

The continental record and the generation of continental crust



INVITED REVIEW

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ABSTRACT

Continental crust is the archive of Earth history. The spatial and temporal distribution of Earth's record of rock units and events is heterogeneous; for example, ages of igneous crystallization, metamorphism, continental margins, mineralization, and seawater and atmospheric proxies are distributed about a series of peaks and troughs. This distribution reflects the different preservation potential of rocks generated in different tectonic settings, rather than fundamental pulses of activity, and the peaks of ages are linked to the timing of supercontinent assembly. The physiochemical resilience of zircons and their derivation largely from felsic igneous rocks means that they are important indicators of the crustal record. Furthermore, detrital zircons, which sample a range of source rocks, provide a more representative record than direct analysis of grains in igneous rocks. Analysis of detrital zircons suggests that at least ~60%–70% of the present volume of the continental crust had been generated by 3 Ga. Such estimates seek to take account of the extent to which the old crustal material is underrepresented in the sedimentary record, and they imply that there were greater volumes of continental crust in the Archean than might be inferred from the compositions of detrital zircons and sediments. The growth of continental crust was a continuous rather than an episodic process, but there was a marked decrease in the rate of crustal growth at ca. 3 Ga, which may have been linked to the onset of significant crustal recycling, probably through subduction at convergent plate margins. The Hadean and Early Archean continental record is poorly preserved and characterized by a bimodal TTG (tonalites, trondhjemites, and granodiorites) and greenstone association that

differs from the younger record that can be more directly related to a plate-tectonic regime. The paucity of this early record has led to competing and equivocal models invoking plate-tectonic- and mantle-plume-dominated processes. The 60%–70% of the present volume of the continental crust estimated to have been present at 3 Ga contrasts markedly with the <10% of crust of that age apparently still preserved and requires ongoing destruction (recycling) of crust and subcontinental mantle lithosphere back into the mantle through processes such as subduction and delamination.

INTRODUCTION

Earth has a bimodal surface elevation reflecting the contrasting chemical-mechanical properties of continental and oceanic crust. The latter is dense, gravitationally unstable, and hence young, whereas continental crust is buoyant and represents the archive of Earth history (Fig. 1), not only of the crust itself but through the rock record of the atmosphere, hydrosphere, and biosphere, and of the mantle through its interactions with the crust. Over 4.5 b.y., the continental crust has evolved to form the environment in which we live and the resources on which we depend. An understanding of the crust and its record is fundamental to resolving questions on the origin of life, the evolution and oxygenation of our atmosphere, past climates, mass extinctions, the thermal evolution of Earth, and the interactions between the surficial and deep Earth. Yet, when and how the continental crust was generated, the volume of continental crust through Earth history, and whether it provides a representative record remain fundamental questions in the earth sciences.

Our aim in this paper is first briefly to outline the general character of the crust and to clarify the terms used, and then to explore the nature of the continental record that is available for study, to evaluate different approaches for when and

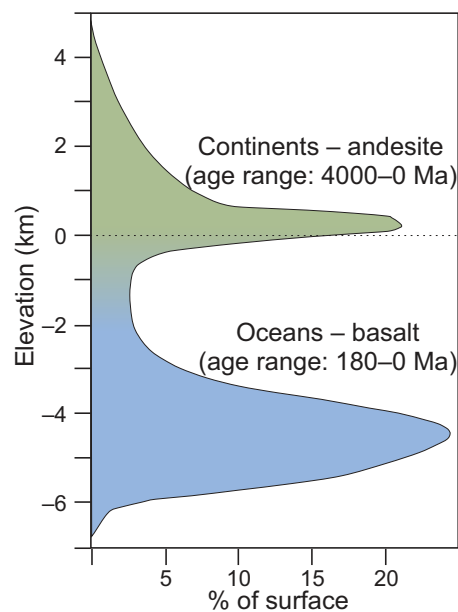


Figure 1. Bimodal distribution of surface elevations on Earth, relative to sea-level, emphasizing the contrasting physical-chemical properties of continental and ocean crust. Figure is adapted from Taylor and McLennan (1985, 1996).

how the continental crust was generated, and to discuss constraints on the rates of continental growth through time and the relationship to tectonic processes, especially the role of the supercontinent cycle in controlling the geological record. We explore the information available from the zircon record, since igneous zircons preserved as detritus in sedimentary rocks are increasingly regarded as a key temporal record of the activity associated with the generation and evolution of the continental crust.

Views on the evolution of the continental crust have changed dramatically as ideas on geological processes have evolved and as methods to interrogate the rock record have

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advanced through developments in stratigraphic analysis, petrography, paleontology, geochemistry, geochronology, geophysics, and modeling. Crucially, our understanding of the processes involved in the generation and the evolution of the continental crust has grown enormously through the latter part of the twentieth and beginning of the twenty-first centuries following on from the development and acceptance of plate-tectonic theory. This has focused our research on plate margins, the sites of continental crust formation and stabilization, and it has resulted in a fundamental change in the way we approach our interrogation of Earth and its record from a descriptive documentation of units and events into investigation into the processes controlling these features. A factor critical to determining these processes is an understanding of rates of change, and this has been facilitated by developments in data collection and analysis. This expansion of knowledge has been particularly important in further understanding not just the exposed surficial rock record, but in gaining insight into the composition and development of the whole crust. In particular, this has led to new ideas into what shaped the record, and how representative, or unrepresentative, it may be.

SOME FACTS AND TERMINOLOGY

The total area of continental crust is 210.4×10^6 km², or some 41% of the surface area of Earth, and the volume is 7.2×10^9 km³, which constitutes some 70% of Earth's crustal volume (Cogley, 1984). The crust extends vertically from the surface to the Mohorovičić discontinuity (Moho) and laterally to the break in slope in the continental shelf (Rudnick and Gao, 2003). The Moho is defined as the jump in seismic primary waves (P-waves) to greater than ~ 7.6 km s⁻¹. This change in seismic-wave velocity is taken as a first-order approximation of the boundary between mafic lower-crustal rock and ultramafic mantle peridotite: the crust-mantle boundary. Petrologic studies from exposed sections of ocean floor (ophiolites) and from xenoliths in continental settings suggest that in some situations, the crust-mantle boundary may differ from the Moho (Malpas, 1978; Griffin and O'Reilly, 1987).

The mean elevation of the continental crust is around 125 m (Fig. 1), and some 31% of the crustal area is below sea level. Continental crustal thickness varies from 20 to 70 km, averaging around 35–40 km (Fig. 2; Mooney et al., 1998). The crust and underlying mantle constitute the lithosphere; the mechanically strong outer layer of Earth that forms the surface plates (Barrell, 1914a, 1914b, 1914c; Daly, 1940; White, 1988). Heat transport in the lithosphere

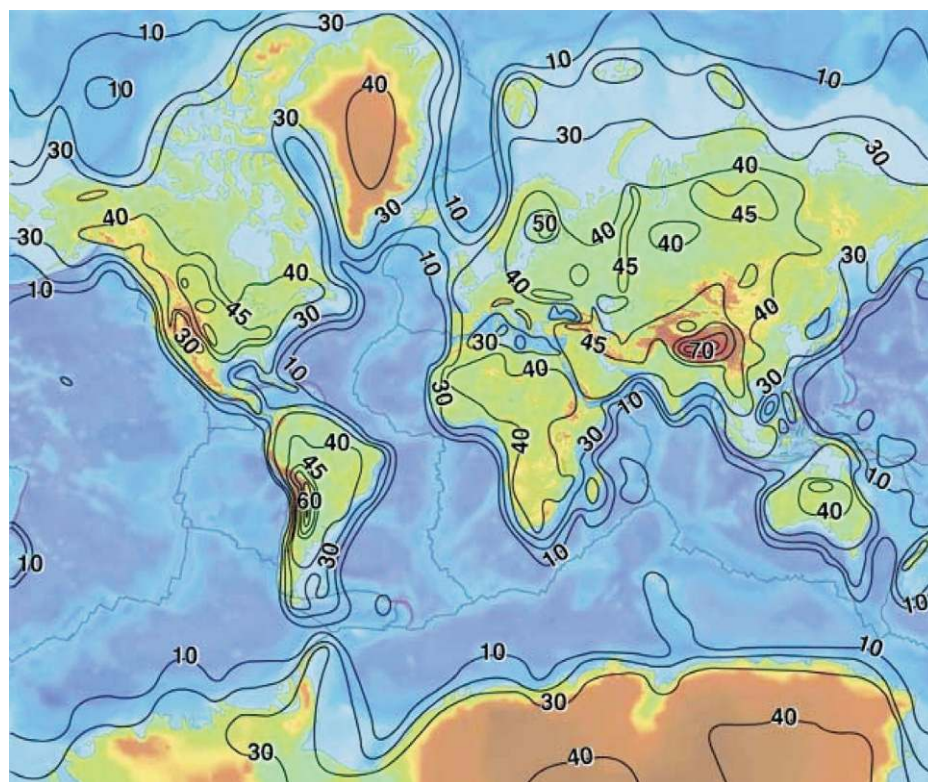


Figure 2. Contour map of the thickness of Earth's crust (developed from model CRUST 5.1). The contour interval is 10 km (45 km contour interval is also shown to provide greater detail on the continents). To a first approximation, the continents and their margins are outlined by the 30 km contour. Source of figure: <http://earthquake.usgs.gov/research/structure/crust/index.php>.

is conductive, and the base is a rheological boundary with the isothermal convecting mantle (Sleep, 2005).

Continents include cratons, areas of stable crust, orogenic belts, and regions of continental extension, which form either intracratonic rifts or develop into zones of continental breakup and thermal subsidence (passive margins). Orogens evolve through one or more cycles of sedimentation, subsidence, and igneous activity punctuated by tectonothermal events (orogenies), involving deformation, metamorphism, and igneous activity, which result in thickening and stabilization of the lithosphere (Figs. 3A and 3B). Cratons are ancient orogens that have generally been undeformed and tectonically stable for long periods of time, often since the Archean, and are divisible into shields, which are regions of exposed crystalline igneous and metamorphic rock, and platforms, where the shield is overlain by a relatively undeformed sedimentary succession (Fig. 4).

Geologic and geophysical data show that the crust is divisible into a felsic upper crust composed largely of sedimentary rock (upper few kilometers) and granite to granodiorite,

a heterogeneous middle crust assemblage of orthogneiss and paragneiss at amphibolite facies to lower granulite facies (Fig. 3C), and a lower crust consisting of granulite-facies country rocks and basic intrusive rocks and/or cumulates (Rudnick and Fountain, 1995; Wedepohl, 1995; Rudnick and Gao, 2003). Thicknesses of the three crustal layers vary, but the upper and middle crustal sections generally form around 30% each of a typical crustal profile, with the lower crust forming the remaining 40% (Fig. 5; Rudnick and Gao, 2003; Hawkesworth and Kemp, 2006a).

