9.07 Mechanisms of Continental Crust Growth

M. Stein, Geological Survey of Israel, Jerusalem, Israel

Z. Ben-Avraham, Tel Aviv University, Ramat Aviv, Israel

9.07.1 Introduction

The continental crust comprises the outermost 20–80 km (average thickness is about 36 km) of the solid surface of the Earth covering ~41% of the Earth surface area. Most of this area (~71%) is currently elevated above sea level while the rest is defined by the topography of the continental shelves. The continental crust density ranges from 2.7 to 2.9 g cm^{-3} and increases with depth. The vertical extent of the continental crust is defined by the compressional seismic wave velocity that jumps from ~7 to >7.6–8 km s^{-1} across the Mohorovicic (or 'Moho') discontinuity. The Moho discontinuity discovered by the Croatian seismologist Andrija Mohorovicic is the boundary between the crust and the mantle, which reflects the different densities of the crust and the mantle, normally occurring at an 'average' depth of ~35 km beneath the continental surfaces and about 8 km beneath the ocean basins. The original recognition and definition of this boundary reflected seismic properties. However, in some regions the

Moho is not always well defined and appears to be transitional in its physical–chemical characteristics (cf. Griffin and O'Reilly, 1987).

The oceanic crust covering ~59% of the Earth's surface is significantly thinner than the continental crust and is characterized by its distinct chemical composition and younger age (the oldest oceanic crust is Jurassic in age). Nevertheless, it appears that growth of continental crust involves accretion of thick magmatic sequences that were formed in the intraoceanic environment (e.g., oceanic plateaus), thus the geodynamic and geochemical evolution of oceanic and continental crusts are related.

The temperature gradient is typically less beneath continents than beneath oceanic basins reflecting the thicker immobile regions underlying the continents. These regions are termed the lithospheric mantle, and they typically form a thick colder 'root' beneath ancient shields and a relatively thinner root beneath younger shields. Yet, in intracontinent rifting regions the asthenospheric mantle is rising and interacts with the lithospheric mantle and continental crust and replaces it with new mantle-derived material. Moreover, it appears that many of the current ocean basins (and oceanic crust) commenced their history within the continental environment calling for some important relationship between stability and magmatic-thermal history of the continental lithosphere and the plate-tectonic cycle. The relationship between asthenospheric and lithospheric mantle, as well as the questions concerning the growth, maturation, and fate of continental lithosphere and crust throughout geological time are fundamental problems in Earth sciences that will be reviewed in this chapter.

The continental crust is of great antiquity and contains the record of most of the geological (physical and chemical) evolution of the Earth. A major question in Earth sciences is when, in what modes, and at what rates did continental crust form and evolve. It is broadly accepted that most of the present-day continental masses are composed of rocks that were produced at the end of the Archean and the beginning of the Proterozoic, within the time period lasting from ~3.2 to ~2.0 Ga. The oldest surviving rocks were found in northwestern Canada and date from about four Ga, 400–500 My after the formation of the Earth (Taylor and McLennan, 1995). Yet, although the continental crust covers about 40% of the terrestrial surface, rocks older than 2.5 Ga account for less than ~14% of the exposed continental area (Windley, 1995). Based on Nd model ages the 'average' age of the continental crust was calculated as ~2.2 ± 0.1 Ga, which indicates that 50–60% of the continental crust was formed before 2.6 Ga (cf. Nelson and DePaolo, 1985; DePaolo *et al.*, 1991; McCulloch and Bennett, 1993; Taylor and McLennan, 1995). These chronological determinations are critical to the long-standing debate on the history and mechanism of crust formation, where models of very early creation and subsequent recycling of continental crust material contrast with continuous/episodic crustal growth (the 'Armstrong–Moorbath debate' elaborated below).

Much of the present-day continental formation appears to occur at convergent plate margins along the subduction zones; however, several studies indicate that large amounts of continent were created in the geological past during short 'superevents' at rates that are difficult to explain by 'typical' subduction activity (Section 9.07.3.3). This would imply that large portions of the continents were formed by other means other than magmatism at convergent margins, probably processes related to hot spot or mantle plume magmatism (e.g., oceanic plateau formation, accretion of juvenile mantle terranes to the continents, and basaltic underplating). How these accreted terranes were transformed into continental crust and retain geophysical properties of crustal material represents another central question that has been open to debate (the 'Ontong Java paradox' outlined below). Crust formation is related to heat generation in the Earth, which has clearly declined since early Archean time through the Proterozoic and Phanerozoic. What was the reflection of this change on the modes and mechanism of crust production, for example, intracrustal or lithospheric melting versus subduction related magmatism and arc accretion?

In the chapter we summarize some of the main results that emerge from the extensive efforts that were devoted over the past decades to determine the composition of the upper and lower and bulk continental crust (this subject and the relevant literature were thoroughly reviewed by Taylor and McLennan, 1995 and recently by Rudnick and Gao, 2003). Then, we go through the Armstrong–Moorbath debate on early versus episodic crustal growth and the Reymer and Schubert dilemma concerning arc-plume modes of crustal growth. We describe mantle overturn and accretion models (cf. Stein and Hofmann, 1994; Ben-Avraham *et al.*, 1981) that invoke episodic activity of large mantle plumes and accretion of juvenile mantle material to the continents, and alternative arc-tectonics models (Patchett and Chase, 2002).

Finally, we describe the production of juvenile continental crust in major segments of the continents (termed major orogenies): the Archean Baltic Shield and Superior province, the Proterozoic Birimian orogen, and the late Proterozoic Arabian–Nubian Shield (ANS).

9.07.2 Structure and Chemical Composition of the Continental Crust

9.07.2.1 General

Based on seismic wave velocities, the continental crust can be vertically divided into two parts: the upper crust between the surface and the Conrad discontinuity, which is poorly defined at 10–20 km from the surface and the lower crust, which principally extends from the Conrad to the Moho discontinuity.

The problem of defining the location of the intracontinental crustal boundaries and obtaining an estimate of the chemical composition of the bulk continental crust reflects the complex history of the crust in different regions of the Earth. This evaluation requires knowledge of the types of the rock comprising the upper and lower crust in various crustal domains. However, only a small portion of the continental crust is exposed (e.g., due to tectonic exhumation of mainly upper-crust segments and deep erosion) and accessible for direct research. The composition and physical properties of the lower crust are mainly evaluated from geophysical data combined with information derived from xenoliths and outcrops of uplifted granulite terranes (cf. Rudnick and Fountain, 1995).

Nevertheless, many researchers devoted extensive efforts over the years to estimate the bulk chemical composition of the crust. This was done by determining the distribution and chemical composition of the dominant rock types that are exposed at the surface (e.g., the pioneering studies of Clarke and co-workers; (Clarke, 1889; Clarke and Washington, 1924) and recent works by Gao *et al.* (1998) (see the comprehensive summaries of this topic by Condie (1993) and Rudnick and Gao (2003)). Another approach was to use sediments as natural samplers of large areas of the upper continental crust. Goldschmidt (1933, 1958) introduced this approach in his pioneering studies on glacial sediments from the Baltic Shield.

