Vol 443|19 October 2006|**doi:10.1038/nature05191**

Evolution of the continental crust

C. J. Hawkesworth[1] & A. I. S. Kemp[1,2]

The continental crust covers nearly a third of the Earth's surface. It is buoyant—being less dense than the crust under the surrounding oceans—and is compositionally evolved, dominating the Earth's budget for those elements that preferentially partition into silicate liquid during mantle melting. Models for the differentiation of the continental crust can provide insights into how and when it was formed, and can be used to show that the composition of the basaltic protolith to the continental crust is similar to that of the average lower crust. From the late Archaean to late Proterozoic eras (some 3–1 billion years ago), much of the continental crust appears to have been generated in pulses of relatively rapid growth. Reconciling the sedimentary and igneous records for crustal evolution indicates that it may take up to one billion years for new crust to dominate the sedimentary record. Combining models for the differentiation of the crust and the residence time of elements in the upper crust indicates that the average rate of crust formation is some 2–3 times higher than most previous estimates.

The Earth's continental crust differs from the crust of other planets in our Solar System. Its formation modified the composition of the mantle and the atmosphere, it supports life, and it remains a sink for CO_2 through weathering and erosion[1,2]. The continental crust has therefore had a key role in the evolution of this planet, and yet when and how it formed remain the topic of considerable debate. Significant obstacles to understanding this evolution include the shortage of rocks from the first billion years of Earth history and that fact that rocks amenable to study on the surface are very different from those that were involved in the generation of new crust. Thus new techniques have been required to model the differentiation and evolution of the continental crust, and to see back through time to the processes involved in its generation.

The broad compositional features of the continental crust are now well established and have been derived using a number of approaches[3]. Less well understood are when and how such continental crust was generated, and how the rates and processes involved in the generation of the crust have changed with time. Overall, it seems likely that the rates at which the crust has been generated have decreased with time, as less heat is generated inside the Earth from the decay of radioactive elements. There is a long-standing debate, however, over whether the volume of continental crust has remained essentially unchanged since the earliest Archaean[4,5], when it might have been more readily destroyed, or whether the volume of crust has grown through time[6–12]. Evaluating the composition of new continental crust can provide important clues as to how and when it may have been generated. To do this in turn requires an understanding of the differentiation processes involved in the generation of the compositionally evolved igneous (granites) and sedimentary rocks that dominate the upper crust, and hence the geological record. Although such issues have been historically problematic, they have been reinvigorated by the development of powerful new techniques for the *in situ* analysis of stable and radiogenic isotopes, leading to more robust models for the evolution of the continental crust. Here we highlight these developments and others that have taken place over the past decade[13]. Our emphasis is on the dynamics of crust generation and its evolution, and the extent to which understanding the rates of crust generation influences models for its formation.

The composition of the continental crust

Although the continental crust represents only 0.57% of the mass of the Earth's mantle, it contains, for example, over 40% of elements such as potassium. The bulk continental crust has 60.6% SiO_2 and 4.7% MgO (Table 1), a composition not likely to have been in equilibrium with the upper mantle. Most models for the generation of the continental crust therefore involve at least two stages of differentiation—the extraction of basaltic magma from the mantle and the remelting or fractional crystallization of that basalt, possibly augmented by sedimentary processes[3,13–19].

The geological record is dominated by rocks of the upper continental crust, and so models for the formation of new crust require an understanding of how the composition of the upper crust is related to that of new continental crust. The upper crust has a relatively low Eu abundance because Eu partitions strongly into plagioclase feldspar (see, for example, ref. 17), a common residual mineral during crustal melting. The median of minor- and trace-element analyses of granitic rocks with the relative Eu content of the average upper crust is very similar to those in the upper crust[19]. The implication is that differentiation of the continental crust is dominated by igneous processes, and that the composition of the upper crust has not been modified significantly by erosion and sedimentation. This knowledge in turn provides the basis for estimating the time-integrated composition of new material added to the continental crust.

There is some debate over the amounts of igneous differentiation that take place below and above the formal base of the crust at the Mohorovičić discontinuity[16,20]. However, generation of continental crust ultimately involves a basaltic precursor, and for the sake of this discussion we shall start with such basaltic compositions and assume that new material added to the continental crust is typically basaltic in composition[13]. In the modern Earth, basaltic magmas are most voluminously emplaced along mid-ocean ridge systems to form the oceanic crust, but these rocks are unlikely to make a significant material contribution to average continental crust. More relevant are the basalts that form in response to subduction of oceanic crust along destructive plate margins at island or continental arcs, and basalts that are generated in intra-plate settings owing to extensional tectonics, and/or to the emplacement of mantle plumes. The wider implication is the extent to which continental crust is generated in

---

[1]Department of Earth Sciences, University of Bristol, Wills Memorial Building, Queens Road, Bristol BS8 1RJ, UK. [2]School of Earth and Environmental Sciences, James Cook University, Townsville, Queensland 4811, Australia.

Table 1 | Average and modelled compositions of the continental crust

	Average continental crust[3]	Average upper continental crust[3]	Upper crust magma composition*	Model new crust[19] (OIB-arc mix)	Residue[19]
Compositions (wt%)					
SiO_2	60.60	66.60	68.02		
TiO_2	0.72	0.64	0.48	0.89	0.95
Al_2O_3	15.90	15.40	15.41		
FeO_t	6.70	5.04	3.30		
MnO	0.10	0.10	0.07		
MgO	4.66	2.48	1.63		
CaO	6.40	3.59	3.62		
Na_2O	3.07	3.27	3.60		
K_2O	1.81	2.80	2.76	0.61	0.27
P_2O_5	0.13	0.15	0.14	0.15	0.16
Compositions (p.p.m.)					
Rb	49	82	80	10.9	0.1
Ba	456	628	568	212	156
Sr	320	320	301	347	354
Nb	8	12	10.3	4.8	3.9
Y	19	21	21	18.0	17.6
Zr	132	193	157	58.4	42.8
Hf	3.7	5.3	ND	1.6	ND
La	20	31	34	7.8	3.7
Ce	43	63	64	17.9	10.6
Nd	20	27	26	11.0	8.6
Sm	3.9	4.7	5.0	2.9	2.5
Eu	1.1	1.0	1.0	1.0	1.0
Gd	3.7	4	4.1	2.8	2.6
Tb	0.6	0.7	ND	0.5	ND
Dy	3.6	3.9	ND	3.0	ND
Er	2.1	2.3	ND	1.8	ND
Yb	1.9	2	1.9	1.7	1.7
Lu	0.3	0.31	ND	0.3	ND
Pb	11	17	12	2.7	1.3
Th	5.6	10.5	9.3	1.5	0.2
U	1.3	2.7	2.1	0.4	0.1
Ratios					
Eu/Eu*	0.93	0.72	0.70	1.06	1.18
ASI	0.85	1.03	0.99	ND	ND
Mg#	55.3	46.7	0.47	ND	ND

