The zircon archive of continent formation through time

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Abstract: The strong resilience of the mineral zircon and its ability to host a wealth of isotopic information make it the best deep-time archive of Earth's continental crust. Zircon is found in most felsic igneous rocks, can be precisely dated and can fingerprint magmatic sources; thus, it has been widely used to document the formation and evolution of continental crust, from pluton- to global-scale. Here, we present a review of major contributions that zircon studies have made in terms of understanding key questions involving the formation of the continents. These include the conditions of continent formation on early Earth, the onset of plate tectonics and subduction, the rate of crustal growth through time and the governing balance of continental addition v. continental loss, and the role of preservation bias in the zircon record.

Supplementary material: A compilation used in this study of previously published detrital zircon U-Pb-Hf isotope data are available at http://www.geolsoc.org.uk/SUP18791

The mineral zircon ($ZrSiO_4$) records information about its growth and thermal history, and is an ideal mineral for geochronology given that it readily incorporates uranium, but very little to no lead during crystallization. Also, the low diffusion rates of these elements allows for further resilience during subsequent diagenesis and metamorphism. Furthermore, zircon can recrystallize under certain conditions, meaning that it often records metamorphic as well as magmatic history. Along with U–Pb, zircon stores additional isotopic and elemental information that can be used to interpret the history of its host rock (e.g. Hf and O isotopes). Zircon is very resistant to physiochemical alteration; this resilience allows detrital zircons to survive multiple sedimentary cycles. Detrital zircon age spectra can be used to determine the provenance of a host sedimentary rock (or river), but of more interest to this review paper, they can be used to sample a broad region of continental upper crust, allowing study of a vast area with only minimal sampling (e.g. Rainbird *et al.* 1992; Fedo *et al.* 2003). For these reasons, zircon arguably forms the best archive of the Earth's continental crust (Hawkesworth *et al.* 2010; Cawood *et al.* 2013).

As evident in many of the papers of this volume (e.g. Boekhout *et al.* 2013; Claesson *et al.* 2014; Kleinhanns *et al.* 2013; Hiess *et al.* 2014; Lancaster *et al.* 2014; Petersson *et al.* 2013), zircon is commonly used to characterize a wide variety of geological scenarios, from individual igneous units to whole terranes and entire orogens. Detrital zircons from sediments form the oldest known crustal materials on Earth (4.37 Ga; Wilde *et al.* 2001; Valley *et al.* 2014), much older than the oldest crust itself; estimated at 3.92 Ga (Mojzsis *et al.* 2014) to 4.03 Ga (Bowring & Williams 1999). Thus, they provide important information regarding Earth's early evolution. Detrital zircons from large modern rivers have been measured in various different studies. These have been used as a broad representation of the continental crust for studying its age and composition, and from this, the patterns of continental growth and evolution (e.g. Wang *et al.* 2009, 2011; Iizuka *et al.* 2010, 2013). In this paper, we review some of the major contributions that zircon studies have made to our understanding of the formation and evolution of continental crust. One of the key questions regarding the use of zircon for this method is the degree to which particular zircon datasets are biased towards particular settings, ages or regions. We include a compilation of *c.* 42 000 hafnium isotope and *c.* 6000 oxygen isotope analyses from zircons; it is assumed that these larger datasets offer less bias than the smaller datasets used in some previous studies (e.g. Condie *et al.* 2005, 2011; Iizuka *et al.* 2005, 2010; Valley *et al.* 2005; Belousova *et al.* 2010; Dhuime *et al.* 2012).

Our compilation of published zircon U–Pb and Hf isotope analyses is shown in Figure 1. As is evident from this plot, much of Earth history has been sampled in this dataset; however, to highlight the potential bias that still exists in the Hadean to Eoarchaean, zircons from Jack Hills are highlighted separately (Amelin *et al.* 1999; Harrison

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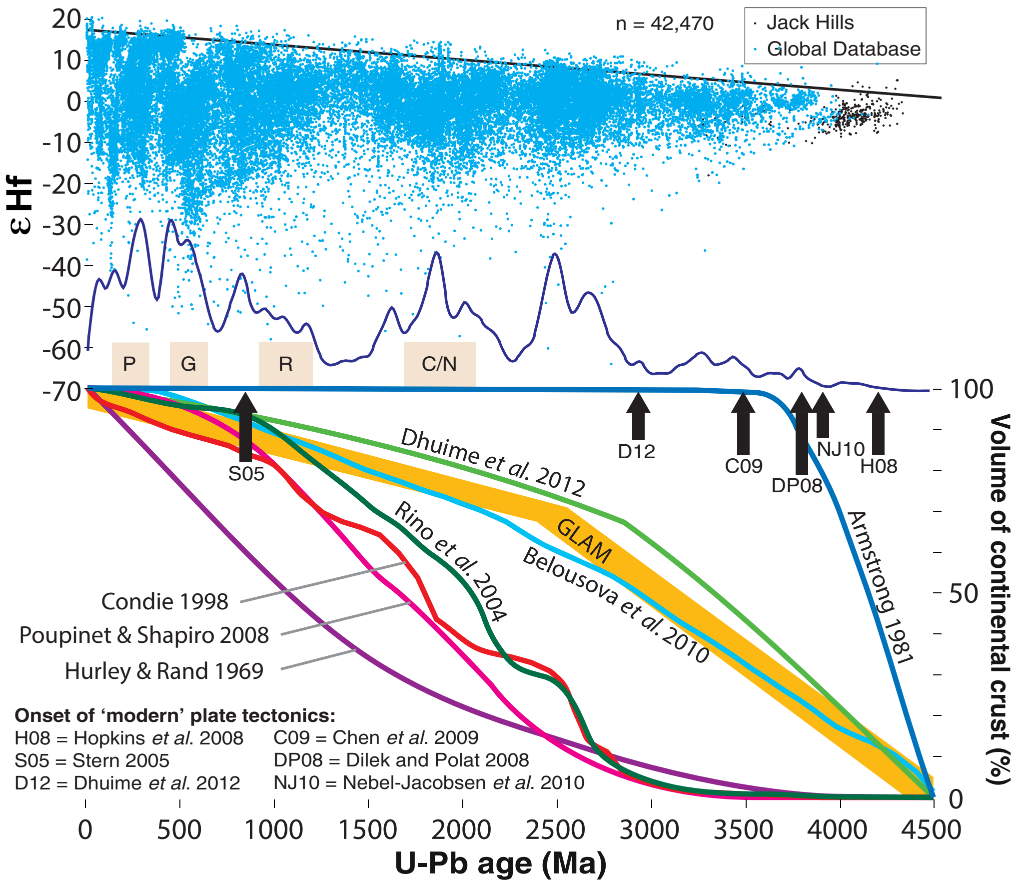


Fig. 1. (**a**) $\varepsilon Hf_{(T)}$ v. time (Ma) for a global database ($n = c.$ 42 000) of zircon data compiled in this study, with those derived from the Jack Hills region of Australia highlighted, and a kernel density estimate of the zircon U–Pb ages (dark blue curve; using Vermeesch 2012). Pale grey bars represent timing of supercontinent formation, P, Pangea; G, Gondwana; R, Rodinia; and C/N, Columbia (a.k.a. Nuna). (**b**) Estimates of the onset of subduction and modern-day-style plate tectonics shown as arrows, and selected published crustal growth curves (GLAM, Global Lithospheric Mapping; Begg *et al.* 2009; Belousova *et al.* 2010).

et al. 2005, 2008; Blichert-Toft & Albarède 2008; Kemp *et al.* 2010). These zircons are from a geographically restricted area and dominate the record for the early Earth. Also shown in Figure 1 are some salient features relating to themes discussed in this paper. These include the rate of crustal growth, determined from the distribution of rock ages (Hurley & Rand 1969; Condie 1998) and lithosphere ages (Poupinet & Shapiro 2009), zircons (i.e. Rino *et al.* 2004; Belousova *et al.* 2010; Dhuime *et al.* 2012) and a hypothetical crustal growth model (Armstrong & Harmon 1981) that posits that the volume of continental crust has remained relatively constant since the early Archaean. The onset of subduction and modern-day-style plate tectonics is still debated, and various suggestions are labelled in Figure 1. Finally, the formation, evolution and preservation of continental crust have been linked to the formation of supercontinents (e.g. Condie 2004; Hawkesworth *et al.* 2009); estimates of supercontinent amalgamation are also shown in Figure 1.

Hf and O isotope systematics

Our understanding of the formation and growth of the continental crust through time has greatly benefited through the development of multi-isotopic analysis in zircon; and more recently through additional accessory minerals such as titanite, monazite, rutile and apatite (e.g. Garçon *et al.* 2011; McFarlane & McCulloch 2007; McFarlane & McKeough 2013; Ewing *et al.* 2014; Hammerli *et al.* 2014). As well as U–Pb to provide a crystallization age, the two most common isotopic systems measured in zircon are lutetium–hafnium and oxygen ($^{18}O/^{16}O$). The following section discusses their systematics and typical uses.

Lu–Hf systematics of zircon

The Lu–Hf isotopic system can be used to track the isotopic differentiation of mantle and crustal reservoirs. ^{176}Lu decays via β-decay to radiogenic ^{176}Hf with a half-life of *c.* 36 billion years (Kossert *et al.* 2013). In magmatic systems, Lu fractionates from Hf. During partial melting of the mantle to form the continental crust, Hf is more incompatible than Lu, so the Lu/Hf ratio in the crust is lowered compared with that of the mantle that the crust was derived from. This subsequently leads to the mantle becoming more radiogenic through time, and the crust becoming less radiogenic. In terrestrial materials (including zircon) the $^{176}Hf/^{177}Hf$ (daughter/stable) ratio is compared with that of the chondritic uniform reservoir (CHUR), which for Hf (a refractory element) has generally been assumed to be a fair representation of the bulk silicate Earth. This was called into question by the work of Caro & Bourdon (2010), who estimated a non-chondritic bulk silicate Earth Hf isotope composition based on Nd isotopes in planetary material, but normalization to CHUR presently remains the norm. The notation εHf refers to this normalization to CHUR (Patchett & Tatsumoto 1981; Blichert-Toft & Albarède 1997), and is based on 10 000 times the difference between the sample and CHUR, that is, 1 epsilon unit is equivalent to 100 ppm change in the $^{176}Hf/^{177}Hf$ ratio. Zircons preferentially incorporate Hf compared with the rare-earth elements such as Lu (Fujimaki 1986) and only a minor amount of ^{176}Lu during crystallization; this means that the ratio measured today is only slightly different from that of the time of zircon crystallization. Additionally, the Hf isotope composition in a zircon is very stable, because Hf is incorporated into the zircon lattice, making it resistant to diffusion and contamination.

A common use of the Hf system in zircon is to date the timing of extraction from a mantle reservoir to form the host magma; this is known as the zircon Hf model age. The mantle reservoir typically used is that of the 'depleted mantle', which itself is based on measurement of primitive mantle-derived magmas (e.g. Griffin *et al.* 2000); the resulting ages are known as depleted mantle model ages (T_{DM}). In addition to the depleted mantle model, the use of other mantle reservoirs has also been proposed, for example 'new crust' model ages (Dhuime *et al.* 2011) are essentially based on a model of enriched mantle derived from island arc magmas.

Calculation of model ages

For zircons, model ages are generally two-stage; this means that the initial $^{176}Hf/^{177}Hf$ ratio at the time of zircon crystallization is determined first, then the age of mantle extraction is determined using a back-projection to the mantle reservoir (see Fig. 2). Three main assumptive variables come into play in calculating Hf-depleted mantle model ages (T_{DM}): (a) a model of the depleted mantle, which typically assumes 4.5 Ga of crustal differentiation (Vervoort & Blichert-Toft 1999; Griffin *et al.* 2000); (b) a ^{176}Lu decay constant (λ), with estimates ranging from 1.867×10^{-11} to 1.983×10^{-11} per year (see Scherer *et al.* 2001; Bizzarro *et al.* 2003; Söderlund *et al.* 2004); and (c) a $^{176}Lu/^{177}Hf$ ratio of the parent magma that defines the isotopic evolution through time in regards to $^{176}Hf/^{177}Hf$ (typically assumed to be between 0.009 and 0.022; e.g. Nebel *et al.* 2007). In regards to the latter, various approaches have been used to make this projection, for example: (a) using the measured Lu/Hf ratio from the zircon (generally between 0.0001 and 0.005); (b) using an average upper crustal value (e.g. 0.015; Griffin *et al.* 2002); (c) using a higher Lu/Hf ratio consistent with mafic lithologies (e.g. 0.022; Nebel *et al.* 2007; Pietranik *et al.* 2008); (d) using the whole rock Lu/Hf from which the zircon was separated (e.g. Andersen *et al.* 2002); or (e) with the assumption of a closed-system single source, using the slope of an array of zircon Hf data in Hf vs

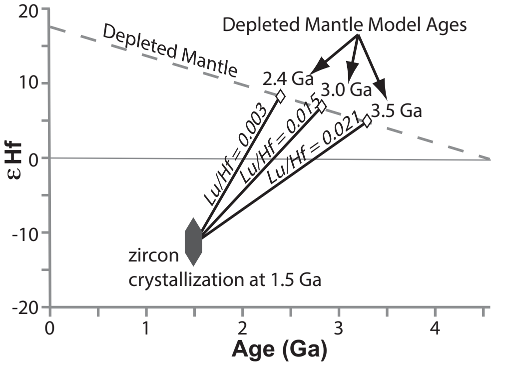


Fig. 2. Calculation of two-stage depleted mantle model ages of a zircon crystallized at 1.5 Ga, using various Lu/Hf ratios. A Lu/Hf ratio of 0.021 corresponds to a mafic host magma, whereas 0.015 relates to more felsic melts and an average of the continental crust, and 0.003 corresponds to an average Lu/Hf ratio of zircon (see text for further explanation).

time space (equivalent to Lu/Hf) to choose the Lu/Hf ratio for model age calculation (e.g. Kemp *et al.* 2006; Pietranik *et al.* 2008; Lancaster *et al.* 2014). It can generally be assumed that the measured Lu/Hf value in zircon and mafic rocks provides the minimum and maximum depleted mantle model ages, respectively (Fig. 2). Lastly, utilizing the recommendations of Ickert (2013), uncertainties for depleted mantle model ages can be determined by incorporating the known uncertainties in the ^{176}Lu decay constant ($\pm$0.47 ε units) and CHUR ($\pm$0.39 ε units at 0 Ma and $\pm$0.18 ε units at 4.5 Ga). This results in a minimum uncertainty of $\pm$0.34 ε units on depleted mantle model ages (equivalent to up to 250 myr).

Use and abuse of model ages

Model ages were first adopted with the Sm–Nd isotopic system (DePaolo 1981), and many studies adopted this approach to produce model age estimates of continental growth of specific regions (Bennett & DePaolo 1987; Dickin & McNutt 1989; DePaolo *et al.* 1991) and of the global sedimentary budget (Allègre & Rousseau 1984; Goldstein & Jacobsen 1988; McLennan *et al.* 1990; Vervoort *et al.* 1999). However, it was noted early on (Arndt & Goldstein 1987) that these model ages represent an average of the sources that fed a crystallizing magma, and that many magmatic systems have multiple sources. Thus, model ages can be thought of as hybrid ages, typically formed from a mixture of at least two mantle- and crustal-derived sources. In sedimentary rocks, model ages represent mixtures of these hybrid ages. For Lu–Hf, the same complexities exist, with average model ages still representing hybrid ages. The advantage of Lu–Hf is that a suite of zircons sampled from a single magmatic unit may exhibit a range of Hf isotope signatures if the zircons crystallized prior to and quicker than homogenization of the magma's chemistry. This means that multiple sources to a magma may be better fingerprinted with zircon than with whole-rock compositions (e.g. Griffin *et al.* 2002; Kemp *et al.* 2007; Roberts *et al.* 2013; Hiess *et al.* 2014).

Despite the caveats of model ages, there remains a popular trend to calculate depleted mantle model ages from zircon suites, and to treat these model ages as crustal growth events. This is based on the premise that this age reflects the age of extraction from the depleted mantle reservoir, and that subsequent zircon crystallization occurred much later; the difference between these events is commonly known as the crustal residence time (e.g. Wang *et al.* 2009; Lancaster *et al.* 2011). Although the underlying reasoning behind this premise remains valid, it remains uncertain how often a host magma will have remained a closed system since extraction from the mantle. For example, during the evolution from a mafic magma extracted from the mantle to the final crystallization of a host intermediate to felsic magma, or during the complex process of crustal melting from migmatization to final emplacement, there are multiple stages at which deviations from this presumed closed system are likely to have occurred.

The depleted mantle model does provide a convenient end-member to work with, but in fact, the mantle source to most magmatic systems is extremely heterogeneous and often enriched through subduction of older crust (sediments) and their partial melting and/or metasomatism (e.g. Armstrong 1971; Hawkesworth *et al.* 1993; Plank & Langmuir 1993; Elliott *et al.* 1997). This enrichment may directly affect a magmatic system, that is, through partial melting of the mantle wedge in a fore-arc region above a subducting slab, or through the long-term entrainment of recycled materials into the mantle and their remobilization into deep-seated mantle melting regions (i.e. 'ocean island basalt' magmas in hotspots). The effect of this sediment recycling has been well documented and studied for several decades, and can be seen as perturbation of radiogenic isotope signatures of mafic magmas in island arcs that have not undergone any crustal contamination (e.g. Hawkesworth *et al.* 1979; Davidson 1983). Where subducted sediments undergo partial melting and when they include zircon, the effect is most intensified. For example, in the Banda arc where magmatism and crustal growth is essentially at 0 Ma, model ages range from 100 to greater than 1000 Ma (see Fig. 3;

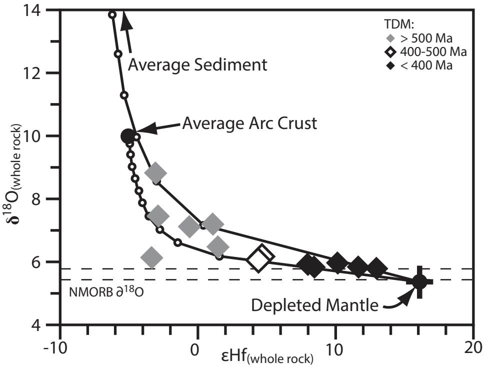


Fig. 3. Whole-rock εHf v. δ^{18}O for Banda arc volcanics, and modelled mixing curves (after Nebel *et al.* 2011). Values are coded so that symbols represent calculated Hf model ages, and highlight that old model ages can be recorded from modern magmas that involve little to no crustal recycling.