Later, other researchers applied similar ideas and methodologies using other samplers including fine-grained sediments such as desert dust or deep-sea terrigenous sediments. Important examples are the works of Taylor and McLennan (1985); Plank and Langmuir (1998); Gallet *et al.* (1998); Barth *et al.* (2000); and Hattori *et al.* (2003). The results of these and other related studies were used to estimate the chemical composition of the upper crust (e.g., Taylor and McLennan, 1995; Rudnick and Gao, 2003).

The estimates on the bulk composition of continental crust require, however, knowledge on the composition of the deeper parts of the crust. This information is provided mainly by xenoliths of mafic granulites that are transported to the surface by intraplate alkali basalts (mostly in rift-related environments) and tectonically uplifted blocks that consist of granulite facies metamorphic terranes. The post-Archean metamorphic granulite terranes appear to represent upper-crustal segments or transitional zones between the upper and lower crust (e.g., depressed in Himalayan-type collision zones (Bohlen and Mezger, 1989; Mezger, 1992). Xenoliths of mafic granulites are considered as representatives of the lower crust. The xenoliths were interpreted as cumulates from basaltic magmas that were rising to lower-crust levels and were later subjected to the granulite facies metamorphism (cf. Kay and Kay, 1981; Rudnick, 1992). Since the xenolith data are rather sporadic, Rudnick and Fountain (1995) combined it with seismic velocities to get an estimate of lower-crust bulk composition. Nevertheless, as noted above, the xenoliths of mafic granulite were mostly transported to the surface by rift-related alkali basalts and they may be associated only with the particular tectonic–magmatic environment that also supports alkali basalt production. Rifting environments are associated with the rise of asthenospheric (mid-ocean ridge basalt (MORB)-like) mantle into the spreading lithosphere (see the example of the opening of the Red Sea below), and thus the production of the source lithologies of the mafic garnulite-xenoliths may be associated with additions of new material to an older continental lithosphere. This scenario is supported by younger Re–Os isotope ages of lower crustal xenoliths compared to the above lying lithosphere (Rudnick, 1992).

Rudnick and Gao (2003) combined the estimates of upper, middle, and lower crust compositions to produce a bulk continental crust composition. Trace-element patterns (normalized to primitive mantle values) of lower, upper, and bulk crust are illustrated in **Figure 1**. The order of appearance of the trace

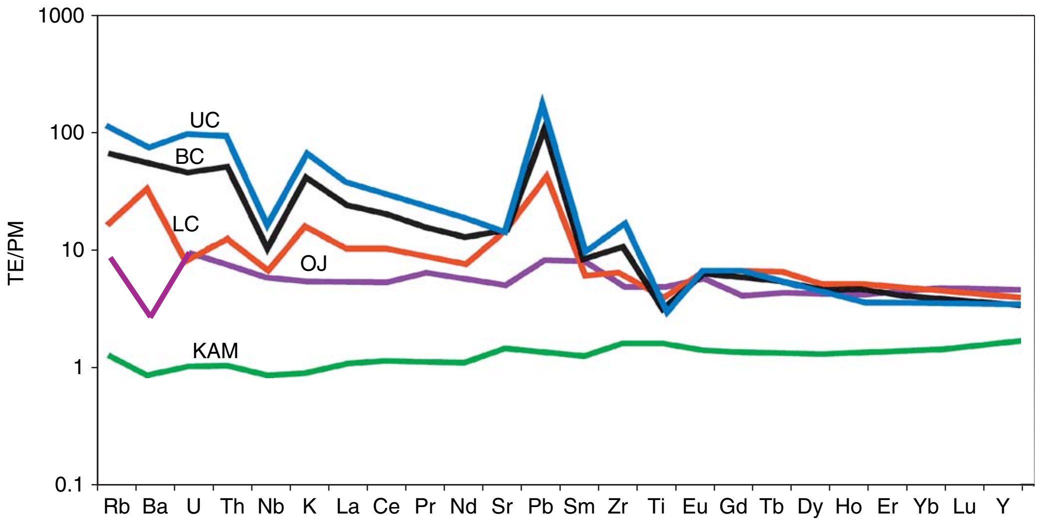

Figure 1 Primitive mantle normalized trace-element patterns of upper, lower, and bulk crust estimates (UC, LC, BC, respectively), the Cretaceous Ontong Java plateau basalt (OJ) and Archean Kambalda komatiite (KAM). Note the overall enrichment of the UC and BC crustal samples in the incompatible trace elements, the distinctive positive Pb and negative Nb anomalies in the crustal samples and the flat patterns of the plateau and almost primitive in KAM. A significant differentiation is required to transform OJ or KAM-type oceanic crust to the upper continental crust. The fractionation of trace elements such as Nb and Pb could reflect processes in the subduction environment. Nevertheless, if accreted plateaus similar to OJ basalts comprise an important part of the continental crust they are either not well represented by the bulk continental crust composition estimates or that they underwent (before or during accretion) internal magmatic differentiation (which is reflected by their continental-like seismic velocities, see **Figure 6** below). Data sources: Rudnick and Gao (2003) and Mahoney *et al*. (1993a).

elements on the abscissa reflects the tendency of the incompatible trace elements to move preferentially from the source to the melt during magma generation. This behavior can be applied to the processes that are involved in the production of continental crust from mantle sources (Hofmann, 1988). Noticeable features in this figure are the strong positive Pb anomalies and the negative Nb anomalies in the lower, upper, and bulk continental crust curves. These anomalies are probably related to mantle–crust differentiation processes that can fractionate Pb and Nb from the rare earth elements (henceforth REE), Th, and U (cf. Hofmann *et al*., 1986). Such fractionation could occur in the subduction environment ('the subduction factory') along with the production of 'arc-type' calc-alkaline magmas (cf. Stein *et al*., 1997). The Mesozoic Ontong Java plateau basalts and the Archean Kambalda komateiite display trace-element patterns that are distinctly different from those of the 'lower, upper, and bulk' continental crust. These types of magmas show almost 'flat' primitive trace-element patterns (although the Ontong Java sample displays small negative Ba anomaly that seems to oppose the strong positive Ba anomaly of the 'average lower crust'). If the Kambalda and Ontong Java-type magmas represent important constituents of continental crust, they are not well represented by the estimated 'bulk crust' patterns.

9.07.2.2 Composition of the Continental Crust: Open Questions

Below we outlined several topics and open questions that emerged from the extensive efforts to evaluate the chemical composition of the continental crust:

9.07.2.2.1 Problems with the simple 'andesite model'

The overall geochemical similarity of continental crust to arc-type magmas (e.g., the estimated average ~60% SiO_2 andesitic composition of bulk continental crust, as well as similar values of indicative trace-element ratios such as Ce/Pb and Nb/Th in arc – crust magmas) led many workers to suggest that the arc-subduction environment where andesitic magmas are produced is the main locus of juvenile crust production (the andesite model of continental crust formation; e.g., Taylor (1967)), and that accretion of arcs at convergent margins is a major means of continental crust growth. Transport and differentiation of melts and fluids that eventually control the composition of continental crust were associated with the subduction environment where mantle-wedge melts and fluids with trace elements move upward, while interacting and mixing with the bulk peridotitic assemblages (e.g., Kelemen, 1995). However, most arcs are too mafic to build continental crust by simple accretion (cf. Anderson, 1982; Nye and Reid, 1986; Pearcy *et al.*, 1990). Moreover, Taylor and McLennan (1985) pointed out several problems with a simplistic application of the andesite model, such as the failure of this model to explain the abundances of Mg, Ni, and Cr and the Th/U ratio in upper crustal rocks. Rudnick (1995) suggested that the lower Sr/Nd and La/Nb ratios in continental crust compared to arc magmas are compatible with involvement of intraplate magmatism in continental crust production. The andesitic magmatism and in a broader sense the intermediate 'grandioritic' bulk composition of the upper crust can be considered as a product of the differentiation processes that are involved in mantle–continental crust transformation. Taylor and McLennan (1985) suggested that only ~20–25% of the crust reflects the currently operating 'andesite growth mode', whereas the dominant mechanism that produced the Archean crust was the derivation of the bimodal basic-felsic magma suites from mantle sources (e.g., formation of Archean tonalitic–trodhjemitic–granodiortic (TTG) magmas). It is argued below that the production of thick oceanic crust and lithosphere by mantle plume activity and their accretion to the continents was an essential part of continental crust production throughout the past 2.7–2.9 Ga of the Earth's history.

