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The origin and evolution of the Earth's continental crust

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The present upper crustal composition of the Earth is attributed largely to intracrustal differentiation, resulting in the production of K-rich granites. The crust grows episodically and it is concluded that at least 60 per cent of it was emplaced by the late Archaean (ca 2.7 Ae). Archaean tonalites and trondhjemites resulted from slab melting of young hot oceanic crust. In contrast, most subduction-related rocks, now the main contributors to crustal growth, are derived from the mantle wedge above subduction zones. The contrast between the processes responsible for Archaean and post-Archaean crustal growth is attributed to faster subduction of younger (hotter) oceanic crust in the Archaean (ultimately due to higher heat flow) compared with subduction of older cooler oceanic crust in more recent times. The terrestrial continental crust appears to be unique compared to crusts on other planets and satellites in the solar system, ultimately, a consequence of the presence of water on Earth.

Introduction

The continental crust constitutes only 0.40 per cent of the mass of the Earth. Although it might seem so small that it could be ignored to a first approximation, the crust contains over 30 per cent of the bulk Earth budget for several of the most incompatible elements, such as Cs, Rb, K, U, Th and La. It is thus a major geochemical reservoir, particularly since the crust is not easily recycled back into the mantle. For these reasons, its composition is a major constraint on all geochemical models of bulk Earth composition and evolution.

41.2 per cent of the surface area of the Earth, or 2.10×10^8 km^2, is occupied by continental crust, of which 71.3 per cent, or 1.50×10^8 km^2, lies above sea level. There are four submerged microcontinents and ten major continental blocks (Cogley 1984). The average elevation of the continents above the mean sea floor (oceanic crust) is about 5 km. The elevation of the area above the 200 m isobath (i.e. the shelf/slope break) is 690 m. The mean elevation of the continental crust above present sea level is 125 m.

Crustal thickness varies between 10 and 80 km, correlating with the size of the continental block and the age of the last tectonic event. The average thickness is 41 km (Christensen & Mooney 1995). The volume of the crust is about 8.3×10^9 km^3 — this includes the submerged continental masses and sediment on the ocean floor derived from the continents, and has an error of about $\pm$ 10 per cent, because it depends on the variations in crustal thickness. The base of the crust is defined as the Mohorovicic Discontinuity (Moho). At this boundary, compressional wave velocities (Vp) increase from about 7 to about 8 km.s^{-1}. The Moho may be absent locally and is often not sharp. Thus, the crust is usually defined as material with seismic shear wave velocity Vs<4.3 km.s^{-1} or Vp<7.8 km.s^{-1}. Because of the possibility of underplating by basic or ultramafic material and probable interlayering of high-velocity mantle material with lower velocity crustal material, the base of the crust is likely to be very complicated in detail.

A mid-crustal boundary, the Conrad Discontinuity, is occasionally observed at a depth of about 10–20 km. It is often absent or poorly constrained by seismic data, especially in shield areas. Sometimes it is gradational over several kilometres. Curiously, super-deep drill holes have failed to identify many of the discontinuities that were based on interpretations of the geophysical evidence.

Estimates of crustal density range from 2.7 to 2.9 gm.cm^{-3} increasing with depth. If an average density of 2.8 gm.cm^{-3} is adopted, a crustal mass of 2.34×10^{25} gm ($\pm$ 7%) is obtained. On this basis, the continental crust forms 0.40 per cent of the mass of the whole Earth and 0.61 per cent of the mass of the crust and mantle.

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Age of the crust

The Sm–Nd isotope system is the most robust approach to dating the crust because it is least prone to resetting during metamorphism. Fractionation of Sm from Nd takes place at the time of mantle melting. Thus the depleted mantle Nd model age (T_{DM}) of crustal igneous rocks reflects their age of extraction from the mantle. However, processes such as intracrustal melting, metamorphic resetting, and assimilation of older material may affect Sm/Nd ratios, and estimates of average crustal age based on the Nd isotope system may still represent a minimum age (McLennan 1988).The average age of continental crust is about 2.0–2.3 Ae (DePaolo et al. 1991; McCulloch & Bennett 1993). If about 60 per cent of the crust was in place by 2.6 Ae, a mean age of 2.4 Ae for the continental crust is obtained.

Composition of the post-Archaean crust

The continental crust is particularly heterogeneous, as a glance at a geological map illustrates. Compositional changes may occur on a scale of metres, and it might be thought difficult, if not impossible, to calculate an average composition. However, the processes of erosion and sedimentation have carried out an efficient sampling of the upper crust and this information is contained in sedimentary rock sequences. Some elements, of which the rare earth elements (REE) are an excellent example, are transferred quantitatively during erosion and sedimentation from parent rocks into clastic sediments because they are not readily partitioned into fluids during weathering or diagenesis.

The REE abundance patterns in post-Archaean clastic sedimentary rocks show extreme uniformity on a global scale. Thus REE patterns for composite shale samples from Europe (ES) and North America (NASC) are similar to those for the post-Archaean Australian average shale (PAAS). These patterns are distinguished by relatively flat heavy-REE abundances — about 10 times chondritic, by light-REE enrichment and a quite uniform depletion in Eu (Eu/Eu* = 0.65). This uniformity extends both within and between continents. It is thus interpreted to represent the REE abundances in the upper continental crust exposed to weathering (Fig. 1).

The concentrations of other insoluble elements, such as Th and Sc, are also, like the REE, a measure of upper crustal abundances. Sc, although trivalent and a member of the same group (III) of the periodic table as the REE, is a much smaller ion, and is concentrated in basic rocks, entering early crystallising pyroxenes. In contrast, Th is typically concentrated in granitic rocks. The Th/Sc ratio in sedimentary rocks thus forms an index of the relative proportions of granitic and basic rocks.

