultramafic komatiitic lavas with eruption temperatures above 1,630° C, versus about 1,400° C for present-day mid-ocean ridge basalts.

There are several lines of evidence which suggest that the greater heat production of the early Earth has also resulted in higher surface temperatures and warmer oceans. Oxygen-isotope data of primary chert precipitates in the nearly 3.8 Ga old Isua supracrustals of West Greenland ($\delta^{18}O = +20.4$) and the 3.4 Ga old Onverwacht Group in South Africa ($\delta^{18}O$ up to +22) are interpreted to signal ocean water temperatures of about 80° C and 70° C, respectively (Knauth & Lowe, 1978). This interpretation is supported by oxygen-isotope studies of Archaean talc-bearing sediments (Costa et al., 1980) and a wide variety of sedimentological observations (Knauth & Lowe, 1978). A progressive decrease of ocean-water temperatures from about 80° C by 3.8 Ga ago should imply that in earlier times the temperatures at the surface of the Earth were too high for the existence of liquid water (Costa et al., 1980, 1981; Lambert, 1982) and that oceans were absent during the period of intense meteoritic impacting. The first half billion years or so of its history the Earth would thus have been enveloped in a hot, steamy, H_2O-rich atmosphere. The development of the hydrosphere appears to coincide with the beginning of the geologic record about 3.9 Ga ago, and it has been suggested that this development was a prerequisite for the preservation of continental crust (Costa et al., 1981). Clearly, development of a liquid hydrosphere must have brought about a much faster heat dissipation from the upper lithosphere, while prevalence of temperatures close to 100° C in oceans and near-surface atmosphere must have had far-reaching consequences for biological and geochemical processes in the early Archaean.

Uniformitarianism?

Abundant giant impacting and much greater thermal output must have resulted in geologic processes for the early Earth vastly different from the conventional plate tectonic regime and magmatism that prevailed for (at least) the last billion years. The recognition of these different conditions has revived once again the debate regarding the validity of the doctrine of uniformitarianism ('the present is the key to the past'), which has been at the heart of geological thinking for the last two centuries. In my view this debate results from too narrow an interpretation of the concept, and a distinction must be made, as argued by Moorbath (1982), between uniformitarian *cause* and non-uniformitarian *effects*.

Although the meteoritic flux decreased steeply about 3.9 Ga ago, it never came to an end and it is nowadays widely accepted that large-scale impacting continued to play a part in the terrestrial evolution throughout geologic time. To which extent this has been (and is) the case, still remains a matter of controversy. Some authors have suggested relationships between large impacts and phenomena such as magnetic reversals and plate movements (Clube & Napier, 1982), but these suggestions remain unproven. More firmly established has become the hypothesis that giant impacts were responsible for a number of global mass-extinctions in the paleontological record (Alvarez et al., 1980; Smit & Hertogen, 1980). From analogies with the other inner planets in the Solar System and the preserved cratering record on Earth, it is estimated that the Earth suffers a giant impacting event (crater >20 km diameter) about every 30 Ma, and some authors have attempted to relate this periodicity to an alleged periodicity in global mass-extinctions in the marine paleontological record (see review by Grieve, 1987).

Anyhow, I consider impacting of extraterrestrial bodies to be an essentially uniformitarian cause with non-uniformitarian effects because of the instantaneous character of the impacting events and the much higher meteoritic flux in the first ~650 Ma of the Earth's history. Similarly, I consider the radiogenic heat production of the Earth to be