9.07.2.2.2 Plumes and arcs

Contribution of juvenile material from 'plume' or 'arc' sources to the continents can be assessed by using trace-element ratios of Ce/Pb and Nb/Th that are distinctly different in arcs and plume-related mantle domains such as OIB, and oceanic plateaus (Hofmann *et al.*, 1986; Stein and Hofmann, 1994; Kerr, 2003). Meta-tholeiites and greenstone sequences from major ancient orogens (e.g., the Superior, Birimian, and ANS), as well as plateau basalts such as the Ontong Java tholeiites display flat (chrondite-like) REE patterns (**Figure 2**) and where data exist show plume-type Nb/Th ratios (**Figures 1** and **3**). It appears that these juvenile crustal units are not well represented by the major sediment samplers. The upper-crust composition is largely dictated by the subduction and post-accretionary processes such as crustal melting (that can lead to granite production), lithospheric

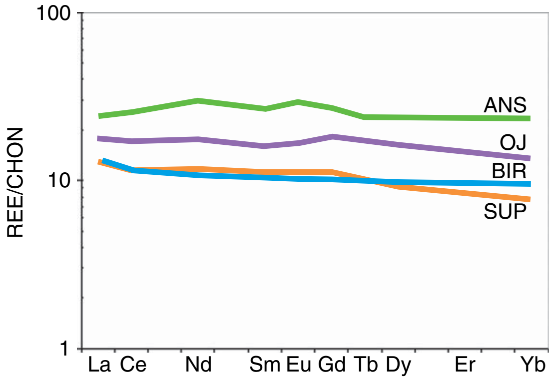

Figure 2 Chondrite normalized REE pattern in oceanic basalt from the Ontong Java plateau (OJ) and several meta-tholeiites that were interpreted as plateaus basalts (from the Arabian-Nubian Shield (ANS); Superior Province (SUP); and Birimian orogen (BIR). All these basalts erupted rapidly and formed thick magmatic sequences marking the early stage in a major orogenic cycle and production within oceanic basin. Data sources: Mahoney *et al*. (1993a), Abouchami *et al*. (1990), Reischmann *et al*. (1983), and Arth and Hanson, (1975).

delamination, and asthenospheric rising. These processes involved the preferential recycling of major elements (Mg, Ca) to the mantle at subduction zones and the removal of other major elements such as Si, Al, Na, K to the continental crust (cf. Albarede, 1998). The subduction factory can be envisaged as a big chromatographic column that fractionates the incompatible trace elements Nb, Pb, Rb from REE and Th, U (Stein *et al.*, 1997). Overall, these processes modified the original geochemical signature of the juvenile mantle material and eventually shifted the upper continental crust to its average andesitic composition (Rudnick, 1995; Stein and Goldstein, 1996; Albarede, 1998).

9.07.2.2.3 The Eu dilemma

The application of the REE to the study of the composition and evolution of continental crust is going back to the works of Goldschmidt (1933) and the subject was thoroughly developed and discussed by Taylor and McLennan (1985). Most of the fine-grain sedimentary rocks that are used to estimate bulk upper-crust compositions typically show a negative Eu anomaly (e.g., loess sample in **Figure 4**), pointing to significant intracrustal melting processes that require the existence of a residual reservoir with an REE pattern showing a positive Eu anomaly (Taylor and McLennan, 1995). However, such a reservoir remains largely unknown,

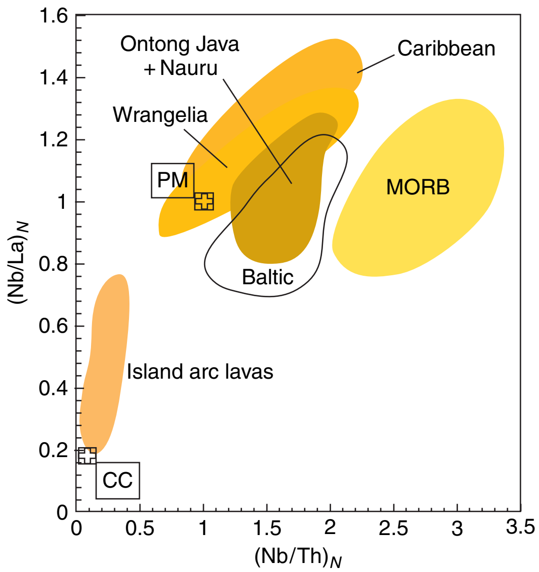

Figure 3 The location of basalts from several plateau basalt provinces on an $(Nb/La)_N$ versus $(Nb/Th)_N$ diagram (after Puchtel *et al.*, 1998). The plateau basalts define fields that are distinctly different from island arc magmas and MORB and lie between primitive mantle compositions (PM) and Nb enriched that are similar to OIB and some rift-related alkali basalts (see **Figure 4**). Hofmann *et al.* (1986) attributed the high Nb/U and Nb/Th ratios to production of a 'residual' mantle after extraction of the low Nb/Th continental crust. Stein *et al.* (1997) proposed that Nb could be fractionated from REE, U, and Th due to production of an amphibole 'front' in the mantle wedge, which they viewed as a giant chromatographic column. The Nb-rich zone is recycled back to the mantle or is frozen at the lower part of the lithospheric mantle after cessation of subduction. Does the convergence of several plateau basalts to primitive mantle ratios indicate that the plume represents mantle with primitive Nb–Th–La ratios? In any event, Archean plateau magmas such as those of the Baltic Shield and the Phanerozoic magmas overlap to some extent in the diagram but are clearly not identical.