By using elemental ratios that either are constant across a wide range of igneous compositions (e.g. K/U) or vary systematically with bulk composition (e.g. K/Rb), it is possible to extrapolate to obtain the upper crustal abundances of a number of other elements. In this manner the abundance of

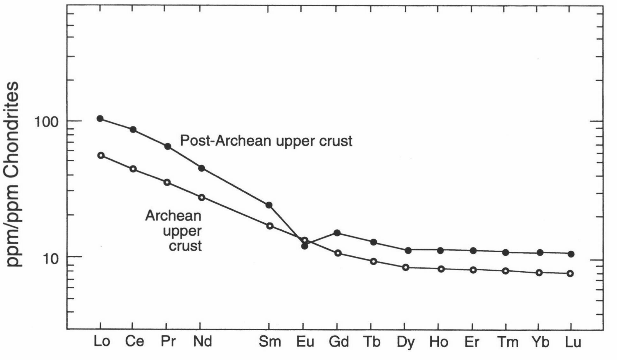


Figure 1. Chondrite-normalised REE diagram, showing estimates of post-Archaean and Archaean upper crust, from Taylor & McLennan (1985).

Rb can be obtained from K/Rb (250); Sr from Rb/Sr (0.3), while U can be obtained from the upper crustal Th/U ratios (3.8) and K from K/U ratios (10,000). The normative mineralogy of the upper crust, based on its major element composition is given in table 4 of Taylor & McLennan (1995). The composition of the post-Archaean *upper* crust is well established, with several estimates converging on a composition close to that of granodiorite (Taylor & McLennan 1995, table 3). Some recent studies employing such approaches include those of Taylor & McLennan (1985), Wedepohl (1991) and Ronov et al. (1992). There is less agreement about the composition of the *bulk* crust (see later).

Origin of post-Archaean upper crust

The upper crustal composition cannot be representative of the whole crust, because of element balance calculations, and heat flow/heat production data. Strong evidence for an intracrustal origin of the upper crust is provided by the Eu-depletion in post-Archaean sedimentary rocks. Eu anomalies rarely occur in igneous rocks derived from the mantle. No primitive mantle-derived volcanic rock shows a relative depletion in Eu. The depletion of Eu that characterises chondrite-normalised REE patterns in clastic sedimentary rocks is not due to surficial processes of oxidation or reduction. This element is present as the trivalent ion in sediments. Under the reducing conditions typical of magmas, much of the Eu is divalent. Thus the depletion in Eu is the signature of an earlier igneous event. The Eu-depleted K-rich granites and granodiorites that now dominate the upper crust were formed by intracrustal melting. The depletion of Eu observed in the upper crust is due to the retention of Eu in residual plagioclase in the lower crust. Plagioclase is only stable to a pressure of 10 kbars (a depth of 40 km on the Earth). This sink for Eu is thus consistent with the experimental studies of intracrustal melting for granite origin (e.g. Wyllie 1983). Two sources of heat are available to initiate intracrustal melting. One is the heat generated by radioactive decay of K, U and Th. The second source is underplating of the crust by basaltic magmas and mantle plumes. This is less easily evaluated, but is needed, since the crustal radioactive sources are probably inadequate.

Information about how long this process has been operating can be obtained from the sedimentary record. Loess from widely scattered localities across the globe has uniform REE patterns. This indicates that during the period represented by the source regions of the loess, the processes producing the upper crust did not change. Loess is derived from source rocks that extend back almost 2 b.y. Loess samples (Taylor et al.1983) from China, Europe, New Zealand and North America have Nd depleted mantle model ages (T_{DM}) extending back to 1700 m.y. Such sedimentary REE patterns also exist for sedimentary rocks with Nd model ages older than about 2 b.y. (e.g. McLennan & Hemming 1992). These data indicate that the composition of the upper continental crust was uniform and produced by similar processes back to at least 2 b.y. The processes producing the upper crustal composition that is being sampled during the formation of sedimentary rocks have, accordingly, been the same since well back into the Proterozoic.

Lower crust

Geophysical data show the very diverse nature of the lower crust. It appears to be at least as heterogeneous as the upper crust, and is likely to be very complex in detail, an example of which appears to be the Ivrea Zone in Northern Italy (Voshage et al. 1990). Owing to the inaccessibility of the lower crust and the absence of some averaging technique, such as provided for the upper crust by sediments, it is much more difficult to arrive at a representative composition for it than for the upper crust. Current understanding of the petrogenesis of most granitic rocks and the ubiquitous presence of negative Eu anomalies in sedimentary rocks indicate that intracrustal partial melting must be a fundamental process governing the composition and chemical structure of the lower continental crust. Xenoliths and granulite facies rocks, both providing enigmatic information, are available as potential samples (Rudnick & Presper 1990).

Xenoliths are frequently found in volcanic pipes and flows and record P–T conditions indicating derivation from the lower crust. They are commonly much more basic in composition than the granulite facies regions, and frequently show a relative enrichment in Eu. The positive Eu anomaly is mostly related to the accumulation of cumulate phases rather than to simple residues from partial melts (e.g. Rudnick & Taylor 1987; Rudnick 1992b).

Granulite facies regions are possible samples of the lower crust and they commonly possess positive Eu anomalies; however, these are typically found in the more acidic compositions rather than in mafic material that could represent residues after partial melting. Many such terranes appear, on compositional grounds, to be upper crust that has been buried in Himalayan-type collisions. Possibly, most regional granulites formed in mid-crust regions, so they are not a good model on which to base lower crustal compositions (e.g. Mezger 1992).