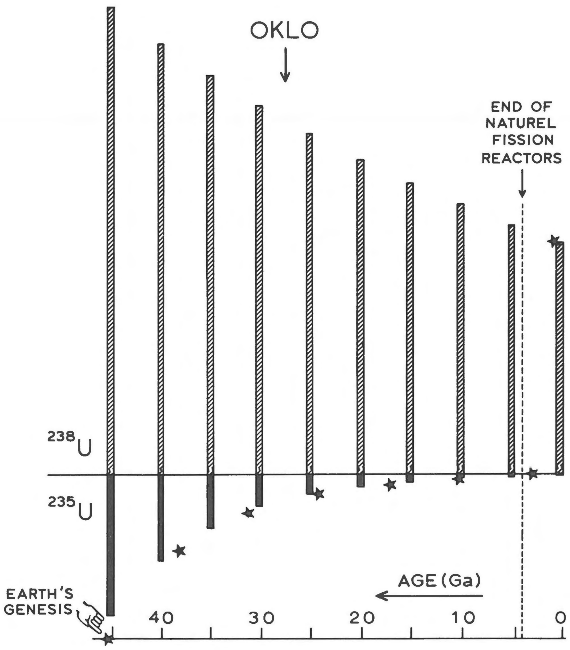

Fig. 7. The isotopic composition of natural uranium changes with geologic time, because of the much faster decay of ^{235}U than of ^{238}U (half-lives of $0.704.10^9$ and $4.468.10^9$ a, respectively; the half-life times since the origin of the Earth are for each isotope represented by asterisks). From an initial abundance of 24.0% 4.55 Ga ago, the ^{235}U isotope has reached its present abundance of 0.72% of natural uranium through an exponentional decline. The minium abundance for the operation of a plausible fission reactor is 1% (Cowan, 1976). This implies that natural fission reactors can have operated through most of geologic history, until about 400 Ma ago, if a number of requirements are fulfilled (Cowan, 1976): the uranium ore deposit should contain at least 10% uranium in a zone of at least 50 cm thick, no excessive amounts should be present of elements that strongly absorb neutrons ('poisons', such as Li, B, and many REE), and a suitable moderator (water) should be present. The Oklo reactors were in operation about 1,780 Ma ago (Lancelot et al., 1975) for a period of some 0.5–1.0 Ma.

a uniformitarian cause underlying all endogenic geologic processes throughout geologic time, but with non-uniformitarian effects because of the decrease in thermal output.

I can think of only one process in the recorded 3,900 Ma of geologic history that could be considered truly non-uniformitarian: the phenomenon of natural fission reactors. The fact that ^{235}U has a shorter half-life than ^{238}U (Table 1) has resulted in a decrease of the ^{235}U content of natural uranium through geologic time (Fig. 7). At the time of the origin of the Earth, ^{235}U accounted for about 24% of natural uranium, versus only 0.72% today. This implies that all uranium in the geologic past, until about 400 Ma ago, was of fuel grade and that initiation of a fission reactor system was possible at every site in the continental crust where geochemical processes led to a sufficient concentration of uranium (see caption Fig. 7). Potentially, many of such sites could be expected in ancient continental crust, but, to date, only one site has been identified with certainty, the uranium mines of Oklo in Gabon, western Africa, where natural fission reactors were in operation about 1,780 Ma ago (Lancelot et al., 1975). How the enormous quantities of released energy (for the Oklo reactors estimated at 15,000 MW years or $4.75.10^{17}$ J, Cowan, 1976) and radioactive waste have affected the continental environment and the biosphere, remains a matter of freewheeling speculation. The discovery of the remains of the Oklo reactors in 1972 made clear, however, that the nuclear fission reactor provided by Enrico Fermi in 1942 in Chicago was certainly not the first on Earth! Anyhow, if we do not consider mankind and his activities as a purely geologic agent (which is, of course, a matter of opinion), the operation of nuclear reactors in the Precambrian might be considered as a non-uniformitarian process in the evolution of the continental crust.

The oldest continental crust

About 3,900 Ma ago, the turbulent infant stage of the Earth came to an end. The oldest preserved terrestrial rocks date from about the same time: the Napier Complex at the northern extremity of Enderby Land in Antarctica, constituted by orthogneisses of trondhjemitic, tonalitic, granodioritic and granitic composition (Black & McCulloch, 1987). Isotopic microanalysis of individual zircons in Canberra (reported by Black & McCulloch, 1987) indicate that the magmatic precursors of these gneisses were emplaced about 3,930 Ma ago. This makes the Napier Complex unquestionably the oldest preserved continental crust presently known. There is evidence, however, for the existence of even older continental crust, with ages rather close to the terrestrial limit of 4.55 Ga. Isotopic microanalysis by the Canberra group of individual zircons from amphibolite-granulite grade quartzites and a greenschist-amphibolite grade quartz-pebble metaconglomerate in the Archaean Yilgarn Block in Western Australia indicates provenance ages of about 4,200 Ma (Froude et al., 1983) and 4,300 Ma (Compston et al., 1985), respectively. The 4,300 Ma old zircons have a morphology indicative of derivation from a granulite-facies terrain, whereas the 4,200 Ma old zircons display a zoning of primary igneous character. It is difficult to decide, of course, whether the crust from which these zircons were derived was acidic or basic, but zircon is a much more characteristic component of acidic rocks.

Although the available evidence is still scanty, it suggests that some true continental crust was already present by about 4,300 Ma ago. Furthermore, if the 4,300 Ma old zircons in Western Australia were indeed derived from a granulite-facies terrain, these rocks must have been formed in the lower part of a continental segment with a substantial thickness, at least of some 20 km. The question is, of course, just how much of the surface of the Earth at that time was covered by continental crust? It is possible that this earliest continental crust had a patchy distribution over the surface of the globe.