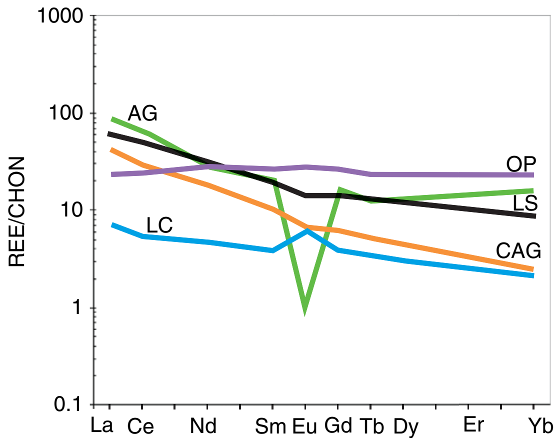

Figure 4 REE (chondrite normalized) pattern of magmas representing main stages in the growth of a 'typical' orogenic cycle: tholeiitic magmas related to oceanic plateaus (OP), calc-alkaline granites (CAG), and alkaline granites (AG), as well as xenoliths of mafic granulites that are interpreted as lower crustal samples (LC) and Late Pleistocene loess material that sample the exposed-eroded upper crust in the Sahara desert (LS). The pattern of the loess material is identical to the estimated bulk upper-crust estimate (not shown). Note the appearance of a pronounced negative Eu anomaly related to plagioclase fractionation in the alkali granites and the almost smooth pattern of the calc-alkaline granite. The production of the alkaline granites in the Arabian–Nubian Shield was attributed to significant fractionation of mafic (Mushkin *et al.*, 2003). The calc-alkaline granites could be the product of mid-crust (amphibolite-type) melting. Both would produce residual assemblages. The topic of production of the vast calc-alkaline granitic batholiths is not well understood. The production of granites requires water ('no water, no granites'), but the source of heat for the melting and the fate of the residual crust are not well known and the whole subject requires further study.

because the xenoliths of mafic granulite that are considered as lower-crust samples, and indeed show positive Eu anomalies are not interpreted as partial melting residues (cf. Rudnick and Taylor, 1987; Rudnick, 1992). It should be noted, however, that among the upper-crust magmatic rock suites, it is mainly the alkaline granites and related volcanics that display distinctive negative Eu anomalies, while most of the calc-alkaline granites from juvenile crustal terranes display REE patterns with small or no Eu anomalies (e.g., the calc-alkaline batholiths of the Birimian and the ANS orogens (**Figure 4**). The alkaline granites and related rhyolitic volcanics typically characterize the closing stages of large orogenic cycles and their petrogenesis reflects fractionation of mafic magmas in shallow crustal levels (e.g., Mushkin *et al.*, 2003 and see elaboration of this topic below). These suites comprise, however, a rather small fraction of the total upper-crust magmatic inventory that is probably irrelevant to the total crustal budget.

9.07.2.2.4 Crustal foundering

The search after 'residual reservoir' is associated with models of crustal foundering. This topic was recently discussed by Plank (2005) in light of the high Th/La ratio in the continental crust (e.g., **Figure 1**). She argued that the high Th/La ratio is unlikely to reflect fractionation processes within the subduction environment, nor could it be related to processes of intraplate magmatism, or to Archean magmatism

(Condie (1993) showed that Archean continents were rather characterized by low Th/La). Alternatively, the Th/La continental crust ratio can be related to production of lower crustal cumulates (with low Th/La ratios), and foundering of such cumulates and partial melting restites (both represent high-density heavy crust) into the mantle. This process would create over geological time bulk continental crust with high Th/La ratio.

9.07.2.2.5 Archean TTG magmatic suites

Archean magmatic suites were characterized by the bimodal mafic-felsic compositions with scarce appearance of andesitic rocks (Taylor and McLennan, 1985). Thus, the simple andesite model cannot be applied to the production of the major inventory of continental crust which is Archean in age. The felsic members of these suites show fractionated heavy rare earth element (HREE) patterns with no Eu anomaly indicating melting in the garnet stability zone and no evidence for intracrustal melting involving plagioclase fractionation. The felsic precursors of these sediments were interpreted to reflect partial melting products of basaltic or TTG magma sources. The TTG magmas for themselves were interpreted as fractionation products of intermediate parental magmas that were derived from enriched and metasomized mantle peridotite (Shirey and Hanson, 1984) or melting products of the subducting slab. For example, high-Al TTG plutons that are characterized by high-La/Yb ratios, like those exposed in the Superior province greenstone-granitoid terranes were interpreted as melting of young, hot subducting oceanic slabs (Drummond *et al.*, 1996). In this relation, the production of modern magnesian andesites, Nb-rich basalts, and adakites could be attributed to subduction of hot oceanic slabs (e.g., central Andes; Stern and Kilian (1996): Papua New Guinea; Haschke and Ben-Avraham (2005)). The processes involved with the TTG magma production required an environment with high heat flow (supporting production of Mg-rich basalts and komatiites) and probably rapid stirring of the mantle. Taylor and McLennan (1985) envisage the formation of the Archean continental nuclei as a process involving eruption piling and sinking of tholeiitic basalts followed by partial melting and formation of the felsic magmas. To a limited extent this scenario resembles the descriptions of fast production of thick oceanic crust in relation to mantle plume activity and the subsequent processing of the juvenile-enriched crust by possible subduction and intraplate processes possible since late Archean time. The major tectonic-magmatic element in this description that apparently was not significant during the Archean is the subduction activity and the production of andesitic-calc-alkaline magmas.

9.07.2.2.6 Post-Archean granites

While the felsic members of the Archean bimodel suites show highly fractionated HREE, the post-Archean granitoids typically show flat or slightly fractionated HREE patterns, indicating a shallower depth of source melting above the garnet stability zone. The depth of melting and the different sources thus represent a fundamental difference between Archean and post-Archean crust formation (Taylor and McLennan, 1985). It should be stressed that TTG magmas are petrogenetically different from the post-Archean granitoids. For example, if the subducting plate melts in the eclogite stability zone, there is a possibility to produce TTG magmas and thus granitic rocks without water. The post-Archean granites are melting and differentiation products of water-bearing mafic sources at shallower depth of the crust. This melting involves fractionation of amphibole instead of clinopyroxene and orthopyroxene. Since the amphibole is low in SiO_2 the residual melt is more silicic driving the differentiation from basalts (the 'normal': melt produced in subduction zones) to silicic melts all the way to granite. The apex of this discussion is the requirement of water carrying subduction processes in the formation of the post-Archean granitic magmas.

Campbell and Taylor (1983) emphasized the 'essential role of water in the formation of granites' and, in turn, continents and introduced to the literature the important concept of 'no water, no granites, no oceans, no continents'. They argued that the Earth, which is the only inner planet with abundant water, is the only planet with granite and continents, whereas the Moon and the other inner planets have little or no water and no granites or continents. This concept could be extended to: 'no water, no plate tectonics, no continents', since with no water plate tectonics is unlikely to work. Thus, there is a fundamental linkage between formation of granitic (granodioritic) continental crust and plate-tectonic mechanism.

9.07.2.3 The Lithospheric Mantle and Continent Stability

Beneath the Moho and the continental crust lies the lithospheric mantle whose thickness reaches 300–400 km beneath old continents and is relatively thin (up to 100–120 km) beneath young continents (e.g., beneath the Canadian and ANS, respectively). Subcontinental lithospheric xenoliths show large variations and heterogeneities in the trace-element abundances that over time can lead to large variations in the radiogenic isotope systems such as Pb, Nd, Sr (cf. McKenzie and O'Nions, 1983; Hawkesworth *et al.*, 1993). The magmatic and thermal history of the lithospheric mantle is related to that of the above-lying continental crust. Based on Pb, Nd, and Sr iotopes in ophiolites, galenas, and basalts, Stein and Goldstein (1996) suggested that the lithospheric mantle beneath the Arabian continent was generated during the events of the late Proterozoic crust formation. The lithospheric mantle can be regarded as a 'frozen' mantle wedge, which previously accommodated the production of calc-alkaline magmas and fractionation–transportation processes of trace elements en route to shallower levels of the crust (the subject is elaborated below). After cessation of subduction, the water-bearing lithospheric mantle could have become a source for alkali magmas that rose to the shallower levels of the continental crust (e.g., alkali granites) or erupted at the surface (cf. Mushkin *et al.*, 2003; Weinstein *et al.*, 2006). Thus, the lithospheric mantle is not only a potential source of information on the 'subduction factory' processes, but can be regarded as an important source of magmas that constitute a part of the continental crust inventory. Yet, the source of heat and conditions of lithospheric melting is not obvious. One possibility is that asthenospheric mantle rises into the rifting continent, supplying heat for melting. The production of magmas may have eventually caused the depletion in the lithospheric mantle of incompatible trace elements and basaltic components, leading to its exhaustion.