In summary, the lower crust appears to be essentially the basic residue left after extraction of the granodioritic upper crust together with additions from underplating by basaltic magmas. Measurements of Poisson's ratio (V_p/V_s; Zandt & Ammon 1995) provide strong support for the basic nature of the lower crust.

The Archaean–Proterozoic boundary

The Archaean–Proterozoic transition marks a major change, both in the volume of crust and intracrustal differentiation. The crustal processes responsible for these changes took place during the late Archaean and are recorded in early Proterozoic sedimentary rocks. The presence of large masses of unsubductable continental crust changed the tectonic regime from the multi-plate Archaean crust and produced the present linear (e.g. South America) or arcuate subduction zones (e.g. western Pacific arcs).

McLennan et al. (1979) found, in the early Proterozoic (2.5–2.2 b.y.) Huronian sedimentary succession of Canada, a progressive change from REE patterns without Eu depletions at the base to typical post-Archaean patterns at the top of the sequence. Similar changes in sedimentary REE patterns have been noted in early Proterozoic successions of the Pine Creek Geosyncline and Hamersley Basin, Australia, and the Slave Province of Canada (see review in Taylor & McLennan 1985). This change reflects an episodic change in upper crustal composition and is related to a massive emplacement of K-rich granitic rocks, depleted in Eu, in the upper crust toward the close of the Archaean. This process ('cratonisation') produces massive intracrustal melting to produce granites, transfers heat-producing elements to the upper crust and generally 'stabilises' the crust. It was non-synchronous over the globe and extended over several hundred million years.

Many geological events correlate with a major change at the Archaean–Proterozoic boundary. The proliferation of banded iron-formations can be related in part to the development of stable shelves during the late Archaean/early Proterozoic. The dramatic increase in ^{87}Sr in marine carbonates from that period (e.g. Veizer 1983) is, similarly, due to an upper crustal enrichment in ^{87}Rb in the K-rich granites that came to dominate the upper crust (coupled with a reduction of the mantle flux of Sr from submarine volcanics, resulting from a cooling Earth). The widespread occurrence of uranium deposits in basal Proterozoic sediments is due to the enrichment of the upper crust in incompatible elements, owing to intracrustal melting in the late Archaean. The first supercontinent probably formed at this time (e.g. Hoffman 1992).

The Archaean crust

The composition of the Archaean upper crust as revealed in the sedimentary record stands in marked contrast to that of the post-Archaean crust (Taylor & McLennan 1995, table 3). A significant difference is shown by the REE patterns in the Archaean sedimentary rocks, which, relative to those of the post-Archaean crust, typically show no Eu anomalies and a lower enrichment in the LREE (Fig. 1). These differences in REE patterns between Archaean and post-Archaean clastic sediments have been documented in many studies (see summary in Taylor & McLennan 1985). They form a crucial observation for models of the evolution of the continental crust.

In detail, there is a great variation in REE patterns in Archaean sediments. This stands in contrast to the very uniform post-Archaean sedimentary REE patterns. Both very steep and flat patterns are locally abundant. The flat patterns are derived from basaltic precursors to first-cycle sediments. The steep patterns occur in first-cycle sediments derived directly from tonalites, trondhjemites and granodiorites (TTG igneous suites). Both REE patterns are derived from the Archaean 'bimodal igneous suite', which largely dominates the Archaean upper crust. The most common patterns bear a superficial resemblance to the REE patterns of island-arc volcanic rocks, such as andesites. This similarity is the result of derivation largely from a mixture of the ubiquitous bimodal suite of felsic igneous rocks (tonalites, trondhjemites, granodiorites or the 'TTG suite' and their volcanic equivalents) and basaltic rocks, which are dominant in many Archaean terrains.

Some workers (e.g. Gibbs et al. 1986; Gao & Wedepohl 1995) have argued that the difference between the REE patterns of Archaean and post-Archaean sediments is without age significance, claiming that it is a consequence of differing tectonic settings. This means that post-Archaean greywackes should be identical to those of Archaean age. Except for those of fore-arc basins of oceanic island arcs, in all post-Archaean tectonic environments, the younger sediments display the crucial signature of Eu depletion. Archaean greywackes are clearly petrographically distinct from their modern counterparts in being plagioclase-rich (tonalitic sources), but having relatively few andesitic rock fragments (McLennan 1984). They appear to have formed in tectonic settings such as back-arc, continental arc, trailing edge and foreland basins. The evidence that Archaean turbidites are derived from the 'bimodal basaltic–TTG suite' rather than from 'island-arc andesites' negates the proposed analogy between conditions in the Archaean and modern arc environments.

Small areas of Archaean crust are preserved in high-grade metamorphic terrains. Such terrains exist in Greenland, India, Montana–Wyoming, Canadian Shield, Western Gneiss Terrain, Australia and the Limpopo Belt, South Africa. These regions form a tectonic environment distinct from that of the greenstone belts. The REE patterns in these sediments form two groups. Commonly, they are highly metamorphosed equivalents of Archaean greenstone belt sediments (e.g. Kapuskasing; Taylor et al. 1986). The second group of sedimentary rocks (Limpopo, Western Gneiss Terrain, Taylor et al. 1986) contains samples with REE patterns indistinguishable from typical post-Archaean patterns. These sediments must have been deposited on a stable shelf environment, most probably occurring on small stable mini-cratons.