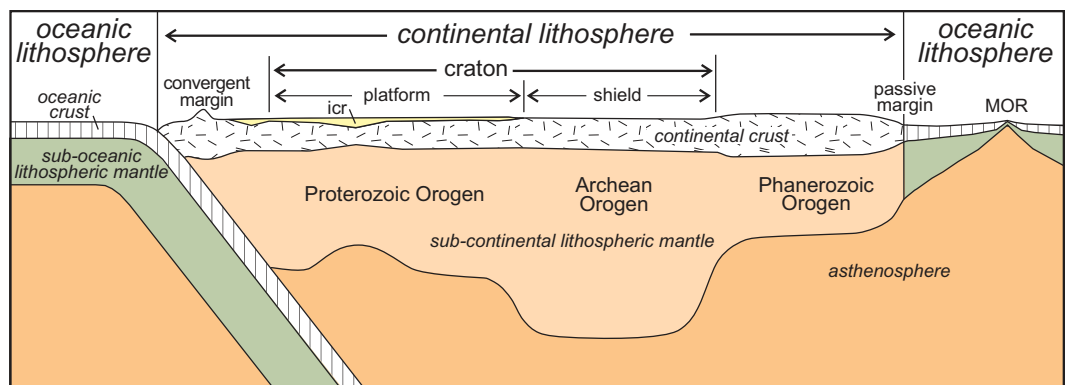
The bulk composition of the crust is equivalent to andesite (Fig. 3D) and requires two stages of formation involving extraction of mafic magmas from the mantle and their differentiation through either fractional crystallization or remelting and return of the cumulate or residue to the mantle (Taylor, 1967; Taylor and McLennan, 1985; Kay and Kay, 1991; Rudnick, 1995; Rudnick and Gao, 2003; Davidson and Arculus, 2006; Hacker et al., 2011).

Surface heat flux of Archean cratons is generally low (30–40 mW m⁻²), Phanerozoic regions show higher heat flux values (>60–80 mW m⁻²),

Figure 3. (A) Cerro Aconcagua (6962 m) in western Argentina is the highest mountain outside Asia and the tallest in the Americas. It lies within the accretionary Andean orogen and formed through noncollisional coupling between the downgoing Nazca oceanic plate and the continental lithosphere of the South American plate. (B) Glencoul thrust on the northern side of Loch Glencoul, NW Highlands, Scotland. Silurian-age thrust faulting of Mesoproterozoic and Neoproterozoic (Lewisian) gneiss over a footwall of Cambrian strata (Eriboll and An-t-Sron Formations) that are resting unconformably on underlying Lewisian basement. (C) Road cut north of Loch Laxford, NW Highlands, Scotland, showing Mesoproterozoic and Neoproterozoic gray gneiss (Lewisian), cut by Paleoproterozoic mafic intrusions (Scourie dikes, now foliated and metamorphosed within the amphibolite facies), and discordant sheets of ca. 1.85 Ga (Laxfordian) granite and pegmatite. (D) Construction of continental crust through andesitic magmatism demonstrated by the Nevados de Payachatas volcanoes: Parinacota (right) and Pomerape (left) in northern Chile. Both were constructed within the last 300 k.y., with Pomerape the older (more eroded) and Parinacota the younger, having undergone postglacial sector collapse to produce a large debris avalanche (foreground hummocks). (E) The Stones of Calanais insert (Callanish) are a Neolithic monument on the island of Lewis, Scotland, composed of Lewisian Gneiss. The stones are arranged in a central circle augmented by linear avenues leading off to the points of the compass. To the north, the avenue comprises a double row of stones. At the heart of the monument, there is a Neolithic burial mound that was raised some time after the erection of the stones. The precise date of construction of the circle is unclear, but the stones were erected ~5000 yr ago. The purpose of sites like this (including Stonehenge) remains a mystery, but research suggests that they constituted important central places for the local farming community and may have been aligned to prominent features of both land and sky. The stones were quarried locally, and their transport and construction would have formed a focal activity at the time. (F) Siccar Point, Berwickshire, NE England. Hutton's iconic angular unconformity between gently dipping Devonian Upper Old Red Sandstone on near-vertical Silurian sandstones and shales. Source of images: A–C, Peter Cawood; D, Jon Davidson, University of Durham; E, Caroline Wickham-Jones; F, British Geological Survey digital data bank <http://geoscenic.bgs.ac.uk/asset-bank/action/viewHome>.



Figure 4. Schematic cross section of types of continental lithosphere emphasizing the thick stable nature of Precambrian cratons. Thickness of lithosphere beneath Archean regions is of the order of 200–250 km and oceanic lithosphere is up to 100 km. Abbreviation: icr—intracratonic rift; MOR—mid-ocean ridge.



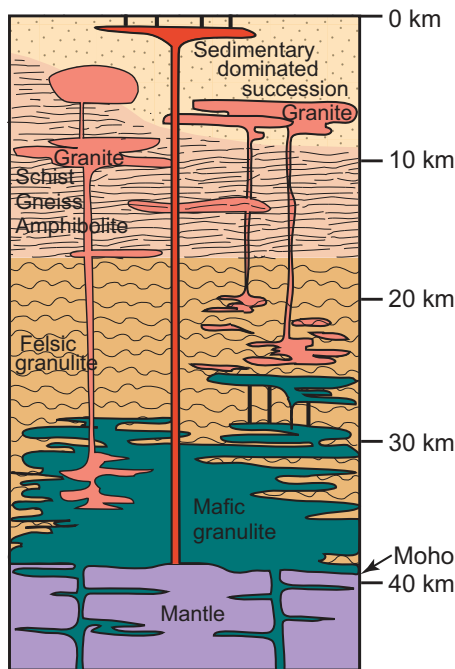


Figure 5. Schematic section of the continental crust, adapted from Hawkesworth and Kemp (2006a).

and Proterozoic regions display intermediate values. This variation appears to correlate with lithospheric thickness, which ranges from some 200–250 km beneath cratons to generally <100 km beneath Phanerozoic regions (Fig. 4; Pollack et al., 1993; Jaupart et al., 1998; Nyblade, 1999; Jaupart and Mareschal, 2003; McLennan et al., 2005). The absence of garnet in mantle xenoliths from Phanerozoic terranes is consistent with a lithospheric thickness of generally less than 60–80 km (Lee et al., 2011). McKenzie and Priestley (2008), using shear-wave velocity data, noted that areas of thick lithosphere do not underlie all cratons, and they also occur under platforms and regions of active deformation. They referred to these regions of continents as cores and suggested that the thick (~250 km) lithosphere beneath the plateaus in Tibet and Iran may represent regions where cratons are now being formed. The subcontinental lithospheric mantle of Archean cratons is composed of dehydrated, highly depleted mantle peridotite (Pollack, 1986; Boyd, 1989; Pearson et al., 1995; Boyd et al., 1997; Begg et al., 2009; Griffin et al., 2009), resulting in it being intrinsically buoyant and strong, and these attributes counteract the destabilizing effect of their cold thermal state (Lee et al., 2011).

Crust generation involves the formation of new crust through the emplacement of new magma from the mantle, and the overall area

of continental crust is inversely proportional to the areas of oceanic and transitional crust on a constant-radius Earth. Crust recycling is taken to be the intracrustal processes involving the return of crust to the mantle. It may occur by sediment subduction and sediment erosion at convergent plate margins, the loss of chemical solute resulting from continental erosion or hydrothermal alteration that is carried in the subducting oceanic crust, and delamination (detachment and sinking) of continental keels at collisional boundaries (Clift et al., 2009; Scholl and von Huene, 2009; Lee et al., 2011). Crustal reworking is used to mean intracrustal in origin, and it involves the remobilization of preexisting crust by partial melting and/or erosion and sedimentation, but all at sites within the continental crust (Hawkesworth et al., 2010). The growth of continental crust is the volume of new crust generated through time less the amount recycled to the mantle. In practice, the growth of continental crust is difficult to tie down, because radiogenic isotopes constrain only the volume of crust that has been stable for long enough for significant differences in isotope ratios to be developed from radioactive decay. However, even the generation of short-lived crust may leave a legacy in the complementary depletion of the upper mantle; depleted mantle is that mantle from which melt that becomes part of the continental crust has been extracted. The assembly of continental crust from different segments that were generated elsewhere and juxtaposed tectonically increases the volume of continental crust in the region being considered, but not the volume of continental crust overall, in the sense that the assembled fragments were already present elsewhere.

Supercontinents are assemblies of all or nearly all Earth's continental blocks, and they have occurred periodically through Earth history (Worsley et al., 1986; Nance et al., 1988; Rogers and Santosh, 2004). Superia and Sclavia are the terms proposed for end-Archean cratonic aggregations (Bleeker, 2003), which were first referred to as Kenorland (Williams et al., 1991), and Nuna, Rodinia, Gondwana, and Pangea are end-Paleoproterozoic, end-Mesoproterozoic, end-Neoproterozoic, and late Paleozoic supercontinents, respectively (Hoffman, 1996). The constituent number and disposition of cratonic blocks or continents within the various supercontinents are best constrained for the younger bodies (e.g., Pangea and Gondwana), and they become progressively more uncertain for older assemblages (e.g., Rodinia, Nuna, Superia; Hoffman, 1991; Williams et al., 1991; Dalziel, 1997; Zhao et al., 2002; Bleeker, 2003; Veevers, 2004; Collins and Pisarevsky, 2005; Li et al., 2008; Murphy et al., 2009; Reddy and Evans,

2009). Although the specific configuration of the early supercontinents is not fully resolved, the truncation of geologic trends at the edges of cratons and ancient orogens provides convincing evidence that they were components in larger continental assemblages (e.g., Hoffman, 1991; Moores, 1991), and this, combined with data on the temporal and spatial distribution of tectonothermal events, indicates that they periodically assembled into supercontinents.

HISTORICAL PERSPECTIVE

Humanity's development has been intimately tied to, and dependent on, Earth and its resources, and hence it is not surprising that most cultures developed mythologies on Earth that often emphasized the interconnected nature of things (e.g., the Greek goddess Gaia—the personification of Earth; Fig. 3E). Some of the earliest recorded observations relevant to understanding of Earth and the origin of the crust were by the Greeks and Romans, who noted the very slow rate of change of Earth's surface with respect to the human time scale, and that the presence of subaerially exposed marine rocks and fossils required the vertical movement of continents with respect to sea level. They also developed criteria that aided in the identification of rocks and minerals. The interplay of scientific ideas and religious doctrine in Europe in the seventeenth to nineteenth centuries influenced the development of geologic thought, focusing ideas on the age and origin of the crust. This led to the concept that rocks, including igneous rocks, formed from minerals that crystallized from water, and in particular from the biblical Great Flood (or Deluge), and reflected catastrophism in Earth history. This concept was termed Neptunism and viewed the entire planet, including the continents, as made up of sedimentary layers. It was popular because it was consistent with the literal interpretation of the Bible, which stated that Earth was only a few thousand years old. However, observations on the contact relations of igneous rocks, which established cross-cutting, intrusive relations, and the recognition of unconformities, which must encompass long periods of time, led to the alternative concept of Plutonism (also known as Vulcanism) in which the interior of Earth was hot, with igneous rocks crystallized from magma. This in turn led to a Uniformitarian view of Earth in which “the present is the key to the past” (Geikie, 1905, p. 299). The permanency of Earth and the immensity of geologic time required by the Uniformitarian view of Earth were encapsulated by James Hutton's (1788, p. 304) phrase “no vestige of a beginning,—no prospect of an end” (Fig. 3F). This powerful and emotive phrase not

only emphasizes the enormity of deep Earth time, but it also provides a counterpoint to the strict biblical interpretation of the age of Earth.