The composition of the average continental crust and the average upper continental crust[3], and the upper crust magma composition calculated from global granite data arrays at Eu/Eu* = 0.7 (ref. 19). Also presented is the proposed composition of model new continental crust (itself calculated as a mixture of 92% average arc and 8% OIB, Fig. 1); this has higher U and Th contents than the average lower crust[19], although these are broadly consistent with continental heat flow data[76]. The residue composition is that formed assuming that the *D* values for Rb and Th are ~0 during the differentiation of the continental crust, whereupon the upper crust magma composition reflects 14% melting (or 86% crystallization) of the newly added crust[19]. (ASI, alumina saturation index; Mg#, magnesium number; Eu/Eu*, europium anomaly, calculated as in ref. 17; ND, not determined.)

response to either deep-seated thermal anomalies within the Earth (as reflected in mantle plumes) or shallow-level plate-tectonic processes. This can be addressed by chemical modelling, as subduction-related basalts and intra-plate basalts typically have contrasting trace-element compositions. A commonly used diagram is Nb/La plotted against Sr/Nd; the former reflects the distinctive negative mantle normalized Nb anomaly of the continental crust (see, for example, refs 13, 21), whereas the latter ratio is elevated in island arc magmas and it fractionates within the crust because Sr is partitioned into plagioclase feldspar (Fig. 1). Plank[18] highlighted that element ratios that are changed by processes within the continental crust cannot be simply used to estimate the proportions of different materials being added to the crust. Nb/La is slightly lower in the average upper crust than in the average lower crust composition (Fig. 1a), but it appears not to fractionate significantly with increasing silica in crustal rock suites, and so it may be acceptable to use it in this way.

It is difficult to constrain the composition of the basaltic end-member magmas early in the Earth's history. Nevertheless, at the present day intra-plate magmas are reasonably represented by average ocean island basalts (OIB)[22] whereas the subduction end-member can be taken as an average of modern primitive island arc basalts (IAB) that contain relatively little contribution from subducted sediment[19]. Arc rocks that include significant contributions from subducted sediment[23,24] have been excluded, because we are concerned with the composition of new material being added to the continental crust, rather than the recycling of pre-existing crustal materials.

Thus the average basaltic protolith to the continental crust (the model new continental crust) should lie at the intersection of arrays between the compositions of magmas generated in subduction and intra-plate settings, and trends for differentiation within the crust (Fig. 1a). It is convenient to calculate that composition in terms of the OIB and IAB end-members (Table 1), but their apparent proportions are much less significant than the proposal that the model new continental crust lies at the intersection of the two arrays. As highlighted by Rudnick[13], the striking conclusion from Fig. 1 is that the bulk continental crust cannot be modelled as a simple mixture between global average OIB and IAB end-members. This agrees with the major-element evidence that the bulk continental crust represents a fractionated, rather than primary, composition. Instead, the composition of model new continental crust coincides with the estimate of average lower continental crust (Fig. 1a and Table 1).

The composition of the lower continental crust is largely estimated by indirect means[3,25], and there is debate over the extent to which lower crustal rocks are residual after melt extraction, and/or have been modified by granulite facies metamorphism. However, Fig. 1 implies that, on average, the lower crust is similar in composition to unmodified basalts, and this conclusion is insensitive to the choice of different within-plate basalt end-members. It has therefore been inferred[19] that the average lower crust is broadly representative of the protolith to the continents.

It has been proposed, and it could be inferred from Fig. 1a, that the model new continental crust may be simply represented as a mixture of ~8% OIB and ~92% IAB[13,21]. This would in turn appear to reinforce the importance of convergent margin processes in the generation of continental crust[14,16]. However, such figures should be regarded with caution. For example, in the Archaean, significant volumes of new crust were associated with the emplacement of the tonalite-trondhjemite-granodiorite (TTG) intrusive suites that are characteristic of this time period[17,26]. The TTG are thought to have been generated by partial melting of hydrated basalt[27], and in such melting processes low Nb/La ratios are associated with low Ti/Zr ratios (see, for example, ref. 17). This differs from recent subduction settings where tholeiitic basalts with low Nb/La are commonly generated without an associated fractionation of Ti. The significance is that in the Archaean, low Nb/La ratios could have resulted from remelting basaltic source rocks even if those rocks were generated in a within-plate setting (Fig. 1a). This highlights the potential pitfalls in assessing how the proportions of new crust generated in different settings may have changed with time[28].

Differentiation of the continental crust

Differentiation of the continental crust is dominated by igneous processes[19]. The degree of differentiation constrains how the rates at which the upper crust is generated relate to those at which new crust is generated, and the amounts of residual crustal material recycled back into the mantle. The degree of differentiation can be estimated from the compositions of model new crust and the upper continental crust[19]. Calculations reveal that the upper crust reflects ~14% partial melting, or the analogous amount of fractional crystallization, of average new basaltic crust (Table 1). Inevitably, the residual composition is very different from the average lower crust in that it has strongly depleted incompatible-element abundances (for example, U, Th, Pb, La, Zr) and relatively high Ti contents; in other words, it is more refractory and genuinely residual. Relative depletions in certain elements during crustal differentiation reflect the principal mineralogical controls in the differentiation of the continental crust. Specifically, Eu and Ti highlight the role of residual plagioclase and a Ti-rich phase, and there is little evidence for residual garnet during crust differentiation, which would lead to lower contents of Y and heavy rare-earth elements. This is consistent with Nd-Hf isotope studies on the lower crust (see, for example, refs 29 and 30), highlighting that most of the continental crust differentiated at depths that were too shallow for significant amounts of residual garnet. Differentiation, and perhaps the generation of significant volumes of new crust, does not appear to have been associated with areas of thickened continental crust.

A consequence of the estimated proportions of new and upper crustal material is the large volume of residual material after differentiation. The upper crust is ~12.5 km thick in crust with an average total thickness of 40 km (ref. 3). If the upper crust is the product of 14% melting, the corresponding residue would be 77 km thick, resulting in a total crustal thickness of ~100 km, including the middle crust. There is no geophysical evidence for a contemporary mafic lower crust of this thickness[3], and the appropriate melt-depleted compositions are uncommon in the spectrum of exposed lower crustal lithologies[25]. Such a thick residual layer would be gravitationally unstable under most geothermal conditions, owing to phase transformations that produce dense minerals like garnet[31–33]. It is therefore concluded that the volumetrically dominant crustal residue of upper crust formation has largely foundered into the mantle[16,17,23,31–33], with the consequence that the residence time of material in the lower crust is much shorter than in the upper crust. This reinforces arguments that delamination of residual lower crustal material is critical for establishing the andesitic composition of the average continental crust (Fig. 1). The residual material returned to the mantle would have had relatively high Sr/Nd and low Rb/Sr, Sm/Nd and Th/La ratios[13,18,32,34,35], and it has been invoked as the cause of trace-element-enriched material in the source of some OIB[35,36].