Both groups of REE patterns occur in close proximity in meta-sediments in Archaean high-grade terrains in India, Greenland and the Limpopo Belt, demonstrating that the sediments were derived from highly localised sources, as is the general case for Archaean sediments. Small areas of stable granitic crust (mini-cratons) in such regions must exist in close association with greenstone belts, but the granites are not contributing significant amounts of erosional debris to the greenstone belt sedimentary basins. The high-grade belts are preferentially preserved, as the greenstone belt environments undergo destruction by erosion and recycling (Veizer & Jansen 1985).

Most of the crust as sampled by the greenstone belt terrains was derived from areas where the bimodal suite of basalt and TTG dominated the land area being eroded to supply the sediments. Despite the close association of greenstone belts with 'granitic' terrains (Bickle et al. 1994), the limited extent of the Archaean granitic terrains with 'post-Archaean' REE signatures is indicated by the absence of such signatures in the greenstone belt sediments. Wind-borne dust should have been widespread in the Archaean, owing to the absence of vegetation. Any granitic terrain should have contributed the characteristic signature of Eu depletion to the greenstone belt sediments. The absence of this signature in the majority of Archaean sediments means that such high-grade terrains represented less than 10 per cent of the exposed Archaean crust (see discussion in Taylor et al. 1986). The zircon data of Stephenson & Patchett (1990) are also consistent with this view of the very limited extent of granitic cratons in the Archaean.

An alternative view has been proposed by Gao & Wedepohl (1995) and Bowring & Housh (1995), who claim that granites with Eu anomalies were formerly widespread in the Archaean, although now missing, owing to erosion. The evidence discussed above makes the former existence of widespread granitic crust in the Archaean unlikely, a view also consistent with the field evidence from Archaean cratons. Basically, what you see is what was there (Trendall 1996).

In modern environments such as arcs or Pacific deep-sea environments, where volcanic provenances dominate, there is ubiquitous evidence for upper crustal material with negative Eu anomalies (e.g. Ben Othmann et al. 1989). In contrast to the situation in post-Archaean time, the isolation of small Archaean cratons enabled the survival of distinct suites of sedimentary rocks with much variation in REE patterns.

Origin of Archaean crust

The Archaean crust probably consisted of many small fast-spreading plates (Pollack 1986). The tonalite–trondhjemite suite was produced by rapid subduction of warm basaltic crust (e.g. Martin 1993, Drummond & Defant 1990).The steep REE patterns of the Archaean TTG suites indicate that garnet was in the residue during partial melting. An origin by melting at mantle depths for the TTG suite is thus indicated, because garnet is only stable in mafic–ultramafic systems at depths below about 40 km. In the Southern Andes, where rapid subduction of young hot oceanic crust occurs today, the slab reaches melting temperatures before complete dehydration occurs, and 'Archaean' tonalites are produced (e.g. Martin 1986).

The Archaean crust thus formed as a mixture of piled-up basalt–komatiite and tonalite–trondhjemite intrusions and extrusives. Sedimentary data suggest that in the upper crust the ratio of basalt to TTG was about equal. Probably this ratio was typical of the Archaean bulk crust, which unlike the present crust, was not vertically zoned in this model. Only minor intracrustal melting occurred in the Archaean. Areas of the crust that had undergone such melting, generating upper crustal negative Eu anomalies, formed cratonic regions of limited extent. Perhaps they were only slightly larger than the present extent of early Archaean terranes in West Greenland–Labrador and the Minnesota River Valley. The limited extent of PAAS-type REE patterns in the Archaean sedimentary record shows that the cratonic regions were very localised. In contrast, the post-Archaean crust is highly stratified into upper and lower crust, owing to extensive intracrustal partial melting.

Bulk crustal composition

There is less of a consensus about both bulk crustal composition and mechanisms of crust formation, since arriving at the average composition of the bulk rather than the exposed crust is complicated. The most important constraint on models of bulk crustal composition relies on the interpretation of continental heat flow data.

Sedimentary rock data provide information only on that portion of the crust exposed to weathering and erosion, but the upper crust is not representative of the entire 41 km thickness of the continental crust. Mass balance calculations show that a 41 km thick crust with K, Th and U abundances equal to that of the present upper continental crust would require about 80–90 per cent of the entire Earth's complement of these elements to be present in the continental crust. Heat flow data show that the upper crust (about 10 km thick) is strongly enriched in the heat-producing elements (K, U, Th).

The present average heat flow from continental areas is about 50 $mW.m^{-2}$ compared to 100 $mW.m^{-2}$ for the oceans and 84 $mW.m^{-2}$ for the whole Earth. However, it is difficult to assign precise values to the contributions from crust and underlying mantle (e.g. Nyblade & Pollack 1993).

Even when the effects of a tectonic heat contribution are removed for younger crust, there is a well-established difference between the heat flow in Archaean and later Precambrian terrains (e.g. Nyblade & Pollack 1993). There appears to be a steep offset in the data at the Archaean–Proterozoic boundary. Because erosional levels are not significantly deeper in Archaean terrains (Watson 1976), the lower heat flow is not due to deeper erosion of the Archaean crust removing a surficial hot layer or an upper granitic layer, as envisaged by Gao & Wedepohl (1995). The difference in heat flow is attributed by Nyblade & Pollack (1993) to an increase in the thickness of the subcrustal lithosphere under Archaean cratons, which lowers surface heat flow by deflecting mantle heat flow around the cratons. We calculate the bulk Archaean crustal values for the heat-producing elements at 1.0% K, 3.8 ppm Th and 1.0 ppm U (McLennan & Taylor 1996).