9.07.3 Models of Crust Formation and Continental Growth

9.07.3.1 General

Models that describe the mechanism of continental crust growth and the evolution of continental masses can be divided into two major groups: (1) models of early crustal growth and subsequent recycling and (2) models of continuous or episodic crustal growth.

The most prominent researcher that pushed the 'early crust growth/recycling' model was R.L. Armstrong. He proposed that soon after the early mantle–crust differentiation and formation of continental crust, the Earth reached a steady state with accretion and destruction of continental masses occurring at approximately equal rate with essentially constant volumes of ocean and crust through geological time (Armstrong, 1981a, 1991). The key process in this model is the recycling of the continental masses back into the mantle. Armstrong based his model on the principles of the constancy of continental freeboard and the uniformity of thickness of stable continental crust with age. These two critical parameters imply negligible continental crustal growth since 2.9 Gy BP. Defending his model, Armstrong addressed some of the arguments that were put against the steady state model, mainly those that relied on the behavior of the radiogenic isotopes (e.g., Nd, Sr, and Pb; Armstrong (1981b)).

Among the pioneering researchers that developed the ideas of continuous/episodic crust growth were W.W. Rubey (who advocated the underplating theory as well), A.E. Engel, P.M. Hurley, and S.R. Taylor. The prominent researcher of this school of thinking was S. Moorbath who pointed out the difficulty of subducting large amounts of continental crust due to its buoyancy and argued that the chronological evidence derived from continental crustal rocks suggests creation of juvenile continental crust throughout Earth's history. Moorbath also developed the concept of continental accretion as a major building process of continental crust (Moorbath, 1975, 1978).

The steady-state-recycling theory of crustal growth requires evidence for substantial sediment recycling. Sediments are clearly recycled in the subduction environment. Sediment involvement in subduction was demonstrated by various geochemical tracers of arc magmas: for example, the radiogenic isotope ratios of Nd, Pb, Sr, and Hf (cf. Patchett *et al.*, 1984), ^{10}Be isotopes (cf. Tera *et al.*, 1986), and trace-element fluxes (Plank and Langmuir, 1998). Plank (2005) introduced the Th/La ratio as a tracer of sediment recycling in subduction zones. Th/La in the subducting sediments reflects the mixing between terrigenous material with high 'upper crustal' Th/La ratios (0.3–0.4) and metalliferous sediments and volcaniclastic sediments with low Th/La ratios (<0.1). White and Patchett (1984) argued that a significant but small (at least 1–2%) contribution of older continental material could be potentially recycled into new continents. Nevertheless, based on mass-age distribution and Nd isotope data of sediments McLennan (1988) estimated the mass of sediment available for subduction to be $<1.6 \times 10^{15}$ g yr^{-1}, and argued that, that is insufficient to support a steady-state crustal mass according to the Armstrong model (see also Taylor and McLennan (1995)).

Hofmann *et al.* (1986) suggested that the Nb/U ratio in mafic–ultramafic magmas can be used to deduce the history of continental crust growth because the Nb/U ratio changed from the primitive mantle ratio of ~30 to a higher ratio ~47 in the 'residual' mantle due to extraction of crust with Nb/U of ~10. Some of the Archean komatiites (e.g., samples from Kambalda and Kostomoksha provinces) show high Nb/U ratios ~47 (Sylvester *et al.*, 1997; Puchtel *et al.*, 1998) that may support the Armstrong steady-state model. Yet, other komatiites and early Proterozoic tholeiites show Nb/U ratios that are similar to the primitive value (**Figure 1**). Puchtel *et al.*, (1998) commented on this topic that: "the Nb/U data alone cannot unequivocally resolve this important controversy."

Although the continents appear to have undergone episodic changes throughout all of geologic time, the available geochronology of crust formation shows time periods when large areas of continental crust were formed at a fast rate, alternating with apparently more quiescent periods of low crust-formation rates (marked in **Figure 5**). It may be that these quiescent periods merely represent missing continental material because the crust was destroyed and recycled into the mantle (according to the Armstrong model) but actual evidence for this process is indirect at best.

9.07.3.2 Timing and Rates of Crustal Growth

Crust formation ages were determined by direct dating of major crustal terranes (numerous works applying the Rb–Sr, Sm–Nd dating methods and U–Pb and zircon ages) and by using the concept of Nd crustal model ages that mark the time of major chemical fractionation and the Sm/Nd ratio change accompanying the extraction of melts from the mantle and their incorporation into the continental crust (e.g., McCulloch and Wasserburg, 1978; Nelson and DePaolo, 1985; DePaolo, 1981; DePaolo *et al.*, 1991). The Sm/Nd crustal formation ages suggest that 35–60% of the currently exposed crustal masses were produced in the Archean (while actual exposed Archean crust accounts only for 14%; Goodwin

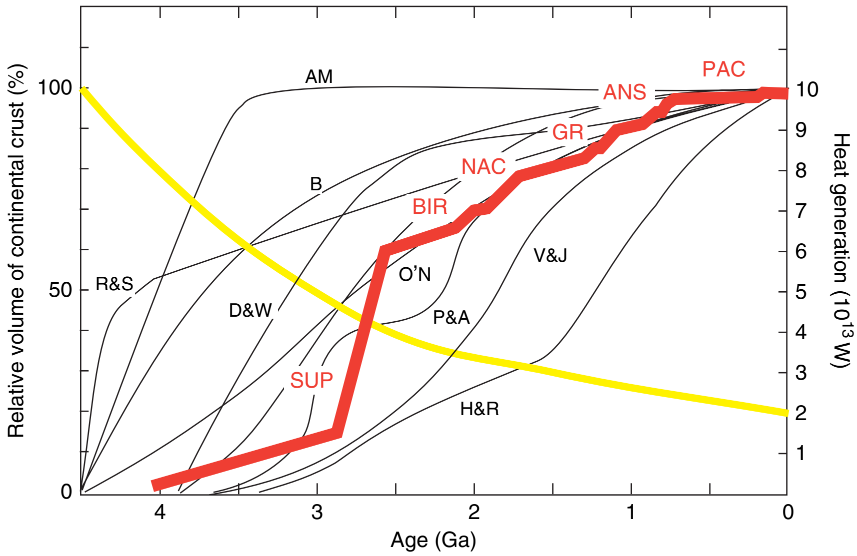

Figure 5 Models of continental crust (cumulative) growth and heat generation (yellow curve) throughout Earth's history (after Reymer and Schubert, 1984; Taylor and McLennan, 1985, 1995; Condie, 1990; Patchett, 1996). Crustal curves: R&S = Reymer and Schubert (1984); AM = Armstrong (1981a,1981b); B = Brown (1979); D&W = Dewey and Windley (1981); O'N = O'Nions *et al*. (1979); P&A = Patchett and Arndt (1986); V&J = Veizer and Jansen (1979); H&R = Hurley and Rand (1969). The thick red curve is after Taylor and McLennan (1995), with marks of the major orogenic episodes that were related by Stein and Hofmann (1994) to major upwelling events and plume head uprise in the mantle (the MOMO model illustrated in **Figure 9**). SUP = Superior province; BIR = Birimian orogeny; NAC = North Atlantic Continent; GR = Grenville orogeny; ANS = Arabian–Nubian Shield; PAC = Pacifica 'super-plume'.