If the average residence times of elements in the lower crust are so much less than those in the upper crust, probably at least five times less[19], what are the geological implications? When is material lost from the lower crust? Crustal differentiation is a logical thermal consequence of crust generation, as seen for example in the Andes[37] and eastern Australia[17,19], particularly where crustal melting is driven by basaltic under/intraplating (see, for example, refs 38–40). Rapid processing of the lower crust, and delamination of the residual material, may therefore largely occur at the sites of active crustal growth[32,33,41], and perhaps less frequently in response to lithospheric thickening linked to compressional tectonics[42].

Reconciling the sedimentary and igneous records

Fine-grained detrital sedimentary rocks such as shales provide the most representative samples of the upper continental crust. They have been used to determine average compositions for the abundances of at least the insoluble elements in the upper crust[14,43] and to constrain the Nd isotope evolution of the continental crust[44–46].

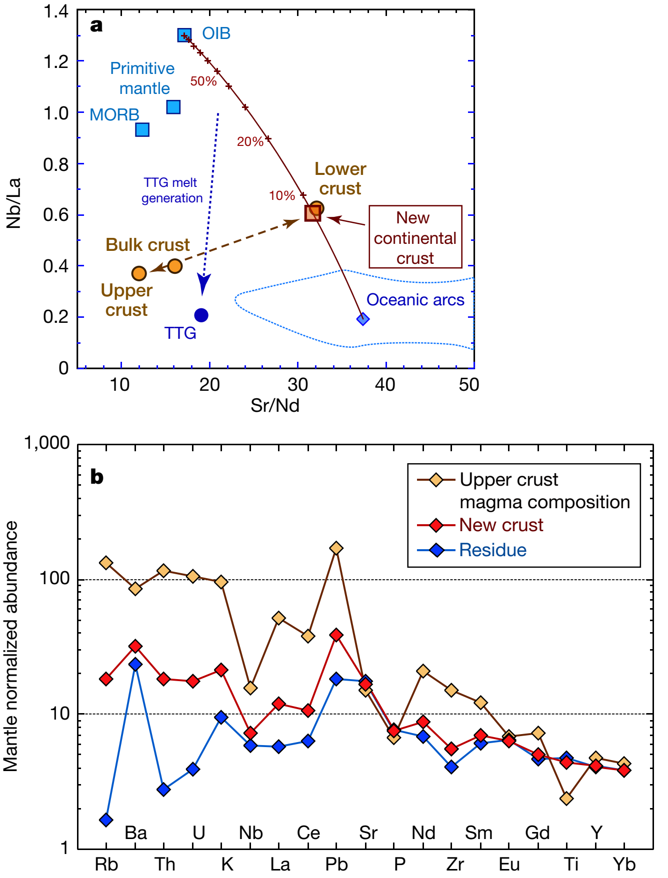

Figure 1 | Trace-element ratios and abundances. a, A plot of the trace-element ratios Nb/La versus Sr/Nd, following refs 13 and 19. It illustrates how the estimated composition of model new continental crust is taken to lie at the intersection of the differentiation trend between the bulk continental crust, the upper and the lower crust, and the array between basalts generated in within-plate and subduction-related settings. The composition of model new continental crust is similar to previous estimates for the lower continental crust[3]. MORB, mid-ocean ridge basalt; OIB, ocean island basalt; and TTG, tonalite-trondhjemite-granodiorite. The steep vector to average TTG illustrates a general model for generating TTG from the remelting of hydrated basalt. **b**, Mantle normalized minor- and trace-element abundances of the estimated composition of model new continental crust and the composition of granites with Eu/Eu* of 0.7 that is very similar to average upper crust[3] (Table 1). The complementary residue composition is also shown. Assuming that the distribution coefficients for elements like Rb and Th are close to zero, the upper crust represents 14% partial melt of new crust, or an analogous degree of fractional crystallization[19].

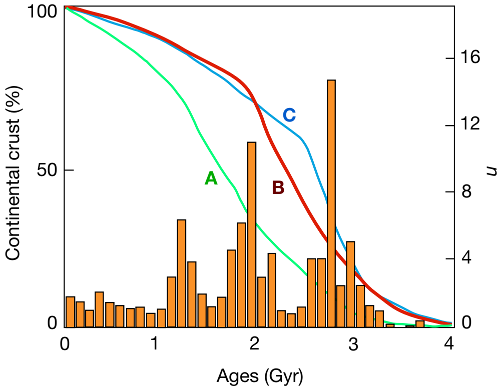

Figure 2 | Generation of continental crust through time as estimated from different approaches. The histogram shows the ages of igneous rocks with mantle-like initial isotope ratios that may therefore be taken to represent the generation of new continental crust (n is the number of samples)[47]. This distribution of ages is contrasted with selected curves for estimates of the increase in the volume of stable continental crust with time. Curve A is based on Nd isotopes in shales[44], B on Pb isotopes[77] and C is from ref. 78.

However, it has proven more challenging to reconcile the picture of crustal evolution obtained from the sedimentary reservoir with the complementary information available from the igneous rock record.

Compilations of emplacement ages of isotopically juvenile (mantle-derived) igneous rocks, exposed on different continents, show striking peaks around 2.7, 1.9 and 1.2 Gyr ago[47,48] (Fig. 2). It is possible that these peaks simply reflect the preservation history of old rocks, but they have been interpreted as periods of accelerated crust formation, each spanning several hundred million years. It is difficult to conceive how global pulses of crustal growth might occur in response to conventional Wilson-style plate tectonics[15], particularly because there are no age peaks in the past billion years when we may surmise that subduction-related magmatism dominated the generation of new crust (Fig. 2). An alternative is that periodicity in igneous activity reflects rapid crust formation during thermal pulses associated with the emplacement of mantle plumes[15,47–50]. There is geological support for this in the plume affinity of basaltic rocks associated with vast tracts of juvenile crust formed over short time periods, such as in the Birimian (2.1 Gyr ago) terranes of West Africa[49] and the Arabian-Nubian Shield[50]. More widely, this reaffirms the distinction between crust generated predominantly in response to deep-seated disturbances in the Earth, and that generated by shallow-level processes, as in some form of plate tectonics.