Uniformitarianism provided a framework in which to study Earth, and researchers focused on understanding the processes that shaped and stabilized continental crust preserved in ancient mountain belts (e.g., Hutton, 1788; Lyell, 1833; Hall, 1859; Dana, 1873; Suess, 1885–1909; Haug, 1900; see discussion and references in Dott, 1974; Şengör, 1982). These early observations on the crust and its origins were based on direct field observations and were focused on Phanerozoic sequences in eastern North America and Western Europe, notably the Appalachian and Alpine orogens, with their contrasting geology influencing the ideas and the theories that were proposed. Early workers on both continents acknowledged that the continental crust in these belts included a very thick accumulation of deformed sediment. North American workers considered these sequences to be shallow-water deposits that accumulated in asymmetric troughs at the margins between continents and ocean basins, but which also included input from an outboard source consisting of a long-established basement high or borderland (Hall, 1859; Dana, 1873; Schuchert, 1910). In contrast, European workers in the Alpine orogen regarded the sediments as being deep-marine deposits that accumulated on an ophiolitic substrate in a symmetrical basin between continents (Suess, 1885–1909; Haug, 1900; Steinmann, 1906). Time-integrated analysis of these orogenic belts led researchers to speculate on the stabilization of continental crust through a tectonic cycle involving geosynclinal, orogenic, and cratonic stages (Haug, 1900; Krynine, 1948; Aubouin, 1965). On a broader

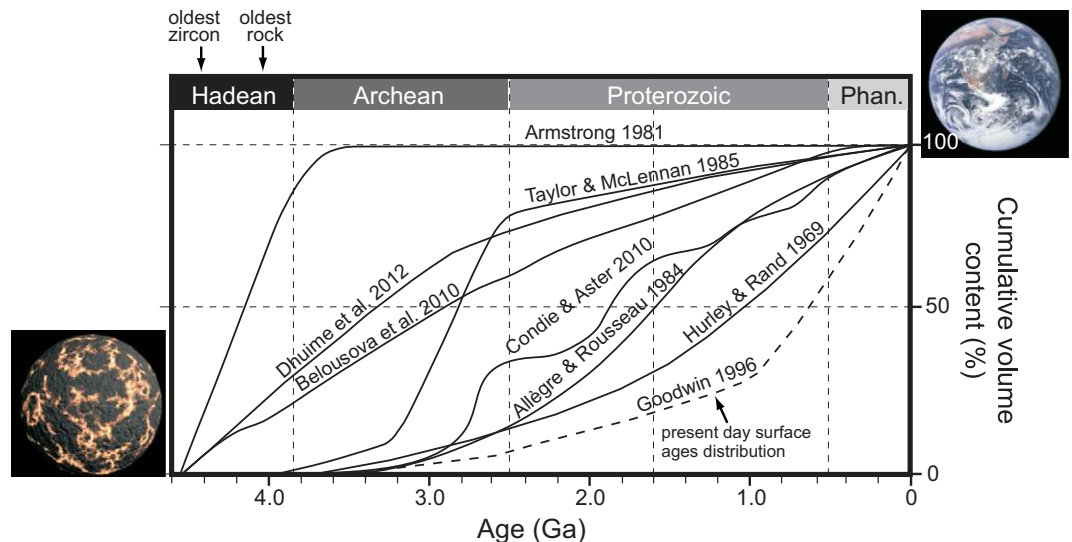
scale, ideas on a tectonic cycle led to the concept of continental accretion through a succession of concentric orogenic belts (Dana, 1873; Suess, 1885–1909; Haug, 1900). This relationship was most readily observed in North America, with its cratonic core and enveloping younger Appalachian and Cordilleran orogens. These ideas on the stabilization of continental crust through mountain-building processes developed in a framework involving the permanency of ocean basins and fixed continents—concepts that in the twentieth century were being increasingly questioned, initially by Western European and Southern Hemisphere workers (Wegener, 1924; Holmes, 1928–1930; du Toit, 1937; Umbgrove, 1947; Carey, 1958), who emphasized the transitory and dynamic nature of Earth's surface. This view of a dynamic Earth was ultimately embraced by the broader geological community and integrated into plate-tectonic theory (Hess, 1962; Vine and Matthews, 1963; Wilson, 1966; Dewey and Bird, 1970). It highlighted the processes through which the continental crust was both generated and destroyed.

Chemical and isotopic data on the composition and age of the continental crust, along with geophysical data on the internal structure of the crust and lithosphere, were then integrated with evolving ideas on tectonic processes to provide further insight into the origin and rate of growth of the crust. Early geochemical data enabled estimates of the average composition of specific rock types/tectonic units and ultimately led to estimations of the average composition of the entire crust (Clarke, 1924; Goldschmidt, 1954; Poldervaart, 1955; Taylor, 1964; Ronov and Yaroshevsky, 1969). This data set has been increasingly refined, as well as integrated with, and fed back into, tectonic models of the crust,

and our understanding of the inferred interrelationship between the crust and the complementary mantle reservoir from which it is derived (Hart, 1969; Taylor and McLennan, 1985; Rudnick, 1995; Rudnick and Fountain, 1995; McLennan and Taylor, 1996; Rudnick and Gao, 2003). These studies helped to establish (1) that the overall composition of the continental crust is similar to calc-alkaline andesite, and (2) the concept that the crust is typically derived in two stages, melting of the mantle to generate mafic magma, which undergoes fractional crystallization, with or without assimilation of preexisting crust, or crystallization, and then remelting to generate average crustal compositions.

Age and radiogenic isotopic data on rocks and minerals have led to a range of models on the rate of growth of the continental crust (Fig. 6). Most models indicate that the continental crust has increased in volume and area through time (Hurley et al., 1962; Hurley and Rand, 1969; Fyfe, 1978; Veizer and Jansen, 1979; Armstrong, 1981; Allègre and Rousseau, 1984; Taylor and McLennan, 1985; Veizer and Jansen, 1985; Armstrong, 1991; Taylor and McLennan, 1996; Belousova et al., 2010; Dhuime et al., 2012). These studies have tended to argue that crustal growth has been continuous, with early models proposing steady or increasing rates of growth through Earth history, and subsequent studies emphasizing an earlier period of more rapid crustal growth, typically in the late Archean or early Proterozoic, followed by decreasing rates of growth to the present day. The early models were often based on the geographic distribution of Rb-Sr and K-Ar isotope ages (Hurley et al., 1962; Hurley, 1968; Hurley and Rand, 1969; Goodwin, 1996), but these are biased by orogenic overprinting from events

Figure 6. Crustal growth models of Hurley and Rand (1969), Armstrong (1981), Allègre and Rousseau (1984), Taylor and McLennan (1985), Condie and Aster (2010), and Dhuime et al. (2012) compared to age distribution of presently preserved crust from Goodwin (1996). Sources of images: early Earth—<http://www.universetoday.com/58177/earth-formation/>; present-day Earth—National Aeronautics and Space Administration.



younger than the age of crust formation, and older crustal terrains tend to be less well preserved than those formed more recently. A key advance was in the development of isotope systems, in particular, Nd and Hf isotopes, which could be used to constrain model ages to indicate when new crust was generated, and which are, for the most part, unaffected by younger orogenic events (McCulloch and Wasserburg, 1978; Patchett et al., 1982). These isotope systems have relatively long half-lives, and so the variations in Nd and Hf isotopes reflect the generation of crust that is relatively long-lived. Material had to be in the crust for long enough in order for significant amounts of the radiogenic isotope to be generated, typically a few hundred million years. New crust that was generated and then destroyed shortly after would not have had time to develop a radiogenic signature, and so it is in effect invisible to these isotope systems. Armstrong (1981, 1991) and Fyfe (1978) proposed an early burst of continental growth that was followed by steady-state or even decreasing crustal volumes. Armstrong (e.g., 1981) argued that constant volume was maintained by crustal recycling, offsetting any additions from crustal generation. He was amongst the first to emphasize the role of recycling (contrast with Moorbath, 1975, 1977) in both understanding crustal volumes and in potentially affecting the composition of the upper mantle. However, because there is little evidence for significant volumes of Hadean and Early Archean crust from radiogenic isotope systems, it implies that much of the proposed recycled crust is too young to have developed distinctive radiogenic isotope signatures (cf. Hurley et al., 1962; Hurley and Rand, 1969; O'Nions et al., 1979; Veizer and Jansen, 1979).

A second aspect is that some models of crustal growth invoke continuous growth, and others indicate that the growth of the crust was in some way episodic. The isotopic models of steady growth of continental crust (plus or minus recycling) would appear to be at odds with map patterns involving discordant data sets of discrete age provinces that are abruptly truncated at their boundaries (Gastil, 1960a, 1960b; Dott, 1964). More recent models of continental growth tend to show pulses of enhanced growth at specific periods in Earth history (e.g., Late Archean; McCulloch and Bennett, 1994; Taylor and McLennan, 1996; Condie, 1998; Rino et al., 2004). Most of the continental growth models have been based on compilations of whole-rock isotopic data, in some cases of fine-grained sediments that may have sampled large source areas. However, Condie (1998) used a limited suite of zircon ages of juvenile continental crust, and Rino et al. (2004) developed a model based on

analysis of detrital zircons of igneous origin in the modern Amazon, Mackenzie, and Mississippi Rivers. Rapidly expanding databases of igneous and detrital zircon data highlight a non-continuous distribution of crystallization ages (see Fig. 7) and have led to further proposals that continental growth has been episodic, corresponding with phases of the supercontinent cycle and/or to mantle-plume activity (Condie, 1998, 2000, 2004; Campbell and Allen, 2008; Voice et al., 2011). However, Gurnis and Davies (1986) noted that young crust is more elevated than old crust and hence more easily eroded (cf. Allègre and Rousseau, 1984), and this will lead to preferential recycling of young crust and can lead to an apparent peak in a continuous crustal growth curve at 2–3 Ga.

THE NATURE OF THE CONTINENTAL RECORD

The rock record is incomplete, which raises questions about the extent to which the continental archive is biased by selective preservation of rocks generated in different tectonic settings, and therefore how best to evaluate temporal changes from the record. Igneous provinces are by definition restricted in space and time, and they therefore provide regional snapshots of the magmatic processes that have occurred in different tectonic settings. The same is broadly true for metamorphic provinces, and the constraints on the thermal histories of different areas provided by pressure-temperature-time (P - T - t) paths. In contrast, sediments contain material from their source rocks irrespective of the conditions under which those rocks were generated. The bulk compositions of detrital sediments have therefore been used to estimate the average composition of the upper crust (Taylor and McLennan, 1981, 1985; Condie, 1993; Taylor and McLennan, 1995; Rudnick and Gao, 2003), and to determine its average Nd isotope ratio and model Nd age (Taylor et al., 1983; Allègre and Rousseau, 1984; Michard et al., 1985). However, the sedimentary record is biased (1) in terms of the different lithologies, and, hence for Phanerozoic strata, the fossil communities that are preserved (e.g., Raup, 1972, 1976; Smith and McGowan, 2007), and (2) because younger source rocks are thought to be more accessible to erosion than older source rocks. Thus, older source rocks may be underrepresented in the bulk compositions of continental detrital sediments (Allègre and Rousseau, 1984). It is unclear whether comparable biases also occur in the igneous and metamorphic rocks of the continental crust.

Data compilations emphasize that the spatial and temporal distribution of rock units and

events is heterogeneous; for example, ages of igneous crystallization, metamorphism, continental margins, mineralization, and seawater and atmospheric proxies are distributed about a series of peaks and troughs. These appear, at least in part, to correspond with the cycle of supercontinent assembly and dispersal (Fig. 7).