Nebel *et al.* 2011). Most island arcs have a degree of enrichment that can be attributed to subducted sediment; for this reason, their model ages average *c.* 200 Ma, rather than 0 Ma (calculated from Dhuime *et al.* 2011).

As described above, another parameter involved in the model age calculation is the Lu/Hf ratio of the host magma. Depending on the age and signature of the zircon, this can lead to variations on the model age of many hundreds of Ma (e.g. Nebel *et al.* 2007). Studies from specific plutons, arcs, orogens or terranes will sometimes use an apparent array of Hf data in εHf-time space, if the Lu/Hf composition of this array represents a typical magmatic value. The Lu/Hf ratio of this array can then be used to calculate the model age of this suite, based on the assumption that the zircons represent closed-system reworking and crystallization of a protolith of a single mantle extraction age (e.g. Kemp *et al.* 2006; Pietranik *et al.* 2008; Petersson *et al.* 2013). In these cases, it can be argued that the model ages are more robust than those derived from scattered data lacking an array. Further comments on model ages are included at pertinent points in later sections.

Oxygen isotopes

The stable oxygen isotopic composition of zircon is preserved in non-metamict crystals from the time of crystallization, and unlike U–Pb and Hf isotopes is independent of time. The oxygen isotope composition, specifically $^{18}O/^{16}O$ (reported as $\delta^{18}O$; normalized to Vienna Standard Mean Ocean Water), is generally unaffected by diagenetic and metamorphic processes; if the zircon domain analysed has suffered radiation damage, deviation of the $\delta^{18}O$ composition can occur (Valley *et al.* 1994, 2005; Valley 2003). The $\delta^{18}O$ of a magma will reflect that of its source rock; fractional crystallization will increase the whole-rock composition ($\delta^{18}O_{WR}$), but the composition of the zircon ($\delta^{18}O_{zircon}$) will remain within *c.* 0.5‰ of the whole rock value since $\Delta^{18}O_{WR-zircon}$ increases at nearly the same rate as $\delta^{18}O_{WR}$ (Valley *et al.* 2005).

The $\delta^{18}O$ composition of the mantle is generally assumed to be fairly homogeneous (Eiler 2001); zircons crystallized from uncontaminated (asthenospheric) mantle should have a value of $+5.3 \pm 0.6$ ‰ (2σ) (Valley *et al.* 1998). In general, low-temperature interaction with meteoric water leads to elevated $\delta^{18}O$ values, and high-temperature interaction leads to low $\delta^{18}O$ values; this is reflected in the range of $\delta^{18}O$ compositions of crustal materials (see Fig. 4). The only low $\delta^{18}O$ materials are meteoric water itself, and materials that have undergone high-temperature alteration with meteoric water or seawater. Subducting oceanic crust can have high $\delta^{18}O_{WR}$ (relative to mantle) values in its upper portion, which reflects low-temperature slab–seawater interaction, and the lower part of the slab is generally thought to have low values owing to high-temperature interaction with percolating water (see Eiler 2001). The continental crust on average has compositions reflecting mixing between igneous mantle-derived material and sediments (see Fig. 4). Sediments that are being subducted and recycled into the mantle will have variably high $\delta^{18}O_{WR}$ depending on the abundance of limestone to silicic material.

Zircon from igneous rocks will have a mantle-like value unless a component of the host magma

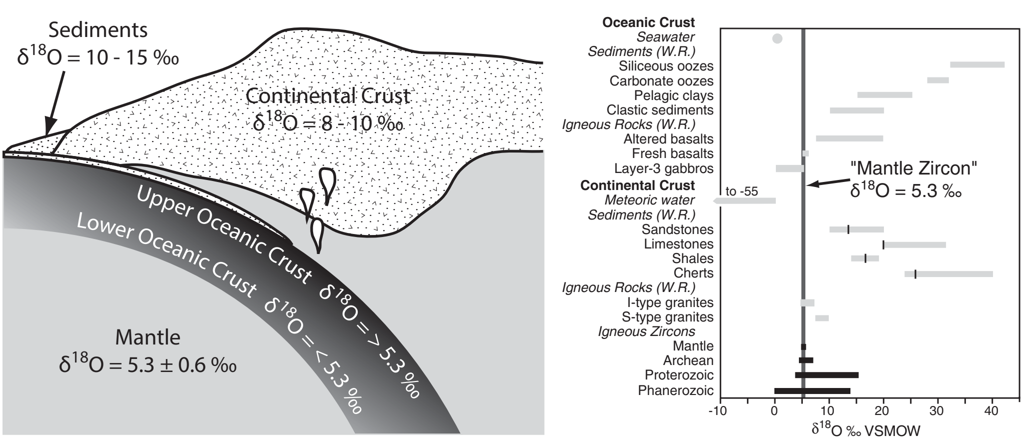


Fig. 4. Average $\delta^{18}O$ of various reservoirs that relate to zircon systematics (modified from King *et al.* 1998) (left), and typical $\delta^{18}O$ of a variety of reservoirs, rocks and sediments (after Eiler 2001; Valley *et al.* 2005; Spencer *et al.* 2014) (right).

has lower or higher than mantle values, that is, reflecting high- or low-temperature alteration, respectively. Since many granites will assimilate some country rock at one or multiple stages of their magmatic evolution, they typically have slightly elevated $\delta^{18}O$ compositions. Granites that have a large component of sedimentary rock in their protolith, typified by S-types (e.g. Chappell & White 1992), have the highest $\delta^{18}O$ values of igneous rocks (Fig. 4). A range of volcanic to hypabyssal rocks have lower than mantle values; these have assimilated material that has undergone high-temperature alteration with meteoric water after it has percolated into the upper crust, and are often associated with caldera forming magmatic suites (e.g. Yellowstone; Bindeman & Valley 2001) and high-latitude areas that have undergone glaciation (e.g. Kamchatka; Bindeman *et al.* 2004).

Use and abuse of $\delta^{18}O$ criteria

As with Hf isotopes and model ages, various assumptions are commonly used in the interpretation of oxygen isotopes, and although some of the effects of these assumptions may be minor, they should still be considered. The main assumption that affects the interpretation of zircon $\delta^{18}O$ is that the mantle source of the magmatic system is homogeneous (i.e. $\delta^{18}O$ $+5.3 \pm 0.6$ ‰ 2σ), and that mantle recycling, that is, subduction of crust and sediments into the mantle, has no marked effect on the mantle signature. Data from many oceanic island arcs, and from mid-ocean ridges, indicate that this is true in many circumstances. The compilation in Eiler (2000*b*) of island arcs suggests that a maximum of 1‰ deviation occurs in this setting, and that this corresponds to a maximum contribution to the magmas of 2.5% fluid flux from high $\delta^{18}O$ sediments. A compilation of mid-ocean ridge (i.e. NMORB) magmas indicates even less deviation in $\delta^{18}O$ composition (Eiler 2000*a*). However, there have now been several studies in more complex arc settings that suggest prolonged enrichment through fluid flux and partial melting of subducting slabs and overlying sediments have led to enrichment of the shallow lithospheric mantle (Bindeman *et al.* 2004; Auer *et al.* 2009; Martin *et al.* 2011). Arc suites that directly involve melting of sediments have the highest recorded $\delta^{18}O$ (e.g. Setouchi, Japan; Bindeman *et al.* 2005). Regions of high $\delta^{18}O$ may result from multiple stages of enrichment, that is, in Kamchatka, where the lower crust and upper mantle are enriched by a prolonged subduction history, and where ascending melts are additionally enriched from a high $\delta^{18}O$ fluid flux (Auer *et al.* 2009). Regardless of these case studies, high $\delta^{18}O$ is typically inferred, particularly in studies addressing crustal growth, to result from intracrustal recycling, that is, addition of high $\delta^{18}O$ material to magmas within the crustal column. The above studies have shown that a degree of mantle recycling may also lead to elevated $\delta^{18}O$ values. The resulting, and as yet unanswered, question is how common are high $\delta^{18}O$ mantle settings temporally and spatially?

Combined Hf and O isotopes

The most pertinent use of Hf and O isotopes is when they are combined, and ideally measured from the same growth zone of a zircon that has also been analysed for U–Pb. The first use of this was the study of detrital, inherited and magmatic zircons from granites and their country rocks within the Tasmanides of southeastern Australia (Kemp *et al.* 2006; Hawkesworth & Kemp 2006; Kemp *et al.* 2007). An important facet of this work is that it highlighted the use of combined Hf–O isotopes to reveal both the relative amount of mantle and crustal compositions involved in granite generation, and perhaps even more pertinently, that the shape of the resultant mixing and AFC trends can be used to infer the chemical composition of the mixing components (Kemp *et al.* 2007). These and subsequent studies (e.g. Appleby *et al.* 2008; Roberts *et al.* 2013), have shown supracrustal sources within I-type magmas that are not observed from whole-rock isotopic signatures.

The onset of continent formation

Given that the detrital zircon record is the only record of Earth's evolution we have before about 4 Ga (Bowring *et al.* 1989; Stern & Bleeker 1998; Bowring & Williams 1999; Mojzsis *et al.* 2014), this record has been crucial to our understanding of Earth's early evolution; because of this importance, however, any conclusions drawn are subject to much scrutiny. As can be seen from Figure 1, nearly the entire record older than 4.0 Ga comes from one source, that of the Jack Hills region in SW Australia (see also Fig. 5). These quartzites of the Narryer Gneiss supracrustals were deposited at around 3 Ga (Spaggiari *et al.* 2007), and contain detrital zircons that are as old as *c.* 4.37 Ga (Wilde *et al.* 2001; Harrison 2009; Holden *et al.* 2009; Valley *et al.* 2014). Just the existence of zircons of this age indicates that zircon-bearing magmas existed during the Hadean. The Jack Hills zircons have been analysed in many studies for a variety of isotopic and elemental signatures, all with the aim of determining the environmental conditions of their formation.

The first *in-situ* Lu–Hf isotopic study of the Jack Hills zircons revealed a range of depleted and

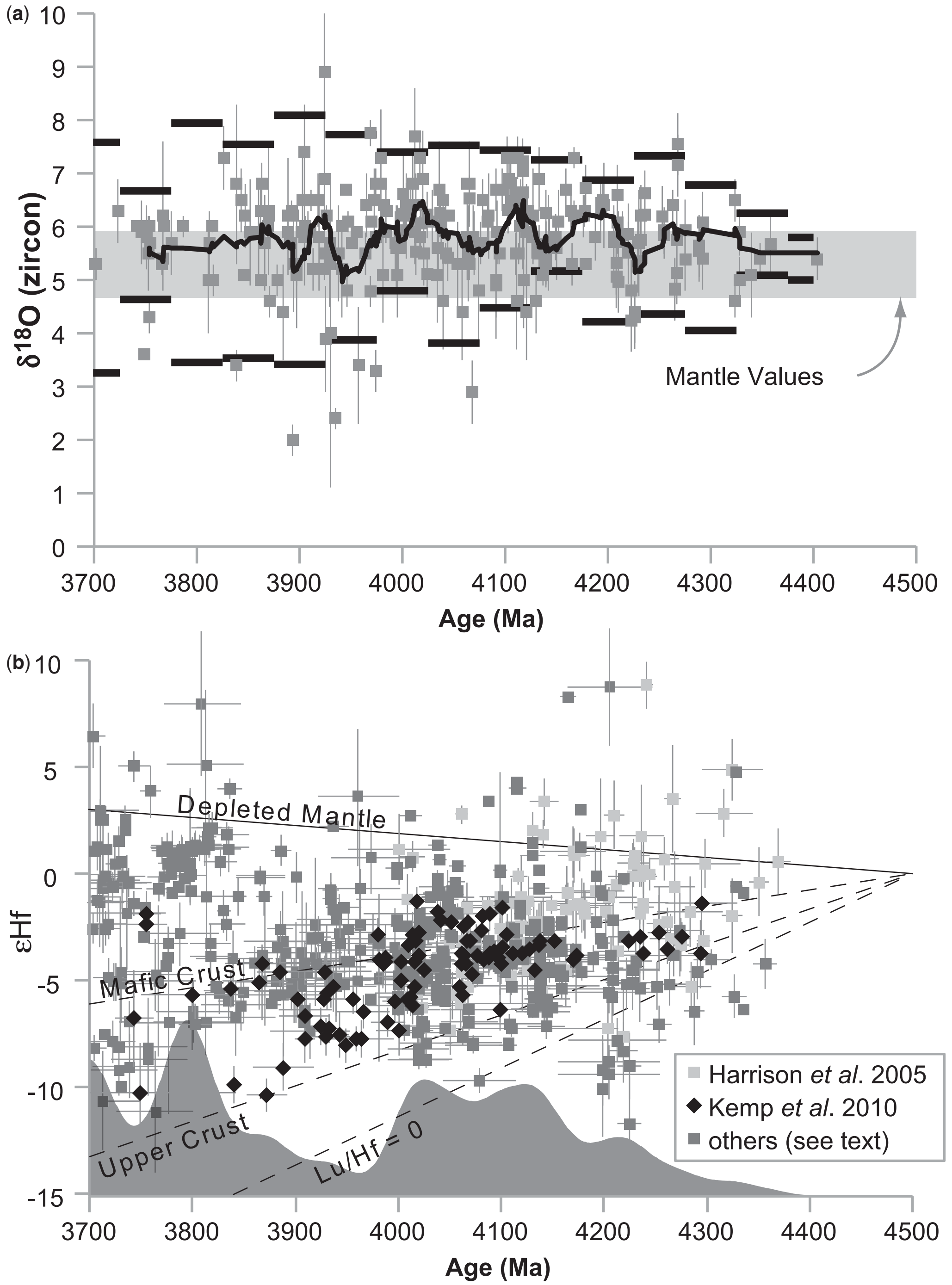


Fig. 5. (**a**) Time v. $\delta^{18}O$ from a global database (Spencer *et al.* 2014) for the 4.5–3.7 Ga time period. Bars bounding the data represent 2σ of 25 Ma bins, and the black line represents the running median. (**b**) Time v. $\varepsilon Hf_{(T)}$ of zircons from the global database (this study) for the 4.5–3.7 Ga time period. The grey solid curve represents a kernel density estimate of the zircon U–Pb ages. Evolution trends for various Lu/Hf ratios are shown as dashed lines. Uncertainty bars on both plots are 2σ.

evolved signatures (see Fig. 5); these were interpreted as evidence for differentiation of the Earth into large volumes of granitic continental crust and a residual depleted mantle (Harrison *et al.* 2005). A later study on the same material using dissolution techniques partially replicated the positive εHf signatures, but with less abundance and a lower range of values (Blichert-Toft & Albarède 2008). Subsequent studies have failed to reproduce any positive εHf signature (Harrison *et al.* 2008; Kemp *et al.* 2010). One reason for this has been the lack of tight age control and the possible assignment of incorrect ages (e.g. Valley *et al.* 2006; Kemp *et al.* 2010; Zeh *et al.* 2014). Systematic imaging and dating of all zircon growth zones should be a prerequisite for work involving rigorous interpretation of U–Pb–Hf and U–Pb–O data, but it does not necessarily mean that the previous studies are invalid. Kemp *et al.* (2010) analysed Jack Hills zircons, and showed that those analyses, which they gave a high confidence of pristine age-Hf signatures, fall on a subchondritic array with a model age of around 4.4–4.5 Ga, and a Lu–Hf ratio similar to mafic crust. These authors ignore the positive εHf signatures of previous studies, and suggest a rather simple model of crust formation in the Hadean, with intracrustal reworking of a mafic protocrust that was extracted from the mantle at 4.4–4.5 Ga. Data from younger Hadean zircons in the Jack Hills and other regions fall on similar subchondritic arrays (i.e. Acasta – Amelin *et al*. 2000; Iizuka *et al.* 2009; Mount Narryer – Kemp *et al.* 2010; Nebel-Jacobsen *et al.* 2010; Limpopo Belt – Zeh *et al.* 2014). Zeh *et al.* (2014) also point out that the superchondritic analyses of previous studies may be spurious, and that their existence negates the advocated mafic protocrust model. This model is based on the subchondritic arrays that have old (4.5–4.4 Ga) model ages, and is similar to that described by Kemp *et al.* (2010). Zeh *et al.* (2014) refer to their model as LOLIHP (Long Lived Hadean Protocrust), in contrast to the plate-tectonic-style models (i.e. Harrison *et al.* 2005, 2008), which they refer to as HACCAP (Hadean Continental Crust Formation and Plate tectonics). It is obvious that the existence of superchondritic Hf signatures in the Hadean is highly contentious, but critical to our understanding of crustal evolution. Guitreau *et al.* (2012) show that the results of the different studies are in fact similar, with the maximum values in gaussian distributions of the εHf data being centred around the same value for each study (Harrison *et al.* 2005, 2008; Blichert-Toft & Albarède 2008; Kemp *et al.* 2010). These authors suggest that the lack of superchondritic values in the studies of Kemp *et al.* (2010) and Harrison *et al.* (2008) is merely due to undersampling.

Oxygen isotopes measured in Jack Hills zircons are consistently above the mantle range (see Fig. 5). Since elevated oxygen values are normally attributed to assimilation of low-temperature altered materials (i.e. supracrustals) by the zircon's host magmas, these data have been interpreted as evidence for felsic continental crust in the Hadean, and the existence of water to produce such alteration (Mojzsis *et al.* 2001; Peck *et al.* 2001; Cavosie *et al.* 2005; Trail *et al.* 2007; Harrison *et al.* 2008). The data do not imply any particular tectonic regime, but indicate that sufficient surface water was available for alteration of supracrustal material, and that such material must have been reworked by sedimentary and subsequent magmatic processes.

Titanium concentration in zircon has been shown to be correlated with the temperature of zircon formation (Watson & Harrison 2005; Watson *et al.* 2006). These studies measured Ti-in-zircon in Hadean zircons, and revealed an average low temperature (*c.* 700 °C) that was interpreted as reflecting evolved granitic magmas and wet melting conditions (Watson & Harrison 2005; Harrison & Schmitt 2007; Harrison 2009). Whether the obtained temperatures can be used to infer melting conditions has been debated (Nutman 2006), and subsequent studies indicate that the temperature range of the Hadean zircons is also attributable to mafic magmas (Coogan & Hinton 2006; Fu *et al.* 2008). Thus, the use of Ti thermometry in this instance remains equivocal. Other trace elements, particularly the rare earth elements, have also been used to infer evolved minimum-melt conditions for Hadean zircons (e.g. Maas *et al.* 1992; Peck *et al.* 2001; Crowley *et al.* 2005; Trail *et al.* 2007), but as with Ti thermometry, the interpretations are debated (e.g. Coogan & Hinton 2006; Grimes *et al.* 2007).