In sharp contrast, models for the volume of the continental crust result in smooth curves through time, with as yet no sign of marked episodicity[4–12] (Fig. 2). The curves are smooth in part because these models are typically based on radiogenic isotope ratios in sediments and sedimentary rocks. Such samples reflect the history of the crust over long periods of time[43–46,51], but they may also result in hybrid crust generation ages that are the result of mixing processes rather than real magmatic events. Igneous rocks preserve the ages of the specific events in which new continental crust may have been generated (as in Fig. 2), although this record is undoubtedly incomplete and may not be representative of the true growth history of the continents. The key therefore is to have an archive in which the evolution of sediments and igneous rocks can be studied separately.

The accessory mineral zircon offers such an archive. Zircons crystallize from medium- to high-silica magmas and are robust under most crustal conditions, being able to survive intense metamorphism and prolonged sedimentary recycling. Detrital zircons in sediments thus encapsulate the geological history of the continental crust[51–54], back even as far as 4.4 Gyr ago[55,56]. Zircons can be dated precisely using U-Pb isotopes[57], and are amenable to *in situ* analysis of both Hf and O isotopes[51,58]. Hf isotopes are analogous to Nd isotopes in that they indicate whether a rock was derived from the mantle or from the pre-existing crust. However, zircons are less prone to isotope disturbance of the kind that has hampered studies of old bulk rocks[59,60], and they preserve ^{176}Hf/^{177}Hf isotope ratios close to the isotope ratio of the magma at the time of crystallization[61]. This initial ratio can then be used to calculate a model Hf crustal residence age, which is the average time at which the source rock(s) of the magma from which the zircon crystallized were derived from the mantle[52–54,62–66]. These model Hf ages are thus commonly referred to as crust formation ages.

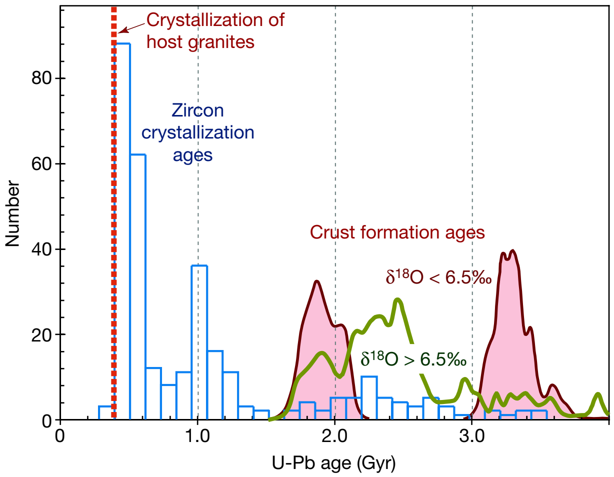

Figure 3 | Histogram of detrital and inherited zircon crystallization ages from southeast Australia overlain with crust formation ages inferred from the Hf-O isotope compositions of the same grains. Data from ref. 66. The thick vertical dashed line shows the crystallization ages of the granitic rocks that contained the inherited zircons. The major magmatic events recorded by the detrital and inherited zircons (blue histogram: ~500 Myr ago and ~1 Gyr ago) were not obviously associated with peaks in the generation of new continental crust, which instead largely occurred in two major episodes at ~1.9 Gyr ago and 3.3 Gyr ago (low-δ^{18}O samples). The samples with higher δ^{18}O (greenish-brown line) are inferred to have hybrid crust formation ages. $\delta^{18}O = [(^{18}O/^{16}O)_{sample}/(^{18}O/^{16}O)_{standard} - 1] \times 1{,}000$; the VSMOW standard is used.

The δ^{18}O values of materials recycled at the Earth's surface are higher than those of igneous rocks that have escaped low temperature interactions with the hydrosphere. Thus oxygen isotopes in zircons can be used to distinguish those that crystallized from magmas that contained a contribution from sediments from those derived from infracrustal igneous source rocks[51,58,65,66] (Fig. 3). The former, those with higher δ^{18}O values, may yield hybrid Hf model ages that are difficult to interpret, and hence may not reflect particular crust-forming events. However, the Hf model ages from zircons with low δ^{18}O values are thought to record the Hf isotope ratios of magmas derived from igneous precursors, whereupon they are much more likely to reflect particular crust forming events.

This approach has recently been illustrated with detrital and inherited zircons from Palaeozoic rocks of southeast Australia[66]. The main observations of this study are summarized in Fig. 3. There were major magmatic episodes at ~0.5 and 1 Gyr ago, but these did not involve the generation of new crust because these crystallization ages do not coincide with the Hf crust formation ages. The zircons with low δ^{18}O are from magmas derived from igneous sources. The Hf model ages of these grains therefore indicate periods of new crustal growth at 1.9 and 3.3 Gyr ago. In contrast, the zircons with high δ^{18}O crystallized from magmas with a sedimentary component, and thus may have mixed age source rocks. The Hf model ages of these zircons peak at 2.2–2.6 Gyr ago, and represent a mixture of the rocks generated at 1.9 and 3.3 Gyr ago. The oxygen isotope data allow us to infer that the 2.1–2.6 Gyr age peak represents mixing in the sedimentary environment, rather than a real crust-forming event. More generally, with this approach we can now contrast the evolution of the igneous and sedimentary reservoirs in the continental crust, and compare the information from zircons with the models for the evolution of the continental crust based on Nd isotope ratios in shales[44].

Figure 4 summarizes the data from zircons and shales from Australia on a graph of the model Nd or Hf age (that is, the crust formation age for each isotope system), plotted against the sedimentation age or the age of the individual zircons, respectively. The decrease in the model Nd age with the age of sedimentation for the shales indicates that significant volumes of new crust were generated as least until the middle Proterozoic. Models for the evolution of the crust were based on assumed episodes of generation of new crust every 0.5 Gyr, and required this new material to have been rapidly eroded into the sedimentary reservoir[44]. Yet the available data from the ages of rocks (Fig. 2) and zircons (Figs 3 and 4) now indicate that the crust may have been generated in fewer and more widespread events, at least in the period 2.9–1.2 Gyr ago.

On Fig. 4, zircons that crystallized from magmas that were new additions to the crust have the same crystallization and crust formation age, and so plot on the black diagonal line. But very few zircons actually plot on that line, which may highlight a time difference between generation of new continental crust and the formation of granitic magmas that can crystallize zircon. Once new crust has been formed, zircons that crystallize from magmas generated from that crust plot on a horizontal array, reflecting the ages of subsequent thermal events. It is very striking that evidence for the generation of new crust at 3.3 Gyr ago is retained in zircons that crystallized over the subsequent 2 Gyr (Fig. 4).

