1. Introduction

The growth rate of Earth’s continental crust remains a topic of much debate; two models that are in accord with the isotope record are: 1) that the average growth rate of continental crust has decreased since the Archaean (Belousova et al., 2010; Hawkesworth et al., 2010), and 2) that continental growth rate has episodically increased (e.g. Condie, 1998, 2004; Kemp et al., 2006; Iizuka et al., 2010). Recently it has been established that apparent episodic increases in growth rate are probably a result of preservation bias during the supercontinent cycle (Hawkesworth et al., 2009; Condie et al., 2011; Lancaster et al., 2011). The apparent episodic nature to continental growth is indicated by compilations of zircon U–Pb and Hf isotope data from individual continents (Rino et al., 2004, 2008; Lizuka et al., 2005, 2010; Kemp et al., 2006; Wang et al., 2009, 2011; Yang et al., 2009). A recent compilation of zircon-Hf data with near-global continental coverage reveals a more protracted history of continental growth during the last 3,000,000,000 years (Belousova et al., 2010); however, peaks in the U–Pb age data indicate a possible role for bias during the supercontinent cycle with preferential preservation of continental crust during supercontinent amalgamation (Condie et al., 2011; Lancaster et al., 2011). Overlap between peaks in both U–Pb ages and Hf model ages have been shown to overlap with periods of supercontinent amalgamation (e.g. Kemp et al., 2006; Lizuka et al., 2010), which has been used in support of a model whereby increased continental growth occurs during supercontinent amalgamation (Condie, 1998; Rino et al., 2004).

Addition of continental crust largely occurs via magmatism driven by subduction in accretionary orogens, with addition from plumes being comparatively minor; continental destruction occurs via subduction of sediments, subduction erosion, and lower crustal delamination (e.g. Clift...
et al., 2009a,b; Keppie et al., 2009, 2011; Hawkesworth et al., 2010; Stern and Scholl, 2010; Stern, 2011). The growth rate of continental crust relies on the balance between addition and loss of continental crust; a concept referred to as Yin and Yang by Stern and Scholl (2010). Mass balance calculations indicate that at present-day, the amount of loss is balanced by that of continental addition (Clift and Vannucchi, 2004; Scholl and von Huene, 2007, 2009; Clift et al., 2009a,b), indicating that the crustal volume is not presently increasing; although a more recent estimation suggests that continental loss outweighs continental addition, implying that the continental crust is currently shrinking (Stern, 2011). The balance between continental addition and loss, and thus continental growth rate, is not a steady-state process, and is controlled by the supercontinent cycle. Continental crust growth rate should increase during supercontinent break-up due to an increased magmatic flux, and conversely, this growth rate should decrease during supercontinent amalgamation due to increased sediment-subduction and tectonic erosion (Santosh, 2010; Stern and Scholl, 2010; Yoshida and Santosh, 2011). To evaluate the dichotomy between previous models of episodic crustal growth and the concept of increased continental growth during supercontinent break-up, this study applies a novel approach to the analysis of a recently published global zircon–Hf database; rather than using estimated model ages, the distribution of Hf data within time-intervals is calculated as an average, and it is the trend of this average that is examined with respect to supercontinent cycles.

2. The zircon record

Detrital zircons, especially those from large rivers, provide a means of sampling continental crust to obtain its age and isotopic composition (e.g. Bodet and Schärer, 2000; Lizuka et al., 2005, 2010; Wang et al., 2009, 2011). A recent compilation of ~15,000 analyses, including those from major rivers on most continents, provides a dataset of igneous and detrital zircon data with near-global continental coverage (Belousova et al., 2010); although this dataset has an Australian bias, it has been demonstrated that this doesn’t significantly change the general distribution of data (Belousova et al., 2010; Condie et al., 2011). These data reveal that juvenile continental crust has been forming continuously since the Hadean, and that 70% of the present-day volume of continental crust has existed since 2.5 Ga (Belousova et al., 2010; Hawkesworth et al., 2010).

Previous models for continental growth typically rely on calculated depleted mantle Nd and Hf model ages; however, rather than being geological events, these are typically mixed ages that average different components within a magma, and for Hf-in-zircon rely on an estimation of the magma’s Lu/Hf ratio. Combining zircon–Hf analyses with oxygen isotopes allows for a better estimation of Lu/Hf (Wang et al., 2009, 2011), and can also be used to screen out non-mantle signatures that derive from sedimentary recycling (Kemp et al., 2006; Lancaster et al., 2011). However, it has been demonstrated that recent magmas in young arcs can have a large spread in Hf model ages even when exhibiting an oxygen isotope signature of mantle-derived zircon (Nebel et al., 2011). In this study, rather than using model ages, the relative distribution of Hf data within individual time intervals is examined, i.e. the extent to which the population of Hf data has departed from the depleted mantle curve within each time interval.