Most of the crust was generated in the Archaean, with lesser amounts from island-arc volcanism being added later to make up the present crust. The overall crustal bulk composition in our models was calculated from a 60/40 mixture of Archaean bimodal and post-Archaean andesitic compositions (Taylor & McLennan 1985). These result in the following concentrations for the heat-producing elements in the bulk continental crust: 1.1% K, 4.2 ppm Th and 1.1 ppm U, which give the crustal component of the heat flow of 29 $mW.m^2$, or slightly over half of the total heat flow measured in the continental crust. Thus, despite a significant difference in the *upper* crustal composition between Archaean and post-Archaean time, there is probably little difference in bulk composition.

In contrast, other recent estimates of bulk crustal composition exceed the heat flow constraint. Thus, the models of Christensen & Mooney (1995) and Wedepohl (1995) predict a greater heat flow than is observed from the continental crust, *even assuming no contribution to the heat flux from the mantle*. The model of Rudnick & Fountain (1995) is only viable if the background mantle contribution to the measured heat flux is less than 10 $mW.m^2$, which we consider too low. In contrast, models that propose a basaltic crustal composition (e.g. Abbott & Mooney 1995) produce too little heat in the crust to account for the observed values (see McLennan & Taylor 1996 for an extended discussion).

Origin of the continental crust

There is a basic distinction between the igneous activity that contributed to the formation of the continental crust in the Archaean and post-Archaean epochs (Figs 2, 3). There were probably many more plates in the Archaean, owing to higher heat flow (e.g. Pollack 1986) resulting in the rapid recycling of young hot oceanic lithosphere. Such basaltic crust reaches melting temperatures before dehydration has occurred. Under these conditions partial melting occurs, leaving a hornblende–garnet residue. The resulting product is the TTG suite with a steep REE pattern and no Eu anomaly.

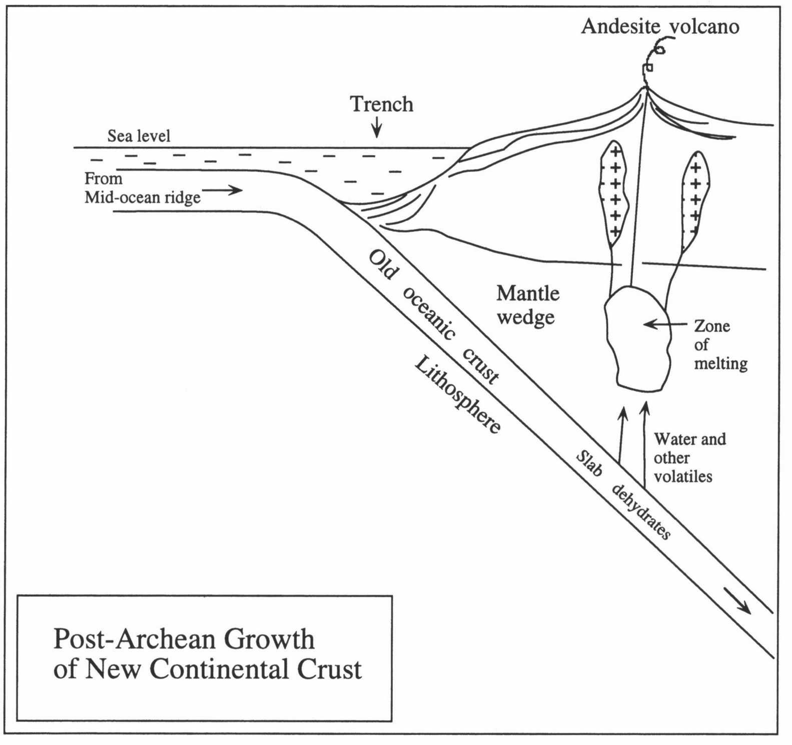
Generation of the high-Mg TTG rocks (tonalites, trondhjemites and granodiorites — the so-called sanukitoid suite; Shirey & Hanson 1984) appears to require low-pressure/high-temperature conditions. Such conditions could have been achieved in the Archaean by mantle melting associated with dehydration and/or partial melting of subducted young and hot oceanic lithosphere (Martin 1993).

The Archaean crust is formed from mixtures of the two dominant ('bimodal') igneous lithologies — Na-rich igneous rocks such as tonalites, trondhjemites and granodiorites or the TTG suite (and their volcanic equivalents), and basalts. REE patterns with Eu depletion similar to PAAS are rarely observed in the sedimentary record. They are restricted to cratonic sediments preserved in high-grade metamorphic terrains. These are interpreted as being derived from mini-cratons (Taylor et al. 1986) that were forerunners of the major development of cratons in the late Archaean.

A massive increase in the growth rate of the continental crust, well documented by Nd-isotope evidence, occurred over an extended period of 3.2–2.6 Ae, but differing for individual cratonic regions (e.g. Galer & Golstein 1991; McCulloch & Bennett 1994). Massive *intracrustal* melting of late Archaean crust produced an upper crust dominated by K-rich granodiorites and granites. This change is reflected in the REE patterns observed in the clastic sediments. These typically display significant depletion in Eu. At this time, the upper crust assumed its present composition, as Archaean-type REE patterns are swamped.

Such intracrustal melting often occurs within 50–100 million years of the derivation of new crust from the mantle (e.g. Moorbath 1978). Mantle plumes arriving beneath the crust are considered to be a prime cause of crustal melting. Possibly, the lithospheric keel beneath older cratons deflects mantle heat flow or mantle plumes towards younger marginal areas.