Igneous and detrital zircon U-Pb ages and Hf isotopic data have identifiable peaks in ages of crystallization (Fig. 7; Condie, 1998, 2000, 2004, 2005; Rino et al., 2004; Groves et al., 2005; Hawkesworth and Kemp, 2006b; Kemp et al., 2006; Campbell and Allen, 2008; Belousova et al., 2010; Voice et al., 2011). Peaks in U-Pb crystallization ages correspond with periods of supercontinent assembly at around 2.7–2.4 Ga (Superia and Sclavia), 2.1–1.7 Ga (Nuna), 1.3–0.95 Ga (Rodinia), 0.7–0.5 Ga (Gondwana), and 0.35–0.18 Ga (Pangea). Peaks in Hf isotope ages were recognized in early and/or regional studies, but expanding data sets suggest a more continuous distribution (Belousova et al., 2010; Hawkesworth et al., 2010; Dhuime et al., 2012). The distribution of the ages of high-grade metamorphic rocks is also episodic (Fig. 7). Brown (2007) categorized high-grade orogenic belts into high-, intermediate- and low-pressure–high-temperature belts. He noted that the high-pressure belts were restricted to the last 600 m.y., and he concluded that they reflect cold subduction as observed at present along convergent margins. Intermediate- to low-pressure–high-temperature rocks are preserved dating back to the Late Archean, and Kemp et al. (2007b) pointed out that their ages are grouped in clusters similar to the peaks of crust generation illustrated in Figure 7. The implication is that periods of granulite-facies metamorphism are in some way linked to the processes of crust generation, as suggested by Kemp et al. (2007b), and/or the peaks of the ages of crust generation and granulite metamorphism are themselves a function of the unevenness of the continental record.

The age pattern of ancient passive margins also reveals major peaks in the Late Archean, late Paleoproterozoic, and late Neoproterozoic to early Paleozoic, which correspond to times of supercontinent aggregation (Fig. 7; Bradley, 2008). The proportion of modern passive margins is somewhat different, correlating with the breakup of Pangea and the resultant increase in margin area (Bradley, 2008). Smith and McGowan (2007) noted that the Phanerozoic diversity of marine fossils is affected by the supercontinent cycle, with marine rocks dominating during rifting phases of supercontinents. Eriksson and Simpson (1998) highlighted the temporal concentration of eolianites at 1.8 Ga and its association with breakup and assembly

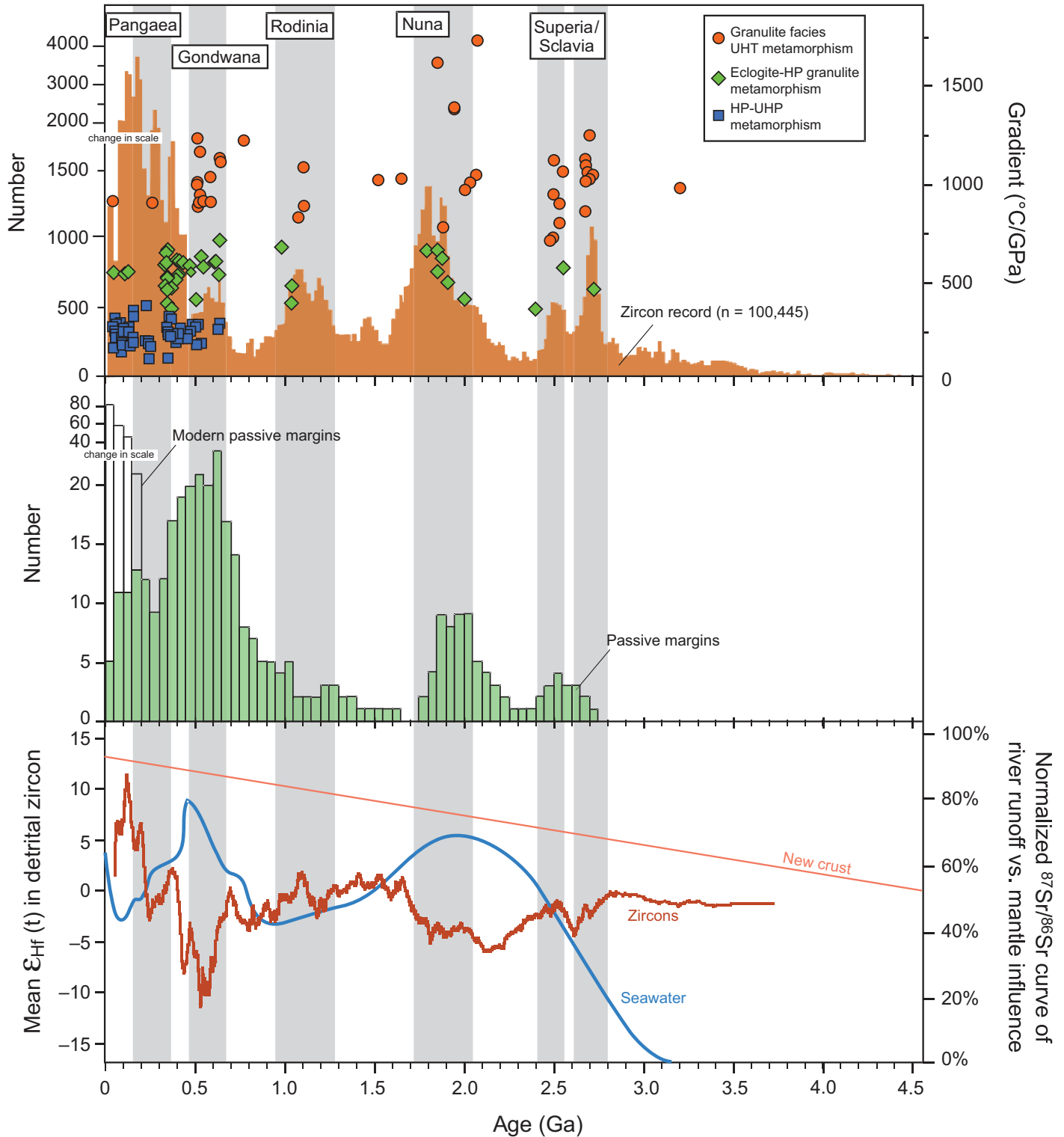


Figure 7. (A) Histogram of over 100,000 detrital zircon analyses showing several peaks in their U-Pb crystallization ages over the course of Earth history (Voice et al., 2011), which are very similar to the ages of supercontinents. Also shown is the apparent thermal gradient versus age of peak metamorphism for the three main types of granulite-facies metamorphic belts (Brown, 2007). UHT—ultrahigh temperature; HP—high pressure; UHP—ultrahigh pressure. (B) Histogram of the ages of ancient and modern passive margins (Bradley, 2008). (C) Normalized seawater $^{87}\text{Sr}/^{86}\text{Sr}$ curve (Shields, 2007) and running mean of initial ϵ_{Hf} in ~7000 detrital zircons from recent sediments (Dhuime, Hawkesworth, and Cawood, personal observation). Low $^{87}\text{Sr}/^{86}\text{Sr}$ values in Archean in part reflect the lack of data and the large proportion of submerged continental crust.

phases of supercontinents. Bradley (2011) recently compiled secular trends in a variety of rock units and events and noted that carbonatites (Woolley and Kjarsgaard, 2008) and greenstone-belt deformation events (Condie, 1994, 1995) also bear the imprint of Precambrian supercontinent cycles.

Mineral deposits are heterogeneously distributed in both space and time, with variations related to long-term tectonic trends associated with the supercontinent cycle and changing environmental conditions such as atmosphere-hydrosphere conditions and thermal history (Meyer, 1988; Barley and Groves, 1992; Groves et al., 2005; Groves and Bierlein, 2007; Bierlein et al., 2009). For example, deposit types associated with convergent plate margins (accretionary orogens), such as orogenic gold and volcanic massive sulfide (VMS) deposits, and to a lesser extent porphyry Cu-Au-Mo and Sn-W, and epithermal Cu-Au-Ag deposits, exhibit well-defined temporal patterns that broadly correlate with supercontinent assembly (Bierlein et al., 2009). However, deposits formed in intracratonic settings and related to mantle processes (e.g., platinum group elements (PGE) deposits) lack such a correlation (Cawood and Hawkesworth, 2012).

GENERATIONAL ARCHIVE OR PRESERVATIONAL BIAS

The evidence for peaks and troughs across the rock record, particularly related to igneous rock generation (Fig. 7), has been used to argue that continental crust formation has been episodic, and that in some way pulses in the formation of continental crust are linked to the development of supercontinents (e.g., Condie, 1998, 2000, 2004, 2005; Rino et al., 2004; Groves et al., 2005; Hawkesworth and Kemp, 2006b; Kemp et al., 2006; Campbell and Allen, 2008; Voice et al., 2011). Punctuated crustal growth remains difficult to explain by global changes in plate-tectonic regimes, which is viewed as a continuous process (but see discussions by O'Neill et al., 2007; Korenaga, 2008; Silver and Behn, 2008a, 2008b), and so it is typically linked to mantle-plume activity (Stein and Hofmann, 1993; Condie, 1998). However, the andesitic composition of continental crust, along with evidence that the plate-tectonic mechanism has been active for much of Earth history and is a major driver for continental assembly and dispersal (Cawood et al., 2006; Shirey and Richardson, 2011), suggests that magmatic arcs are the major source of continental growth (cf. Taylor, 1967; Taylor and McLennan, 1985; McCulloch and Bennett, 1994; Rudnick, 1995; Davidson and Arculus, 2006; Hawkesworth

and Kemp, 2006b). Recently, Stern and Scholl (2009) argued that peaks of ages reflect periods of increased magmatic activity associated with increases in the volumes of subduction-related magmas that are generated during continental breakup. The peaks of igneous crystallization ages correspond however, with the time of maximum supercontinent aggregation, not with their breakup. Also the preservation potential of intra-oceanic arcs, which are largely submarine, is poor (Condie and Kröner, 2012). From a comparison of U-Pb ages and Hf isotope model ages for ~5100 detrital zircons, Voice et al. (2011) proposed that the age-frequency distribution of detrital grains reflects predominantly episodic crustal recycling (i.e., destruction) rather than crustal growth. Recycling of crust is indeed an important issue in evaluating growth models of continental crust (cf. Armstrong, 1981), but its role can only be evaluated after the validity of the record, and the extent to which it has been influenced by preservation processes, has been established.

Analysis of the rock record at modern convergent plate margins has established that they are not only major sites for the generation of continental crust but also for its removal and recycling back into the mantle (Fig. 8). Global compilations of sites of continental addition and removal (Scholl and von Huene, 2007, 2009; Clift et al., 2009; Stern, 2011) highlight that crustal growth rates in continental collision zones, sites of continental aggregation and supercontinent assembly, are low and insufficient to generate the present volume of continental crust over the history of Earth. Although

these compilations reach different conclusions on the values for addition and removal of continental crust for individual tectonic settings, reflecting the different data sets and proxies used, they reach a similar overall conclusion: that, on a global scale, the processes of continental addition along destructive plate margins are counterbalanced by those of continental removal, resulting in no net growth in the current volume of continental crust.