Fig. 1 shows a compilation of global Hf data (data from Belousova et al., 2010 and Lancaster et al., 2011) plotted as $\varepsilon_{Hf(t)}$ versus time, with the mean, median and interquartile range shown and based on 50 m.y. intervals and a 100 m.y. moving window. The mean and median are similar throughout, only in the Phanerozoic where there is greatest variation, does the median depart from the mean, indicating only moderate skew to the population. Over time the depleted mantle curve increases into positive $\varepsilon_{Hf(t)}$ space, and the minimum value of $\varepsilon_{Hf(t)}$ (since a magma should not have a model age older than ~4.56 Ga) decreases into negative $\varepsilon_{Hf(t)}$ space; thus, the possible range in $\varepsilon_{Hf(t)}$ for any one time interval will increase over time. For this reason, the mean curve is recalculated so that it represents the percentage of depleted mantle input within an available range, with this range being based on the depleted mantle value plus two epsilon units for the maximum, and an evolution line from 4.56 Ga using a crustal Lu/Hf value of 0.015 for the minimum (mantle–input curve in Fig. 1); this mantle input curve is plotted with the mean of the Hf data as zero. The population density in the Palaeoarchaean and Hadean is suggested to be too small and biased by too few samples to represent a reliable average of global continental crust, thus, the mantle–input curve is only shown up to 3300 Ma.

3. An oscillating growth curve

The Hf data is used as a proxy for continental growth in the following ways: positive excursions (from the mean) indicate increased mantle input and continental addition, negative excursions indicate decreased mantle input and increased continental loss. The population of Hf data is limited in the Hadean and Palaeoarchaean, but from ~4.5 to ~2.0 Ga, can be described as a broad triangular swath that is rather symmetrical about the CHUR evolution. Although the median and mean trends show variation during this period, the record for Mesaoarchaean and older is likely biased by sampling, with most samples coming from only a few localities; thus, no significant interpretation of the shape of the mantle–input curve is made for this period. After ~2.0 Ga the record is more varied with significant excursions of the data into positive and negative $\varepsilon_{Hf(t)}$ space. Positive excursions occur at ~1.7–1.2 Ga, ~0.85–0.65 Ga, and ~0.45–0.35 Ga, whilst negative excursions occur at ~1.1–0.9 Ga, ~0.65–0.50 Ga and ~0.2–0.1 Ga. The excursions correlate for most of the last ~2.0 Ga with the supercontinent cycle, with negative excursions at ~1.0 Ga and ~0.55 Ga overlapping formation of the Rodinia and Gondwana supercontinents respectively, and positive excursions at ~1.7–1.2 Ga, ~0.75 Ga, and ~0.40 Ga overlapping the break-up of Columbia (Nuna), Rodinia and Gondwana respectively. If the excursions are representative of global continental addition versus continental loss, then for the last ~2.0 Ga the data point to increased continental growth rate during periods of supercontinent break-up, and decreased growth rate during supercontinental formation.

4. Controls on continental growth rate

Addition and loss of continental material are controlled by both the type of orogen, and by the geodynamic regime within the orogen. Internal orogens, i.e. continental collision belts such as the Alpine–Himalayan–Indonesian system, feature both addition and loss of continental crust, but with recycling of pre-existing crust probably more dominant than addition of juvenile crust. This is represented in the zircon Hf record as deviation to both positive and negative $\varepsilon_{Hf}$ values that progressively fan out over time (Fig. 2), due to lower crust and lithospheric mantle being replaced by continental lithosphere from accreting continents (Collins et al., 2011). External orogens, i.e. accretionary or circum-Pacific orogens, feature a trend to positive $\varepsilon_{Hf}$ values from their onset until termination of the orogenic system; this is related to removal of lower crust and lithospheric mantle by newly formed crust which melts to produce rocks with positive $\varepsilon_{Hf}$ signatures (Collins et al., 2011). The pattern of zircon–Hf data from across the globe at any one time can therefore be thought of as representing the balance between crustal production in internal orogens versus that in external orogens. Internal orogens will outbalance external orogens during periods of supercontinent amalgamation, whereas external orogens will dominate when the continents are dispersed. The balance between internal versus external orogens during the supercontinent cycle thus provides a first order control on the continental crust growth rate.