The number of plates became fewer as global heat flow diminished in the Late Archaean, and modern-style plate tectonics became the dominant tectonic theme. Oceanic crust was both older and colder by the time it reached the subduction zone. Older oceanic crust returns to the deep mantle without being remelted, and fluids from dehydration of the slab rise into the overlying mantle wedge, where they induce melting. This results in the production of the present subduction zone calc-alkaline suite and addition of this material to the crust.



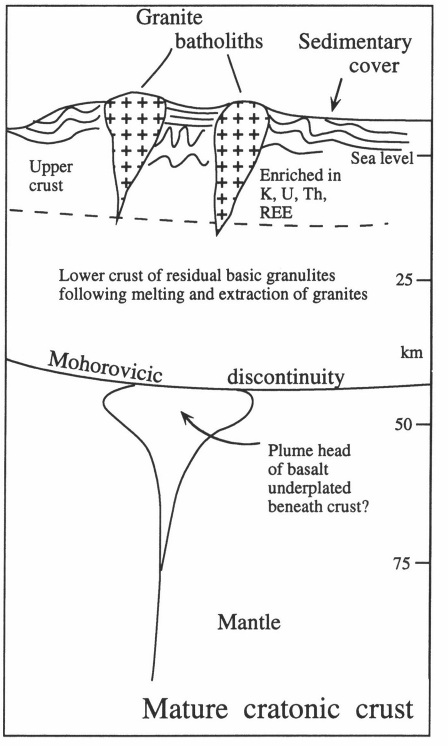


Figure 2. Schematic models of post-Archaean crustal development at subduction zones.

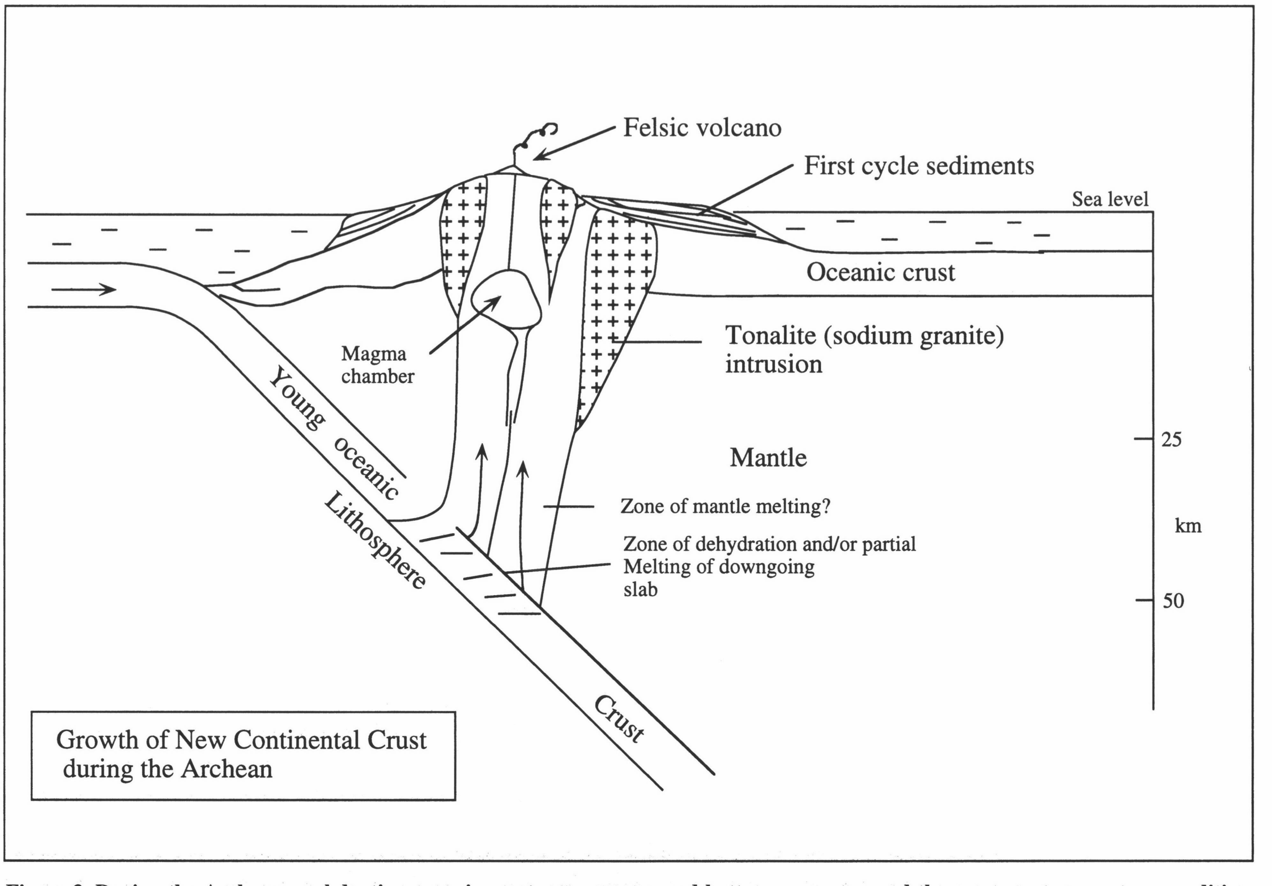


Figure 3. During the Archaean, subducting oceanic crust was younger and hotter on average and the pressure–temperature conditions of dehydration/melting of the subducting slab and overlying mantle differed considerably.