Integration of data on the rates and sites of continental generation and recycling (Fig. 8) with the observed punctuated rock record (Fig. 7) suggests that peaks in age data may not represent episodic growth but instead reflect the greater preservation potential of rocks formed during the latter stages of ocean closure and collision, and that the record is therefore biased by the construction of supercontinents (Hawkesworth et al., 2009, 2010; Condie et al., 2011). Thus, the observed rock record of igneous crystallization ages is the integration of the volumes of magma generated during the three phases of the supercontinent cycle (subduction, collision, and breakup), and their likely preservation potential within each of these phases. This is illustrated schematically in Figure 9: Magma volumes are high in subduction settings but low during continental collision and breakup (Fig. 8). In contrast, the preservation potential of rocks in convergent and breakup settings is poor, whereas the preservation potential of late-stage subduction prior to collision, and collisional settings is high (Fig. 8). Peaks in crystallization ages that are preserved (shaded area under the curves in Fig. 9) would then

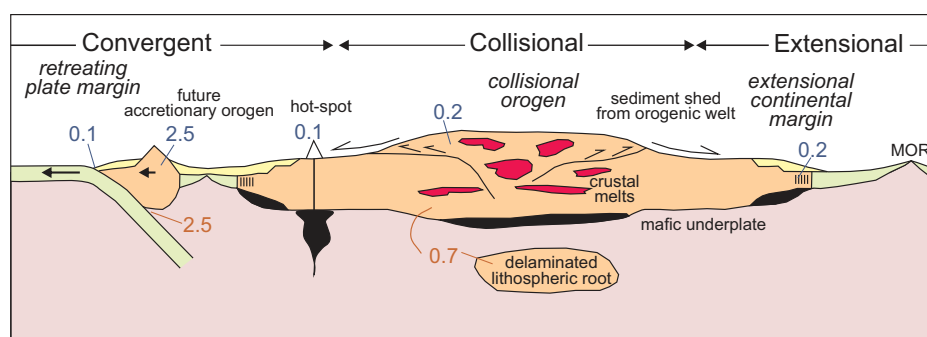


Figure 8. Schematic cross section of convergent, collisional, and extensional plate boundaries associated with supercontinent cycle showing estimated amounts (in $\text{km}^3 \text{yr}^{-1}$) of continental addition (numbers in blue above Earth surface) and removal (numbers in red below surface). Data are from Scholl and von Huene (2007, 2009). The volume of continental crust added through time via juvenile magma addition is approximately compensated by the return of continental and island-arc crust to the mantle, implying that there is no net growth of continental crust at the present day. Data compilations from Clift et al. (2009) and Stern (2011) calculated values of crustal recycling that are greater than the volume of juvenile crustal addition, requiring a decrease in present-day continental volumes. MOR—mid-ocean ridge.

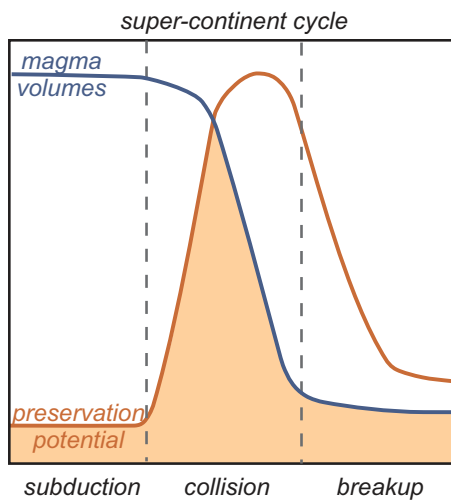


Figure 9. The volumes of magma generated (blue line), and their likely preservation potential (red line) based on relations outlined in Figure 8, vary through the three stages associated with the convergence, assembly, and breakup of a supercontinent. Peaks in igneous crystallization ages that are preserved (shaded area) reflect the balance between the magma volumes generated in the three stages and their preservation potential, and will result in an episodic distribution of ages in the rock record.

reflect the balance between the magma volumes generated in the three stages of supercontinent evolution and their preservation potential. As such, they would be unrelated to any underlying variation in the rate of crust generation. Note that the resultant peak in Figure 9 corresponds to the collisional phase of the supercontinent cycle, even though this is not a major phase of crustal generation (compare with data in Figs. 7 and 8), rather than periods of subduction- and extension-related magmatism (Hawkesworth et al., 2009). In detail, magmas generated in subduction-zone settings crystallize less zircon per volume of magma than those in collision settings (e.g., Moecher and Samson, 2006; Dickinson, 2008). Nonetheless, given the differences in the volumes of magma generated, the number of zircons crystallized in the former is many orders of magnitude higher than those generated during the collision stage (Fig. 10). Thus, the greater abundance of zircons for which ages correspond with the time of supercontinent assembly (Fig. 7) is only possible due to their higher preservation potential and cannot be simply related to the volumes generated (Fig. 10). Furthermore, the flux of sediment from the source region to depositional basin reflects the influence and feedback among relief, cli-

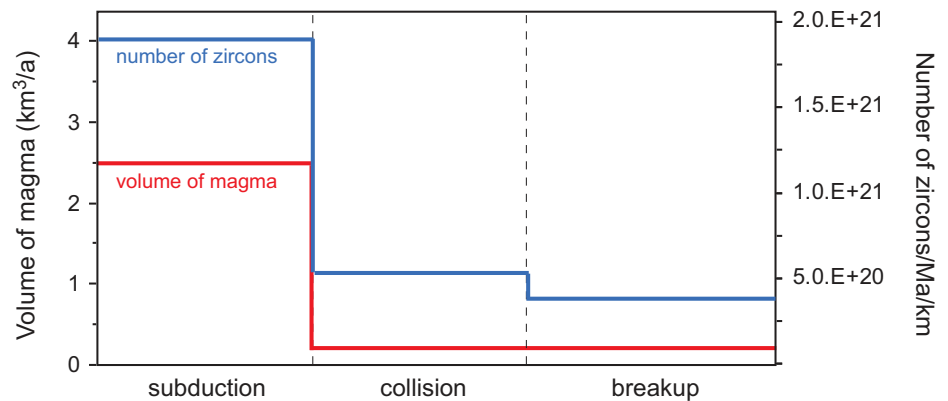


Figure 10. Volumes of magma generated (in $\text{km}^3 \text{yr}^{-1}$) during the three stages of the supercontinent cycle (from Scholl and von Huene, 2007, 2009; see also Fig. 8) compared to estimates of number of zircons likely to crystallize for each setting (per m.y., per km). The volume of zircon generated per million years was calculated from the average Zr content in magmas, using the relationship: $\text{vol}\% \text{ zircon} = 1.15 \text{ wt}\% \text{ Zr}$ (Dickinson, 2008); and $\text{Zr} = 150 \text{ ppm}$, 520 ppm , and 375 ppm for subduction, collision, and anorogenic magmas, respectively (Dickinson, 2008). Average zircon dimensions of $150 \times 60 \times 60 \mu\text{m}$ were used to convert zircon volumes into number of zircons. The number of zircons that crystallized per million years is normalized to the total length of convergent margins ($42,000 \text{ km}$; Scholl and von Huene, 2009; Stern, 2011), in order to obtain zircon generation rates in million years per km.

mate, and tectonic setting. High runoff in zones of crustal thickening and uplift will result in rapid exhumation, erosion, and high sediment flux (Koons, 1995; Clift, 2010). Basins adjacent to zones of continental collision or convergent Andean margins that receive high orographic/monsoonal rainfall feed high-flux river systems. The Yellow, Amazon, and Brahmaputra Rivers are the top three sediment-producing rivers in the world, whereas rivers draining low-relief, arid environments have low sediment flux (Summerfield and Hulton, 1994). Thus, the high sediment flux within collision zones is likely to further accentuate the preservation bias-induced, episodic zircon record.

Preservation bias also explains other secular trends related to the supercontinent cycle. The peaks in passive margin ages at around 2.5 Ga, 2.0 Ga, and 0.5 Ga are consistent with selective preservation. If passive-margin distribution were related to the time at which they were developed, they should follow a predictable pattern related to changes in area of continental margins through time, with a minimum number of margins corresponding to the peak in supercontinent aggregation when continental margin area is reduced relative to the area of the individual constituent continents. In detail, their distribution during a supercontinent cycle should be characterized by: (1) a decrease in global population of passive margins during supercontinent assembly; (2) few passive margins when the supercontinent is fully assembled; and (3) an

increase in number of passive margins during supercontinent breakup as surface area of continental margins increases (Bradley, 2008). This is not what is typically observed, and only the most recent supercontinent, Pangea, and its subsequent breakup record, represented by the distribution of modern margins, appear to follow this trend. The difference in passive-margin distribution associated with Pangea breakup relative to those of earlier supercontinents can be explained by the fact that the next supercontinent after Pangea has not yet formed (termed Amasia by Hoffman, 1992), and hence any preservation bias in the record will not be apparent until then. Unlike the relationship between peaks in passive-margin ages that correspond to the Superia/Sclavia, Nuna, and Gondwana supercontinents, there is no peak associated with Rodinia. A possible explanation is that closure of the ocean related to Rodinia assembly did not involve passive margins draining older source regions, but rather was bounded by convergent plate margins (e.g., like the current circum-Pacific “Ring of Fire”).

The time interval corresponding with the Rodinia supercontinent also lacks anomalies in the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio seawater record or in the average ϵ_{Hf} values from large zircon compilations (Fig. 7). The Sr isotope ratio is different from inferred proxies of continental growth, such as the U-Pb detrital zircon record, in that it is unlikely to have been influenced by preservation bias in the geological record. It is taken as a measure of

the relative amount of continental versus mantle input, and positive excursions, such as those corresponding with the Gondwana and Nuna supercontinents, are often considered indicative of uplift and erosion of continental basement during continental collision (e.g., Bartley et al., 2001; Prokoph et al., 2008). Similarly, the negative values of ϵ_{Hf} during these time intervals are taken to indicate greater crustal reworking associated with collision-related crustal thickening. For Rodinia time, the absence of preserved passive margins, of an enriched, low ϵ_{Hf} signal, and a spike in seawater $^{87}\text{Sr}/^{86}\text{Sr}$ ratios suggests that margins juxtaposed during supercontinent assembly consisted of largely juvenile material, perhaps analogous to those rimming the modern Pacific.

EROSION AND THE CONTINENTAL RECORD

The clastic sedimentary record, which samples a range of source rocks, and may provide the only record of a source that has been lost due to erosion, dismemberment, or overprinting, is widely thought to provide a more representative record of the evolution of the continental crust than the present outcrop of igneous and metamorphic rocks (Taylor and McLennan, 1995). The sedimentary record is however biased; plate margins are dominated by young rocks, which are therefore more prone to erosion than older rocks. Thus, the bulk compositions of sediments are biased toward the younger material in the source terrains (Allègre and Rousseau, 1984).

Erosion-induced bias is expressed through the erosion factor K , as defined by Allègre and Rousseau (1984). K contrasts the relative proportions of rocks of different ages in the catchment area with the proportion of those source rocks present in the sediments analyzed (Fig. 11). This has proven difficult to measure in natural systems, and values ranging between 2 and 3 have been commonly assumed in previous studies (Garrels and Mackenzie, 1971; Allègre and Rousseau, 1984; Goldstein and Jacobsen, 1988; Jacobsen, 1988; Kramers and Tolstikhin, 1997; Kramers, 2002). These values for K would suggest that ~25%–30% of the present volume of the continental crust had been generated by 3 Ga (Fig. 12). Dhuime et al. (2011b) developed an approach to measure K in a modern river system using Hf isotopes in detrital zircons and Nd isotopes in fine-grained sediments from the Frankland River in southwest Australia. They demonstrated that K was variable (4–17), and that it increased downstream with water volume and with topographic relief.

The running mean ϵ_{Hf} value for zircons ranging in age from Hadean to Cenozoic is ~0 (e.g.,

Figure 11. Schematic representation of the “preferential erosion” model used by Allègre and Rousseau (1984) to model the growth of the continental crust through time. The “erosion factor,” K , is defined by $K = (y/[1 - y])/(x/[1 - x])$.