The oldest-preserved continental crust which has been studied in detail is the Godthåbsfjord area in West Greenland and the opposite coastal area of Labrador on the other side of the Davis Strait (see, for example, the reviews by Moorbath, 1977a, and Moorbath & Taylor, 1984). Both areas were joined before the opening of the Davis Strait about 60 Ma ago. This ancient crustal segment consists of an extensive high-grade gneiss complex (in West Greenland designated as the Amîtsoq gneisses), dominated by tonalitic, trondhjemitic and granodioritic orthogneisses and containing enclaves of ultramafic, mafic and acidic metavolcanics as well as volcanoclastic and chemical metasediments. The largest of these enclaves (the Isua supracrustal belt) is a banded iron formation at the head of Godthåbsfjord. A great number of isotopic age measurements have assigned ages of about 3,750 Ma both to the igneous precursors of the orthogneisses and to the volcanics, indicating that the deposition of the supracrustals and the subsequent intrusion of the granitoid magmas occurred within the space of a few million years. Although the whole association is high-grade metamorphic (up to the amphibolite-granulite facies), these gneisses and supracrustals have become one of the most critical regions for the understanding of the early Earth. For example, a variety of studies of the metasedimentary rocks has shown that the deposition was remote from sialic crust (Nutman et al., 1984), that the ocean-water temperature may have been as high as about 80° C (Knauth & Lowe, 1978), and that, although earlier claims to the occurrence of microfossils are disputed (Lambert, 1982), the observed carbon-isotope compositions ($\delta^{13}C$ values of -2.3 in carbonate and -13.0 in graphite of possibly organic derivation) indicate a biological signature (Schidlowski, 1987). The Sr, Nd and Pb isotopic compositions of the volcanic rocks and orthogneisses limit crustal residence to at most 100–200 Ma (Moorbath & Taylor, 1984), which, in combination with the absence of terrigenous sediments, indicates that the magmas were mantle-derived and represent juvenile continental crust. Moreover, Rb-Sr and Pb-Pb age measurements in the small sector of the Amîtsoq gneisses that is in the granulite facies indicate that the metamorphism occurred about 3,600 Ma ago, while the metamorphic mineral assemblages require a crustal thickness of at least 20 km and geothermal gradients equal to, or less than, 30° C/km (Griffin et al., 1980).

In view of the much greater radiogenic heat production in Archaean times than that of today, it may be a surprise that the magnitude of the geothermal gradient in the Godthåbsfjord continental crust 3,600 Ma ago was similar to that of the present day. The same conclusion has been reached in other Archaean high-grade terrains. No rock types have been observed in any of the Archaean remnants that could not be matched in the present-day environment. Also, the thickness of the continental crust and the distribution of the heat-producing elements were much the same as they are now.

Although the ancient rocks in the Archaean remnants are themselves amazingly normal, the lithological associations are characteristic and differ from those in the later Proterozoic and younger continental crust. Archaean terrains are mainly made of high-grade gneiss-granulite belts, greenstone-granite (gneiss) belts of mostly low grade, and cratonic sedimentary sequences. Particularly the greenstone-granite belts, supracrustal volcano-sedimentary sequences intruded by plutons ranging in composition from tonalitic to granitic (such as the Godthåbsfjord area in West Greenland) have received much attention. This is particularly inspired by their economic importance, as greenstone belts belong to the main depositories of elements such as Au, Ag, Cu, Zn, Cr and Ni, and also contain the Algoma-type banded iron formations (BIF). I will not go into the conflicting models here about the development of greenstone belts, but refer to the recent reviews by, for example, Kröner (1985), Taylor & McLennan (1985), Windley (1986), and to several papers in Kröner (ed., 1981). Anyhow, it is widely accepted that plate tectonics of one kind or another were in operation already in the beginning of recorded geologic history, and most models try to explain the formation of greenstone belts in terms of plate-tectonic mechanisms. Other proposed mechanisms, not involving plate tectonics, are based, for example, on the assumption that hot-spot tectonics similar to those in operation on Venus today (Morgan & Phillips, 1984) were the dominant regime on the Archaean Earth (Fyfe, 1978).