Inclusions in zircons have also been used to reveal information on the source magma composition. Quartz, potassium feldspar, albitic feldspar and muscovite inclusions in Hadean zircons (Harrison *et al.* 2005, 2008; Hopkins *et al.* 2008, 2010) are used as additional evidence to argue that these zircons are derived from granitic magmas – and thus indicate the presence of evolved continental crust. Quartz and feldspar inclusions found within mafic rocks, as shown by units of the Sudbury igneous complex (Darling *et al.* 2009), and a possible metamorphic origin of mineral inclusions (Rasmussen *et al.* 2011), have led to ambiguity in their interpretation.

Finally, the lithium isotopic system, which is less studied in zircon in general, has been used to indicate that the protoliths of Hadean melts are not purely mantle-derived, but include a weathered supracrustal component (Ushikubo *et al.* 2008), for example, similar to the findings of oxygen isotope studies.

In summary, there is a wealth of evidence that zircons in the Hadean are derived from evolved felsic melts. However, each line of evidence has been strongly debated. Elevated oxygen and lithium isotope signatures suggest recycling of crust and water–rock interaction. Hf isotope signatures are compatible with reworking of a mafic protocrust extracted early in the Hadean (>4.4 Ga), but the minor yet possible existence of superchondritic signatures suggests that more complex crust–mantle dynamics may also have existed. Trace element compositions, Ti thermometry and mineral inclusions can be used to indicate that the Jack Hills zircons were derived from evolved melts, but equally, each interpretation has been suggested to be compatible with an origin in mafic crust. Overall, a geodynamic regime involving burial and internal reworking of crust is required, and this may have been mafic in nature, in line with the models of Kemp *et al.* (2010) and Zeh *et al.* (2014). To alleviate all of the major observations, Nebel *et al.* (2014) propose that the dominant composition of the crust in the Hadean was mafic, but that small pockets of low-degree anatectic melt provided the evolved compositions in which zircon was crystallized. Finally, the record of magmatism, as indicated by the zircon age population from Jack Hills, is apparently continuous and non-episodic. This characteristic is not particularly exclusive to either a LOLIHP or a HACCAP model.

The onset of plate tectonics

The onset of plate tectonics has been a subject of great debate since the initial inception of this geodynamic concept. As mentioned in the previous section, estimates for its onset begin in Earth history as early as the Hadean (e.g. Harrison 2009). Plate tectonics would not have simply turned on instantly, and therefore the definition of plate tectonics needs to be provided with any one particular estimate. One of the particular nuances of this topic is whether one is referring to modern-day-style plate tectonics, that is, with subduction of cold lithosphere and steep geotherms, or to any form of horizontal plate motion with interaction along plate margins. The timing of the onset of plate tectonics has largely been derived from the observed geological record, and through field, chemical and lithological comparisons with modern settings (for review see Furnes *et al.* 2013; Kusky *et al.* 2013). Other estimates are based on more theoretical modelling (e.g. Labrosse & Jaupart 2007; O'Neill *et al.* 2007; van Hunen & van den Berg 2008; Korenaga 2011), and of interest to this review, zircon data has also had its part to play in this debated topic.

The youngest estimates of modern-day-style plate tectonics are in the Neoproterozoic, and are based on the first existence of lithological associations that we find in the recent past (Stern 2005, 2008; Hamilton 2011), such as ophiolites, blueschists and accretionary mélanges. Contrary to these young estimates, the postulated existence of ophiolite units and of accreted arcs in the Neoarchaean (e.g. Kusky *et al.* 2001; Polat *et al.* 2005; Santosh *et al.* 2013; Dey *et al.* 2013), Mesoarchaean (e.g. Polat *et al.* 2007; Smithies *et al.* 2007; Windley & Garde 2009; de Wit *et al.* 2011) and Eoarchaean (e.g. Polat *et al.* 2002; Furnes *et al.* 2007, 2013; Kusky *et al.* 2013; Nutman *et al.* 2013) provides evidence for much earlier forms of plate convergence. As a consequence, the Archaean remains the most popular timeframe for the onset of plate tectonics. Numerical modelling is yet to provide an answer to this long-standing question. It is generally thought that a hotter mantle in the past (e.g. Herzberg *et al.* 2010) would have limited slab subduction (e.g. O'Neill *et al.* 2007; Sizova *et al.* 2010). A gradual change from shallow subduction in the Archaean, to steep subduction after 1 Ga, is compatible with the record of metamorphic gradients (Brown 2014; Sizova *et al.* 2014). As discussed by Korenaga (2013), however, the role of water in the mantle is not always accounted for in numerical models, and if dry enough, a hot mantle may also allow for slab subduction.

Evidence from zircon data agrees with an Archaean or even Hadean onset of plate tectonics. Although a Hadean onset has been hypothesized, the evidence for this remains equivocal (see previous section). Næraa *et al.* (2012) propose that the pattern exhibited by zircon data from SW Greenland indicates reworking of crust from 3.9 to 3.2 Ga, and that significant juvenile signatures with fluctuating trends after 3.2 Ga correspond to the onset of a geodynamic regime featuring arc formation and accretion (see Fig. 6). However, the data from Greenland are not representative of other exposed Archaean cratons. As can be seen from Figure 6, zircons from other cratons feature continued juvenile growth throughout the Eo- and Mesoarchaean. With all the global data plotted, there is no clear break or change in the pattern of εHf–U–Pb data. Thus, it may be that the Greenland data represents only a sampling bias, or that the onset of accretion (and preservation) of juvenile crust occurred diachronously across different regions. Regardless, the patterns in zircon εHf data do not provide robust evidence for any significant change in crustal growth mechanisms during the Eo- to Mesoarchaean.

Perhaps more significant than a supposed marked change during the Mesoarchaean is the onset of a

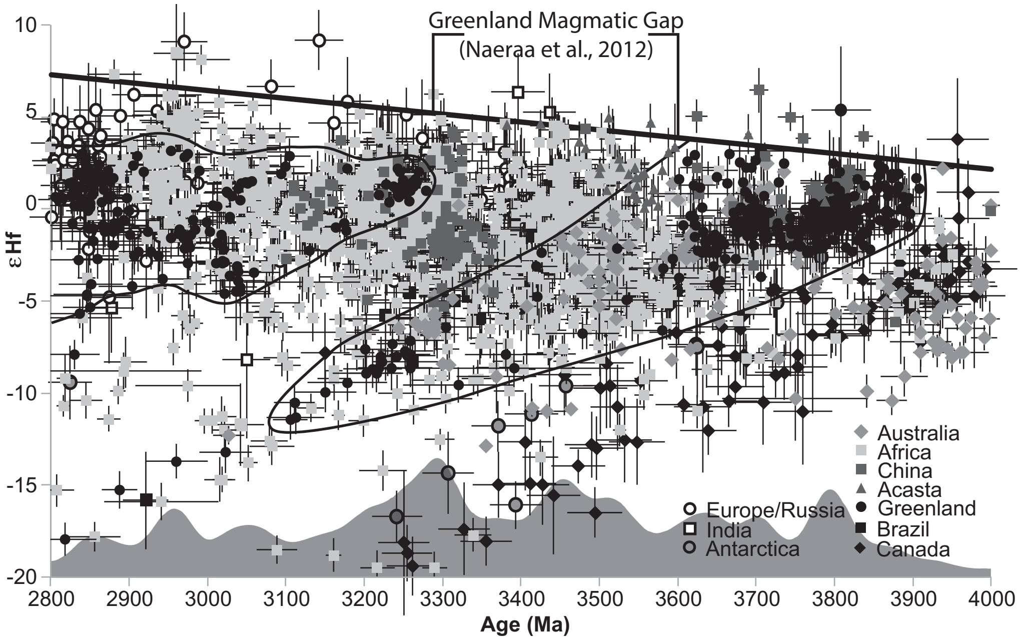


Fig. 6. Global compilation of zircon $\varepsilon Hf_{(T)}$ data (this study) for the 4.0–2.8 Ga time period. Data from different localities are highlighted. The solid grey curve shows a kernel density estimate of the zircon U–Pb ages. Uncertainty bars are 2σ. Outlined are the data from Greenland and the magmatic gap highlighted by Næraa *et al.* (2012). Although there is little data from Greenland within this gap, other localities worldwide do not demonstrate a similar gap.

continued trend of juvenile addition (i.e. zircons with εHf near the depleted mantle) that begins around 3.9 Ga (see Figs 1 & 6). Zeh *et al.* (2014) noted this marked change, interpreting it as the large-scale formation of new crust after 4.1 Ga, with the prior period featuring internal reworking of crust formed at 4.5 Ga. These authors also noted the possibility that this change around the Hadean/Archaean boundary may reflect the onset of plate tectonics, but are careful to not make any bold claims. This time period was also described by Kamber *et al.* (2005) as a transitional period from a different style of crustal formation in the Hadean to that of the Archaean. Nebel *et al.* (2014) more recently reviewed the Hadean dataset, and proposed that *c.* 4.0 Ga may mark a change in geodynamics that corresponds to the onset of subduction. It is worth noting that this time period also marks the point where we move from a detrital zircon record only to one with both rocks and zircons. Thus, establishing the nature of the oldest crust, for example, Acasta gneiss (Bowring *et al.* 1989; Guitreau *et al.* 2014; Mojzsis *et al.* 2014), and improving the records that straddle this boundary (e.g. from detrital zircons) are crucial to move our understanding further on.

Crustal growth models will be discussed in a later section; however, of note here is the interpretation of such models. Belousova *et al.* (2010) and Dhuime *et al.* (2012) both use large global databases to study continental growth. The derived growth curves (see Figs 1 & 10) feature more rapid crustal growth early on in Earth history. Dhuime *et al.* (2012) note an inflection in their growth rate at 3 Ga, which they speculate marks the onset of a geodynamic regime dominated by subduction. A growth curve using the same reworking indices (see later section on ‘Crustal growth rate’), but using a larger (*c.* 42 000 compared with *c.* 6000 analyses) U–Pb–Hf zircon database, exhibits no such inflection, but a more continuous slowing of the growth rate through time. Thus, it is suggested that any such inflections around this time are subject to bias in the database selection, and may also be artefacts of the calculations themselves. Additionally, these growth models are based on model ages, and although Dhuime *et al.* (2012) filter their model ages based upon $\delta^{18}O$ compositions, they are likely to be mixed ages that average multiple components (see previous discussion in the ‘Use and abuse’ section).

Oxygen isotopes in zircon highlight significant sedimentary input to magmas. Valley *et al.* (2005) compiled a database ($n = 1200$) of all the published zircon $\delta^{18}O$ data at that time, and noted a broad similarity of mantle-like to moderately elevated

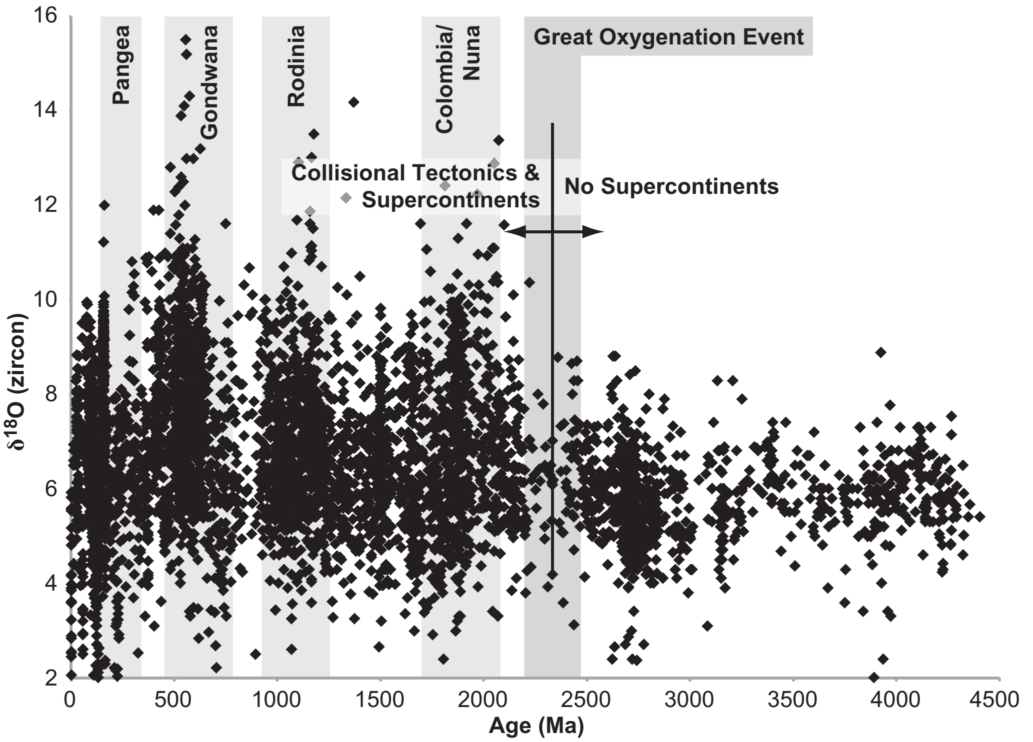


Fig. 7. Global compilation of zircon $\delta^{18}O$ data ($n = c.$ 6000; after Spencer *et al.* 2014) plotted v. time. Light grey bars represent supercontinent amalgamation events, and the dark grey bar correlates with estimates of timing of the Great Oxygenation Event (Campbell & Allen 2008).

$\delta^{18}O$ in Archaean rocks, and a steadily increasing maximum $\delta^{18}O$ throughout the post-Archaean. These authors relate this change to the secular evolution of Earth processes, with increases in sedimentary recycling, burial and intracrustal reworking occurring during the Proterozoic. Spencer *et al.* (2014) re-evaluate the published zircon oxygen database ($n = 6000$). These authors make particular reference to the correlation between periods with elevated $\delta^{18}O$ and those of supercontinent amalgamation (Fig. 7), and suggest that the increase in $\delta^{18}O$ in the Proterozoic reflects the onset of collisional tectonics that will play host to significant crustal thickening and associated erosion, burial and reworking of high $\delta^{18}O$ sedimentary rocks. In this interpretation, Columbia at 2.1–1.7 Ga (a.k.a Nuna) reflects the first true supercontinent. Interestingly, the timeframe of increased $\delta^{18}O$ in zircon also correlates with the Great Oxygenation Event (Bekker *et al.* 2004); thus, the role of this increase in atmospheric oxygen v. that of a geodynamic change is worthy of further exploration.

So when did plate tectonics begin? Alone, the zircon archive is unlikely to be able to unequivocally answer this question. However, there appears to be some convergence amongst many models that propose the Mesoarchaean to Neoarchaean. Some key constraints from other types of data also agree with this estimate, for example, diamond compositions suggest a change in geodynamics at 3 Ga (Shirey & Richardson 2011), the metamorphic record at 2.7 Ga (Brown 2006, 2007) and the geochemistry of the rock record at 2.5 Ga (Keller & Schoene 2012). The many publications that cite the lithological and geochemical signatures of Archaean rocks as being related to subduction provide ambiguous evidence to build on this topic, as the origins of these have multiple interpretations (see Van Kranendonk *et al.* 2014).

In summary, the weight of evidence, and arguably, much consensus from previous studies, suggest the following: (a) collisional tectonics with similarities to the modern day were well underway by the early Proterozoic; (b) the Archaean hosted a transitional regime ranging from plume-related and vertical tectonic processes to those of modern-day-style horizontal tectonics; (c) the Hadean eon was probably dominated by internal reworking of a (mafic) 'protocrust' and (d) subduction and accretionary orogenesis possibly started in the Eoarchaean, but more likely not until the Mesoarchaean, or perhaps even the Neoarchaean.

Regional isotopic trends

The evolution of regional magmatism, and hence continent formation through time, has been extensively studied using whole-rock geochemistry and radiogenic isotope signatures (e.g. Haschke *et al.* 2002; Kay *et al.* 2005; Mamani *et al.* 2010). Magmatic zircon can also be used in a similar fashion, given that its isotopic signature (i.e. Hf and O) relates to the host magma composition. Combined U–Pb and Hf data provide an insight to the proportion of crust and mantle input to magmas. However, crustal input involves a variety of different sources and pathways; therefore, modelling requires assumptions about the composition of the crust and mantle end-members. Ideally these will have some prior constraints from other isotopic and geochemical data. Hf data alone, for example, cannot differentiate whether crustal contamination occurs through mantle recycling (source contamination) or through intracrustal recycling (crustal contamination). The interpretation of isotopic trends through time generally assumes that these variables stay the same, and that the amount of crustal input is varying.

The utility of temporal Hf isotope trends was demonstrated in the accretionary orogen of the Tasmanides, SE Australia (Kemp *et al.* 2009; see Fig. 8a). The data reveal fluctuations in εHf through time that are interpreted as varying degrees of mantle input. These variations are not well constrained, but it is assumed that a variety of processes will lead to the same result, that is, that extensional regimes will lead to an increased mantle input, and that contractional regimes will lead to an increase in crustal contamination. This can occur through mantle recycling, as sediment input to the mantle through subduction erosion and sediment subduction will typically be higher in a contractional regime; leading to a greater 'crustal' signature in subsequent magmatism. A contractional regime will also lead to a thicker crustal column, which will result in a greater 'crustal' signature. Alternatively, an extensional regime may feature magmatism that intrudes and differentiates without significant crustal contamination, and will also lead to thinning of the lithosphere and a possible greater mantle input to the crust.

Boekhout *et al.* (2013), applied a similar technique to determine the degree of juvenile crustal growth in the proto-Andean accretionary orogen in Peru. In this example, detrital and igneous data were combined (Fig. 8b). Modelling of the mantle input through time was estimated using a crustal end-member that is derived from an 'integral crust' calculation, as applied to the global crust by Belousova *et al.* (2010). This integral crust starts with the assumed end-member crustal contamination, in this case the Palaeoproterozoic Arequipa basement, and over time the average signature of new crust is added successively. The results show that fluctuations in εHf through time can be correlated with periods of arc migration, and compression/extension of the over-riding plate. As in the Tasmanides example, extensional periods feature greater mantle input. This study revealed that overall crustal growth in the region should have increased through time in the Palaeozoic. Given that the location of the subduction margin is more or less similar to where it was in the Carboniferous, this increased growth may be explained by an overall crustal thickening since that time.