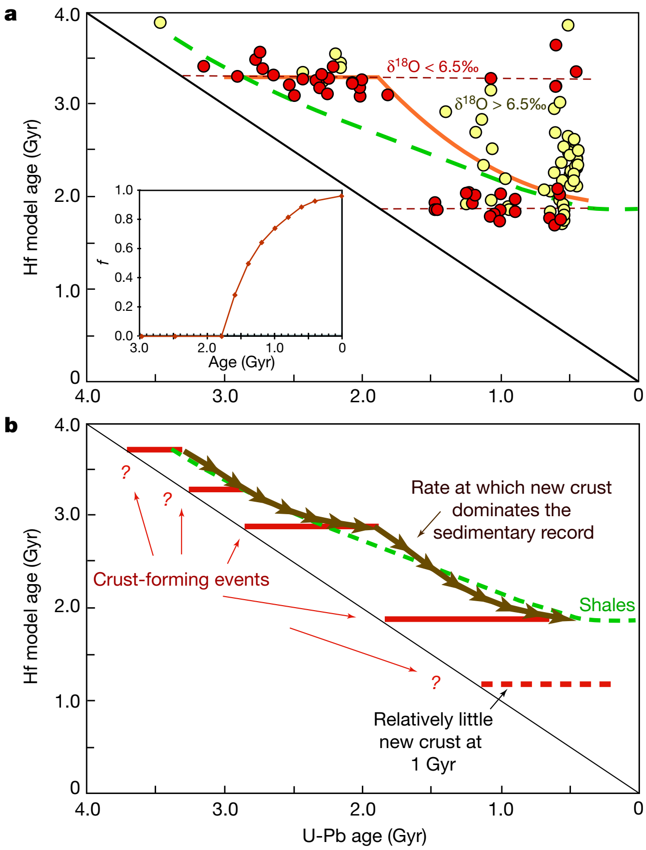

Figure 4 | Hf and U-Pb model ages. a, **b**, Hf model ages, or the Hf crust formation ages, plotted against the U-Pb crystallization ages of different zircon suites[66]. The dashed horizontal lines represent particular times of crust formation, and the diagonal black line indicates 1:1 correspondence between zircon U-Pb crystallization ages and the 'model' crust formation ages inferred from Hf isotopes. **a**, The red circles represent low-$\delta^{18}O$ zircons, and the yellow circles are the high-$\delta^{18}O$ zircons. The latter are inferred to contain a contribution from sediments, and as such their model ages may be hybrids that reflect mixing rather than specific crust forming events. The green dashed line is from a plot of Nd model age versus sedimentation age for average Australian shale compositions[44]. The inset shows the fraction (f) of 1.9-Gyr crust contributing to the sedimentary reservoir as a function of time, as modelled from the high $\delta^{18}O$ zircon trend (orange curve). **b**, An alternative interpretation of the evolution of the upper crust, in which new crust can take up to 1 Gyr to dominate the sedimentary record (arrowed brown curve). The solid red bars mark times of crust generation, whereas the dashed red bar represents a magmatic event in which little new crust formed.

As with Fig. 3, the low-$\delta^{18}O$ zircons in Fig. 4 define the crust-forming events at 1.9 Gyr ago and 3.3 Gyr ago. To reproduce the Nd isotope curve in sediments from only a few crust-forming events implies that the slope of that curve in part reflects the rate at which new crust enters the sedimentary record. Although the data set is still small, the results from the high-$\delta^{18}O$ zircons also indicate that it may take a long time, perhaps up to ~ 1 Gyr, for new crustal material to dominate the sedimentary record (Fig. 4). Such relative rates of erosion of different aged source terrains are difficult to reconcile with the development and envisaged rapid erosion of mountainous areas, as typically form along destructive plate boundaries. The implication is that much new crust is emplaced within the pre-existing crust by under- or intraplating[38,67]. This delays its erosion and contribution to the sedimentary record, and it is consistent with significant volumes of crust being generated in within plate settings.

Rates of generation of the continental crust

Radiogenic isotopes constrain the volumes of continental crust that were stable for long enough for the radiogenic isotopes to change in response to radioactive decay, and the rates of increase in the volume of such stable crust. They do not constrain the overall rates of generation of continental crust. The latter have instead been investigated using magma addition rates at volcanic arcs and intra-plate hotspots, and the residence times of elements in the upper crust. The estimated rates at which basaltic magma is added to the continental crust range from 1.65 $km^3 yr^{-1}$ (ref. 68) to 3.7 $km^3 yr^{-1}$ (ref. 69) if destruction of continental crust by forearc erosion and sediment subduction is taken into account[70]. Such data are inevitably restricted to the recent geologic past, and it is widely accepted that the rates of crust generation were higher earlier in the history of the Earth[71].

Another approach links the annual flux of new material into the continental crust to the residence times of elements in the crust

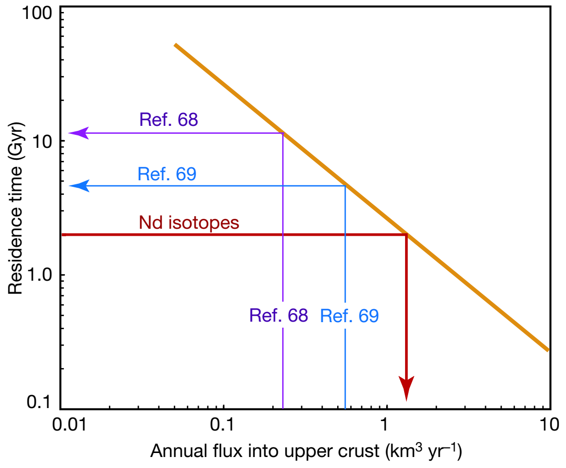

Figure 5 | The relationship between the residence time of elements in the upper crust and the annual rates of crust generation. This was calculated using the present-day volume of continental crust, and assuming that the rate of upper crust generation was 15% of the rate of new crustal addition. For a residence time of 2 Gyr in the upper crust (the model Nd age of the upper crust), the rates of crust generation are >6 times those in the recent geologic past, and 2-3 times greater than the average rates inferred from radiogenic isotopes. Also shown are residence times calculated from crust generation rates obtained from refs 68 and 69.