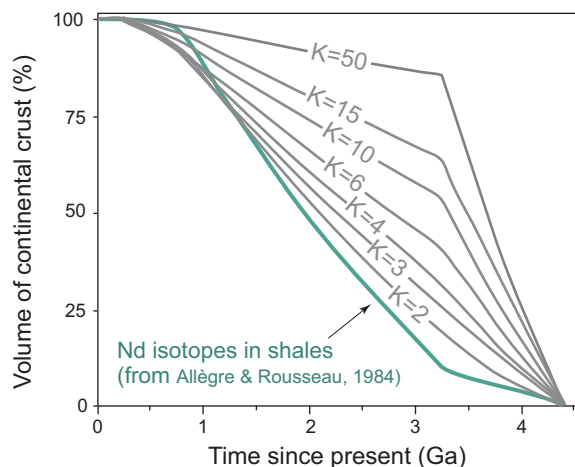
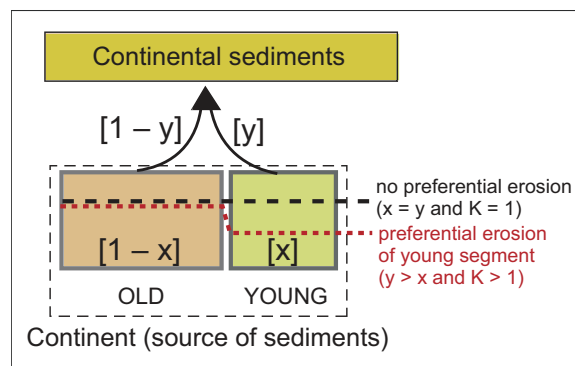


Figure 12. Continental growth curves for the Gondwana supercontinent, calculated from the Nd isotope data for Australian shales (Allègre and Rousseau, 1984). The variation of the erosion factor K has a dramatic influence on the shape of the growth curves. If $K = 1$ (i.e., no preferential erosion of the different lithologies producing the sediment), then 30% of the continental crust was generated by the end of the Archean, but this increases to 75% if $K = 15$.

Fig. 7; Belousova et al., 2010; Roberts, 2012), which indicates that there was a significant crustal component in the magmas from which those zircons crystallized. Crustal melting is in turn often associated with crustal thickening, and hence with areas of high relief. Denudation rates increase with increasing average relief (Summerfield and Hulton, 1994), and so the bias in the sedimentary record may be dominated by erosion in areas of high relief. If so, and this has yet to be tested rigorously, values of $K = 10$ –15 may be more appropriate for erosion and deposition of sediments that dominate the geological record. It follows that models for the evolution of the continental crust based on analysis of sedimentary rocks should be based on the values of K that characterize areas of crustal thickening and hence marked topographic relief. Assuming that they are applicable back into the Archean, such values of K would suggest that 60%–70% of the continental crust had been generated by 3 Ga (Fig. 12). Invoking a lower value of K for the Archean (e.g., $K \sim 2$), to reflect possible lower topographic relief for the major areas of continental erosion, but maintaining a high value for post-Archean units ($K \sim 15$), generates a similar growth curve

to that for a single-stage model with a uniform value of $K \sim 15$. This is because the age contrast between new crust and preexisting crust in the binary model of Allègre and Rousseau (1984) is small in the Archean.

UNRAVELING THE RECORD OF CONTINENTAL GROWTH

Zircons yield high-precision U-Pb crystallization ages. Zircons can also be analyzed for Hf and O isotopes, and for trace elements, and they contain silicate inclusions, which can be used to constrain the nature of the host magma from which they crystallized (e.g., Jennings et al., 2011).

Hafnium isotope studies of magmatic and detrital zircons have been utilized to explore the petrogenesis of granitic rocks (Belousova et al., 2006; Kemp et al., 2007a) and to unravel crustal evolution (Patchett et al., 1982; Amelin et al., 1999; Vervoort et al., 1999; Griffin et al., 2004). Hf is concentrated in zircon (~1 wt%), and so zircons have very low Lu/Hf, and their measured $^{176}\text{Hf}/^{177}\text{Hf}$ ratio approximates that of the host magma at its time of generation. The Hf isotope ratios are a measure of the crustal

residence age, i.e., the average time since the sources of the igneous rocks from which the zircons crystallized were extracted from the mantle (the Hf “model” age). The calculations of Hf model ages from zircon are more difficult to tie down than, for example, model Nd ages from Nd isotope ratios in sediments. This is primarily because the Lu/Hf ratio of zircon is much lower than that in the host magma, or more critically of that in the crustal precursor to the granitic melts from which the zircon crystallized. The Lu/Hf ratio of the crustal precursor therefore either has to be inferred from trends of initial Hf isotope ratios versus age, or assumed using average Lu/Hf ratios for mafic rocks or the bulk continental crust. This inevitably introduces uncertainty, and our approach is to plot distributions of calculated model ages so that the relative variations can be evaluated.

The second major assumption involved in the calculation of model ages is that of the Hf isotope ratios of new continental crust. The depleted mantle has been regarded as the complementary reservoir to the continental crust, and because the depleted mantle is thought to reside at relatively shallow levels in Earth’s mantle, it has also been regarded as the source of the basaltic magma involved in the generation of new continental crust (Jacobsen and Wasserburg, 1979; O’Nions et al., 1980; Allègre et al., 1983). Depleted mantle compositions (as measured in mid-ocean-ridge basalt [MORB]) have therefore typically been used in model age calculations, but new continental crust is mostly generated along destructive plate margins (e.g., Taylor, 1967; Rudnick, 1995; Davidson and Arculus, 2006; Scholl and von Huene, 2007, 2009). Such magmas are more enriched isotopically than MORB, since they tend to contain a contribution from recycled sediments (White and Patchett, 1984; Plank, 2005; Chauvel et al., 2008). The implication is that new crust, i.e., the magmas that cross the Moho, typically have lower Nd and Hf radiogenic isotope ratios than magmas generated from the depleted mantle (DePaolo, 1981; Vervoort and Blichert-Toft, 1999; Dhuime et al., 2011a). The weighted mean of ϵ_{Hf} in modern island-arc basaltic lavas is 13.2 ± 1.1 (Dhuime et al., 2011a), lower by $\sim 3\text{--}4$ ϵ units than the average ϵ_{Hf} of contemporary MORB (e.g., Salters and Stracke, 2004; Workman and Hart, 2005). If model ages are calculated using the new continental crust evolution line, anchored by the present day ϵ_{Hf} value of 13.2, they are up to 300 m.y. younger than those calculated using depleted mantle compositions (Dhuime et al., 2011a).

A key aspect in developing improved models for the generation and evolution of the continental crust is the ability to interrogate the informa-

tion now available in the hundreds of thousands of zircon analyses. One difficulty is that many crustally derived magmas contain a contribution from sedimentary source rocks, and such sediments typically include material from a number of different source rocks. Thus, the Hf isotope ratios and the model Hf ages of such magmas, and the zircons that crystallized from them, are likely to yield “mixed” source ages that do not provide direct evidence for when new crust was generated from the mantle. It is therefore necessary to find ways to strip off the zircons that may contain a contribution from sedimentary source rocks; this is most easily done using oxygen isotopes, and then crust generation models can be constructed on the basis of the zircon record that may more faithfully record when new crust was generated.

Hf isotope analyses of zircons are increasingly being combined with oxygen isotopes (i.e., $^{18}\text{O}/^{16}\text{O}$, expressed as $\delta^{18}\text{O}$ relative to the Vienna standard mean ocean water [VSMOW] standard). The latter are fractionated by surficial processes, and so the $\delta^{18}\text{O}$ value of mantle-derived magmas ($5.37\text{‰}\text{--}5.81\text{‰}$ in fresh MORB glass; Eiler et al., 2000) contrasts with those from rocks that have experienced a sedimentary cycle, which have generally higher $\delta^{18}\text{O}$ values. This is reflected in the high $\delta^{18}\text{O}$ value of the crystallizing zircons, and it is a “fingerprint” for a sedimentary component in granite genesis, and thus for the reworking of older crust. The in situ measurement of O isotope ratios from the same zircon zones analyzed for Lu-Hf is therefore used to distinguish those zircons from magmas that include a sedimentary component, which might in turn have hybrid model ages. In practice, there are large numbers of zircon analyses that are not accompanied by O isotope data. Thus, Dhuime et al. (2012) recently explored the extent to which the variations between Hf isotopes and $\delta^{18}\text{O}$ in zircons might be generalized in order to evaluate the changes in the proportions of reworked and new crustal material in zircons of different ages. The changing proportions of model ages thought to represent the generation of new crust (new crust formation ages) and of hybrid model ages (those with high $\delta^{18}\text{O}$) are represented by the black curve in Figure 13A. This was then used to recalculate the distribution of new crust formation ages (Fig. 13B, green curve) from the distribution of model ages for ~ 7000 detrital zircons for which O isotope data are not available (Fig. 13B, black histogram). The shape of the green curve suggests that new continental crust formation (crust generation) is a continuous process. A new model for the evolution of the continental crust was then established from the changes in the proportions of new and re-