Both the proto-Andes (Boekhout *et al.* 2013) and Tasmanide (Kemp *et al.* 2009) examples reflect the evolution of the Gondwana margin. This margin can be viewed as an external orogeny, that is, one that forms on the exterior of a (super)continental mass. These orogens are in contrast to interior orogens that form during the closure of an internal ocean bounded by cratonic blocks. Collins *et al.* (2011) noted that, using Phanerozoic orogenic systems as a proxy, the modern external and internal orogenic belts produce different patterns in the εHf-time variations of zircons derived from them. Interior orogens (e.g. Central Asian Orogenic Belt, Himalaya) feature an εHf array with an increasing crustal component along with continued juvenile input after the onset of orogenesis (e.g. fanning out array; Fig. 9a). This is interpreted by Collins *et al.* (2011) as resulting from formation within a single mantle convection cell, which limits the degree to which the subcontinental lithospheric mantle is removed. In contrast, εHf values of external orogens (e.g. convergent systems surrounding the Pacific Ocean) shift to more primitive compositions after the onset of orogenesis, which is interpreted as the result of formation at the boundary of two large mantle convection cells, and thus enhancing the removal of the subcontinental lithospheric mantle.

Spencer *et al.* (2013*b*) explored these differences in ancient orogenic systems that were associated with supercontinent amalgamation, namely the *c.* 1.0 Ga Grenville orogeny associated with the formation of Rodinia, and the *c.* 0.6 Ga Pan-African orogeny involved in Gondwana formation. These authors showed that zircons derived from the Pan African orogenies reveal similar isotopic characteristics to the modern internal orogenic systems (Fig. 9b). Likewise in contrast to the Pan-African

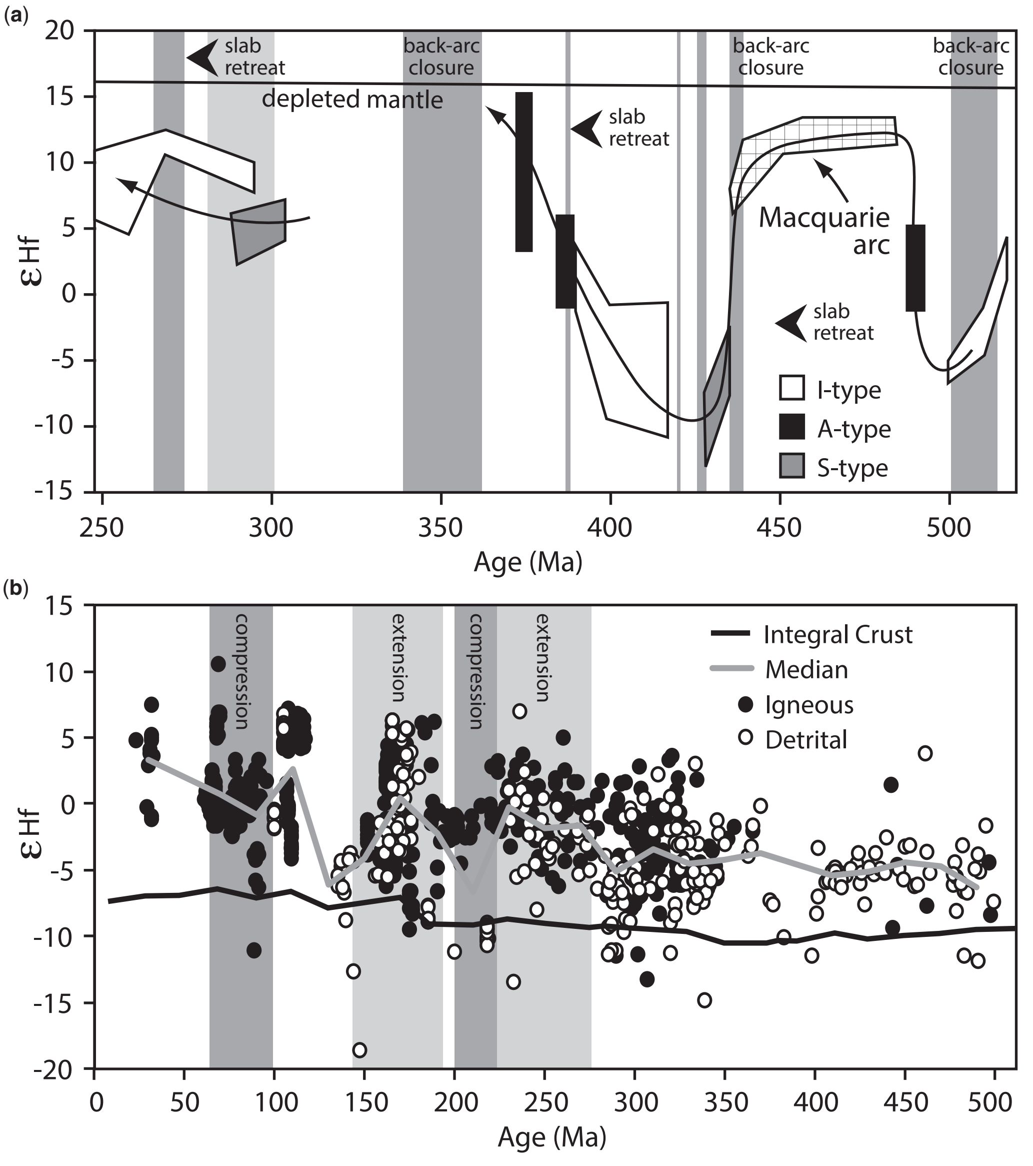


Fig. 8. εHf of zircon within (**a**) the Tasmanide orogenic belt (modified from Kemp *et al.* 2009) and (**b**) proto-Andes (modified from Boekhout *et al.* 2013), outlining the response of zircon Hf to contractional and extensional tectonic regimes. εHf values in zircon become increasingly depleted (higher εHf) during periods of extension and/or slab retreat and enriched (lower εHf) during contractional phases.

external orogens, the Grenville Orogeny displays similar isotopic characteristics to the external orogens of the modern Pacific Ocean. These authors relate the distinctive isotopic characteristics of the contrasting orogenic systems to the geological age of converging continental margins, and infer that continental margins of older geological ages will produce magmatism that is more radiogenically enriched, and those with young converging margins will give a more depleted signature (Fig. 9b).

Global isotopic trends

It is assumed that the same trends that occur on a regional or orogenic scale can also be examined

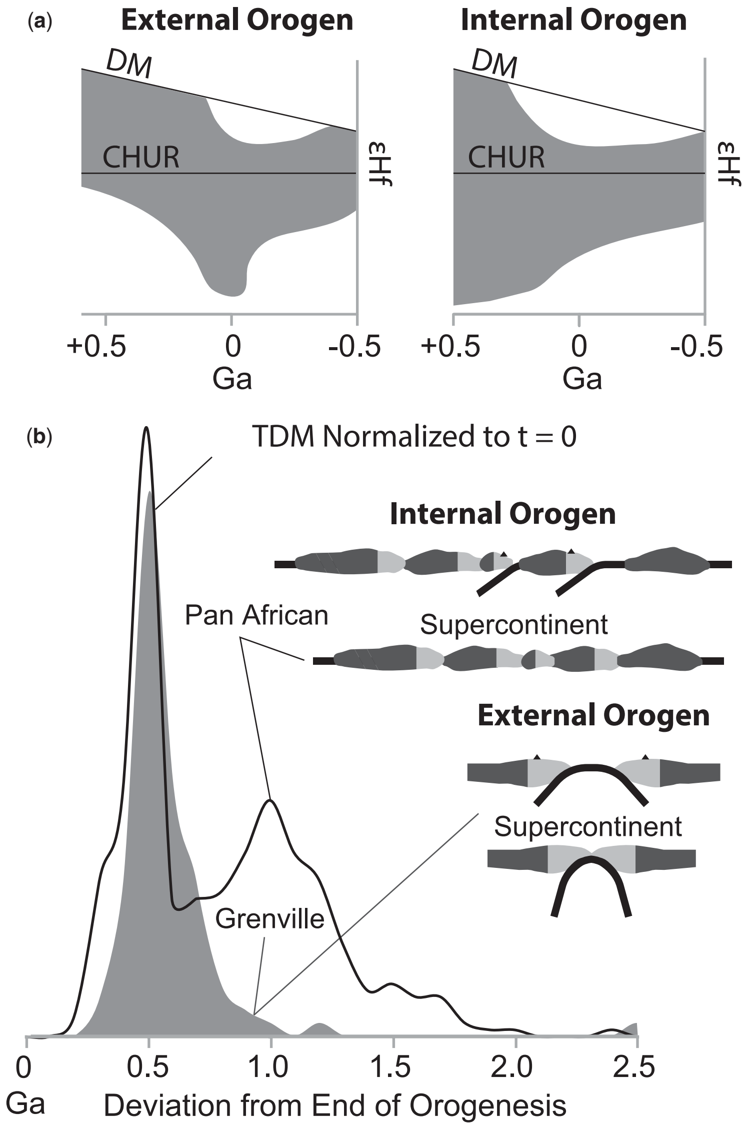


Fig. 9. (**a**) Patterns of Hf isotopes in zircon within external (circum-Pacific) and internal (Eurasian) orogenic systems (after Collins *et al.* 2011). The increasingly depleted isotopic signature post-0.5 Ga is attributed by Collins *et al.* (2011) to represent progressive removal of the subcontinental lithospheric mantle (SCLM) and the influx of more radiogenically depleted material. Conversely, the internal orogenic system continuously reworks the SCLM driving the later stages of magmatism to more enriched compositions. DM, Depleted mantle. (**b**) KDEs of Hf-depleted mantle model ages (normalized to the end of orogenesis) of detrital zircons derived from the Pan-African and Grenville orogenies (*sensu stricto*) that respectively represent internal and external orogenies, and their associated schematic cross-sections (after Spencer *et al.* 2013*b*). The key difference in the external (Grenville) and internal (Pan-African) orogenies is the additional older component in the internal orogenies.

globally. If a representative dataset is used, then these trends may reflect global balances in magmatism. Roberts (2012) used the same assumptions as those in the previously discussed regional examples to investigate the balance of crustal growth and recycling through time. An additional assumption important to this particular study is that deviations to evolved Hf signatures not only reflect greater crustal recycling, but will also indicate periods of crustal destruction, that is, loss to the mantle. The overall validity of this assumption has not been tested in terms of overall volumes or the consistency across different geodynamic settings. For example, in subduction settings, when arc magmatism is reworking pre-existing crust through crustal contamination and crustal melting, sediment subduction and subduction erosion (and possibly crustal delamination) will also lead to crustal destruction (Roberts 2012). In continent–continent collision settings, crustal destruction will occur through lower crustal delamination. It is the mass balance between crustal destruction and crustal reworking that will help better constrain crustal growth models. The findings of Roberts (2012) indicate that relative crustal growth rate can be correlated with the supercontinent cycle, with periods of supercontinent breakup featuring increased continental growth, and periods of supercontinent amalgamation featuring decreased growth. Such a balance of the destruction and growth of continental crust through time was conceptualized by Stern & Scholl (2010), and termed the Yin and Yang of the Earth's crust.

A major question asked of all global crustal growth models is whether the data used is representative. Early studies using zircon utilized the catchment of large rivers (e.g. Condie *et al.* 2005; Iizuka *et al.* 2005, 2010, 2013; Wang *et al.* 2009) and now there are various databases that allow analysis of much greater zircon populations with coverage of a greater area of the Earth's crust. Most databases feature similar age distribution patterns (see Fig. 10b), indicating that, at a broad level, they probably represent the exposed crust. Condie & Aster (2013) compared some of the different published databases, and noted differences for the last 1 Ga in particular. These authors suggested that discrepancies were due to geographic variations, particularly since some of the databases were small and restricted to particular rivers. In Figure 10a we plot the median trend through the data. Compared with that of the database used by Roberts (2012), our trend, based on a much larger database, seems to show less fluctuation overall, and in general, there is some filling of gaps in the overall fan of εHf data in this type of plot. It could be argued in this case that continued addition of new data will eventually fill in more gaps, and that patterns such as varying εHf and episodic growth may be smoothed out. However, we argue against this, as although greater coverage has led to the filling of some gaps in the array, the bivariate kernel density estimation (2D KDE) of Figure 10a shows that considerable heterogeneity in the distribution remains.

The 2D KDE shown in Figure 10a highlights the heterogeneous patten in εHf vs time space. Compared with previous studies (e.g. Belousova *et al.* 2010; Condie *et al.* 2011; Iizuka *et al.* 2010), features that remain in this large database, include a positive εHf trend after Columbia formation at 1.7 Ga, and up to Rodinia formation at 1.3 Ga. This was noted by Roberts (2012), and ascribed by Roberts (2013) to continuous continental growth on the margin of the Columbia supercontinental 'lid'. Another significant feature is the large negative εHf trend at 550 Ma that correlates with Gondwana formation and reworking of much older crust (see section on 'Regional isotopic trends'; Spencer *et al.* 2013*b*). The ability to pick out such evolution trends in the Archaean is hindered partly by data coverage, that is, does the drop in U–Pb crystallization ages around 2.3 Ga reflect tectonic shutdown (Condie *et al.* 2009), an under-representation in the data of actual events (Partin *et al.* 2014) or an artefact of preservation (see section 'Preservation of the geological record')? The lack of evolution trends may also reflect differing tectonic processes in the Archaean, perhaps suggesting greater balance between crustal growth and loss.

Crustal growth rate

One of the topics of continent formation most discussed in recent literature, which has been a matter of debate for half a century now, is the growth rate of continental crust. Models for continental growth rate can be categorized into those that feature early growth of the continental crust and subsequent steady-state recycling and growth (e.g. Armstrong & Harmon 1981; Fyfe 1976), and those that feature increasing growth through time (O'Nions *et al.* 1979; Veizer & Jansen 1979; Hurley & Rand 1969). The latter may feature episodic growth (McCulloch & Bennett 1994; Condie 1998; Rino *et al.* 2004; Condie & Aster 2010), or rapid growth in the Archaean and subdued growth thereafter (Belousova *et al.* 2010; Dhuime *et al.* 2012).

Armstrong is the most famous advocate in favour of near-steady-state continental volume throughout most of Earth history (e.g. Armstrong & Harmon 1981; Armstrong 1991), using consistency of continental freeboard and crustal thickness as part of his evidence. Fyfe (1976) had also commented on this topic, and had in fact suggested that the continental volume may have even decreased through time. We refer to the 'Armstrong model' as one in which the current volume of crust had been reached by 3.5 Ga, and where continental growth and destruction have been roughly balanced since that time.

Models that appear to be generally more accepted within the literature feature increasing continental volume through time. These may have a rate that has increased through time (e.g. Veizer & Jansen 1979), but more commonly feature a rate

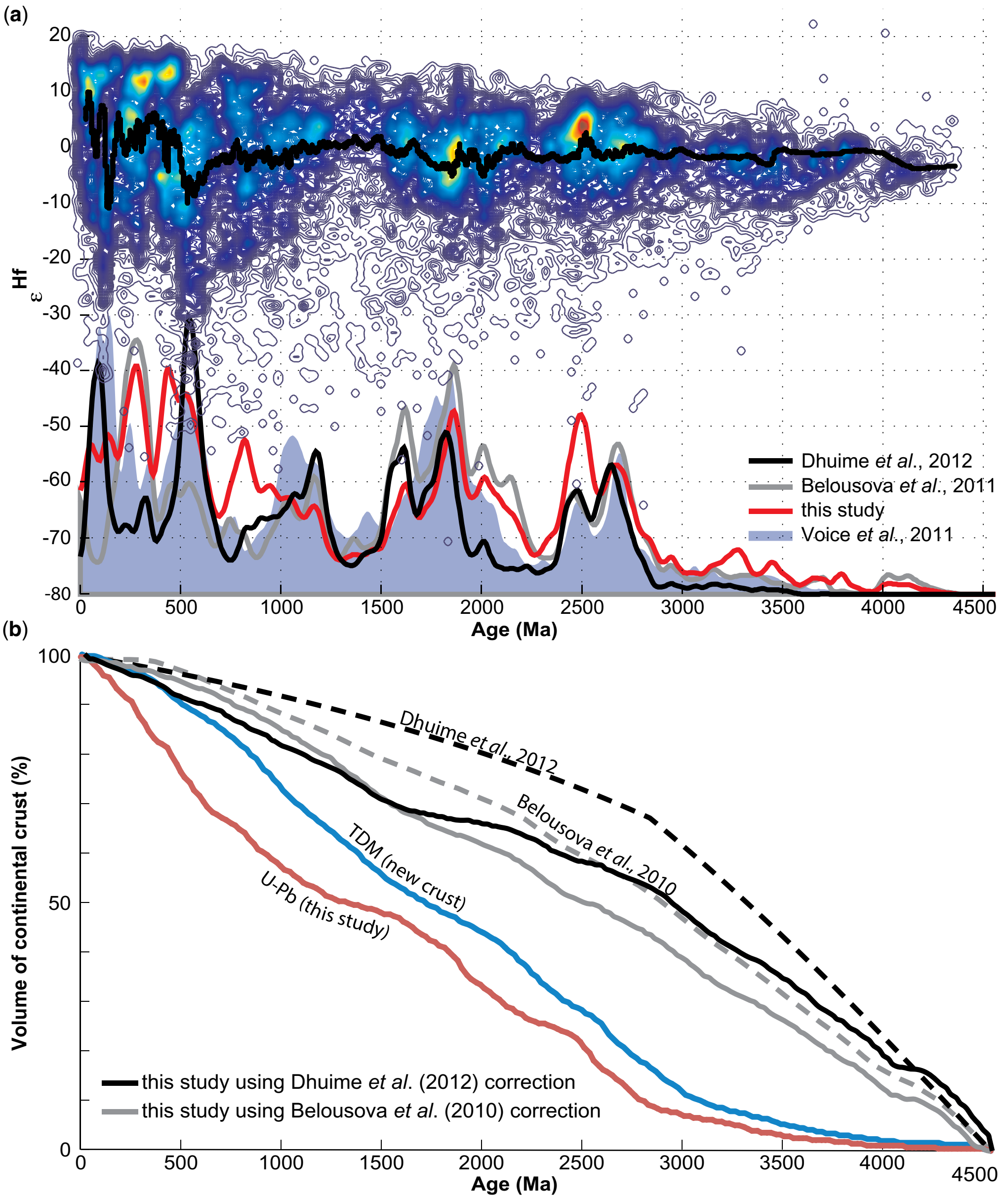


Fig. 10. (**a**) Bivariate kernel density estimation of εHf data presented in this study. Purple to red colour spectrum represent data density (purple = low, red = high). Superimposed is a black line representing the running mean of the dataset. Plus KDEs of zircon U–Pb ages for various data compilations. (**b**) Crustal growth curves of Dhuime *et al.* (2012); Belousova *et al.* (2010) (dashed black and grey lines) compared with the crustal growth curves of this study that utilize the calculation(s) of the aforementioned studies (solid black and grey lines). Also plotted are the cumulative new crust model age curve (Dhuime *et al.* 2011) and the cumulative U–Pb age curve.

of growth that has overall slowed over time (e.g. Dhuime *et al.* 2012); the latter are compatible with a secular cooling of the Earth, if mantle heat is linked in some way to continental generation. Compatible, and often linked with these models of increasing growth, is the idea of episodic growth. This concept has also been around for a long time, and is based on the episodic nature of the rock record, for example mineral deposit ages (Gastil 1960), zircon U–Pb crystallization ages (Condie 1998; Rino *et al.* 2004) and ages of He isotope (Parman 2007) and Re–Os derived mantle depletion events (Pearson *et al*. 2007). In terms of crustal growth curves, the episodicity is generally seen as an oscillation of the growth rate through time that is a secondary feature on top of the overall increasing continental volume; this leads to stepwise growth rate curves (see Fig. 1; Condie 1998; Rino *et al.* 2004). An overlap between the timing of supercontinent formation, and peaks in the crustal growth record (i.e. recorded in zircon U–Pb ages) has led to the question of causality between this relationship. Most hypotheses that advocate episodicity include the influence of mantle geodynamics (see O'Neill *et al.* 2013 for a review), that is, mantle upwellings, superplumes and slab avalanches.