Continental growth

How much of the surface of the Earth was covered by an (at least) 20 km thick continental crust 3,600 Ma ago, the time of the granulite-facies metamorphism in the Godthåbsfjord area? This is, of course, the fundamental question in the growth-versus-recycling controversy. Those who favour a 'no-growth of continent' history postulate that virtually the entire mass of the present-day continental crust was differentiated from the mantle very early in the Earth's evolution, perhaps even by about 4,000 Ma ago (e.g., Hargraves, 1976; Fyfe, 1978; Armstrong, 1981). The fundamental assumption in this model is that the continental crust is continuously recycled in bulk through the mantle, by means of subduction of continental material and complete mixing between returned crust and mantle, and that the rates of continental growth and recycling through the mantle reached a near-steady state in the first 1,000 Ma of the Earth's history. The oldest preserved continental segments, such as in Enderby Land and the Godthåbsfjord area, should then represent only a tiny fraction of the continental crust present at that time, which has escaped the recycling process. Advocates of the continental growth model, on the other hand, assume that the preserved segments of ancient continental crust represent a major part of the areal extent and total mass of the continental crust at that early time. Moorbath (1977a, 1977b) estimates that by about 3.5 Ga ago only 5–10% of the present volume of continental crust was in existence. Throughout the next 3,500 Ma the remaining part of the present-day continental mass was formed by irreversible chemical differentiation of the mantle. This process of continental growth is usually explained in terms of modern-style (or some primitive variation of) plate tectonics, the principal mechanism since the late Archaean being the generation of juvenile continental crust along plate margins through andesitic volcanism in island arcs and 'underplating' of dioritic to tonalitic intrusions in Andean-type orogenic zones. For the Phanerozoic, the net growth rate (balancing additions through magmatism and losses through erosion and sediment subduction) is estimated at an average of 0.6 km^3/a (Dewey & Windley, 1981) to 1.0 km^3/a (Reymer & Schubert, 1984).

I will not enter into all arguments in this controversy. However, if we accept, as I do, that newly generated continental crust always maintains a memory of its origin in the radiogenic isotopes, the conclusion is inevitable that the distribution of isotopic ages and initial radiogenic isotope signatures favour continental growth by the addition of juvenile material to the continents throughout geologic time (e.g., De Paolo, 1980; Moorbath, 1977a, 1977b, 1982; Moorbath & Taylor, 1981, 1984). As I explained before, there is irrefutable isotopic evidence that some recycling of continental constituents through the mantle does take place through the medium of subducted oceanic crust along with continent-derived sediments, but these recycled components can in no way be regarded as quantitatively representative of bulk continental crust (Moorbath, 1982).

The radiogenic isotope record also reveals that continental growth was not continuous through geologic time, but episodic. As we have seen, radiogenic isotopes enable us to date continental rocks of all ages, to decide between mantle and ancient-crust origin for continental igneous rocks, and to define the crustal residence time for any sector of the continental crust. Many data have been, and still are being, accumulated from large parts of the continents. From these, it is concluded that a high proportion of the shield areas has been derived from source regions with mantle-type Sr, Nd and Pb isotope compositions, and that for the formation of any continental segment the whole sequence of events starting with differentiation of magma from the mantle, through intracrustal magmatic differentiation and mostly metamorphism, and ending in the stabilization of the new continental crust, occurred within a time-span not exceeding 100–200 Ma. Such 'immense, relatively brief, continent-forming events seem to occur throughout the Earth's history at different times in different places' (Moorbath, 1982), and are labelled as Crustal Accretion-Differentiation Superevents, or CADS (Moorbath, 1977a, 1977b, 1978, 1982; Moorbath & Taylor, 1981, 1984). The biggest CADS in Earth's history on a world-wide basis took place in the late Archaean, between about 3,000 and 2,500 Ma ago. It is in this period that our planet witnessed the most rapid and voluminous production of mantle-derived, juvenile continental crust in its history. Estimates of the continental areas in existence by about 2,500 Ma ago vary between 50 and 85% of the present-day continental crustal mass (Windley, 1986).

Why there was this spectacular peak of continental crustal growth between about 3,000 and 2,500 Ma ago, is still a matter of speculation. Anyhow, the peak represents an important global marker of the late Archaean, although not isochronous on a global scale, but occurring at somewhat different times within the 3,000 – 2,500 Ma interval in different regions. The massive increase in continental growth towards the end of the Archaean is reflected in many ways in the geologic record. For example, the sedimentary record reveals a sudden change in REE patterns (including the appearance of a negative Eu anomaly, which is absent in Archaean rocks) and a major increase (by 0.0025) in the $^{87}Sr/^{86}Sr$ ratio of carbonates (reflecting sea-water composition). This is explained by the intracrustal generation (by partial melting) and upward movement of magmas of granodioritic composition (enriched in light REE, depleted in Eu and with higher Rb/Sr than the source rocks) during the late Archaean CADS, so that towards the end of the Archaean the continental crust had assumed its present character (Taylor & McLennan, 1985). Also, the emergence of large volumes of continental crust towards the end of the Archaean can be related to the change from immature to mature sedimentary facies, resulting from the expanding epicontinental seas (Windley, 1986). Moreover, it may be taken to have led to the initiation of modern linear subduction regimes through the raising of massive barriers to the oceanic lithosphere spreading from ridges (Taylor & McLennan, 1985).