(Fig. 5)[19]. For the rare earth elements, the residence time can be taken to be the average model Nd age of the upper crust, ~2 Gyr (ref. 44). A critical step is to determine how the rates of generation of the upper crust compare with those for the generation of bulk new continental crust. As discussed above, the annual flux into the upper crust is ~14% of the rate at which new crust is generated. Thus the estimated rates of magma addition to the continental crust of 1.65 and 3.7 $km^3 yr^{-1}$ (refs 68, 69) are equivalent to values of 0.23 and 0.56 $km^3 yr^{-1}$ for the rate of formation of new upper crust. These correspond to upper crustal residence times of 11 Gyr and 4.7 Gyr, respectively, which are unrealistic as they exceed the age of the Earth. These calculations highlight how the rates of continental crust generation, and of the fluxes of differentiated magma into the upper crust, must have been greater earlier in Earth history than in the relatively recent geologic past. In contrast, a residence time of 2 Gyr for elements in the upper crust converts to an average rate of generation of 1.3 $km^3 yr^{-1}$ for the upper crust (Fig. 5), and of 9.3 $km^3 yr^{-1}$ for the bulk continental crust. This value is substantially higher than the average rate of crustal growth if the present volume has been generated since 3.5 Gyr ago, which is 2.2 $km^3 yr^{-1}$. But given that crustal material is recycled into the mantle, the true rates of crust generation are inevitably higher than the rates at which volumes of continental crust were stabilized.

Future directions

There is increasing evidence that the rates of generation of the continental crust, and the ways in which it has been generated, have changed with time. The average rates were higher in the past, and the data from the rock record and from zircons also indicate that there were periods of relatively rapid crustal growth, and periods when much less crust was generated (Fig. 2). The implication is that the pulses of rapid crustal growth were linked to thermal instabilities within the Earth, perhaps manifest as superplumes. Such pulses are much less obvious in the past 1 Gyr of Earth history when it is presumed that the generation and reworking of continental crust were more closely associated with plate tectonics and magmatism along destructive plate margins.

Developments in the *in situ* analysis of zircon have made it possible separately to investigate the igneous and sedimentary reservoirs in the continental crust. These also indicate that significant volumes of continental crust were generated in major episodes in the Archaean and the early Proterozoic, and they highlight that long periods of time may be required before the new crust begins to dominate the sedimentary record. This modifies models for erosion of the continental crust. It is argued that the residence times of elements in the upper crust are much greater than in the lower crust. The annual flux of material into the upper crust can in principle be inferred from its volume and the residence times of elements in the upper crust. A maximum value of the latter is provided by the model Nd age of the upper crust of 2 Gyr, and this indicates that the average rates of crust generation (~9.3 $km^3 yr^{-1}$) are in excess of six times those in the recent geologic past, and 2–3 times greater than the rates inferred from radiogenic isotopes. Simple calculations based on this establish that more than half the K, and a quarter of the Li, in the silicate Earth may have been processed through the continental crust over the past 4 Gyr (ref. 19).

Such arguments bring us back to the difficulties in assessing the past volumes of continental crust, particularly in the Archaean. Relict outcrops of rocks containing exceptionally old zircons, such as at Jack Hills in Australia[55,56], provide little constraint as there is no indication of how representative these zircons may be. The isotope evidence for depleted upper mantle by 3.8 Gyr ago and possibly 4 Gyr ago[72,73] indicates that crust had been extracted by that time, and the parent/daughter element ratios are more readily fractionated in the generation of continental rather than basaltic oceanic crust. Widespread depleted mantle therefore offers encouragement to ideas of significant volumes of continental crust, although it may be residual after mafic crust formation provided that the degree of melting is reasonably small (<10%). Preliminary thermal models indicate that it is difficult to envisage a hotter Earth in which basaltic crust is not remelted to form high-silica crust. New insights may be available from the Nd and Ca isotope ratios in other well dated mineral archives, such as titanite[74], in the Archaean, and the isotope composition of elements fractionated in the hydrosphere and returned to the mantle, such as Li (ref. 75). The ways forward include improving those models for melt generation in the early Archaean, and determining the distribution of crystallization and crust formation ages as they were at different times in the Archaean from integrated Hf-O isotope studies of detrital and inherited zircons in old rocks.