(1991)). **Figure 5** shows that the most dramatic shift in the generation of continental crust happened at the end of the Archean, 2.5–2.7 Gya. The late Archean, early Proterozoic pulse of crustal growth is represented in all continents (Condie, 1993). Very little continental crust was preserved before 3 Ga. The oldest preserved terrestrial rocks are the 3.96 Ga Acasta Gneiss in the Northwest Territories of Canada (Bowring *et al.*, 1990) and detrital zircons from the Yilgarn Block, Western Australia, were dated to a much older age of 4.4 Ga (Wilde *et al.*, 2001).

9.07.3.3 The Reymer and Schubert Dilemma

Reymer and Schubert (1984, 1986) evaluated the rate of growth of major continental crust segments (e.g., the Superior province in North America, and the ANS) and showed that they significantly exceed the rate of continental addition that prevailed along the subduction margins during Phanerozoic time (~1 $km^3\,yr^{-1}$). The Phanerozoic rate was based on estimates of crustal additions along Mesozoic–Cenozoic arcs, hot spots, and some other additional sources (e.g., underplating). They calculated a total addition rate (mainly along the arcs) of 1.65 and total subtraction rate 0.59 $km^3\,yr^{-1}$, yielding a net growth rate of 1.06 $km^3\,yr^{-1}$. In addition, Reymer and Schubert calculated a growth rate by an independent model based on the constancy of freeboard relative to mantle with declining radiogenic heat production.

The term freeboard was derived from civil engineering, describing the additional height above a normal operating water level and the top of the water-holding structure. The geological–geophysical use of this concept encompasses the complicated relation between continental crustal thickness, volume, mantle temperature, and Earth's heat budget. Wise (1974) tied the freeboard concept to crustal growth rates, by suggesting that approximately constant continental crustal volumes and areas pertained since the Archaean–Proterozoic boundary, thus maintaining approximately the same elevation or freeboard of the continents above mean sea level. The freeboard calculation of Reymer and Schubert yielded a Proterozoic–Phanerozoic growth rate of 0.9 $km^3\,yr^{-1}$, very similar to the above-calculated value. The Phanerozoic growth rate is 3 times less than the average Archean growth rate. Assuming formation between 900 and 600 Ma during the Pan-African orogeny, Reymer and Schubert showed that if the ANS was generated by convergent margin magmatism it required an addition rate of 310 km^3-km of convergence^{-1}-Ma^{-1}, as compared to a global total of ~40 km^3-km^{-1}-Ma^{-1} (**Figure 6**). Thus, they speculated that an additional large-scale mechanism, probably plume-related magmatism, played an important role in the formation of the ANS, as well as in other major orogenic events.

At the very least, the results of Reymer and Schubert indicate that the local crust-formation rates of the major continental provinces are as high as or higher than the present-day rate averaged over the entire globe. As a consequence of their analysis, Reymer and Schubert proposed that juvenile additions to continental crust require additional mechanism besides the arc magmatism and this mechanism could involve activity of hot spots or mantle plumes. The geodynamic conditions during these intervals of juvenile crust creation may resemble to some extent the Archean conditions.

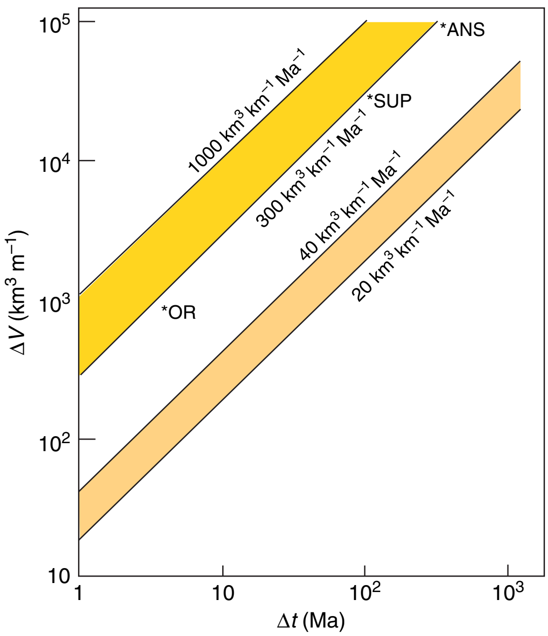

Figure 6 Arc-crust (volume) addition curves (after Reymer and Schubert, 1984) showing the large difference between Phanerozoic arc activity (20–40 $km^3\,km^{-1}\,Ma^{-1}$) and the growth of some major orogens (e.g., SUP = Superior; ANS = Arabian–Nubian Shield ~300 $km^3\,km^{-1}\,Ma^{-1}$; OR = Oregon coastal range). The figures demonstrate the Reymer and Schubert dilemma that points to the necessity to invoke additional sources and mechanism to support the production of continental crust during the episodes of major orogens production.

Galer (1991) performed another evaluation of the interrelationships between continental freeboard, tectonics, and mantle temperature. He concluded that before ~3.8 Ga the potential temperature exceeded 1600°C in the shallow mantle and therefore little or no hypsometric or tectonic distinction existed between 'continental' and 'oceanic' regimes. Thus, the pre-3.8 Ga period was subjected to tectonic and sedimentary processes that were distinctly different to those characterized by the Earth surface since mid-Archean time. An important component in his evaluation is the assumption that the early Archean oceanic crust was significantly thicker than typical present crust. The Reymer and Schubert (1984, 1986) assessment indicates that mantle plumes could play an important role in post-Archean crustal history of the Earth. Below we discuss the possible role of thick oceanic crust (e.g., the oceanic plateaus) in continental crustal growth. We can ask the question whether post-Archean episodes of crustal growth have some important components that 'mimic' to some extent the early Archean conditions.

9.07.3.4 Continental Growth and Heat Production

The heat production in the Earth, which is the main driving force for mantle convection and upwelling processes has declined by a factor of 5 since the Early Archean time (see yellow curve in **Figure 5**). This decrease reflects the decay of the long-living radiogenic isotopes. **Figure 5** indicates that continental crust accumulation does not reflect the Earth thermal history very well, particularly in its earliest Archean history since only in the late Archean does a rough correspondence appear between the heat generation and crustal growth (Patchett, 1996). The heat engine of the mantle should have supported extensive mantle movements in the Archean such as plume head upwelling. Yet, the low cumulative crustal growth in the Archean may indicate that the system was as efficient at destroying continental crust as making it (Taylor and McLennan, 1995).