Towards the end of the Archaean, the Earth thus acquired many of the features which make our planet unique in the Solar System. The surface of the Earth changed from one of essentially oceanic aspect, with a few continental islands, to one characterized by the topographic dichotomy of today. There is thus every reason to consider the Archaean-Proterozoic transition as the most important threshold in the evolution of the Earth, although this transition is not sharply defined in time on a global scale but spreads over a time-span of several hundred millions of years.

After the dramatic increase in continental volume towards the end of the Archaean, a smaller part of the present-day continental mass has been generated in post-Archaean times. Most of this continental growth appears to have taken place through generation of juvenile continental crust along plate margins, followed by accretion to the continents. Three periods of CADS can be distinguished on the basis of radiogenic isotope evidence, two in the Proterozoic at about 1.9 – 1.6 Ga and 1.2 – 0.9 Ga ago, and the third covering the whole Phanerozoic, the last 600 Ma of Earth's history (Moorbath, 1976, 1977a).

Precambrian perspectives

In the foregoing I have broadly sketched some aspects of the key role that the isotopic record in continental rocks plays in gaining insight into the processes that governed the formation of the continental crust, and in setting time constraints to all aspects of the evolution of the Earth. Compared to the impressive knowledge gained in the 1960s and 1970s with regard to the building and evolution of the oceanic crust, the continental masses are still poorly understood. The work of the geologists and geophysicists who studied the ocean floor has established, as firmly as any concept can be in geology, the operation of the plate-tectonic regime on Earth. In comparison to the oceanic crust, however, it is poorly understood how plate tectonics governed the origin and evolution of the continents and how these processes have changed in the past. Since the late Archaean, juvenile continental crust appears to have been added to the continents mainly by accretion at the continental margins, but how did it form in the early Archaean when only tiny patches of continental crust were present on an Earth of oceanic character? How did the distinct periods of accelerated continental growth originate? What are the relative proportions in the continental crust of juvenile mantle-derived material and reworked ancient crust?

The answers to all these questions are contained in the continental geologic record. This record, particularly the part that is preserved in the Archaean segments of the Precambrian shields, can explain how our planet developed as such a strange anomaly in the Solar System. The ancient record can also provide the clues to the understanding of how terrestrial life, and the environment that supports it, came into being. At present, the oldest undisputed traces of biological activity are the ~3,500 Ma old stromatolitic cherts at North Pole in Western Australia (Lambert, 1982), but simpler ancestors must have existed earlier than these already rather complex structures. In fact, the carbon-isotope compositions of the ~3,750 Ma old Isua metasediments in West Greenland do suggest a biological signature (Schidlowski, 1987). Did life with a fair degree of biological sophistication already exist at that early stage of the Earth's history? If so, biological activity would already have been flourishing on Earth within some ten millions of years after the waning of the hostile environmental conditions of high meteoritic flux and surface temperatures above the boiling point of water. Very little time would then have been available for the sequence of chemical and biological processes which is thought to have been instrumental in the origin of life. In that case, the question arises whether life, rather than being of exclusively terrestrial derivation, could have originated on Earth from abiotic, complex organic molecules of extraterrestrial origin, formed by chemical reactions in the deep cold of interstellar space and deposited on the early Archaean Earth (Hoyle & Wickramasingh, 1977; Goldanskii, 1986).

Clearly, the search for the oldest traces of life must continue, along with the search for ancient rocks. Answers to the questions regarding the earliest history of the Earth and the evolution of the continental crust are crucial to the understanding of the origin and evolution of the unique geosphere-biosphere system, but also to gaining insight into the evolution of planetary bodies in general, which explains the present-day appeal of the Archaean Earth to planetary geoscientists. These are the challenges presented to earth science in the 1980s. After two decades of successful ocean-floor exploration, the new target of international earth-science collaboration in the 1980s has become the continents, in particular the Precambrian shields. The intensive continental exploration will certainly be continued through the year 2000, before the understanding of the continental crust is brought to the level achieved for the oceanic crust. It is essential for the future of the earth sciences in The Netherlands that here also more vigorous efforts should be made to meet these challenges for the remainder of this century and the early decades of the next, instead of continuing to pursue the earth-science targets of the past!

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