1. Lowe, D. R. & Tice, M. M. Geologic evidence for Archaean atmospheric and climatic evolution: fluctuating levels of CO_2, CH_4 and O_2, with an overriding tectonic control. *Geology* **32**, 493–496 (2004).
2. Kramers, J. D. Global modelling of continent formation and destruction through geological time and implications for CO_2 drawdown in the Archaean Eon. *Spec. Publ. Geol. Soc. Lond.* **199**, 259–274 (2002).
3. Rudnick, R. L. & Gao, S. in *The Crust* (ed. Rudnick, R. L.) 1–64 (Treatise in Geochemistry, Vol. 3, Elsevier, Amsterdam, 2003).
4. Armstrong, R. L. Radiogenic isotopes: the case for crustal recycling on a near steady state no-continental-growth model. *Phil. Trans. R. Soc. Lond. A* **301**, 443–472 (1981).
5. Fyfe, W. S. The evolution of the Earth's crust: modern plate tectonics to ancient hot spot tectonics? *Chem. Geol.* **23**, 89–114 (1978).
6. Hurley, P. M. & Rand, J. R. Pre-drift continental nuclei. *Science* **164**, 1229–1242 (1969).
7. Allègre, C. J. & Lewin, E. Chemical structure and history of the Earth: evidence from global non-linear inversion of isotopic data in a three-box model. *Earth Planet. Sci. Lett.* **96**, 61–88 (1989).
8. Moorbath, S. Age and isotope evidence for the evolution of the continental crust. *Phil. Trans. R. Soc. Lond. A* **288**, 401–413 (1978).
9. O'Nions, R. K., Evensen, N. M. & Hamilton, P. J. Geochemical modelling of mantle differentiation and crustal growth. *J. Geophys. Res.* **84**, 6091–6101 (1979).
10. Jacobsen, S. & Wasserburg, G. J. The mean age of mantle and crustal reservoirs. *J. Geophys. Res.* **84**, 7411–7427 (1979).
11. Kramers, J. D. & Tolstikhin, I. N. Two major terrestrial Pb isotope paradoxes, forward transport modelling, core formation and the history of the continental crust. *Chem. Geol.* **139**, 75–110 (1997).
12. Albarède, F. & Brouxel, M. The Sm/Nd secular evolution of the continental crust and the depleted mantle. *Earth Planet. Sci. Lett.* **82**, 25–35 (1987).
13. Rudnick, R. L. Making continental crust. *Nature* **378**, 573–578 (1995).
14. Taylor, S. R. & McLennan, S. M. *The Continental Crust: Its Composition and Evolution* (Blackwell, Malden, MA, 1985).
15. Albarède, F. The growth of continental crust. *Tectonophysics* **296**, 1–14 (1998).
16. Arculus, R. J. Origins of the continental crust. *J. Proc. R. Soc. New South Wales* **132**, 83–110 (1999).
17. Kemp, A. I. S. & Hawkesworth, C. J. in *The Crust* (ed. Rudnick, R. L.) 349–410 (Treatise in Geochemistry, Vol. 3, Elsevier, Amsterdam, 2003).
18. Plank, T. Constraints from thorium/lanthanum on sediment recycling at subduction zones and the evolution of continents. *J. Petrol.* **46**, 921–944 (2005).
19. Hawkesworth, C. J. & Kemp, A. I. S. The differentiation and rates of generation of the continental crust. *Chem. Geol.* **226**, 134–143 (2006).
20. Turner, S. P. *et al.* Time-scales of magma formation, ascent and storage beneath subduction-zone volcanoes. *Phil. Trans. R. Soc. Lond. A* **358**, 1443–1464 (2000).
21. Barth, M., McDonough, W. F. & Rudnick, R. L. Tracking the budget of Nb and Ta in the continental crust. *Chem. Geol.* **165**, 197–213 (2000).
22. Sun, S.-s & McDonough, W. F. Chemical and isotopic systematics of oceanic basalts: implications for mantle composition and processes. *Spec. Publ. Geol. Soc. Lond.* **42**, 313–345 (1989).
23. Plank, T. & Langmuir, C. H. The geochemical composition of subducting sediment and its consequences for the crust and mantle. *Chem. Geol.* **145**, 325–394 (1998).
24. Hawkesworth, C. J., Turner, S. P., McDermott, F., Peate, D. W. & van Calsteren, P. U-Th isotopes in arc magmas: implications for element transfer from the subducted crust. *Science* **276**, 551–555 (1997).
25. Rudnick, R. L. & Fountain, D. Nature and composition of the continental crust: a lower crustal perspective. *Rev. Geophys.* **33**, 267–309 (1995).
26. Martin, H. in *Archaean Crustal Evolution* (ed. Condie, K. C.) 205–260 (Elsevier, Amsterdam, 1994).
27. Rapp, R. P. & Watson, E. B. Dehydration melting of metabasalt at 8–32 kbar: implications for continental growth and crust-mantle recycling. *J. Petrol.* **36**, 891–931 (1995).
28. Condie, K. C. in *Archaean Crustal Evolution* (ed. Condie, K. C.) 85–120 (Elsevier, Amsterdam, 1994).
29. Vervoort, J. D. & Patchett, P. J. Behaviour of hafnium and neodymium isotopes in the crust: constraints from Precambrian crustally derived granites. *Geochim. Cosmochim. Acta* **60**, 3717–3723 (1996).
30. Vervoort, J. D. *et al.* Hf-Nd isotopic evolution of the lower crust. *Earth Planet. Sci. Lett.* **181**, 115–129 (2000).