Zircon U–Pb ages do not necessarily reflect continental growth, as they can refer to ages of metamorphism, or merely reflect magmatism that has reworked pre-existing crust. Various attempts have been made to remove the effect of crustal reworking, and retain the record of crustal generation. Using a simple approach, Kent Condie has compiled ages of granitoids that have juvenile radiogenic isotopic signatures (Condie 1998; Condie & Aster 2010), with subsequent histograms still providing peaks of growth at similar ages to those based purely on U–Pb ages, but with the size of peaks differing between datasets. A common approach used by many has been to use Hf model ages of zircons rather than the U–Pb ages to provide histograms of crustal growth (e.g. Iizuka *et al.* 2005, 2010; Yang *et al.* 2009).

Given the complexities discussed in the earlier section on Hf–O isotopes, model ages do not account for mixed sources to magmas, and thus are also not representative of the record of crustal growth. For this reason, recent studies have attempted more complex ways of removing the record of crustal reworking and recycling in an attempt to observe the record of continental growth. Kemp *et al.* (2007) were the first to use oxygen isotopes to discriminate zircon Hf data representing crustal reworking from juvenile crust generation. They excluded zircons with oxygen isotopes higher than 6.5‰, suggesting that these will have a supracrustal input and that model ages will subsequently be mixed ages. From a study using zircons from the Gondwana margin of Australia, they revealed two major growth periods, at 3.3 and 1.9 Ga. This oxygen-isotope based criterion has subsequently been used in many other studies (e.g. Wang *et al.* 2009, 2011; Lancaster *et al.* 2011), typically revealing that crustal growth in any one particular region occurs in discrete episodes. Along with Hf isotopes, oxygen isotopes in zircon provide equivocal information, and thus their use as a binary discriminator is less than ideal (i.e. Fig. 4, where rocks with 'mantle-like' oxygen isotope signatures still record mixed model ages). Studies that combine isotopic systems (i.e. U–Pb, Hf and O) are arguably still the most robust and informative zircon-based datasets for answering questions of crustal evolution, as long as they are used with an understanding of their limitations.

Along with model ages with or without oxygen isotopes as discriminators, recent studies have used alternative approaches of calculating crustal growth from zircon data. Again, the aim of these has been to correct for zircon data that represent crustal reworking, so that the distribution of crustal growth alone can be determined. Belousova *et al.* (2010) studied a global compilation of detrital zircon data; these authors indicated that juvenile compositions only represent *c.* 10% of the total record, and they calculated an index of juvenile growth that tries to account for reworking. For this index, the data are binned by age, and then for each bin the proportion of zircons with model ages corresponding to that bin is compared with the number of zircons with U–Pb ages (for the same bin); oxygen isotope analyses are not used to screen or characterize the data. The results show that the amount of reworking increased throughout Earth history, with stepped increases at *c.* 2.2 and *c.* 0.6 Ga. A growth curve through time was obtained from the reworking index, and showed that, by 2.5 Ga, 62% of the total volume of continental crust had been formed (grey dashed curve; Fig. 10b). We have recalculated the global reworking index and associated growth curve using the database compiled in this study (grey curve; Fig. 10b). The new curve follows a broadly similar shape, but is lower than the Belousova *et al.* (2010) curve, which implies that the larger database of this study comprises a greater record of younger growth relative to older growth. It should be noted, though, that within the study of Belousova *et al.* (2010), the modelling of growth curves and proportions of reworked material is derived from model ages, and as discussed previously, these are based on many assumptions that do not hold true.

Dhuime *et al.* (2012) produced an index of reworking to correct a global compilation of detrital zircon data using an alternative method that also made use of oxygen isotope data. Their correction involved two parts; firstly, the amount of reworked (hybrid model ages) v. juvenile magmatism (new crust formation ages) through time is estimated using a $\delta^{18}O$ compilation. This relationship is then used to make a correction to the distribution of new crust (Dhuime *et al.* 2011) model ages. A second index of reworking is calculated using the distribution of zircon data with new crust model ages that are older than their corresponding U–Pb crystallization ages. The rate of reworking through time is then calculated using the combination of this latter reworking index with the distribution of new crust model ages calculated from the first index of reworking. The crustal growth curve is defined by this reworking rate through time, and is shown in Figure 10b (black dashes curve); a curve using the database compiled in this study is also shown (black curve). Dhuime *et al.* (2012) make a strong interpretation of the inflection that is recorded at *c.* 3.0 Ga, suggesting that it marks a change in global geodynamics (see section on 'The onset of plate tectonics'). The approach taken by Dhuime *et al.* (2012) involves two 'corrections' to the data, and could be considered overly convoluted. It also uses oxygen isotope compositions as a binary discriminator, which, as discussed in the previous 'Use and abuse' section, is a flawed approach. The ideologies of Belousova *et al.* (2010) and Dhuime *et al.* (2012) are nearly the same; both studies attempt to correct the observed distribution of crustal generation recorded by model ages (i.e. the blue curve in Fig. 10b), for crustal reworking, so that a crustal generation curve can be calculated (i.e. the grey and black curves in Fig. 10b). The effect on the data using both studies is the same, that is, the crustal growth curve is shifted so that more growth occurs earlier on in Earth history; this reflects the abundance of crustal reworking that occurs in the younger part of Earth history, but is also partly an artefact of using depleted mantle or new crust model ages.

Preservation of the geological record

Although continental crust is the only long lasting repository of Earth history, it is vulnerable to the destructive processes associated with subduction erosion (Scholl *et al.* 1980; Stern 1991; von Huene & Scholl 1993; Clift *et al.* 2009; Stern 2011) and lower crustal delamination (Bird 1979; Houseman *et al.* 1981; Kay & Mahlburg Kay 1993; Houseman & Molnar 1997; Schott & Schmeling 1998; DeCelles *et al.* 2009). The locus of crustal recycling is primarily at convergent plate margins, where the subducting oceanic slab both removes ocean floor sediments and volcanic material, as well as tectonically eroding crustal material from the overriding plate into the mantle. Paradoxically, convergent plate margins are the locus of the vast majority of continental crust growth (e.g. Clift *et al.* 2009; Scholl & von Huene 2009). Scholl & von Huene (2009) hypothesize that currently the ratio of crustal formation and destruction is nearly one-to-one, resulting in a zero net gain of continental crustal volume. Stern (2011) actually estimates the total current rate of crustal destruction to be greater than the rate at which the crust is being replaced by magmatic activity, and therefore the present total volume of continental crust may be decreasing.

How representative the current distribution of continental crust is to the amount initially generated remains an unresolved issue (e.g. Bowring & Housh 1995; Hawkesworth *et al.* 2009, 2010; Condie *et al.* 2011; Cawood *et al.* 2013). The issue at hand is whether the temporal heterogeneity of presently exposed continental crust is a primary or secondary feature. That is, do the peaks of continental crust formation (i.e. those recorded with zircon U–Pb crystallization ages) represent episodic growth of continental crust (Condie 1998; Yin *et al.* 2012; Walzer & Hendel 2013; Arndt & Davaille 2013; Rino *et al.* 2004) or biased crustal preservation (Hawkesworth *et al.* 2009; Condie *et al.* 2011; Lancaster *et al.* 2011; Roberts 2012).

Proponents of episodes of enhanced magmatism call upon a dramatic influx of mantle-derived material (Condie 1998; Rino *et al.* 2004; Komiya 2007; Arndt & Davaille 2013). However, the geochemical composition of the bulk continental crust (Taylor & McLennan 1985; Rudnick & Gao 2003) implies that crustal growth occurs primarily along convergent margins (see McCulloch & Bennett 1994; Davidson & Arculus 2006; Hawkesworth & Kemp 2006; Cawood *et al.* 2013). Alternatively, Hawkesworth *et al.* (2009) hypothesize that the peaks and troughs of zircon ages are the result of enhanced preservation associated with the assembly of supercontinents (see Fig. 11), wherein the latest stage of subduction zone magmatism and that involved with collision itself, are isolated in the interior of continents following collisional

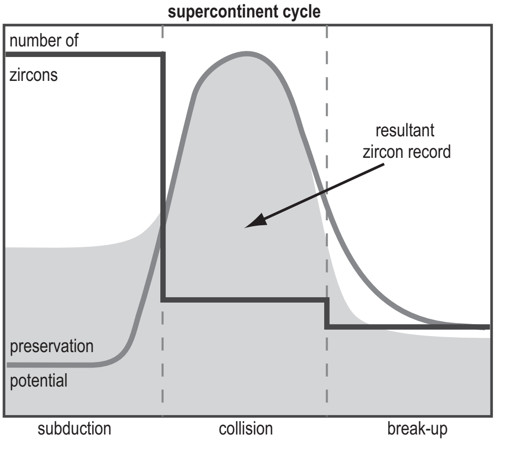


Fig. 11. Conceptual model of preservation bias during continental collision (modified after Hawkesworth *et al.* 2009; Cawood *et al.* 2013).

orogenesis (Hawkesworth *et al.* 2009; Condie *et al.* 2011; Cawood *et al.* 2013). The concept of preservation bias has yet to be fully tested, and ideally will involve, for a particular orogenic cycle and study region, calculation of the volumes of crust generated and destroyed compared with that which is preserved. Preliminary studies provide evidence in favour of a record biased through preservational processes (Roberts 2014; Spencer *et al.* 2013*a*).

Measuring Earth's Yin and Yang

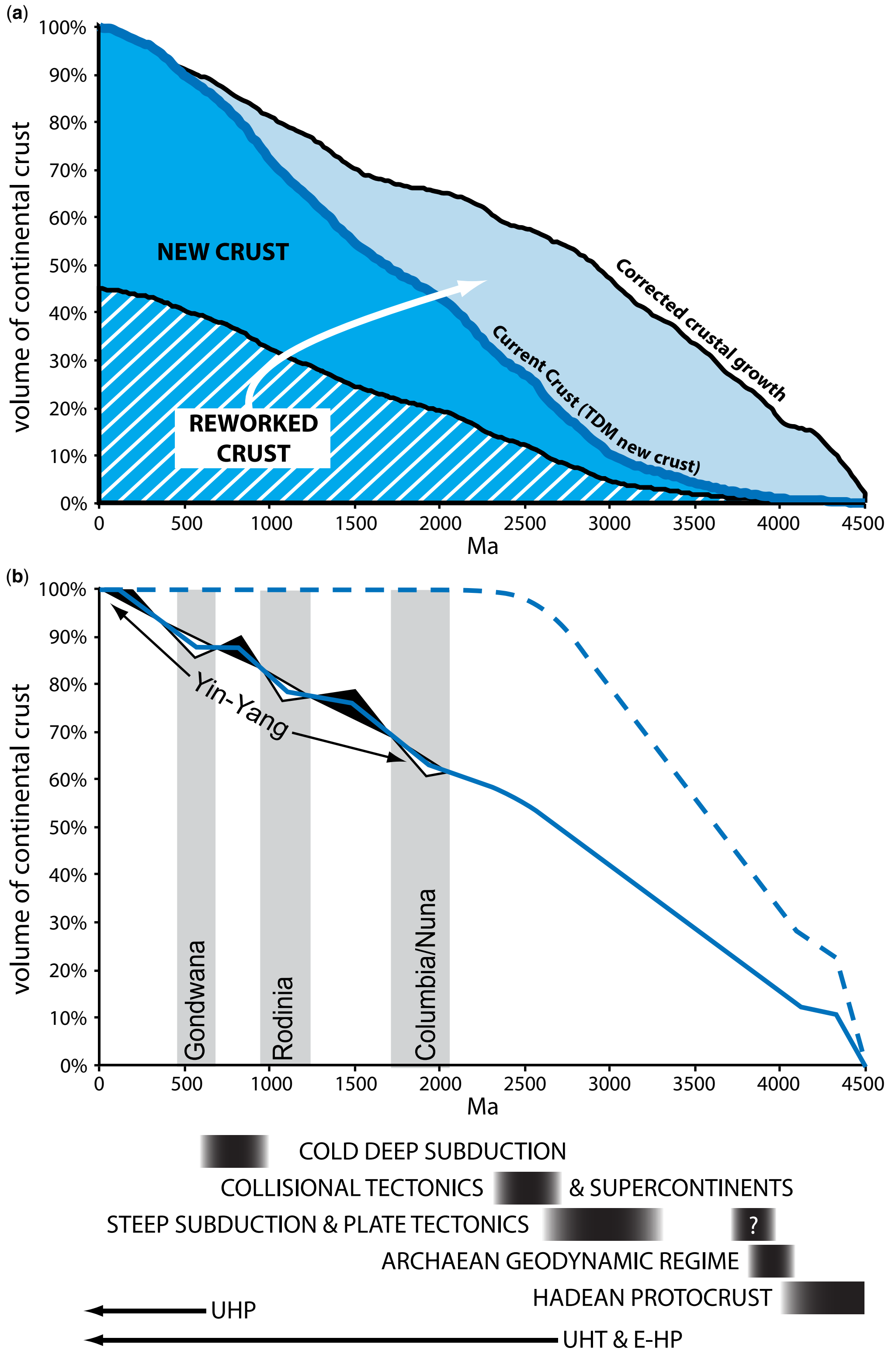
Figure 10b presents crustal growth curves based on the U–Pb–Hf detrital compilation that are corrected for crustal reworking. This correction takes an estimate of the current age distribution of continental crust (i.e. the blue curve and shaded area in Fig. 12a), and recalculates it to account for greater reworking in later Earth history so that the resulting curve features greater growth in earlier Earth history (black curve in Fig. 12a). Using the approach of Dhuime *et al.* (2012), 50% of the volume of continental crust was formed by 3 Ga, and 65% by 2.5 Ga. Two processes are not specifically accounted for by these estimates of crust generation – preservation and crustal destruction.

As discussed in the previous section, the idea of preservation bias is still conceptual and not fully tested; however, the postulate is that an increased population of zircon data in the global record does not represent an increase in crust generation, but an artefact of selective preservation during the supercontinent cycle. This effect will, however, partly be accounted for by the reworking indices of Dhuime *et al.* (2012) that are used to calculate the growth curve in Figure 12. This is because the abundance of reworked crust increases during times of supercontinent assembly, and the bias of this reworked component is corrected for in the calculations. Thus, by accounting for reworked crust, the role of selective preservation is also partly accounted for. Until the true role of preservation bias can be determined in Earth history, correcting for such a process in a quantitative manner cannot be achieved. It would also be preferable, for reasons discussed in the 'Use and abuse' section, not to derive growth models upon model age calculations.

Also discussed in the previous section is the process of crustal destruction (i.e. through subduction erosion, sediment subduction and delamination). The volume of continental crust at any one time will depend on the volume of crustal generation compared with the volume of crustal destruction (i.e. return to the mantle). This concept of the balance between Earth's Yin and Yang (growth and destruction; Stern & Scholl 2010) was shown to be compatible with trends in the global zircon Hf database (Roberts 2012). Whether the balance has a large variation through time or only results in small variations is uncertain. The resulting change in crustal growth volume is conceptualized in Figure 12b; here the effect of varying Yin–Yang is imposed upon an overall increasing continental volume, with increased growth rate between supercontinent formation, and decreased growth rate (increased crustal destruction) during supercontinent formation.

The solid blue curve shown in Figure 12b is based on the calculated growth curve in Figure 10b, that is, is based upon the reworking-corrected distribution (using Dhuime *et al.* 2012) of crust using the global zircon Hf detrital database of this study. This curve does not account for crustal destruction. At present, crustal destruction is estimated to be roughly balanced with crustal growth (see previous section), such that the crustal growth curve should lie flat. A major question then arises as to how far back in time this balance has been maintained. The modelled curves suggest that growth has outweighed destruction for most of Earth history, whereas the Armstrong model (see Crustal Growth Rate section) implies that growth has been balanced by destruction for the last 3.5 Ga. Since the growth curves derived from zircon data do not take into account the volumes of crustal destruction, there still remains a big question mark over their accuracy.