Episodic growth of the crust

One hypothesis proposes that the present mass of the crust formed very early in the Earth's history and has been recycled through the mantle. New additions are balanced by losses, resulting in a steady-state system (Armstrong 1991). The second proposes that the crust has grown throughout geological time in major episodic pulses (e.g. Moorbath 1978; Taylor & McLennan 1981, 1985). The REE abundances enable some testing of these models. They show a major change in the sedimentary record around the Archaean–Proterozoic transition (ca 2.5 Ae), which has been correlated with an earlier major pulse of crustal growth during the late Archaean.

Evidence from ^{10}Be and Pb isotopes shows that sediments have been recycled into the mantle (see review by McLennan 1988). The evidence from ^{10}Be and other geological, geochemical and isotopic constraints limits the amount of subducted sediments to a few per cent. Those models that propose massive recycling of the crust through the mantle encounter various difficulties. Data from long-lived radiogenic isotopes for mantle-derived rocks provide no independent constraints on this problem (Armstrong 1990). During sedimentary recycling processes, the sedimentary mass is largely cannibalistic, with little new material being added from the mantle (McLennan 1988). Thus, the mass of sediment available for subduction is $<1.6\times10^{15}$g.yr^{-1} (about 0.5 km^3.yr^{-1} of crust). This amount does not provide sufficient material to support a steady-state crustal mass.

For the past two billion years, sea level has been within about one kilometre of the present level. This constitutes the so-called freeboard constraint and is evidence that the volume of continents relative to oceans has been roughly constant over this period (e.g. Kasting & Holm 1992). The sedimentary record in the Archaean is much too fragmentary to make freeboard a restriction on crustal volume. The freeboard constraint appears valid at least back into the Early Proterozoic, but provides no constraints of crustal volume before the Proterozoic. Most models accommodate the constraint by proposing that most crustal growth had occurred by the late Archaean, with lesser additions since the Early Proterozoic.

Two observations inform us that early 'granitic' crusts were very limited in extent. There was no land vegetation in the Archaean (Cloud 1988). Hence, there must have been wide exposures of bare rock, with the consequent formation of dust and its transport by wind. Present-day mid-ocean sediments, derived largely by wind transport, show the tell-tale signature of Eu depletion of the upper continental crust. This signature is missing from virtually all Archaean sediments except for those found in the local cratonic areas discussed above. Most of the exposed crustal rocks from which the Archaean sediments were derived were basalts and the TTG suite, rather than granites and granodiorites.

The second critical observation is that of Stephenson & Patchett (1990): they analysed zircons, mainly from quartzites, from the Canadian Shield and the Wyoming, North Atlantic (Labrador, Greenland, Scotland) and Kaapvaal (Southern Africa) cratons. Zircon, being a highly durable mineral, survives many cycles of weathering, erosion and deposition of sediments. Stephenson & Patchett (1990) found that the age of zircon populations, dated by the ^{176}Lu–^{176}Hf technique, in Early Archaean quartzites, is mostly the same as that of the terrain in which they are found, and that there is a scarcity or absence of significantly older zircons. If there was an early sialic crust, then a massive population of ancient zircons derived from it by erosion should have survived and been recycled into younger Archaean sediments. Nutman (personal communication 1994) using U–Pb dating by ion-microprobe, likewise has found no older zircon populations in 3.8 Ae metasediments from Isua, Greenland.

In summary, the growth of the continental crust has proceeded in an episodic fashion throughout geological time (e.g. Moorbath 1978; Taylor & McLennan 1981, 1985), with a major increase in growth rate in the late Archaean (Fig. 4). The present crust has continued to grow by island-arc volcanism and related magmatism, followed by episodes of intracrustal melting (e.g. eastern Australia). There has been no significant change in this process, detectable in the sedimentary REE record, since Archaean–Proterozoic boundary time.

Subcrustal lithosphere

Are there deep roots to the continents? If so, what is their relationship to the overlying crust? There is good evidence that fast seismic P-wave velocities extend to depths of 220 km. Lateral variations in S-wave velocities extend much deeper, to 400 km. Thus, there appear to be deep keels, which are relatively cold and seem to be refractory in composition under the continents (Boyd & Gurney 1986, Jordan 1988). The heat-flow data (Nyblade & Pollack 1993) are consistent with the lithosphere beneath Archaean cratons being much thicker