31. Kay, R. W. & Kay, S. M. Creation and destruction of lower continental crust. *Geol. Rundsch.* **80**, 259–278 (1991).
32. Kay, R. W. & Kay, S. M. Delamination and delamination magmatism. *Tectonophysics* **219**, 177–189 (1993).
33. Jull, M. & Kelemen, P. B. On the conditions for lower crustal convective instability. *J. Geophys. Res.* **106**, 6423–6446 (2001).
34. Ellam, R. M. & Hawkesworth, C. J. Is average continental crust generated at subduction zones? *Geology* **16**, 314–317 (1988).
35. Arndt, N. T. & Goldstein, S. L. An open boundary between lower continental crust and mantle: its role in crust formation and crustal cycling. *Tectonophysics* **161**, 201–212 (1989).
36. Hofmann, A. W. Mantle geochemistry: the message from oceanic volcanism. *Nature* **385**, 219–229 (1997).
37. Petford, N. & Atherton, M. Na-rich partial melts from newly underplated basaltic crust: the Cordillera Blanca Batholith, Peru. *J. Petrol.* **37**, 1491–1521 (1996).
38. Voshage, H. *et al.* Isotopic evidence from the Ivrea Zone for a hybrid lower crust formed by magmatic underplating. *Nature* **347**, 731–736 (1990).
39. Annen, C. & Sparks, R. S. J. Effects of repetitive emplacement of basaltic intrusions on thermal evolution and melt generation in the crust. *Earth Planet. Sci. Lett.* **203**, 937–955 (2002).
40. Bergantz, G. W. & Dawes, R. in *Magmatic Systems* (ed. Ryan, M. P.) 291–317 (Academic, San Diego, CA, 1994).
41. Zandt, G. *et al.* Active foundering of a continental arc root beneath the southern Sierra Nevada in California. *Nature* **431**, 41–46 (2004).
42. Moore, V. M. & Wiltschko, D. V. Syncollisional delamination and tectonic wedge development in convergent orogens. *Tectonics* **23**, doi:10.1029/2002TC001430 (2004).
43. McLennan, S. M. Relationships between the trace element composition of sedimentary rocks and upper continental crust. *Geochem. Geophys. Geosyst.* **2**, doi:10.1029/2000GC000109 (2001).
44. Allègre, C. J. & Rousseau, D. The growth of continents through geological time studied by Nd isotope analysis of shales. *Earth Planet. Sci. Lett.* **67**, 19–34 (1984).
45. O'Nions, R. K., Hamilton, P. J. & Hooker, P. J. A Nd isotope investigation of sediments related to crustal development in the British Isles. *Earth Planet. Sci. Lett.* **63**, 229–240 (1983).
46. Davies, G., Gledhill, A. & Hawkesworth, C. Upper crustal recycling in southern Britain: evidence from Nd and Sr isotopes. *Earth Planet. Sci. Lett.* **75**, 1–12 (1985).
47. McCulloch, M. T. & Bennett, V. C. Progressive growth of the Earth's continental crust and depleted mantle: geochemical constraints. *Geochim. Cosmochim. Acta* **58**, 4717–4738 (1994).
48. Condie, K. C. Episodic continental growth and supercontinents: a mantle avalanche connection? *Earth Planet. Sci. Lett.* **163**, 97–108 (1998).
49. Boher, M., Abouchami, W., Michard, A., Albarede, F. & Arndt, N. T. Crustal growth in West Africa at 2.1 Ga. *J. Geophys. Res.* **97**, 345–369 (1992).
50. Stein, M. & Hofmann, A. W. Mantle plumes and episodic crustal growth. *Nature* **372**, 63–68 (1994).
51. Valley, J. W. *et al.* 4.4 billion years of crustal maturation: oxygen isotope ratios of magmatic zircon. *Contrib. Mineral. Petrol.* **150,** 561–580 doi:10.1007/s00410-005-0025-8 (2005).
52. Amelin, Y., Lee, D-G., Halliday, A. N. & Pidgeon, R. T. Nature of the Earth's earliest crust from hafnium isotopes in single detrital zircons. *Nature* **399**, 252–255 (1999).
53. Bodet, F. & Scharer, U. Evolution of the SE-Asian continent from U-Pb and Hf isotopes in single grains of zircon and baddeleyite from large rivers. *Geochim. Cosmochim. Acta* **64**, 2067–2091 (2000).
54. Vervoort, J. D., Patchett, P. J., Gehrels, G. E. & Nutman, A. P. Constraints on early Earth differentiation from hafnium and neodymium isotopes. *Nature* **379**, 624–627 (1996).
55. Wilde, S. A., Valley, J. W., Peck, W. H. & Graham, C. M. Evidence from detrital zircons for the existence of continental crust and oceans on the Earth 4.4 Gyr ago. *Nature* **409**, 175–178 (2001).
56. Harrison, T. M. *et al.* Heterogeneous Hadean hafnium: Evidence of continental crust at 4.4 to 4.5 Ga. *Science* **310**, 1947–1950 (2005).
57. Ireland, T. R. & Williams, I. S. in *Zircon* (eds Hanchar, J. M. & Hoskin, P. W. O.) 215–238 (Reviews in Mineralogy and Geochemistry Vol. 53, Mineralogical Society of America, Washington DC, 2003).
58. Valley, J. W., Chiarenzelli, J. R. & McLelland, J. M. Oxygen isotope geochemistry of zircon. *Earth Planet. Sci. Lett.* **126**, 187–206 (1994).
59. Blichert-Toft, J., Albarède, F., Rosing, M., Frei, R. & Bridgwater, D. The Nd and Hf isotopic evolution of the mantle through the Archean. Results from the Isua supracrustals, West Greenland, and from the Birimian terranes of West Africa. *Geochim. Cosmochim. Acta* **63**, 3901–3914 (1999).
60. Moorbath, S., Whitehouse, M. J. & Kamber, B. S. Extreme Nd-isotope heterogeneity in the early Archaean — fact or fiction? *Chem. Geol.* **135**, 213–231 (1997).
61. Patchett, J. P., Kouvo, O., Hedge, C. E. & Tatsumoto, M. Evolution of continental crust and mantle heterogeneity: evidence from Hf isotopes. *Contrib. Mineral. Petrol.* **78**, 279–297 (1981).
62. Knudsen, T-L., Griffin, W. L., Hartz, E. H., Andresen, A. & Jackson, S. E. In situ hafnium and lead isotope analysis of detrital zircons from the Devonian sedimentary basin of NE Greenland: a record of repeated crustal reworking. *Contrib. Mineral. Petrol.* **141**, 83–94 (2001).
63. Griffin, W. L., Belousova, E. A., Shee, S. R., Pearson, N. J. & O'Reilly, S. Y. Archean crustal evolution in the northern Yilgarn Craton: U–Pb and Hf-isotope evidence from detrital zircons. *Precambr. Res.* **131**, 231–282 (2004).
64. Condie, K. C., Beyer, E., Belousva, E., Griffin, W. L. & O'Reilly, S. Y. U-Pb isotopic ages and Hf isotopic composition of single zircons: the search for juvenile Precambrian continental crust. *Precambr. Res.* **139**, 42–100 (2005).
65. Hawkesworth, C. J. & Kemp, A. I. S. Using hafnium and oxygen isotopes in zircons to unravel the record of crustal evolution. *Chem. Geol.* **226**, 144–162 (2006).
66. Kemp, A. I. S., Hawkesworth, C. J., Paterson, B. A. & Kinny, P. D. Episodic growth of the Gondwana supercontinent from hafnium and oxygen isotope ratios. *Nature* **439**, 580–583 (2006).
67. Rudnick, R. L. Growing from below. *Nature* **347**, 711–712 (1990).
68. Reymer, A. & Schubert, G. Phanerozoic addition rates to the continental crust and crustal growth. *Tectonics* **3**, 63–77 (1984).
69. Clift, P. & Vannuchi, P. Controls on tectonic accretion versus erosion in subduction zones: Implications for the origin and recycling of the continental crust. *Rev. Geophys.* **42**, 1–31 (2004).
70. von Huene, R. & Scholl, D. W. Observations at convergent margin concerning sediment subduction, subduction erosion and the growth of the continental crust. *Rev. Geophys.* **29**, 279–316 (1991).
71. Davies, G. F. Punctuated tectonic evolution of the Earth. *Earth Planet. Sci. Lett.* **136**, 363–379 (1995).
72. Bowring, S. A. & Housh, T. The Earth's early evolution. *Science* **269**, 1535–1540 (1995).
73. Vervoort, J. D. & Blichert-Toft, J. Evolution of the depleted mantle: Hf isotope evidence from juvenile rocks through time. *Geochim. Cosmochim. Acta* **63**, 533–556 (1999).
74. Storey, C. D., Jeffries, T. E. & Smith, M. Common lead-corrected laser ablation ICP-MS U-Pb systematics and geochronology of titanite. *Chem. Geol.* **227**, 37–52 (2006).
75. Elliott, T. E., Jeffcoate, A. & Bouman, C. The terrestrial Li isotope cycle: light weight constraints on mantle convection. *Earth Planet. Sci. Lett.* **220**, 231–245 (2004).
76. Jaupart, C. & Mareschal, J-C. in *The Crust* (ed. Rudnick, R. L.) 65–84 (Treatise in Geochemistry, Vol. 3, Elsevier, Amsterdam, 2003).
77. Collerson, K. D. & Kamber, B. S. Evolution of the continents and the atmosphere. *Science* **283**, 1519–1522 (1999).
78. Taylor, S. R. & McLennan, S. M. The geochemical evolution of the continental crust. *Rev. Geophys.* **33**, 241–265 (1995).

Acknowledgements We thank J. Blundy, T. Elliott, K. Gallagher, G. Helffrich, J. Kramers and S. Turner for many discussions on these and related topics, and for reading early versions of the manuscript. R. Arculus and J. Vervoort provided comments and suggestions that greatly improved the paper. The analytical work relied heavily on the enthusiasm and expertise of the Edinburgh Ion Microprobe Facility, and of C. Coath and G. Foster in Bristol.

Author Contributions Both authors contributed equally to this paper**.**

Author Information Reprints and permissions information is available at www.nature.com/reprints. The authors declare no competing financial interests. Correspondence should be addressed to C.J.H. (c.j.hawkesworth@bristol.ac.uk).