The dashed curve in Figure 12b represents an alternative global growth curve that takes into account various considerations. First, Hadean protocrust was grown in the first 200–300 million years of Earth history, and was subsequently reworked throughout the Hadean and early Archaean. A second phase of crustal growth started around 4.0 Ga. This pattern of growth recorded in the zircon database (see Figs 1, 5 & 6) is presumed to be a primary signal, and is thus included within the shape of the global growth curves. Second, if plate tectonics similar to the modern day, with supercontinents, collisional orogenesis and subduction, is likely to have been around since at least the end of the Archaean (i.e. Brown 2007; Sizova *et al.* 2010; Spencer *et al.* 2014), then there is potential that the processes and thus rates of crustal growth and destruction have been maintained throughout this time. Hence, a flat growth curve is derived for the last 2.5 Ga. Third, the crustal volume must have increased to its present-day value at some point since 4.5 Ga. For this to happen, growth must have outweighed destruction; this may have been assisted through the existence of differing tectonic processes compared with today. An increase in the role of plume-related and/or other vertical tectonic processes in the Archaean (see Van Kranendonk *et al.* 2014), particularly the earlier Archaean, may be


(a)
volume of continental crust
100%
90%
80%
70%
60%
50%
40%
30%
20%
10%
0%
NEW CRUST
REWORKED CRUST
Corrected crustal growth
Current Crust (TDM new crust)
0
500
1000
1500
2000
2500
3000
3500
4000
4500
Ma
(b)
volume of continental crust
Yin-Yang
Gondwana
Rodinia
Columbia/Nuna
Ma
COLD DEEP SUBDUCTION
COLLISIONAL TECTONICS & SUPERCONTINENTS
STEEP SUBDUCTION & PLATE TECTONICS
?
ARCHAEAN GEODYNAMIC REGIME
HADEAN PROTOCRUST
UHP
UHT & E-HP

responsible for enhanced growth. A lack of steep subduction in the earlier Archaean (Sizova *et al.* 2010; Nutman *et al.* 2013), which may limit crustal destruction, is also another factor governing growth rate.

In summary, at some point in Earth history, the crustal growth rate must have changed from a regime dominated by growth to a regime with balanced growth and destruction. We estimate that a change in geodynamic process, that is, the onset of plate tectonics and/or the onset of steep subduction, and/or the onset of thicker cratons enabled through colder geotherms, will have coincided with this change in growth rate. We suggest that such a change will not be sudden, but gradual, and that 3.2–2.5 Ga seems the best estimate of this change with current knowledge. Understanding of the rates of crustal growth and destruction will therefore also benefit from a better understanding of tectonic processes through Earth history. We suggest that the dashed curve in Figure 12b represents a conservative upper-bound of crustal growth, whereby the rate of crustal destruction has been increasing through time, but has been more prevalent since the end of the Archaean. The true rate of crustal growth may lie somewhere between the two curves.

Summary and conclusions

This paper discusses a number of topics that are hotly debated and still contentious, such as the onset of continent formation, plate tectonics and crustal growth rate. Hypotheses and models that are compatible with the zircon data represented in this study are as follows:

(1) The Hadean featured a dominantly mafic protocrust with internal reworking and the formation of low-degree evolved melts in the presence of water. Low-temperature weathering of crust occurred, and this was reworked into magmas through burial by some form of geodynamic process. Juvenile crust generation in the Hadean was not continuous, but magmatism was.

(2) A transition to an Archaean geodynamic regime occurred around 4.0 Ga, which featured renewed juvenile crust generation. It is not clear whether this regime featured subduction, but it certainly featured continuous juvenile crust generation throughout the Archaean across many continents.

(3) A change in the abundance of crustal reworking occurred around 2.5 Ga; this is interpretable from the zircon oxygen isotope record, and is indicative of the onset of collisional orogenesis and formation of supercontinents.

(4) The rate of crustal growth varies within orogens through time, and is controlled by tectonic and magmatic processes. The same processes control the global balance of crustal growth.

(5) Crustal growth rate is affected by the amount of crust generation v. that of crust destruction; this probably varies throughout Earth history, and throughout supercontinent cycles.

(6) The record of continent formation is probably biased by selective preservation during different geodynamic regimes; thus, the episodic nature of the zircon record is partly an artefact of this process.

(7) At least 50% of the current volume of continental crust existed by the end of the Archaean; since this estimate does not account for crustal destruction through time, this is likely to be an underestimate.

Figure 12c highlights some of the changes in Earth history that are discussed in this paper; these estimates are assisted through the knowledge gained from the zircon archive, but also require lines of evidence outside of zircon data. A transformation from the Hadean to Archaean is seen in the zircon archive at *c.* 4.0 Ga. Does this change reflect the onset of plate tectonics, subduction or some other geodynamics? Although some authors advocate plate tectonics in the early Archaean (see above), there is evidence that geodynamics is likely to have been different. For example, a hotter mantle will probably have resulted in flatter subduction and underthrusting at plate boundaries, with greater internal continental deformation (Sizova *et al.* 2010, 2014), and the role of vertical v. horizontal tectonic processes was probably greater (Van

Fig. 12. (**a**) Estimates of crustal growth. The blue curve is based on new crust model ages using the global database of this study, and the black curve is after correction for crustal reworking using the approach of Dhuime *et al.* (2012). The reworking correction adds on *c.* 45% of the current volume of continental crust, with this amount demonstrated by the shaded region. (**b**) The solid blue curve is an estimate of crustal growth based upon the black curve in (a). The concept of Yin and Yang, that is, the balance between crustal growth (black regions) and crustal destruction (white regions), during supercontinent cycles (grey bars) is imposed upon this curve. The dashed curve represents a hypothetical crustal growth curve allowing for balanced crustal growth and destruction since *c.* 2.5 Ga. The black bars below represent hypothesized timing of global geodynamic regimes based upon this review, and arrows show the record of metamorphic conditions: UHP, ultra-high pressure; UHT, ultra-high temperature; E-HP, eclogite-high pressure.

Kranendonk *et al.* 2014; Van Kranendonk 2010). After *c.* 4.0 Ga, there is no obvious marked change in a global record until *c.* 3.0 Ga that may reflect the onset of subduction and/or plate tectonics. Modelling of mantle temperatures suggests that steeper subduction may have begun to occur around 3.2–2.5 Ga (Sizova *et al.* 2010). The metamorphic record shows that eclogite-high pressure and ultra-high temperature metamorphism started around *c.* 2.7 Ga (Brown 2006, 2007, 2014). Diamond inclusion compositions show a change around 3 Ga that is interpreted as the onset of formation of eclogite through subduction and its capture via continental collision (Shirey & Richardson 2011). Thus, the Neoarchaen is perhaps the best estimate of a geodynamic change to a regime with modern-day-style subduction and plate motions. The formation of supercontinents and collisional orogenesis will have followed this, probably occurring after 2.5 Ga. In this instance, the significant record of crustal growth in the Neoarchaean at 2.9–2.7 Ga (see Figs 1 & 10a) may represent the onset of stabilization of Archaean crust (Hawkesworth *et al.* 2009), and/or may also represent a true episode of increased crustal growth (Condie 1998; Arndt & Davaille 2013). Concurrent with the preservation of ultra high-pressure metamorphism, the onset of cold, deep and steep subduction like the present-day probably occurred in the Neoproterozoic (Stern 2005, 2008; Brown 2006, 2007, 2014; Sizova *et al.* 2010, 2014).

So where now? Questions remain over most of the key topics discussed here. These include two great unknowns: what is the role of preservation bias in the geological record, and what is the extent of crustal destruction through time? Knowledge of these is implicit in fully understanding the evolution of the continents, and thus Earth history as a whole. It is unlikely that zircon data alone will solve these mysteries, but along with the integration of multiple isotopic proxies, numerical modelling and geophysical constraints, zircon will no doubt at least play its part.

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References

Allègre, C. J. & Rousseau, D. 1984. The growth of the continent through geological time studied by Nd isotope analysis of shales. *Earth and Planetary Science Letters*, **67**, 19–34.

Amelin, Y., Lee, D. C., Halliday, A. N. & Pidgeon, R. T. 1999. Nature of the Earth's earliest crust from hafnium isotopes in single detrital zircons. *Nature*, **399**, 252–255.

Amelin, Y., Lee, D. C. & Halliday, A. N. 2000. Early-middle Archaean crustal evolution deduced from Lu–Hf and U–Pb isotopic studies of single zircon grains. *Geochimica et Cosmochimica Acta*, **64**, 4205–4225.

Andersen, T., Griffin, W. L. & Pearson, N. J. 2002. Crustal evolution in the SW part of the Baltic Shield: the Hf isotope evidence. *Journal of Petrology*, **43**, 1725–1747.

Appleby, S. K., Graham, C. M., Gillespie, M. R., Hinton, R. W. & Oliver, G. J. H. 2008. A cryptic record of magma mixing in diorites revealed by high-precision SIMS oxygen isotope analysis of zircons. *Earth and Planetary Science Letters*, **269**, 105–117.

Armstrong, R. L. 1971. Isotopic and chemical constraints on models of magma genesis in volcanic arcs. *Earth and Planetary Science Letters*, **12**, 137–142.

Armstrong, R. L. 1991. The persistent myth of crustal growth. *Australian Journal of Earth Sciences*, **38**, 613–630.

Armstrong, R. L. & Harmon, R. S. 1981. Radiogenic isotopes: the case for crustal recycling on a near-steady-state no-continental-growth Earth [and discussion]. *Philosophical Transactions of the Royal Society of London. Series A, Mathematical and Physical Sciences*, **3011**, 443–472.

Arndt, N. & Davaille, A. 2013. Episodic Earth evolution. *Tectonophysics*, **609**, 661–674.

Arndt, N. T. & Goldstein, S. L. 1987. Use and abuse of crust-formation ages. *Geology*, **15**, 893–895.

Auer, S., Bindeman, I., Wallace, P., Ponomareva, V. & Portnyagin, M. 2009. The origin of hydrous, high-δ18O voluminous volcanism: diverse oxygen isotope values and high magmatic water contents within the volcanic record of Klyuchevskoy volcano, Kamchatka, Russia. *Contributions to Mineralogy and Petrology*, **157**, 209–230.

Begg, G. C., Griffin, W. L. *et al.* 2009. The lithospheric architecture of Africa: seismic tomography, mantle petrology, and tectonic evolution. *Geosphere*, **5**, 23–50.

Bekker, A., Holland, H. D. *et al.* 2004. Dating the rise of atmospheric oxygen. *Nature*, **427**, 117–120.

Belousova, E. A., Kostitsyn, Y. A., Griffin, W. L., Begg, G. C., O'Reilly, S. Y. & Pearson, N. J. 2010. The growth of the continental crust: constraints from zircon Hf-isotope data. *Lithos*, **119**, 457–466.

Bennett, V. C. & DePaolo, D. J. 1987. Proterozoic crustal history of the western United States as determined by neodymium isotopic mapping. *Geological Society of America Bulletin*, **99**, 674–685.

Bindeman, I. N. & Valley, J. W. 2001. Low-δ18O rhyolites from Yellowstone: magmatic evolution based on analyses of zircons and individual phenocrysts. *Journal of Petrology*, **42**, 1491–1517.

Bindeman, I. N., Ponomareva, V. V., Bailey, J. C. & Valley, J. W. 2004. Volcanic arc of Kamchatka: a province with high-$\delta^{18}O$ magma sources and large-scale $^{18}O/^{16}O$ depletion of the upper crust. *Geochimica et Cosmochimica Acta*, **68**, 841–865.

Bindeman, I. N., Eiler, J. M. *et al.* 2005. Oxygen isotope evidence for slab melting in modern and ancient subduction zones. *Earth and Planetary Science Letters*, **235**, 480–496.

Bird, P. 1979. Continental delamination and the Colorado Plateau. *Journal of Geophysical Research: Solid Earth (1978–2012)*, **84**, 7561–7571.

Bizzarro, M., Baker, J. A., Haack, H., Ulfbeck, D. & Rosing, M. 2003. Early history of Earth's crust–mantle system inferred from hafnium isotopes in chondrites. *Nature*, **421**, 931–933.

Blichert-Toft, J. & Albarède, F. 1997. The Lu–Hf isotope geochemistry of chondrites and the evolution of the mantle-crust system. *Earth and Planetary Science Letters*, **148**, 243–258.

Blichert-Toft, J. & Albarède, F. 2008. Hafnium isotopes in Jack Hills zircons and the formation of the Hadean crust. *Earth and Planetary Science Letters*, **265**, 686–702.

Boekhout, F., Roberts, N. M., Gerdes, A. & Schaltegger, U. 2013. A Hf-isotope perspective on continent formation in the south Peruvian Andes. *In*: Roberts, N. M. W., Van Kranendonk, M., Parman, S., Shirey, S. & Clift, P. D. (eds) *Continent Formation Through Time*. Geological Society, London, Special Publications, **389**, first published online September 9, 2013, http://dx.doi.org/10.1144/SP389.6

Bowring, S. & Housh, T. 1995. The Earth's early evolution. *Science*, **269**, 1535–1540.

Bowring, S. A. & Williams, I. S. 1999. Priscoan (4.00–4.03 Ga) orthogneisses from northwestern Canada. *Contributions to Mineralogy and Petrology*, **134**, 3–16.

Bowring, S. A., Williams, I. S. & Compston, W. 1989. 3.96 Ga gneisses from the Slave province, Northwest Territories, Canada. *Geology*, **17**, 971–975.

Brown, M. 2006. Duality of thermal regimes is the distinctive characteristic of plate tectonics since the Neoarchean. *Geology*, **34**, 961–964.

Brown, M. 2007. Metamorphic conditions in orogenic belts: a record of secular change. *International Geology Review*, **49**, 193–234.

Brown, M. 2014. The contribution of metamorphic petrology to understanding lithosphere evolution and geodynamics. *Geoscience Frontiers*, **5**, 553–569.

Campbell, I. H. & Allen, C. M. 2008. Formation of supercontinents linked to increases in atmospheric oxygen. *Nature Geoscience*, **1**, 554–558.

Caro, G. & Bourdon, B. 2010. Non-chondritic Sm/Nd ratio in the terrestrial planets: consequences for the geochemical evolution of the mantle–crust system. *Geochimica et Cosmochimica Acta*, **74**, 3333–3349.

Cavosie, A. J., Valley, J. W. & Wilde, S. A. 2005. Magmatic $\delta^{18}O$ in 4400–3900 Ma detrital zircons: a record of the alteration and recycling of crust in the Early Archean. *Earth and Planetary Science Letters*, **235**, 663–681.

Cawood, P. A., Hawkesworth, C. J. & Dhuime, B. 2013. The continental record and the generation of continental crust. *Geological Society of America Bulletin*, **125**, 14–32.

Chappell, B. W. & White, A. J. R. 1992. *I- and S-Type Granites in the Lachlan Fold Belt*. Geological Society of America, Boulder, CO, Special Papers, **272**, 1–26.

Claesson, S., Bibikova, E., Shumlyanskyy, L., Dhuime, B. & Hawkesworth, C. J. 2014. The oldest crust in the Ukrainian Shield – Eoarchaean U–Pb ages and Hf–Nd constraints from enderbites and metasediments. *In*: Roberts, N. M. W., Van Kranendonk, M., Parman, S., Shirey, S. & Clift, P. D. (eds) *Continent Formation Through Time*. Geological Society, London, Special Publications, **389**, first published online January 3, 2014, http://dx.doi.org/10.1144/SP389.9

Clift, P. D., Vannucchi, P. & Morgan, J. P. 2009. Crustal redistribution, crust–mantle recycling and Phanerozoic evolution of the continental crust. *Earth-Science Reviews*, **97**, 80–104.

Collins, W. J., Belousova, E. A., Kemp, A. I. & Murphy, J. B. 2011. Two contrasting Phanerozoic orogenic systems revealed by hafnium isotope data. *Nature Geoscience*, **4**, 333–337.

Condie, K. C. 1998. Episodic continental growth and supercontinents: a mantle avalanche connection? *Earth and Planetary Science Letters*, **163**, 97–108.

Condie, K. C. 2004. Supercontinents and superplume events: distinguishing signals in the geologic record. *Physics of the Earth and Planetary Interiors*, **146**, 319–332.

Condie, K. C. & Aster, R. C. 2010. Episodic zircon age spectra of orogenic granitoids: the supercontinent connection and continental growth. *Precambrian Research*, **180**, 227–236.

Condie, K. C. & Aster, R. C. 2013. Refinement of the supercontinent cycle with Hf, Nd and Sr isotopes. *Geoscience Frontiers*, **4**, 667–680.

Condie, K. C., Beyer, E., Belousova, E., Griffin, W. L. & O'Reilly, S. Y. 2005. U–Pb isotopic ages and Hf isotopic composition of single zircons: the search for juvenile Precambrian continental crust. *Precambrian Research*, **139**, 42–100.

Condie, K. C., O'Neill, C. & Aster, R. C. 2009. Evidence and implications for a widespread magmatic shutdown for 250 My on Earth. *Earth and Planetary Science Letters*, **282**, 294–298.

Condie, K. C., Bickford, M. E., Aster, R. C., Belousova, E. & Scholl, D. W. 2011. Episodic zircon ages, Hf isotopic composition, and the preservation rate of continental crust. *Geological Society of America Bulletin*, **123**, 951–957.

Coogan, L. A. & Hinton, R. W. 2006. Do the trace element compositions of detrital zircons require Hadean continental crust?. *Geology*, **34**, 633–636.

Crowley, J. L., Myers, J. S., Sylvester, P. J. & Cox, R. A. 2005. Detrital zircon from the Jack Hills and Mount Narryer, Western Australia: evidence for diverse > 4.0 Ga source rocks. *The Journal of Geology*, **113**, 239–263.

Darling, J., Storey, C. & Hawkesworth, C. 2009. Impact melt sheet zircons and their implications for the Hadean crust. *Geology*, **37**, 927–930.

Davidson, J. P. 1983. Lesser Antilles isotopic evidence of the role of subducted sediment in island arc magma genesis. *Nature*, **306**, 253–256.

Davidson, J. P. & Arculus, R. J. 2006. The significance of Phanerozoic arc magmatism in generating continental crust. *In*: Brown, M. & Rushmer, T. (eds) *Evolution and Differentiation of the Continental Crust.* Cambridge Press, Cambridge, 135–172.

DeCelles, P. G., Ducea, M. N., Kapp, P. & Zandt, G. 2009. Cyclicity in Cordilleran orogenic systems. *Nature Geoscience*, **2**, 251–257.

DePaolo, D. J. 1981. Neodymium isotopes in the Colorado Front Range and crust–mantle evolution in the Proterozoic. *Nature*, **291**, 193–196.

DePaolo, D. J., Linn, A. M. & Schubert, G. 1991. The continental crustal age distribution: methods of determining mantle separation ages from Sm–Nd isotopic data and application to the southwestern United States. *Journal of Geophysical Research: Solid Earth (1978–2012)*, **96**, 2071–2088.

de Wit, M. J., Furnes, H. & Robins, B. 2011. Geology and tectonostratigraphy of the Onverwacht Suite, Barberton greenstone belt, South Africa. *Precambrian Research*, **186**, 1–27.