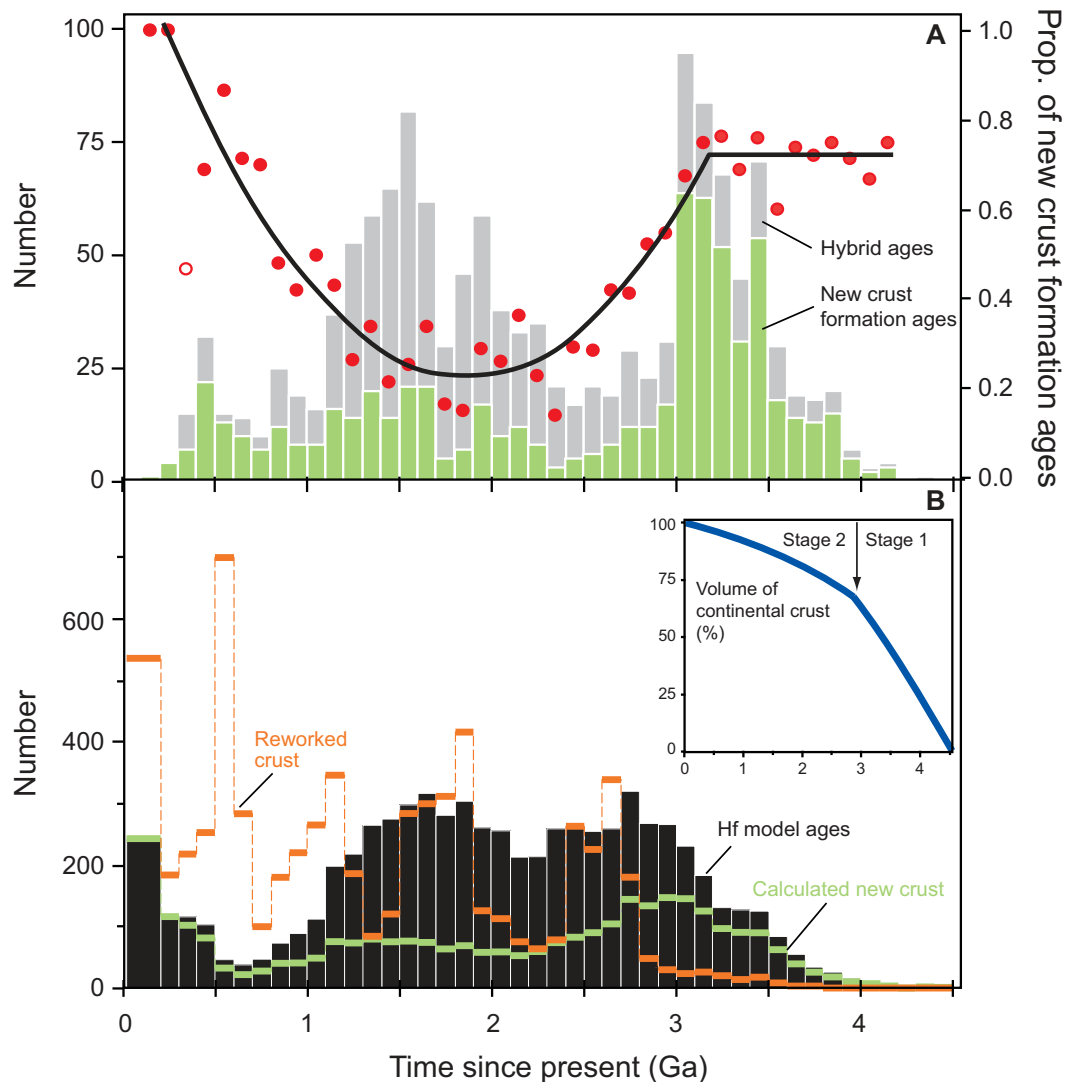
worked crust calculated from the Hf and O isotope data (Fig. 13B, green and orange histograms, respectively). This model suggests that $\sim 65\%$ of the present-day volume of the continental crust was already established by 3 Ga (Figs. 6 and 13B, inset), and it is striking that this figure of $\sim 65\%$ is similar to that independently estimated from Nd isotopes in sediments if K is ~ 15 (Fig. 12). The average growth rate of the continental crust during the first ~ 1.5 b.y. of Earth’s history is estimated to be ~ 3 km³ of crust added to the continental mass each year (Fig. 13B, inset, stage 1). Intriguingly, this is similar to the rates at which new crust is generated (and destroyed) at the present time (Scholl and von Huene, 2007, 2009). There was then a reduction in the net rates of growth of the continental crust at ca. 3 Ga, and subsequently the rate of crustal growth has been calculated at ~ 0.8 km³ of new crust added each year (Fig. 13B, inset, stage 2). This reduction in the average growth rate may primarily reflect an increase in the rates at which continental crust is destroyed (recycled), linked perhaps to the suggested onset of subduction at ca. 3 Ga (Cawood et al., 2006; Condie and Kröner, 2008; Shirey and Richardson, 2011; Dhuime et al., 2012).

TECTONIC SETTING AND CRUSTAL RECORD

The distribution of U-Pb crystallization ages in detrital sediments varies with respect to basin type, and, like the modal abundances of the clastic fill (cf. Dickinson and Suczek, 1979; Garzanti et al., 2007), it varies in response to tectonic setting (Cawood et al., 2012). The observed zircon record within a basinal succession represents the summation of two major variables: the presence or absence of synsedimentation magmatic activity and the overall spread of ages recorded. For those basins that contain igneous zircons with ages close to the time of sediment accumulation, this also reflects the setting of the magmatic activity, for example, forearc, trench, and backarc basins at convergent plate margins. Older grains reflect the prehistory of the basin’s distributive province and will likely show an episodic pattern of peaks and troughs reflecting preservational bias within the supercontinent cycle (Fig. 9; Hawkesworth et al., 2009; Condie et al., 2011). These variables can be represented graphically by plotting the distribution of the difference between the measured crystallization age for a detrital zircon grain and the depositional age of the succession in which it occurs (Fig. 14).

On this basis, detrital zircon data can be grouped into three main tectonic settings: convergent, collisional, and extensional. The

Figure 13. (A) Distribution of Hf model ages in 1376 detrital and inherited zircons sampled worldwide, from which O isotopes have been measured (from Dhuime et al., 2012, and references therein). O isotopes were used by Dhuime et al. (2012) to screen “hybrid model ages” (gray bins) and model ages that represent true periods of crust generation (green bins) in the global model age distribution. The proportion of new crust formation ages versus hybrid model ages is represented by the red dots. These dots define a systematic variation with time, represented by the black curve. (B) The systematic relationship defined by the black curve in A was used by Dhuime et al. (2012) to calculate the distribution of new crust formation ages (green histogram) from the distribution of model ages in ~7000 detrital zircons worldwide. From the variations in the proportions of the new crust (green histogram) and the reworked crust (orange histogram, which represents the distribution of the crystallization ages of zircons with Hf model ages greater than their crystallization ages), a continental growth curve has been calculated (blue curve, inset).



detrital zircon patterns plotted in this way show a spectrum of results with overlap between the three broad basin types, particularly for increasing age of detritus with respect to depositional age (Fig. 14). There is, however, a significant change in the proportion of ages associated with the youngest and older magmatic events between settings. Convergent margins settings are dominated by detrital zircon ages close to the depositional age of the sediment, and some arc-trench basins display unimodal age spectra. Zircons from collisional basin settings contain a lower proportion of grains with crystallization ages approaching the depositional ages, but they still contain a significant proportion of grains with ages within 150 m.y. of the host sediment. This pattern reflects the input of material from the magmatic arc that existed during ocean closure prior to collision and the variable amounts of syncollision magmatism, along with a spectrum of older ages

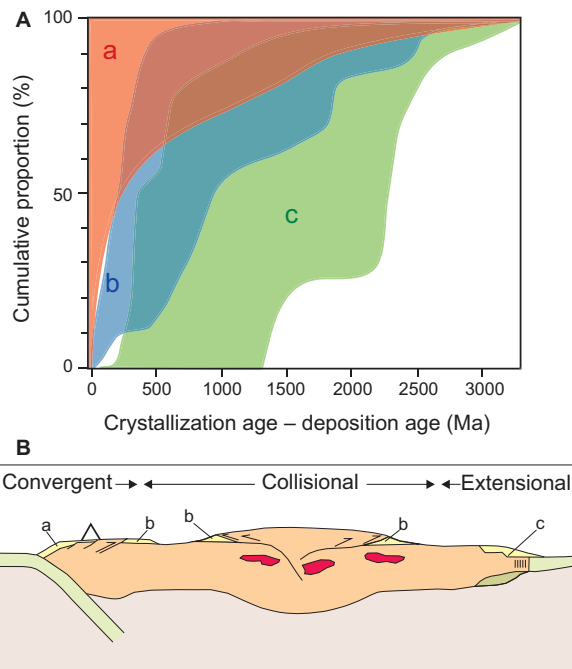
from cratonic blocks caught with the collision zone. Detrital zircon age patterns from extensional basins are dominated by grains that are significantly older than the depositional age of the basin. Syndepositional magmatism in extensional settings, such as volcanic rifted margins, is largely of mafic composition with a low yield of zircon (e.g., Fig. 10). The potential of this approach is that it can now be applied to sedimentary sequences in old terrains to evaluate tectonic settings in areas and periods where they are not well constrained. This may be of particular use in studies of the Archean, where tectonic events may have masked original settings. Initial studies indicate that the early Archean sedimentary sequences at Isua Greenland accumulated at a convergent plate margin, whereas the ancient zircons at Jack Hills Australia that occur in Late Archean and younger strata developed in extensional environments (Cawood et al., 2012).

TECTONIC CONTROLS ON THE GENERATION OF NEW CONTINENTAL CRUST

Present-day continental crust is predominantly generated through plate tectonics at convergent plate margins (Taylor, 1967; Rudnick, 1995; Davidson and Arculus, 2006), stabilized through orogenesis (Cawood and Buchan, 2007; Cawood et al., 2009, 2011), and preferentially preserved in the long-term geological archive through the supercontinent cycle (Hawkesworth et al., 2009, 2010). It is this combination of crust generation and subsequent processes that determines the preservation of the crust and is responsible for the episodic rock record of the continental archive.

Analysis of Hf isotopic data from detrital zircon data sets (Belousova et al., 2010; Dhuime et al., 2012) suggests that new continental crust has been generated continuously through time,

Figure 14. (A) Summary plot of variation of the difference between the measured crystallization age for a detrital zircon grain and the depositional age of the succession in which it occurs based on cumulative proportion curves, and displayed as a function of three main tectonic settings: convergent setting (a), collisional setting (b), and extensional setting (c). (B) Schematic cross section of convergent (a), collisional (b), and extensional (c) plate boundaries associated with supercontinent cycle showing simplified basinal settings for accumulation of detrital zircons.



with the bulk of the continental crust generated prior to the Proterozoic, and a progressive decrease in the rate of continual growth since the Archean. The volumes of continental crust estimated by this approach are minimum values, as some crust is rapidly recycled such that it does not leave any isotopic record (cf. Armstrong, 1981), whereas other crust generates few zircons (e.g., mafic crust). Upper limits on the volume of crust on early Earth are assumed to be at or near the current volume (Fyfe, 1978; Armstrong, 1981; see also Fig. 6).

Differences between the predicted minimum volumes of continental crust and actual preserved volumes through time are significant (Fig. 6). Dhuime et al. (2012) argued that some ~65% of the current-day crustal volume was present by the end Archean, yet compilations of present age distributions equate to less than 5% of the current volume at that time (Goodwin, 1996). The striking implication is that much of the crust that was generated in the Archean, and since then, has been destroyed and recycled back into the mantle.

The overall calc-alkaline andesitic composition of continental crust suggests that most of the crust was generated by processes similar to modern-day convergent plate margins (e.g., Rudnick, 1995; Davidson and Arculus, 2006). Our calculations on rates of generation and growth of continental crust of up to ~3 km³ per annum (Dhuime et al., 2012) are comparable to rates of magmatic addition at convergent plate margins (e.g., Scholl and von Huene, 2007, 2009; Clift et al., 2009). The plate-tectonic mechanism is a

response to the thermal state of the planet. Establishing how long plate tectonics have been the modus operandi of continental growth remains difficult to tie down, with suggestions ranging from not long after initial formation of the lithosphere in the Hadean (e.g., Sleep, 2007; Harrison et al., 2008) to the Neoproterozoic (e.g., Stern, 2005). Increasing evidence from a variety of sources, including geological, paleomagnetic, geochemical, geophysical, and mineralization patterns, suggests that plate tectonics have been active since at least the Late Archean (Cawood et al., 2006; Condie and Kröner, 2008; Rey and Coltice, 2008; Sizova et al., 2010). Extrapolation of the role of plate tectonics further back into the Archean or into the Hadean is hindered by the paucity of the rock record, apart from a few regional remnants (e.g., Isua greenstone belt, Acasta Gneiss, Nuvvuagittuq greenstone belt), along with mineral fragments (detrital zircons and their inclusions) of appropriate age (e.g., Jack Hills of the Narryer Gneiss terrane).

Models of the tectonic setting(s) in which early continental crust was generated must consider its complementary subcontinental lithospheric mantle, which also shows an episodic and for the most part coeval age distribution (Pearson, 1999; Pearson et al., 2007), albeit based on a small data set. This implies that the lithosphere as a whole, not just the crust, is affected by postgenerational tectonic processes. The buoyancy and strength of Late Archean cratonic lithosphere (Lee et al., 2011), assuming that the currently preserved crust is representative of crust generated at that time, suggest

that the paucity of an earlier record may reflect secular changes in generation and destruction processes, probably linked to thermal evolution of the mantle.