Accretionary (i.e. external) orogens can be divided into retreating- or advancing-types, based on the relative movement of the subduction trench relative to the over-riding plate, and will typically oscillate over the timespan of a subduction-zone margin (e.g. Collins, 2002). Isotope evidence from arc magmas in the Tasmanides indicate that retreating-
mode favours increased mantle input, i.e. continental crust addition, whereas advancing-mode features decreased mantle input (Kemp et al., 2009; Phillips et al., 2011); this is due to a variety of factors, such as attenuated thinner crust in retreating mode, allowing greater influx of mantle material; and in advancing mode, thicker crust, increased sediment input and increased subduction erosion leading to increased crustal contamination. Examples of this relationship have been documented for Andean and Japanese subduction zones using various isotopic systems (Takagi, 2004; Kay et al., 2005; Stern et al., 2011); in all cases the isotope data point to an increased crustal signature when the volume of subducted continental material is increased. Since only 10 to 20% of subducted material is recycled back into arc magmas (Clift et al., 2009a,b; Stern, 2011), deviation from depleted mantle values for isotopes within arc magmas can be regarded not only as continental recycling, but as a proxy for continental loss. The balance between accretionary orogens in retreating mode versus those in advancing mode provides a second order control on continental growth rate.

5. Controls on preservation of continental crust

It has been argued that the record of preserved continental crust is biased by variation in preservation during the supercontinent cycle (Hawkesworth et al., 2009; Condie et al., 2011; Lancaster et al., 2011). Hawkesworth et al. (2009) suggest that crust formed preceding supercontinent formation has a moderate preservation potential, crust formed during supercontinent break-up has a weak preservation potential, and crust formed during supercontinent amalgamation has the greatest preservation potential; this is due to capturing of crust during continent–continent collisions and stabilisation within cratons. This process may be responsible for peaks in U–Pb age histograms, i.e. an apparent episodic nature to continental growth (e.g. Rino et al., 2008;
The effect of preservation during the supercontinent cycle is conceptualised in Fig. 3. If continental addition equals continental loss throughout Earth history, then the Hf data will comprise an array to increasingly positive and negative εHf values over time (Fig. 3, 1A). If continental addition versus continental loss is always balanced, but preservation of continental crust is increased during supercontinent amalgamation, then the data will fill the same array, but will be concentrated in time periods of supercontinent amalgamation (Fig. 3, 1B). The same array will occur if continental addition versus continental loss is balanced, but addition of continental crust is increased during supercontinent amalgamation. If supercontinent amalgamation increases the preservation of ‘recycled’ crust compared to ‘juvenile’ crust (i.e. low εHf crust compared to high εHf), but continental addition equals continental loss, then an array like that in Fig. 3 (4a) can be envisaged; in this case, the mantle input curve will vary over time, as is displayed in Fig. 1. As has been previously discussed, continental addition has likely outweighed continental loss during certain times within earth history; on the εHf-time plot, this will be represented as an array that oscillates over time (Fig. 3, 2A–3B). Increased continental addition during supercontinent amalgamation will produce positive excursions in the mantle-input curve during these periods, and vice versa. The combination of the U–Pb histogram, the Hf-time array, and the mantle input curve, can thus be used to make inferences about the role of preferential preservation during the supercontinent cycle, and the balance between continental addition versus continental loss. The examined zircon–Hf dataset exhibits a scenario similar to that depicted in Fig. 3 (3B), whereby continental loss outweighs continental addition during supercontinent amalgamation.

Fig. 3. The concept of increased preservation potential during supercontinent formation, and the concept of continental addition versus continental loss, are displayed using four hypothetical supercontinents at 0.5, 1.0, 2.0 and 3.0 Ga, with darker shading representing a greater population of data. 1A, 2A and 3A feature no preferential preservation of crust; 1B, 2B and 3B feature preferential preservation of all crust (‘juvenile’ and ‘recycled’); 4A features preferential preservation of ‘recycled’ crust compared to ‘juvenile’ crust. The impact on a U–Pb age histogram, and a calculated mantle input curve similar to that in Fig. 1 is also depicted. The mantle input curve is similar in both 3B and 4A, but the shape of the data array in εHf-time space is indicative of the processes involved.
and whereby these periods feature either preferential preservation of continental crust, or an increased production of continental crust; the data do not allow discrimination between the latter two processes.