Dey, S., Nandy, J., Choudhary, A. K., Liu, Y. & Zong, K. 2013. Neoarchaean crustal growth by combined arc–plume action: evidence from the Kadiri Greenstone Belt, eastern Dharwar craton, India. *In*: Roberts, N. M. W., Van Kranendonk, M., Parman, S., Shirey, S. & Clift, P. D. (eds) *Continent Formation Through Time*. Geological Society, London, Special Publications, **389**, first published online September 2, 2013, http://dx.doi.org/10.1144/SP389.3

Dhuime, B., Hawkesworth, C. & Cawood, P. 2011. When continents formed. *Science*, **331**, 154–155.

Dhuime, B., Hawkesworth, C. J., Cawood, P. A. & Storey, C. D. 2012. A change in the geodynamics of continental growth 3 billion years ago. *Science*, **335**, 1334–1336.

Dickin, A. P. & McNutt, R. H. 1989. Nd model age mapping of the southeast margin of the Archean foreland in the Grenville province of Ontario. *Geology*, **17**, 299–302.

Eiler, J. M. 2001. Oxygen isotope variations of basaltic lavas and upper mantle rocks. *Reviews in Mineralogy and Geochemistry*, **43**, 319–364.

Eiler, J. M., Crawford, A., Elliott, T., Farley, K. A., Valley, J. W. & Stolper, E. M. 2000*a*. Oxygen isotope geochemistry of oceanic-arc lavas. *Journal of Petrology*, **41**, 229–256.

Eiler, J. M., Schiano, P., Kitchen, N. & Stolper, E. M. 2000*b*. Oxygen-isotope evidence for recycled crust in the sources of mid-ocean-ridge basalts. *Nature*, **403**, 530–534.

Elliott, T., Plank, T., Zindler, A., White, W. & Bourdon, B. 1997. Element transport from slab to volcanic front at the Mariana arc. *Journal of Geophysical Research: Solid Earth (1978–2012)*, **102**, 14 991–15 019.

Ewing, T. A., Rubatto, D. & Hermann, J. 2014. Hafnium isotopes and Zr/Hf of rutile and zircon from lower crustal metapelites (Ivrea–Verbano Zone, Italy): implications for chemical differentiation of the crust. *Earth and Planetary Science Letters*, **389**, 106–118.

Fedo, C. M., Sircombe, K. N. & Rainbird, R. H. 2003. Detrital zircon analysis of the sedimentary record. *Reviews in Mineralogy and Geochemistry*, **53**, 277–303.

Fu, B., Page, F. Z. *et al.* 2008. Ti-in-zircon thermometry: applications and limitations. *Contributions to Mineralogy and Petrology*, **156**, 197–215.

Fujimaki, H. 1986. Partition coefficients of Hf, Zr, and REE between zircon, apatite, and liquid. *Contributions to Mineralogy and Petrology*, **94**, 42–45.

Furnes, H., de Wit, M., Staudigel, H., Rosing, M. & Muehlenbachs, K. 2007. A vestige of Earth's oldest ophiolite. *Science*, **315**, 1704–1707.

Furnes, H., Dilek, Y. & de Wit, M. 2013. Precambrian greenstone sequences represent different ophiolite types. *Gondwana Research*, http://dx.doi.org/10.1016/j.gr.2013.06.004

Fyfe, W. S. 1976. Hydrosphere and continental crust: growing or shrinking? *Geoscience Canada*, **3**, 82–83.

Garçon, M., Chauvel, C. & Bureau, S. 2011. Beach placer, a proxy for the average Nd and Hf isotopic composition of a continental area. *Chemical Geology*, **287**, 182–192.

Gastil, R. G. 1960. The distribution of mineral dates in time and space. *American Journal of Science*, **258**, 1–35.

Goldstein, S. J. & Jacobsen, S. B. 1988. Nd and Sr isotopic systematics of river water suspended material: implications for crustal evolution. *Earth and Planetary Science Letters*, **87**, 249–265.

Griffin, W. L., Pearson, N. J., Belousova, E., Jackson, S. E., Van Achterbergh, E., O'Reilly, S. Y. & Shee, S. R. 2000. The Hf isotope composition of cratonic mantle: LAM-MC-ICPMS analysis of zircon megacrysts in kimberlites. *Geochimica et Cosmochimica Acta*, **64**, 133–147.

Griffin, W. L., Wang, X., Jackson, S. E., Pearson, N. J., O'Reilly, S. Y., Xu, X. & Zhou, X. 2002. Zircon chemistry and magma mixing, SE China: in-situ analysis of Hf isotopes, Tonglu and Pingtan igneous complexes. *Lithos*, **61**, 237–269.

Grimes, C. B., John, B. E. *et al.* 2007. Trace element chemistry of zircons from oceanic crust: a method for distinguishing detrital zircon provenance. *Geology*, **35**, 643–646.

Guitreau, M., Blichert-Toft, J., Martin, H., Mojzsis, S. J. & Albarède, F. 2012. Hafnium isotope evidence from Archean granitic rocks for deep-mantle origin of continental crust. *Earth and Planetary Science Letters*, **337**, 211–223.

Guitreau, M., Blichert-Toft, J., Mojzsis, S. J., Roth, A. S., Bourdon, B., Cates, N. L. & Bleeker, W. 2014. Lu–Hf isotope systematics of the Hadean–Eoarchean Acasta Gneiss Complex (Northwest Territories, Canada). *Geochimica et Cosmochimica Acta*, **135**, 251–269.

Hamilton, W. B. 2011. Plate tectonics began in Neoproterozoic time, and plumes from deep mantle have never operated. *Lithos*, **123**, 1–20.

Hammerli, J., Kemp, A. I. S. & Spandler, C. 2014. Neodymium isotope equilibration during crustal metamorphism revealed by in situ microanalysis of REE-rich accessory minerals. *Earth and Planetary Science Letters*, **392**, 133–142.

Harrison, T. M. 2009. The Hadean crust: evidence from >4 Ga zircons. *Annual Review of Earth and Planetary Sciences*, **37**, 479–505.

Harrison, T. M., Blichert-Toft, J., Müller, W., Albarede, F., Holden, P. & Mojzsis, S. J. 2005. Heterogeneous Hadean hafnium: evidence of continental crust at 4.4 to 4.5 Ga. *Science*, **310**, 1947–1950.

Harrison, T. M. & Schmitt, A. K. 2007. High sensitivity mapping of Ti distributions in Hadean zircons. *Earth and Planetary Science Letters*, **261**, 9–19.

Harrison, T. M., Schmitt, A. K., McCulloch, M. T. & Lovera, O. M. 2008. Early ($\geq$ 4.5 Ga) formation of terrestrial crust: Lu–Hf, $\delta^{18}O$, and Ti thermometry results for Hadean zircons. *Earth and Planetary Science Letters*, **268**, 476–486.

Haschke, M., Siebel, W., Günther, A. & Scheuber, E. 2002. Repeated crustal thickening and recycling during the Andean orogeny in north Chile (21–26 S). *Journal of Geophysical Research: Solid Earth (1978–2012)*, **107**, ECV-6.

Hawkesworth, C. J. & Kemp, A. I. S. 2006. Using hafnium and oxygen isotopes in zircons to unravel the record of crustal evolution. *Chemical Geology*, **226**, 144–162.

Hawkesworth, C., Cawood, P., Kemp, T., Storey, C. & Dhuime, B. 2009. Geochemistry: a matter of preservation. *Science*, **323**, 49–50.

Hawkesworth, C. J., O'Nions, R. K. & Arculus, R. J. 1979. Nd and Sr isotope geochemistry of island arc volcanics, Grenada, Lesser Antilles. *Earth and Planetary Science Letters*, **45**, 237–248.

Hawkesworth, C. J., Gallagher, K., Hergt, J. M. & McDermott, F. 1993. Mantle and slab contribution in arc magmas. *Annual Review of Earth and Planetary Sciences*, **21**, 175–204.

Hawkesworth, C. J., Dhuime, B., Pietranik, A. B., Cawood, P. A., Kemp, A. I. S. & Storey, C. D. 2010. The generation and evolution of the continental crust. *Journal of the Geological Society*, **167**, 229–248.

Herzberg, C., Condie, K. & Korenaga, J. 2010. Thermal history of the Earth and its petrological expression. *Earth and Planetary Science Letters*, **292**, 79–88.

Hiess, J., Yi, K., Woodhead, J., Ireland, T. & Rattenbury, M. 2014. Gondwana margin evolution from zircon REE, O and Hf signatures of Western Province gneisses, Zealandia. *In*: Roberts, N. M. W., Van Kranendonk, M., Parman, S., Shirey, S. & Clift, P. D. (eds) *Continent Formation Through Time*. Geological Society, London, Special Publications, **389**, first published online June 2, 2014, http://dx.doi.org/10.1144/SP389.10

Holden, P., Lanc, P., Ireland, T. R., Harrison, T. M., Foster, J. J. & Bruce, Z. 2009. Mass-spectrometric mining of Hadean zircons by automated SHRIMP multi-collector and single-collector U/Pb zircon age dating: the first 100 000 grains. *International Journal of Mass Spectrometry*, **286**, 53–63.

Hopkins, M., Harrison, T. M. & Manning, C. E. 2008. Low heat flow inferred from> 4 Gyr zircons suggests Hadean plate boundary interactions. *Nature*, **456**, 493–496.

Hopkins, M. D., Harrison, T. M. & Manning, C. E. 2010. Constraints on Hadean geodynamics from mineral inclusions in >4 Ga zircons. *Earth and Planetary Science Letters*, **298**, 367–376.

Houseman, G. A. & Molnar, P. 1997. Gravitational (Rayleigh–Taylor) instability of a layer with non-linear viscosity and convective thinning of continental lithosphere. *Geophysical Journal International*, **128**, 125–150.

Houseman, G. A., McKenzie, D. P. & Molnar, P. 1981. Convective instability of a thickened boundary layer and its relevance for the thermal evolution of continental convergent belts. *Journal of Geophysical Research: Solid Earth (1978–2012)*, **86**, 6115–6132.

Hurley, P. M. & Rand, J. R. 1969. Pre-drift continental nuclei. *Science*, **164**, 1229–1242.

Ickert, R. B. 2013. Algorithms for estimating uncertainties in initial radiogenic isotope ratios and model ages. *Chemical Geology*, **340**, 131–138.

Iizuka, T., Hirata, T., Komiya, T., Rino, S., Katayama, I., Motoki, A. & Maruyama, S. 2005. U–Pb and Lu–Hf isotope systematics of zircons from the Mississippi River sand: implications for reworking and growth of continental crust. *Geology*, **33**, 485–488.

Iizuka, T., Komiya, T., Johnson, S. P., Kon, Y., Maruyama, S. & Hirata, T. 2009. Reworking of Hadean crust in the Acasta gneisses, northwestern Canada: evidence from in-situ Lu–Hf isotope analysis of zircon. *Chemical Geology*, **259**, 230–239.

Iizuka, T., Komiya, T., Rino, S., Maruyama, S. & Hirata, T. 2010. Detrital zircon evidence for Hf isotopic evolution of granitoid crust and continental growth. *Geochimica et Cosmochimica Acta*, **74**, 2450–2472.

Iizuka, T., Campbell, I. H., Allen, C. M., Gill, J. B., Maruyama, S. & Makoka, F. 2013. Evolution of the African continental crust as recorded by U–Pb, Lu–Hf and O isotopes in detrital zircons from modern rivers. *Geochimica et Cosmochimica Acta*, **107**, 96–120.

Kamber, B. S., Whitehouse, M. J., Bolhar, R. & Moorbath, S. 2005. Volcanic resurfacing and the early terrestrial crust: zircon U–Pb and REE constraints from the Isua Greenstone Belt, southern West Greenland. *Earth and Planetary Science Letters*, **240**, 276–290.

Kay, R. W. & Mahlburg Kay, S. 1993. Delamination and delamination magmatism. *Tectonophysics*, **219**, 177–189.

Kay, S. M., Godoy, E. & Kurtz, A. 2005. Episodic arc migration, crustal thickening, subduction erosion, and magmatism in the south-central Andes. *Geological Society of America Bulletin*, **117**, 67–88.

Keller, C. B. & Schoene, B. 2012. Statistical geochemistry reveals disruption in secular lithospheric evolution about 2.5 Gyr ago. *Nature*, **485**, 490–493.

Kemp, A. I. S., Hawkesworth, C. J., Paterson, B. A. & Kinny, P. D. 2006. Episodic growth of the Gondwana supercontinent from hafnium and oxygen isotopes in zircon. *Nature*, **439**, 580–583.

Kemp, A. I. S., Hawkesworth, C. J. et al. 2007. Magmatic and crustal differentiation history of granitic rocks from Hf–O isotopes in zircon. *Science*, **315**, 980–983.

Kemp, A. I. S., Hawkesworth, C. J., Collins, W. J., Gray, C. M. & Blevin, P. L. 2009. Isotopic evidence for rapid continental growth in an extensional accretionary orogen: the Tasmanides, eastern Australia. *Earth and Planetary Science Letters*, **284**, 455–466.

Kemp, A. I. S., Wilde, S. A. et al. 2010. Hadean crustal evolution revisited: new constraints from Pb–Hf isotope systematics of the Jack Hills zircons. *Earth and Planetary Science Letters*, **296**, 45–56.

King, E. M., Valley, J. W., Davis, D. W. & Edwards, G. R. 1998. Oxygen isotope ratios of Archean plutonic zircons from granite–greenstone belts of the Superior Province: indicator of magmatic source. *Precambrian Research*, **92**, 365–387.

Kleinhanns, I. C., Fullgraf, T., Wilsky, F., Nolte, N., Fliegel, D., Klemd, R. & Hansen, B. T. 2013. U–Pb zircon ages and (isotope) geochemical signatures of the Kamanjab Inlier (NW Namibia): constraints on Palaeoproterozoic crustal evolution along the southern Congo craton. *In*: Roberts, N. M. W., Van Kranendonk, M., Parman, S., Shirey, S. & Clift, P. D. (eds) *Continent Formation Through Time*. Geological Society, London, Special Publications, **389**, first published online August 22, 3013, http://dx.doi.org/10.1144/SP389.1

Komiya, T. 2007. Material circulation through time: chemical differentiation within the mantle and secular variation of temperature and composition of the mantle. *In*: Yuen, D., Maruyama, S., Karato, S. & Windley, B. D. (eds) *Superplumes: Beyond Plate Tectonics*. Springer, The Netherlands, 187–238.

Korenaga, J. 2011. Thermal evolution with a hydrating mantle and the initiation of plate tectonics in the early Earth. *Journal of Geophysical Research: Solid Earth (1978–2012)*, **116**.

Korenaga, J. 2013. Initiation and evolution of plate tectonics on Earth: theories and observations. *Annual Review of Earth and Planetary Sciences*, **41**, 117–151.

Kossert, K., Jörg, G. & Gostomski, C. L. V. 2013. Experimental half-life determination of ^{176}Lu. *Applied Radiation and Isotopes*, **81**, 140–145.

Kusky, T. M., Li, J. H. & Tucker, R. D. 2001. The Archean Dongwanzi ophiolite complex, North China Craton: 2.505-billion-year-old oceanic crust and mantle. *Science*, **292**, 1142–1145.

Kusky, T. M., Windley, B. F., Safonova, I., Wakita, K., Wakabayashi, J., Polat, A. & Santosh, M. 2013. Recognition of ocean plate stratigraphy in accretionary orogens through Earth history: a record of 3.8 billion years of sea floor spreading, subduction, and accretion. *Gondwana Research*, **24**, 501–547.

Labrosse, S. & Jaupart, C. 2007. Thermal evolution of the Earth: secular changes and fluctuations of plate characteristics. *Earth and Planetary Science Letters*, **260**, 465–481.

Lancaster, P. J., Storey, C. D., Hawkesworth, C. J. & Dhuime, B. 2011. Understanding the roles of crustal growth and preservation in the detrital zircon record. *Earth and Planetary Science Letters*, **305**, 405–412.

Lancaster, P. J., Storey, C. D. & Hawkesworth, C. J. 2014. The Eoarchaean foundation of the North Atlantic Craton. *In*: Roberts, N. M. W., Van Kranendonk, M., Parman, S., Shirey, S. & Clift, P. D. (eds) *Continent Formation Through Time*. Geological Society, London, Special Publications, **389**, first published online August 18, 2014, http://dx.doi.org/10.1144/SP389.11

Maas, R., Kinny, P. D., Williams, I. S., Froude, D. O. & Compston, W. 1992. The Earth's oldest known crust: a geochronological and geochemical study of 3900–4200 Ma old detrital zircons from Mt. Narryer and Jack Hills, Western Australia. *Geochimica et Cosmochimica Acta*, **56**, 1281–1300.

Mamani, M., Wörner, G. & Sempere, T. 2010. Geochemical variations in igneous rocks of the Central Andean orocline (13 S to 18 S): tracing crustal thickening and magma generation through time and space. *Geological Society of America Bulletin*, **122**, 162–182.

Martin, E., Bindeman, I. & Grove, T. L. 2011. The origin of high-Mg magmas in Mt Shasta and Medicine Lake volcanoes, Cascade Arc (California): higher and lower than mantle oxygen isotope signatures attributed to current and past subduction. *Contributions to Mineralogy and Petrology*, **162**, 945–960.

McCulloch, M. T. & Bennett, V. C. 1994. Progressive growth of the Earth's continental crust and depleted mantle: geochemical constraints. *Geochimica et Cosmochimica Acta*, **58**, 4717–4738.

McFarlane, C. R. & McKeough, M. 2013. Petrogenesis of the Kulyk Lake monazite-apatite-Fe (Ti)-oxide occurrence revealed using in-situ LA-(MC)-ICP-MS trace element mapping, U–Pb dating, and Sm–Nd isotope systematics on monazite. *American Mineralogist*, **98**, 1644–1659.

McFarlane, C. R. M. & McCulloch, M. T. 2007. Coupling of in-situ Sm–Nd systematics and U–Pb dating of monazite and allanite with applications to crustal evolution studies. *Chemical Geology*, **245**, 45–60.

McLennan, S. M., Taylor, S. R., McCulloch, M. T. & Maynard, J. B. 1990. Geochemical and Nd/Sr isotopic composition of deep-sea turbidites: crustal evolution and plate tectonic associations. *Geochimica et Cosmochimica Acta*, **54**, 2015–2050.

Mojzsis, S. J., Harrison, T. M. & Pidgeon, R. T. 2001. Oxygen-isotope evidence from ancient zircons for liquid water at the Earth's surface 4,300 Myr ago. *Nature*, **409**, 178–181.

Mojzsis, S. J., Cates, N. L., Caro, G., Trail, D., Abramov, O., Guitreau, M. & Bleeker, W. 2014. Component geochronology in the polyphase *c.* 3920 Ma Acasta Gneiss. *Geochimica et Cosmochimica Acta*, **133**, 68–96.