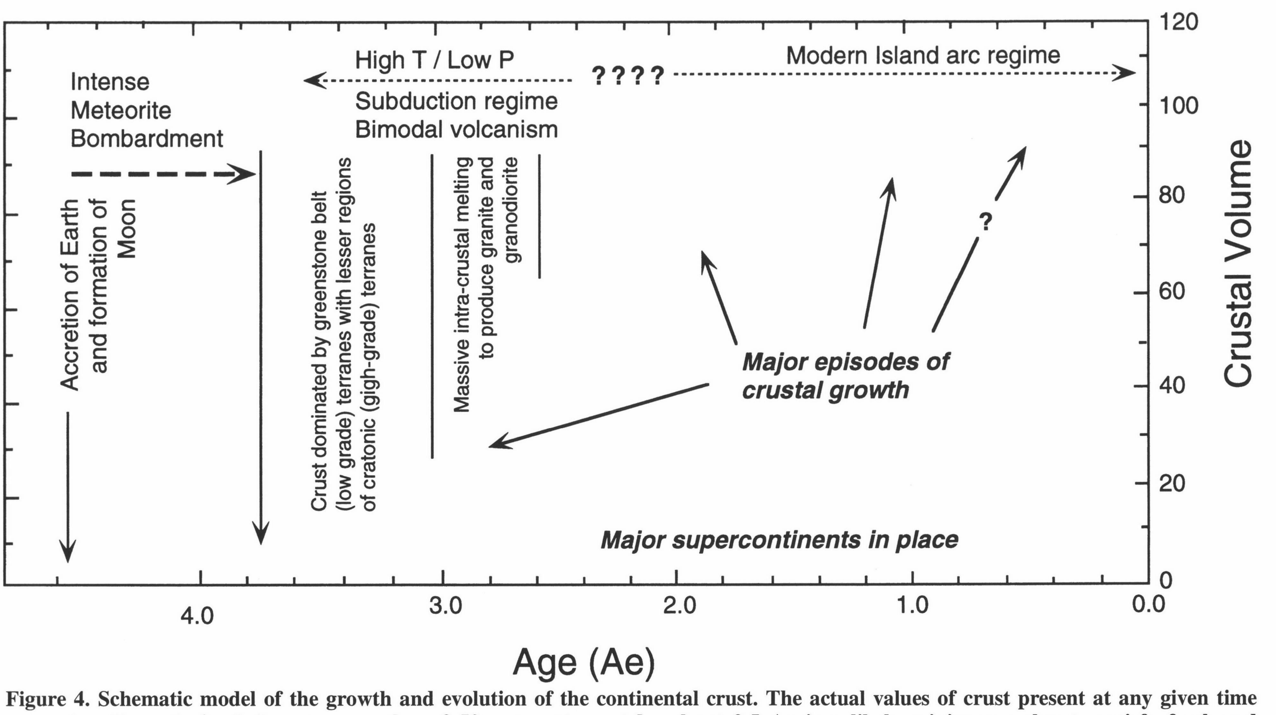


Figure 4. Schematic model of the growth and evolution of the continental crust. The actual values of crust present at any given time are not well constrained; however, a value of 50 per cent crust by about 2.5 Ae is a likely minimum value to satisfy freeboard constraints. Although major global episodes of crustal growth and differentiation are well documented during the late Archaean and at about 2.1–1.7 Ae, it is less clear if crustal growth has been episodic on a global scale during younger times.

than under younger cratons. Studies of inclusions in diamonds indicate that low temperatures have persisted beneath shields for up to 3000 m.y. (Pearson et al. 1995). Possibly, these deep roots represent an Mg-enriched zone of refractory, lower density material, a residue remaining from the extraction of an Fe-rich partial melt. Whether the roots are residual from the extraction of basaltic oceanic crust or are associated with the extraction of continental crust is an unresolved question. However, the presence of these ancient roots, apparently welded to the base of the crust, raises clear problems with those models that invoke delamination of basic lower crusts (e.g. Rudnick & Fountain 1995).

Early crusts

It is often thought that, by analogy with the Moon, the Earth formed an early anorthositic crust. Several reasons make this unlikely. Firstly, the Moon is richer in Ca and Al than the terrestrial mantle (Taylor 1987; Lucey et al. 1995), leading to the early appearance of plagioclase during crystallisation of the lunar magma ocean. Secondly, plagioclase is unstable at shallow depths (40 km) in the Earth, transforming to garnet and, thus, locking up Ca and Al in a dense phase. In contrast, plagioclase will be stable in the Moon to depths of several hundred kilometres.

Plagioclase will not float in a wet terrestrial basaltic magma. The oldest terrestrial anorthosites are not distinct from and closely resemble younger Archaean examples (John Myers, pers. comm.). They must share the same petrogenesis, which makes derivation from a primordial magma ocean of the older examples less likely. Finally, there is no sign of an ancient reservoir of Eu or of primitive $^{87}Sr/^{86}Sr$, which would have resided in an early Sr-rich Rb-poor anorthositic crust.

The conditions for the production of massive granitic crusts are probably unique to the Earth and require three or more stages of derivation from a primitive mantle composition. The Earth has transformed less than 0.5 per cent of its volume to continental crust of intermediate composition and less than 0.2 per cent of its volume into granitic continental crust (ie. upper continental crust) in over 4000 million years, so the process is inefficient. The highland feldspathic crust of the Moon, about 12 per cent of lunar volume, formed, in contrast, within a few million years, during crystallisation from a magma ocean.

No good evidence exists for the enduring geological myth of a primordial world-encircling crust of 'sial' or granite. Such models originated through false analogies with the production of a silicic residuum during crystallisation of basaltic magmas and conditions in an early molten Earth, and a lack of appreciation of the difficulties of producing granite. The absence of evidence for such a crust is discussed above . A primitive 'sialic' crust is refuted by, amongst other evidence, the absence of an old zircon population in younger sediments (Stephenson & Patchett 1990). The oldest preserved terrestrial rock is the 3.96 Ae Acasta Gneiss, in the Northwest Territories of Canada (Bowring et al. 1990).

The formation of crusts distinct from the bulk planetary composition is a common planetary phenomenon. However, crusts similar to the Earth's continental crust do not seem to have formed on the other terrestrial planets or on the icy satellites of the giant planets. On the inner planets, the crusts are basaltic, which are the typical partial melts derived from Fe–Mg silicate mantles. On the satellites of the giant planets, crusts involving water, methane or ammonia ices form by melting of ice–rock mixtures.

The significant feature about the Earth, in contrast to the other terrestrial planets, appears to be the presence of liquid water at the surface, coupled with plate tectonics and subduction, which enable recycling of subducted basaltic crust through the mantle. It is these processes that permit the slow production of continental crust (e.g. Campbell & Taylor 1983). The absence of subduction leads to the persistence of barren basaltic plains, such as we observe on other planetary bodies and the Moon.

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