6. The supercontinent cycle

Evidence for a single supercontinent existing in the Archaean is lacking (Senshu et al., 2009), although there is evidence for formation of ‘supercontinents’ around 2700 Ma (Rogers, 1996; Bleeker, 2003; Eriksson et al., 2009). The first true supercontinent is that of Columbia (Rogers and Santosh, 2002); the maximum packing of this supercontinent based on timing of collisional orogenies is around 1.9–1.85 Ga (Zhao et al., 2004; Rogers and Santosh, 2009; Yakubchuk, 2010). The next supercontinent to form was Rodinia at ~1.1–1.0 Ga (e.g. Dalziel, 1991; Hoffman, 1991; Li et al., 2008). Later supercontinents, although well studied, still feature debate on their timing in regard to true supercontinent formation. General consensus is that Gondwana formed around 550 Ma, and that Pangea formed around 250 Ma (e.g. Veevers, 2004; Collins and Psarevsky, 2005; Meert and Liebermann, 2008; Murphy et al., 2009). Senshu et al. (2009) define Gondwana and Pangea as a single supercontinent, with Gondwana only being a significant landmass that formed prior to later supercontinent amalgamation; based on mantle dynamics these workers advocate the timing of supercontinent formation to be 340 Ma.

According to an idealised model of a supercontinent cycle, i.e. the Wilson cycle, the time period between the aforementioned supercontinents should see a dispersal of continents referred to as supercontinent break-up (e.g. Nance et al., 1988). In reality, the formation of supercontinents does not seem to fit such a pattern. Supercontinents can form by two end-member processes: introversion (i.e. ‘inside-in’), where oceanic spreading along interior orogens is transformed to collision along the same orogens, and extroversion (i.e. ‘inside-out’) where exterior accretionary orogens are transformed into interior collisional orogens (Hoffman, 1991; Murphy and Nance, 2003, 2008; Murphy et al., 2009). The latter has been proposed for formation of Columbia, Rodinia and Gondwana (e.g. Hoffman, 1991; Rogers and Santosh, 2002; Murphy and Nance, 2003; Rino et al., 2008), whereas introversion is indicated for formation of Pangea (Murphy and Nance, 2003, 2008; Murphy et al., 2009). The degree that continents were dispersed prior to the maximum packing of supercontinents has varied over time. For example, there appears to be a lack of evidence indicating break-up of Columbia; although some large radial dyke swarms suggest rifting amongst continental blocks (e.g. Ernst et al., 2008; Hou et al., 2008; Goldberg, 2010), palaeomagnetic evidence allows for the continents to be connected throughout most of the ~1.8–1.2 Ga period (Yakubchuk, 2010). Rather than break-up completely, it appears that large continental blocks rotated from their position within Columbia to new positions within Rodinia, a process that transformed exterior orogens into interior orogens such as the Grenville Orogen (e.g. Johansson, 2009; Li et al., 2008), i.e. the process of extroversion. The break-up of Rodinia appears to have seen more significant dispersal of continents prior to formation of Gondwana and subsequently Pangea (Li et al., 2008), than that of Columbia, although this may be a consequence of more reliable palaeomagnetic data from this time period. The break-up of Gondwana involved separation of several continental blocks via spreading in the interior Rheic ocean, but was halted by a geodynamic change that led to reamalgamation along this interior orogen (e.g. Murphy and Nance, 2008; Murphy et al., 2009); again, only partial dispersal of large continental blocks occurred during supercontinent break-up. The break-up of Pangea has been occurring since 200–180 Ma (e.g. Veevers, 2004; Santosh et al., 2009), and has led to the current dispersal of continents that we see today.

It appears that the supercontinent cycle has not involved complete separation of continents prior to reamalgamation into single supercontinents, but rather a cycle whereby large groups of continents are rifted, rotated and reamalgamated. Even with this being the case, the effect on global addition and loss of continental crust will still be the same: interior orogens will be most prominent during the maximum packing of supercontinents, and exterior orogens will be most prominent during maximum dispersal of supercontinents. During supercontinent break-up, the balance between exterior orogens that are retreating compared to those that are advancing, may be largely controlled by the rate of oceanic spreading, i.e. a global increase in oceanic spreading should lead to a greater dominance of advancing accretionary orogens. Whether retreating or advancing orogens are more dominant will therefore depend on mantle convection rates as well as plate movements.