Næraa, T., Scherstén, A., Rosing, M. T., Kemp, A. I. S., Hoffmann, J. E., Kokfelt, T. F. & Whitehouse, M. J. 2012. Hafnium isotope evidence for a transition in the dynamics of continental growth 3.2 Gyr ago. *Nature*, **485**, 627–630.

Nebel, O., Nebel-Jacobsen, Y., Mezger, K. & Berndt, J. 2007. Initial Hf isotope compositions in magmatic zircon from early Proterozoic rocks from the Gawler Craton, Australia: a test for zircon model ages. *Chemical Geology*, **241**, 23–37.

Nebel, O., Vroon, P. Z., van Westrenen, W., Iizuka, T. & Davies, G. R. 2011. The effect of sediment recycling in subduction zones on the Hf isotope character of new arc crust, Banda arc, Indonesia. *Earth and Planetary Science Letters*, **303**, 240–250.

Nebel, O., Rapp, R. P. & Yaxley, G. M. 2014. The role of detrital zircons in Hadean crustal research. *Lithos*, **190**, 313–327.

Nebel-Jacobsen, Y., Münker, C., Nebel, O., Gerdes, A., Mezger, K. & Nelson, D. R. 2010. Reworking of Earth's first crust: constraints from Hf isotopes in Archean zircons from Mt. Narryer, Australia. *Precambrian Research*, **182**, 175–186.

Nutman, A. P. 2006. Comment on 'Zircon thermometer reveals minimum melting conditions on earliest Earth' II. *Science*, **311**, 779b–779b.

Nutman, A. P., Bennett, V. C. & Friend, C. R. 2013. The emergence of the Eoarchaean proto-arc: evolution of a c. 3700 Ma convergent plate boundary at Isua, southern West Greenland. *In*: Roberts, N. M. W., Van Kranendonk, M., Parman, S., Shirey, S. & Clift, P. D. (eds) *Continent Formation Through Time*. Geological Society, London, Special Publications, **389**, first published online September 9, 2013, http://dx.doi.org/10.1144/SP389.5

O'Neill, C., Lenardic, A., Moresi, L., Torsvik, T. H. & Lee, C. T. 2007. Episodic precambrian subduction. *Earth and Planetary Science Letters*, **262**, 552–562.

O'Neill, C., Lenardic, A. & Condie, K. C. 2013. Earth's punctuated tectonic evolution: cause and effect. *In*: Roberts, N. M. W., Van Kranendonk, M., Parman, S., Shirey, S. & Clift, P. D. (eds) *Continent Formation Through Time*. Geological Society, London, Special Publications, **389**. First published online November 6, 2013, http://dx.doi.org/10.1144/SP389.4

O'Nions, R. K., Evensen, N. M. & Hamilton, P. J. 1979. Geochemical modeling of mantle differentiation and crustal growth. *Journal of Geophysical Research: Solid Earth (1978–2012)*, **84**, 6091–6101.

Parman, S. W. 2007. Helium isotopic evidence for episodic mantle melting and crustal growth. *Nature*, **446**, 900–903.

Partin, C. A., Bekker, A., Sylvester, P. J., Wodicka, N., Stern, R. A., Chacko, T. & Heaman, L. M. 2014. Filling in the juvenile magmatic gap: evidence for uninterrupted Paleoproterozoic plate tectonics. *Earth and Planetary Science Letters*, **388**, 123–133.

Patchett, P. J. & Tatsumoto, M. 1981. Lu/Hf in chondrites and definition of a chondritic hafnium growth curve. *In*: *Lunar and Planetary Science Conference*, **12**, 819–821.

Pearson, D. G., Parman, S. W. & Nowell, G. M. 2007. A link between large mantle melting events and continent growth seen in osmium isotopes. *Nature*, **449**, 202–205.

Peck, W. H., Valley, J. W., Wilde, S. A. & Graham, C. M. 2001. Oxygen isotope ratios and rare earth elements in 3.3 to 4.4 Ga zircons: ion microprobe evidence for high $\delta^{18}O$ continental crust and oceans in the Early Archean. *Geochimica et Cosmochimica Acta*, **65**, 4215–4229.

Petersson, A., Scherstén, A., Andersson, J. & Möller, C. 2013. Zircon U–Pb and Hf–isotopes from the eastern part of the Sveconorwegian Orogen, SW Sweden: implications for the growth of Fennoscandia. *In*: Roberts, N. M. W., Van Kranendonk, M., Parman, S., Shirey, S. & Clift, P. D. (eds) *Continent Formation Through Time*. Geological Society, London, Special Publications, **389**, first published online August 22, 2013, http://dx.doi.org/10.1144/SP389.2

Pietranik, A. B., Hawkesworth, C. J., Storey, C. D., Kemp, A. I. S., Sircombe, K. N., Whitehouse, M. J. & Bleeker, W. 2008. Episodic, mafic crust formation from 4.5 to 2.8 Ga: new evidence from detrital zircons, Slave craton, Canada. *Geology*, **36**, 875–878.

Plank, T. & Langmuir, C. H. 1993. Tracing trace elements from sediment input to volcanic output at subduction zones. *Nature*, **362**, 739–743.

Polat, A., Hofmann, A. W. & Rosing, M. T. 2002. Boninite-like volcanic rocks in the 3.7–3.8 Ga Isua greenstone belt, West Greenland: geochemical evidence for intra-oceanic subduction zone processes in the early Earth. *Chemical Geology*, **184**, 231–254.

Polat, A., Kusky, T., Li, J., Fryer, B., Kerrich, R. & Patrick, K. 2005. Geochemistry of Neoarchean (ca. 2.55–2.50 Ga) volcanic and ophiolitic rocks in the Wutaishan greenstone belt, central orogenic belt, North China craton: implications for geodynamic setting and continental growth. *Geological Society of America Bulletin*, **117**, 1387–1399.

Polat, A., Kusky, T. & Li, J. 2007. Geochemistry of Neoarchean (*c.* 2.55–2.50 Ga) volcanic and ophiolitic rocks in the Wutaishan greenstone belt, central orogenic belt, North China craton: implications for geodynamic setting and continental growth: reply. *Geological Society of America Bulletin*, **119**, 490–492.

Poupinet, G. & Shapiro, N. M. 2009. Worldwide distribution of ages of the continental lithosphere derived from a global seismic tomographic model. *Lithos*, **109**, 125–130.

Rainbird, R. H., Hearnan, L. M. & Young, G. 1992. Sampling Laurentia: detrital zircon geochronology offers evidence for an extensive Neoproterozoic river system originating from the Grenville orogen. *Geology*, **20**, 351–354.

Rasmussen, B., Fletcher, I. R., Muhling, J. R., Gregory, C. J. & Wilde, S. A. 2011. Metamorphic replacement of mineral inclusions in detrital zircon from Jack Hills, Australia: implications for the Hadean Earth. *Geology*, **39**, 1143–1146.

Rino, S., Komiya, T., Windley, B. F., Katayama, I., Motoki, A. & Hirata, T. 2004. Major episodic increases of continental crustal growth determined from zircon ages of river sands; implications for mantle overturns in the Early Precambrian. *Physics of the Earth and Planetary Interiors*, **146**, 369–394.

Roberts, N. M. W. 2012. Increased loss of continental crust during supercontinent amalgamation. *Gondwana Research*, **21**, 994–1000.

Roberts, N. M. 2013. The boring billion?– Lid tectonics, continental growth and environmental change associated with the Columbia supercontinent. *Geoscience Frontiers*, **4**, 681–691.

Roberts, N. M. W. 2014. Preservation bias in the geological record: a Fennoscandian case study. *In*: *31st Nordic Geological Winter Meeting*, Lund.

Roberts, N. M. W., Slagstad, T., Parrish, R. R., Norry, M. J., Marker, M. & Horstwood, M. S. 2013. Sedimentary recycling in arc magmas: geochemical and U–Pb–Hf–O constraints on the Mesoproterozoic Suldal Arc, SW Norway. *Contributions to Mineralogy and Petrology*, **165**, 507–523.

Rudnick, R. L. & Gao, S. 2003. Composition of the continental crust. *Treatise on Geochemistry*, **3**, 1–64.

Santosh, M., Shaji, E., Tsunogae, T., Ram Mohan, M., Satyanarayanan, M. & Horie, K. 2013. Suprasubduction zone ophiolite from Agali hill: petrology, zircon SHRIMP U–Pb geochronology, geochemistry and implications for Neoarchean plate tectonics in southern India. *Precambrian Research*, **231**, 301–324.

Scherer, E., Münker, C. & Mezger, K. 2001. Calibration of the lutetium–hafnium clock. *Science*, **293**, 683–687.

Scholl, D. W. & von Huene, R. 2009. Implications of estimated magmatic additions and recycling losses at the subduction zones of accretionary (non-collisional) and collisional (suturing) orogens. Geological Society, London, Special Publications, **318**, 105–125.

Scholl, D. W., von Huene, R., Vallier, T. L. & Howell, D. G. 1980. Sedimentary masses and concepts about tectonic processes at underthrust ocean margins. *Geology*, **8**, 564–568.

Schott, B. & Schmeling, H. 1998. Delamination and detachment of a lithospheric root. *Tectonophysics*, **296**, 225–247.

Shirey, S. B. & Richardson, S. H. 2011. Start of the Wilson cycle at 3 Ga shown by diamonds from subcontinental mantle. *Science*, **333**, 434–436.

Sizova, E., Gerya, T., Brown, M. & Perchuk, L. L. 2010. Subduction styles in the Precambrian: insight from numerical experiments. *Lithos*, **116**, 209–229.

Sizova, E., Gerya, T. & Brown, M. 2014. Contrasting styles of Phanerozoic and Precambrian continental collision. *Gondwana Research*, **25**, 522–545.

Smithies, R. H., Van Kranendonk, M. J. & Champion, D. C. 2007. The Mesoarchean emergence of modern-style subduction. *Gondwana Research*, **11**, 50–68.

Söderlund, U., Patchett, P. J., Vervoort, J. D. & Isachsen, C. E. 2004. The ^{176}Lu decay constant determined by Lu–Hf and U–Pb isotope systematics of Precambrian mafic intrusions. *Earth and Planetary Science Letters*, **219**, 311–324.

Spaggiari, C. V., Pidgeon, R. T. & Wilde, S. A. 2007. The Jack Hills greenstone belt, Western Australia: part 2: lithological relationships and implications for the deposition of $\geq$4.0 Ga detrital zircons. *Precambrian Research*, **155**, 261–286.

Spencer, C. J., Cawood, P. A., Hawkesworth, C. & Roberts, N. M. W. 2013*a*. The isotopic artifacts of enhanced crustal preservation in collisional orogenesis. *Mineralogical Magazine*, **77**, 2245

Spencer, C. J., Hawkesworth, C., Cawood, P. A. & Dhuime, B. 2013*b*. Not all supercontinents are created equal: Gondwana–Rodinia case study. *Geology*, **41**, 795–798.

Spencer, C. J., Cawood, P. A., Hawkesworth, C. J., Raub, T. D., Prave, A. R. & Roberts, N. M. W. 2014. Proterozoic onset of crustal reworking and collisional tectonics: reappraisal of the zircon oxygen isotope record. *Geology*, **42**, 451–454.

Stern, C. R. 1991. Role of subduction erosion in the generation of Andean magmas. *Geology*, **19**, 78–81.

Stern, C. R. 2011. Subduction erosion: rates, mechanisms, and its role in arc magmatism and the evolution of the continental crust and mantle. *Gondwana Research*, **20**, 284–308.

Stern, R. A. & Bleeker, W. 1998. Age of the world's oldest rocks refined using Canada's SHRIMP: the Acasta Gneiss Complex, Northwest Territories, Canada. *Geoscience Canada*, **25**.

Stern, R. J. 2005. Evidence from ophiolites, blueschists, and ultrahigh-pressure metamorphic terranes that the modern episode of subduction tectonics began in Neoproterozoic time. *Geology*, **33**, 557–560.

Stern, R. J. 2008. Modern-style plate tectonics began in Neoproterozoic time: an alternative interpretation of Earth's tectonic history. *When Did Plate Tectonics Begin on Planet Earth*, 265–280.

Stern, R. J. & Scholl, D. W. 2010. Yin and yang of continental crust creation and destruction by plate tectonic processes. *International Geology Review*, **52**, 1–31.

Taylor, S. R. & McLennan, S. M. 1985. *The Continental Crust: Its Composition and Evolution*. Blackwell Scientific, Oxford, 1–312.

Trail, D., Mojzsis, S. J., Harrison, T. M., Schmitt, A. K., Watson, E. B. & Young, E. D. 2007. Constraints on Hadean zircon protoliths from oxygen isotopes, Ti-thermometry, and rare earth elements. *Geochemistry, Geophysics, Geosystems*, **8**, Q06014, http://dx.doi.org/10.1029/2006GC001449

Ushikubo, T., Kita, N. T., Cavosie, A. J., Wilde, S. A., Rudnick, R. L. & Valley, J. W. 2008. Lithium in Jack Hills zircons: evidence for extensive weathering of Earth's earliest crust. *Earth and Planetary Science Letters*, **272**, 666–676.

Valley, J. W. 2003. Oxygen isotopes in zircon. *Reviews in Mineralogy and Geochemistry*, **53**, 343–385.

Valley, J. W., Chiarenzelli, J. R. & McLelland, J. M. 1994. Oxygen isotope geochemistry of zircon. *Earth and Planetary Science Letters*, **126**, 187–206.

Valley, J. W., Kinny, P. D., Schulze, D. J. & Spicuzza, M. J. 1998. Zircon megacrysts from kimberlite: oxygen isotope variability among mantle melts. *Contributions to Mineralogy and Petrology*, **133**, 1–11.

Valley, J. W., Lackey, J. S. *et al.* 2005. 4.4 billion years of crustal maturation: oxygen isotope ratios of magmatic zircon. *Contributions to Mineralogy and Petrology*, **150**, 561–580.

Valley, J. W., Cavosie, A. J., Fu, B., Peck, W. H. & Wilde, S. A. 2006. Comment on 'Heterogeneous Hadean hafnium: evidence of continental crust at 4.4 to 4.5 Ga'. *Science*, **312**, 1139–1139.

Valley, J. W., Cavosie, A. J. *et al.* 2014. Hadean age for a post-magma-ocean zircon confirmed by atom-probe tomography. *Nature Geoscience*, **7**, 219–223.

van Hunen, J. & van den Berg, A. P. 2008. Plate tectonics on the early Earth: limitations imposed by strength and buoyancy of subducted lithosphere. *Lithos*, **103**, 217–235.

Van Kranendonk, M. J. 2010. Two types of Archean continental crust: plume and plate tectonics on early Earth. *American Journal of Science*, **310**, 1187–1209.

Van Kranendonk, M. J., Smithies, R. H. *et al.* 2014. Making it thick: a volcanic plateau origin of Paleoarchean continental lithosphere of the Pilbara and Kaapvaal cratons. *In*: Roberts, N. M. W., Van Kranendonk, M., Parman, S., Shirey, S. & Clift, P. D. (eds) *Continent Formation Through Time*. Geological Society, London, Special Publications, **389**, first published online September 3, 2014, http://dx.doi.org/10.1144/SP389.12

Veizer, J. & Jansen, S. L. 1979. Basement and sedimentary recycling and continental evolution. *The Journal of Geology*, 341–370.

Vermeesch, P. 2012. On the visualisation of detrital age distributions. *Chemical Geology*, **312**, 190–194.

Vervoort, J. D. & Blichert-Toft, J. 1999. Evolution of the depleted mantle: Hf isotope evidence from juvenile rocks through time. *Geochimica et Cosmochimica Acta*, **63**, 533–556.

Vervoort, J. D., Patchett, P. J., Blichert-Toft, J. & Albarède, F. 1999. Relationships between Lu–Hf and Sm–Nd isotopic systems in the global sedimentary system. *Earth and Planetary Science Letters*, **168**, 79–99.

Von Huene, R. & Scholl, D. W. 1993. The return of sialic material to the mantle indicated by terrigeneous material subducted at convergent margins. *Tectonophysics*, **219**, 163–175.

Walzer, U. & Hendel, R. 2013. Real episodic growth of continental crust or artifact of preservation? A 3-D geodynamic model. *Journal of Geophysical Research: Solid Earth*, **118**, 2356–2370.

Wang, C. Y., Campbell, I. H., Allen, C. M., Williams, I. S. & Eggins, S. M. 2009. Rate of growth of the preserved North American continental crust: evidence from Hf and O isotopes in Mississippi detrital zircons. *Geochimica et Cosmochimica Acta*, **73**, 712–728.

Wang, C. Y., Campbell, I. H., Stepanov, A. S., Allen, C. M. & Burtsev, I. N. 2011. Growth rate of the preserved continental crust: II. Constraints from Hf and O isotopes in detrital zircons from Greater Russian Rivers. *Geochimica et Cosmochimica Acta*, **75**, 1308–1345.

Watson, E. B. & Harrison, T. M. 2005. Zircon thermometer reveals minimum melting conditions on earliest Earth. *Science*, **308**, 841–844.

Watson, E. B., Wark, D. A. & Thomas, J. B. 2006. Crystallization thermometers for zircon and rutile. *Contributions to Mineralogy and Petrology*, **151**, 413–433.

Wilde, S. A., Valley, J. W., Peck, W. H. & Graham, C. M. 2001. Evidence from detrital zircons for the existence of continental crust and oceans on the Earth 4.4 Gyr ago. *Nature*, **409**, 175–178.

Windley, B. F. & Garde, A. A. 2009. Arc-generated blocks with crustal sections in the North Atlantic craton of West Greenland: crustal growth in the Archean with modern analogues. *Earth-Science Reviews*, **93**, 1–30.

Yang, J., Gao, S. *et al.* 2009. Episodic crustal growth of North China as revealed by U–Pb age and Hf isotopes of detrital zircons from modern rivers. *Geochimica et Cosmochimica Acta*, **73**, 2660–2673.

Yin, Q. Z., Wimpenny, J. *et al.* 2012. Crustal evolution of the South Mayo Trough, western Ireland, based on U–Pb ages and Hf–O isotopes in detrital zircons. *Journal of the Geological Society*, **169**, 681–689.

Zeh, A., Stern, R. A. & Gerdes, A. 2014. The oldest zircons of Africa – their U–Pb–Hf–O isotope and trace element systematics, and implications for Hadean to Archean crust–mantle evolution. *Precambrian Research*, **241**, 203–230.