The intensive pulses of continental crust formation in the late Archean and early Proterozoic probably took place in a mantle that convected less vigorously than at earlier times. One possibility is that the tectonic regime became more similar to modern plate tectonics, where water played an essential role. By late Archean and early Proterozoic times we might enter the regime of 'water, plate tectonics, continents'.

9.07.4 Oceanic Plateaus and Accreted Terranes

9.07.4.1 General

With the advances made in geophysical techniques over the last 40 years, a different and new picture of the Earth began to emerge. For the first time, it was possible to image and sample the planet's crust and even layers deeper down. During the 1970s it started to become clear that the ocean floor is not homogenous, but contains areas of significantly thicker crust than the norm. The discovery of one such area, the Ontong Java plateau with a crustal thickens of over 30 km, gave rise to the term 'oceanic plateau' (Kroenke, 1974). Today, about 100 such anomalous regions ranging in size from 1000 km to a few kilometers have been documented (**Figure 7**).

The formation of these plateaus is still not clearly understood. They were considered to represent remnants of extinct arcs, abandoned spreading ridges, detached and submerged continental fragments, anomalous volcanic piles, mantle plume head traces, and uplifted oceanic crust. However, both isotope dating and modeling support the theory that they formed rapidly, often in less than 2–3 My (Richards *et al.*, 1989), with the majority forming at or near mid-ocean ridges (Kerr, 2003). These areas are conducive to large decompression melting (Eldholm and Coffin, 2000), providing a source of magma needed to form these features.

At least from a geophysical point of view, values of crustal seismic velocities, V_p, in the upper 5–15 km (where V_p ranges from 6.0 to 6.3 km s^{-1}, insert in **Figure 7**), together with the anomalous thickness, relief, and gravity data, would indicate a continental structure for some of these plateaus (Nur and Ben-Avraham, 1977, 1978; Ben-Avraham *et al.*, 1981; Carlson *et al.*, 1980). Other plateaus are from clearly oceanic origin, probably the result of continuous extrusion of basalts, probably from an active hot spot or mantle plume head.

Nur and Ben-Avraham (1977) and Ben-Avraham *et al.* (1981) suggested that the plateaus with the 'crustal seismic velocities' (e.g., Ontong Java plateau) represent continental fragments rifted away from a parent continent that they named Pacifica. Stein and Hofmann (1994) and Abbott and Mooney (1995) on the other hand suggested that the plateaus were first extracted from the mantle within the oceanic basins above mantle plumes and were later accreted to the continents. In this context, it is interesting to note the

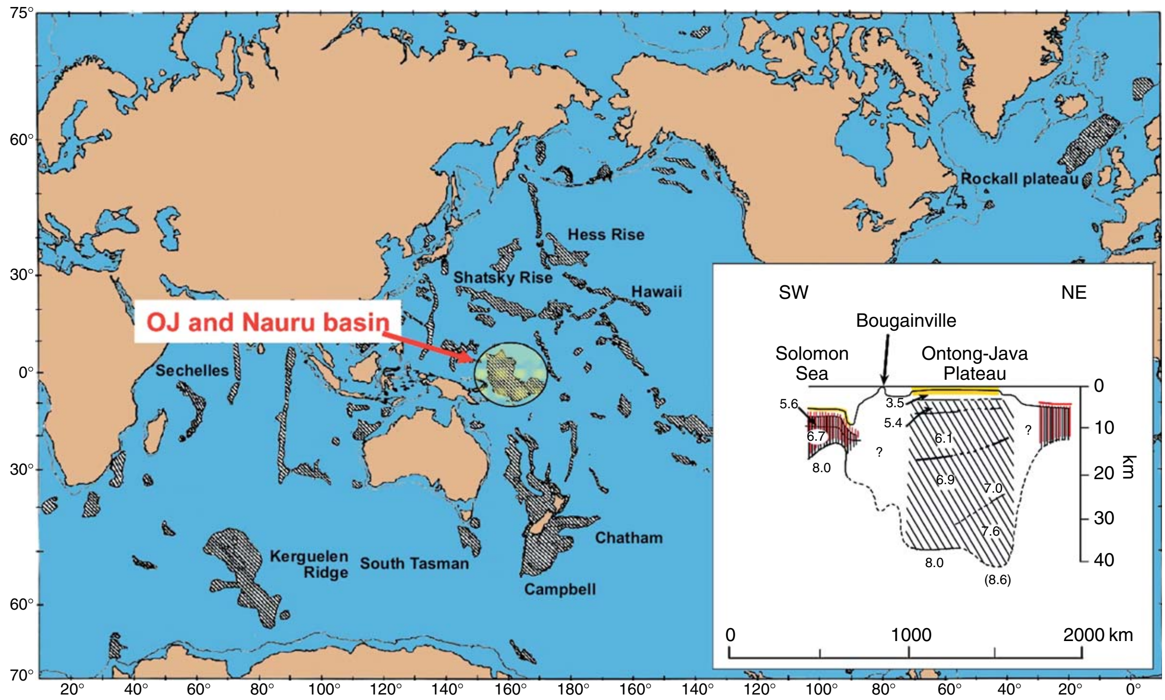

Figure 7 Distribution of oceanic plateaus and location of the Ontong Java plateau and Nauru Basin in the Pacific (after Ben-Avraham *et al*., 1981). The Mesozoic Pacific was the locus of enhanced magmatism that was related to plume activity ("The last pulse of the Earth." Larson (1991)). It also appears that shortly after the production of the thick oceanic plateaus subduction zones were developed on their margins. The Nauru Basin comprised oceanic crust that erupted in an ocean-spreading environment but shows geochemical similarities to the adjacent Ontong Java plateau. Thus, it appears that both regions derived their magmas from similar enriched sources. This configuration supports the existence of an enriched upper mantle beneath the Pacific. Stein and Hofmann (1994) argued that the Nd–Sr isotope composition of this mantle is similar to the PREMA isotope composition (e.g., εNd $\sim +6$ and ^{87}Sr/^{86}Sr $\sim$ 0.7035 as defined by Worner *et al*. (1986) for various volcanic rocks around the world). It appears that the Pacific-PREMA composition characterized the entire upper mantle during the time of the production of the Ontong Java and Nauru Basin basalts and thus could mark the overturn or plume uprise events according to the MOMO model. Anderson (1994) proposed an alternative interpretation to the production of the Mesozoic Pacific plateau that requires no plume activity. He suggests that the upper mantle of the Mesozoic Pacific and the Indian Ocean is hot because these regions have not been cooled by subduction for more than 200 My. Anderson argues that the timing of formation and location of the plateaus are related to the organization of the tectonic plates. This would require that the PREMA-type enriched isotope compositions characterized the upper mantle rather than MORB, but magmatism in evolved continental rift environments (e.g., Ross Sea, Antractica; and Red Sea, Arabia) clearly show transition from enriched isotope composition to typical depleted MORB (e.g., Stein and Hofmann (1992); Rocholl *et al*. (1995) and see Figure 11 and 'Red Sea model' in **Figure 12** below). (b) Seismic velocities across the Ontong Java plateau (after Ben-Avraham *et al*., 1981). Surprisingly, the velocities suggest continental crust composition and structure. This observation illustrates a kind of 'paradox' since it is expected that Ontong Java will show seismic structure of a thick oceanic crust. Does this mean that crustal formation processes operate in the mid-oceanic environment? We regard the 'Ontong Java paradox' as a major open question that should be addressed by future studies.

co-appearance of the basaltic sequences of the Ontong Java plateau and those of the adjacent Nauru Basin in the Pacific Ocean (cf. Mahoney *et al*., 1993a, 1993b). The tectonic setting of the Nauru Basin is clearly of a mid-oceanic spreading center type, with the typical appearance of magnetic anomalies and eruption of tholeiitic basalts. The important point is that the Ontong Java and Nauru Basin basalts display similar geochemical characteristics such as Nd and Sr isotope ratios that are enriched relative to the depleted MORB-type mantle (e.g., the basalts from the East Pacific Rise). Thus, it appears that the basalts erupted in the Nauru Basin and the Ontong Java plateau represent a similar uppermost mantle that is enriched relative to the 'normal' depleted asthenospheric mantle. This enriched asthenosphere could be related to a large plume head (or a mantle upwelling) event that brought enriched material to the shallower mantle.