Figure 15 presents a temporal evolution of early Earth lithosphere (S. Foley, 2012, personal commun.). The section is deliberately generic, avoiding explicit representation of plate-tectonic processes, although these are implicit in the proposed development of accretionary and then collisional orogens after 3.0 Ga. The diagram highlights one of the critical features of the preserved early crust, which differentiates it from more modern crust, the change from an initial bimodal association of TTG (tonalites, trondhjemites, and granodiorites) and greenstone into a regime producing continental crust of andesitic composition through orogenic cycles. This results in Archean crust having marked bimodal distributions in silica (Martin, 1993; Rollinson, 2010; Van Kranendonk, 2010). Such bimodal distributions are a feature of intraplate continental flood basalt provinces (e.g., Mahoney and Coffin, 1997), and this has led to suggestions that much of the Archean crust was generated in intraplate settings, which in many instances have been linked to mantle plumes (e.g., Smithies et al., 2005; Bédard, 2006; Bédard et al., 2012). Archean crust and younger crust both have marked negative Nb anomalies, and these are typically taken as a strong indication of magmas generated in subduction-zone settings (Pearce, 1982; Wilson, 1989; Pearce and Peate, 1995). However, such trace-element discriminants have been established for mafic rocks, and they work best in rocks that can be related reasonably directly to their mafic precursors—as in modern destructive plate-margin settings (Wilson, 1989; Macdonald et al., 2000). In association with a marked silica gap, negative Nb anomalies could have been developed during melting of the mafic source rocks, or they could be a feature of those source rocks, and so they may not be a reliable indicator of the setting in which the high-silica rocks were generated. The negative Nb anomalies of the TTG, and hence the average Archean crust, have been attributed to small amounts of residual rutile (Rollinson, 2010), and the presence of amphibole (Foley et al., 2002), during partial melting of hydrated source rocks, and as such they do not require a subduction-related setting. A greater challenge, however, is the presence of water in the source rocks for the TTG (Rapp, 1997), and their marked positive Pb anomalies (Rollinson, 2010), and how these might be introduced in an intraplate setting. These equivocal geochemical signatures have resulted in subduction-related and plume-related interpretations, often for similar regions/periods (cf. Bédard, 2006; Wyman, 2012).

The Jack Hills zircons have received a great deal of attention, and yet it is unclear how representative they are of magmatic process in the Hadean and Early Archean (Compston and Pidgeon, 1986; Kober et al., 1989; Mojzsis et al., 2001; Wilde et al., 2001). Kemp et al. (2010) highlighted how different the Hf isotope-crystallization age trends were compared with the magmatic records of destructive plate-margin associations. Instead, they invoked an enriched mafic crust that might have formed by terrestrial magma ocean solidification, and interaction with the nascent hydrosphere at low temperatures. Foundering of such a hydrated basaltic shell to deeper crustal levels would in turn result in partial melting and the generation of the TTG (Kamber et al., 2005) and crystallization of the zircons now found in the Jack Hills sediments. Hadean and Early Archean detrital zircons from Wyoming show similar features that are related to anhydrous melting of primitive mantle in a plume-like setting (Mueller and Wooden, 2012).

Models of the evolution of continental crust, particularly early ones, focused on processes of crustal generation. Armstrong (1991, and references therein) referred to this as the myth of crustal growth, and he was one of the first to emphasize that crustal growth depends on the balance between the rates at which new crust is generated and the rates at which it is destroyed by weathering and erosion, and ultimately returned to the mantle. Such processes are clearly observed at the present day as plate tectonics involve the generation and recycling at convergent plate margins through arc magmatism and sediment recycling through sediment subduction and subduction erosion (von Huene and Scholl, 1991; Scholl and von Huene, 2007, 2009; Stern, 2011). Residual mafic material is recycled back into the mantle in most models for the generation of the relatively evolved bulk crustal compositions, and such processes have presumably been active for as long as average continental crust has been generated, i.e., ca. 3.5 Ga. However, it is much more difficult to establish how long significant volumes of crustal material have been destroyed by erosion and subduction, which in turn require an active plate-tectonic process.

The period from 3 Ga to the end of the Archean is increasingly viewed as a time of marked change in the dominant global tectonic regime(s) (Fig. 15). Shirey and Richardson (2011) noted that eclogitic mineral inclusions in diamonds, which come from subcontinental mantle lithospheric keels, are only present after 3.0 Ga, and they proposed that this reflects the onset of subduction and continental collision, in ways comparable to that at the present day. Some models of crustal growth also indicate a

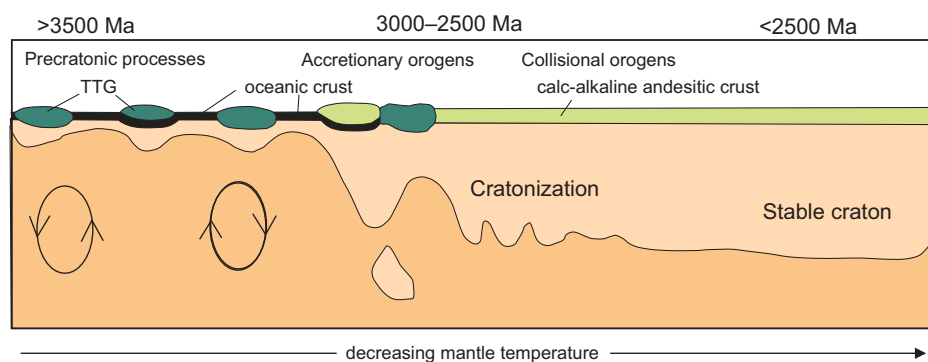


Figure 15. Temporal evolution of the lithosphere associated with the thermal evolution of the mantle from early Earth (adapted from S. Foley, 2012, personal commun.) involving a bimodal association of TTG (tonalites, trondhjemites, and granodiorites) and greenstone evolving through development of accretionary and collisional orogenic processes into a regime producing continental crust of andesitic composition by at least 3.0 Ga.

change in the average rate of continental crustal growth ca. 3 Ga (Fig. 6; Taylor and McLennan, 1985; Dhuime et al., 2012), as do models that involve erosion factors (K) of ~ 15 in the interpretation of the Nd isotope ratios in shales (Fig. 12). Figure 15 illustrates the possible changes in the nature of continental crust and lithosphere that may have taken place toward the end of the Archean in response to an evolving thermal regime (S. Foley, 2012, personal commun.). An early lithosphere consisting of a bimodal association of TTG and greenstone evolved through development of accretionary and collisional orogenic processes into a regime producing andesitic crust by at least 3.0 Ga. A compilation of $\delta^{18}\text{O}$ versus age for zircons shows uniform values for Archean and older grains but an overall increasing range of values for younger grains, which Valley et al. (2005) related to crustal reworking. This is consistent with data, largely from sedimentation patterns, for increasing continental emergence in the Late Archean and early Paleoproterozoic (Reddy and Evans, 2009, and references therein).

CONCLUSION

The continental crust is the archive of Earth history, and the present record shows an episodic distribution of rock units and events. There is increasing evidence that this distribution is not a primary feature reflecting processes of generation, as has been assumed by many, but is a consequence of secondary processes in which plate tectonics resulted in a biased preservational record. The importance of understanding the role of tectonics in unraveling the time-integrated history of Earth's continental growth is that it provides a mechanism for the ongoing continuous generation of average continental crustal

compositions of calc-alkaline andesite, and the destruction and recycling of crust on both short- and long-term time scales through convergent plate interaction and delamination of gravitationally unstable crust (cf. Armstrong, 1981).

The history of Earth is primarily read through the continental record, particularly prior to that preserved in the oceans. Advances in analytical techniques are allowing unprecedented interrogation of the record, but many uncertainties and exciting challenges remain in our understanding of the generation of the continental archive, particularly for early Earth, including:

(1) Differentiating primary and secondary signals in the rock record. It remains a high priority to distinguish those signals that are sensitive to the biases introduced by tectonic processes from those that are not; whether the episodic age distribution is a generational or preservational feature (Fig. 7) is a first-order example of this issue. Components of the record that preserve a temporally related frequency distribution, such as igneous crystallization ages, ages of metamorphism, or of passive margins, are here regarded as secondary signals modified by the proportions of rocks and minerals of a specific age that are preserved. Such preservation-related biases are most pronounced for those components of the record related to top-down, plate-margin-driven processes such as those controlled by the supercontinent cycle (e.g., Fig. 9). Frequency data driven by bottom-up, deep Earth processes may be more independent of such secondary controls and tend to display a primary signal. For example, PGE mineral deposits related to layered intrusion occur within stable continental crust generally removed from plate margins, and hence their distribution is less likely to have been modified by the supercontinent cycle. Similarly, data sets defined by

the first occurrence of a phase or form (e.g., evolution of life), or temporal changes independent of the number of data points, such as those related to seawater and atmospheric proxies (e.g., Sr in seawater; Fig. 7), are also more likely to preserve primary signals.

(2) The mafic record. Current models of continental volumes (Belousova et al., 2010; Dhuime et al., 2012) are based largely on proxies related to the generation of crust with an overall felsic composition (e.g., zircons) and, hence, are minimum volumes. These models do not take into account volumes of crust unrelated to such proxy calculations, even though part of the resultant volume is dependent on production of an intermediate mafic phase. Calculating the volumes of mafic and ultramafic crust generated within the continental record may be possible by tracking the proportion of U-bearing mineral phases, such as baddeleyite and zirconolite, that occur within such lithologies or their derived sediments (Bodet and Schärer, 2000; Rasmussen and Fletcher, 2004; Heaman, 2009; Voice et al., 2011). As with zircon, these minerals can be analyzed for both U-Pb crystallization ages and Lu-Hf isotopes to constrain their time of extraction from the mantle. However, their applicability to the problem of continental crust generation is yet to be evaluated and may be limited by their general paucity both in igneous rocks and also in derived sediments due to their lower physio-chemical resilience than zircon through the rock cycle.

(3) Processes involved in generation of early continental crust. Tectonic models for the generation of continental crust on early Earth remain controversial, and data from geochemical analyses and geodynamic modeling remain equivocal. This reflects the significant differences between the character of the early and current crusts of Earth, which are largely tied to the Earth's evolving thermal state, and the resultant lack of modern analogues for comparing process and constraining assumptions. Basic questions to be resolved include when and how did the continental crustal composition evolve from a bimodal to a more continuous distribution with respect to silica and other elements? The distribution pattern of U-Pb crystallization ages in detrital sediments, which can differentiate convergent, collisional, and extensional basin settings (Fig. 14), may provide a new way to constrain tectonic environment and process of generation of Archean greenstone sequences. Its applicability will be dependent on sequences in which the depositional age of the basin is reasonably well constrained, such as the Isua belt, which is rich in volcanic detritus with zircon U-Pb ages close to the depositional age of the basin and consistent with a convergent plate-

margin setting (Nutman et al., 2009; Cawood et al., 2012).

(4) Processes of crustal recycling. The striking contrast in crustal volumes between the preserved exposed record and those predicted by the latest models on crustal growth (Fig. 6) requires that crustal recycling has been active throughout Earth history (Armstrong, 1981). Recycling on current Earth takes place via sediment subduction and subduction erosion at sites of ongoing convergent plate interaction, through slab breakoff (which may contain a component of continental lithosphere) during the early stages of continental collision, and delamination during collision and possibly during crustal differentiation. Were these processes also the method of crustal recycling in Archean and older rocks and what controlled the rates at which crustal recycling took place?

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