Peaks in the U–Pb histogram occur at 2700, 2450, 2000, 1850, 1600, 1150, 1000, 500, and 250 Ma. These may reflect increased preservation of continental crust at these times, or increased production of crust. These peaks largely overlap periods in time where supercontinents are amalgamating, particularly the more significant peaks at 2700, 1850, 1100–1000, 500 and 250 Ma. The εHf array, and associated mantle-input curve (Fig. 1), suggest increased continental addition (i.e. increased continental crust growth rate) during the periods subsequent to supercontinent amalgamation, i.e. at ~1.7–1.2 Ga, ~0.85–0.65 Ga, and ~0.45–0.35 Ga. Periods of increased continental loss (i.e. decreased continental crust growth rate), occurring at ~1.1–0.9 Ga, and ~0.65–0.50 Ga overlap the periods of supercontinent amalgamation for Rodinia and Gondwana respectively. The amalgamation of Columbia features significant crustal loss, as indicated by the low εHf values, but also features substantial concurrent juvenile growth (high εHf). This pattern is also seen during amalgamation of Pangea at 250 Ma. A possible explanation for this irregularity in the pattern of continental growth, is that of the different processes of supercontinent formation. Exterior orogens that initiated during Gondwana formation, i.e. the Terra Australis orogen (Cawood and Buchan, 2007), allowed significant continental growth during break-up of Gondwana, but rather than being shutdown, these orogens continued to host continental growth during formation of Pangea, since they were still on the exterior of the supercontinent. Based on the εHf array, it is suggested that introversion may have been responsible for the formation of Columbia at ~1.9–1.85 Ga. The supercontinents that are well documented to have occurred by extroversion, i.e. Rodinia and Gondwana (e.g. Hoffman, 1991; Li et al., 2008), have the greatest amplitude in variations in the εHf array, suggesting there may be a greater imbalance between continental loss and continental addition during the process of extroversion compared to that of introversion.

7. Conclusions

Based on the assumption that εHf of detrital zircons provide a proxy record for the amount of continental addition versus continental loss, the zircon–Hf data presented here, indicate that the balance of addition and loss of continental crust has varied throughout Earth’s history. Periods of supercontinent amalgamation feature increased continental loss, whereas increased continental addition occurs in the periods subsequent to supercontinent amalgamation; these are partly correlative with supercontinent break-up, but this term is avoided since complete break-up appears to be rare. The data confirm the non steady-state nature of continental growth, and are compatible with the concept of Yin and Yang, i.e. continental growth rate increases during supercontinent break-up (Stern and Scholl, 2010); however, it should be noted that since the overall growth rate of continental crust has likely been decreasing since the Archaean (Hawkesworth et al., 2009; Belousova et al., 2010), the variations in growth rate indicated in this study do not confirm actual increases in continental crustal volume, but changes in the balance between addition and loss of continental crust.

The balance between continental loss and addition is controlled by: 1) the balance between interior orogens and exterior orogens, with the latter hosting greater continental addition, and 2) the balance within exterior orogens between advancing and retreating modes, with the latter also hosting greater continental addition. The process of
supercontinent formation also controls the balance between these orogens, introversion allows continuous continental growth along exterior orogens concurrent with continental loss within interior orogens, whilst extroversion features termination of growth along exterior orogens as they are transformed into interior orogens. A greater imbalance between continental addition and continental loss seems to occur during cycles of extroversion, i.e. during the formation and break-up of Rodinia and Gondwana. Pangea formed by introversion (Murphy and Nance, 2003, 2008; Murphy et al., 2009), and by analogy, the concurrent continental addition with continental loss during amalgamation of Columbia, suggests that this supercontinent may also have formed by introversion. Preferential preservation of continental crust during the supercontinent cycle has likely affected the detrital U–Pb and Hf record, with increases in U–Pb crystallisation resulting from greater preservation potential during supercontinent amalgamation (Hawkesworth et al., 2009); however, by examining the distribution of Hf data, it is suggested that this bias has not significantly affected the general pattern of positive and negative εHf excursions, and therefore the interpretation of continental loss versus continental loss remains applicable. Although the interpretation of data in this study uses a qualitative approach, it presents a novel analysis of continental growth; more rigorous testing of crustal growth models in the future will be amenable with the constantly increasing detrital zircon database.

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