Recently, Richardson *et al.* (2000) produced a three-dimensional tomographic model of the seismic structure beneath the Ontong Java plateau that indicated the existence of a low-velocity mantle 'root' reaching the depth of ~300 km. This was interpreted as a remnant of the Cretaceous Ontong Java plume that was attached to the plateau.

Ishikawa *et al.* (2004) address the question of this seismically anomalous low-velocity root beneath the Ontong Java Plateau and its lithospheric mantle composition and structure by studying suite mantle xenoliths (peridotites and pyroxenites) from Malaita, Solomon Islands. The shallower mantle (Moho to 95 km) is composed of variably metasomatized peridotite with subordinate pyroxenite derived from metacumulates, while the deeper mantle (95–120 km) is represented by pyroxenite and variably depleted peridotites. The shallower and deeper zones are separated by a garnet-poor zone (90–100 km), which is dominated by refractory spinel harzburgites. Ishikawa *et al.* attributed this depth-related variation to different degree of melting for a basalt–peridotite hybrid source at different level of arrival depth within a single adiabatically ascending mantle plume: the lack of pyroxenites at shallower depths was related to extraction of hybrid melt from completely molten basalt through the partially molten ambient peridotite, which caused the voluminous eruption of the Ontong Java plateau basalts. The authors concluded that the lithosphere forms a genetically unrelated two-layered structure, comprising shallower oceanic lithosphere and deeper impinged plume material, which involved a recycled basaltic component, now present as a pyroxenitic heterogeneity. The possible relevancy of this model to the shallower seismic structure of the Ontong Java plateau (**Figure 7**) could be that evolved magmas were produced within the shallower layers.

Regardless of the mechanisms responsible for the formation of oceanic plateaus, they are assumed by many researchers to have an important role in the formation of new continental crust. Because they are part of moving plates and are embedded in them, they are destined to arrive at subducting plate boundaries and to accumulate, at least in some cases, onto the continental margin as accreted allochthonous terranes.

The potential of oceanic crust to be accreted to the continents rather than being subducted depends on its magmatic and thermal history and is reflected in its buoyancy. Under conditions of 'normal seafloor spreading operation' the oceanic lithosphere begins its history as a buoyant hot plate that cools with time and becomes susceptible to subduction. The buoyancy of the oceanic lithosphere depends on its age and density distribution (Oxburgh and Parmentier, 1977). The density distribution reflects the composition and the thickness of the crustal and mantle layers of the lithosphere, which in turn are controlled by the mantle temperature at the place of crust formation. A hotter mantle will produce a thicker oceanic crust with more depleted lithospheric mantle, which can become less dense than more fetile mantle and thus, more buoyant (McKenzie and Bickle, 1988; Oxburgh and Parmentier, 1977; Langmuir *et al.*, 1992). This is probably the case in the production of oceanic plateaus by rising plume heads. The plumes are ~200–300°C hotter than an ambient mantle and can produce 15–40 km thick oceanic plateaus (compared to ~7 km thickness of normal oceanic crust). Therefore, some oceanic plateaus are too buoyant to subduct and they can either obduct onto the continent (e.g., Caribbean and Wrangelia; Kerr *et al.* (1997); Lassiter *et al.* (1995)) or when arriving to the subducting magins start subduction in the opposite direction (e.g., Ontong Java plateau; Neal *et al.* (1997)). The Archean mantle whose temperature was a few hundred degrees higher than the post-Archean mantle could produce thicker oceanic crust of 20–25 km (Sleep and Windley, 1982) and the plume-related plateaus could reach an average thickness of ~30 km (e.g., Puchtel *et al.*, 1997). Some of these thick oceanic plateaus might be too warm and too buoyant when they reach the subduction margins and remain susceptible to subduction. Cloos (1993) and Abbott and Mooney (1995) argued that oceanic plates with crust thickness over ~25 km are not subductable regardless of their age.

Nevertheless, several oceanic plateaus are currently being consumed, either by collision or by subduction. Among these are the Nazca and Juan Fernandez Ridges off South America and the Louisville and Marcus-Necker Rises in the western Pacific (Pilger, 1978; Cross and Pilger, 1978; Nur and Ben-Avraham, 1981). These plateaus, which are presently being subducted, can clearly be associated with gaps in seismicity of the downgoing slab (e.g., Nur and Ben-Avraham, 1982). However, even more pronounced than these gaps are gaps in volcanism, associated with these plateaus. Close spatial association was found between zones of collision or subduction of plateaus and gaps in volcanism in the Pacific (McGeary *et al.*, 1985). Thus, these currently active areas of continental accretion are clearly identifiable.

Large portions of the Pacific margins, especially the northeast margin, are made of accreted or allochthonous terranes (**Figure 8(a)**). The best-understood accreted terranes are those in the northern cordillera of western North America, particularly in southern Alaska and British Columbia (Coney *et al.*, 1980; Monger, 1993). Paleomagnetic evidence indicate that many of the North Pacific accreted terranes in Alaska and northeast Asia migrated several thousand kilometers over periods of tens of millions of years prior to their accretion to the margins. The accreted terranes comprising the Canadian–Alaskan cordillera are predominantly juvenile in composition (e.g., Samson and Patchett (1991) estimated that ~50% by mass of the Canadian segment of the cordillera was juvenile crustal material). This evaluation is based on the initial $^{87}Sr/^{86}Sr$ and ε_{Nd} values of lithological assemblages from both outboard and inboard terrains (e.g., Alexander, Cache Creek, and Slide Mountain), as well as arc assemblages like Quesnel (Armstrong (1988); Armstrong and Ward (1993); Samson *et al.* (1989, 1990, 1991); and other references listed by Patchett and Samson (2003) in their recent review on this topic).

Future allochthonous terranes may be found in the oceans, in the various plateaus, which are present on the ocean floor (Ben-Avraham *et al.*, 1981). The plateaus are modern accreted terranes in migration, moving with the oceanic plates in which they are embedded and fated eventually to be accreted to continents adjacent to subduction zones. The present distribution of oceanic plateaus in the Pacific Ocean and the relative motion between the Pacific Plates and the Eurasian Plate suggest that the next large episode of continental growth will take place in the northwest Pacific margin, from China to Siberia. Similar situations may have occurred in the past in other parts of the world.

Thus, it may be possible to identify ancient accreted terranes in the geological record, based on

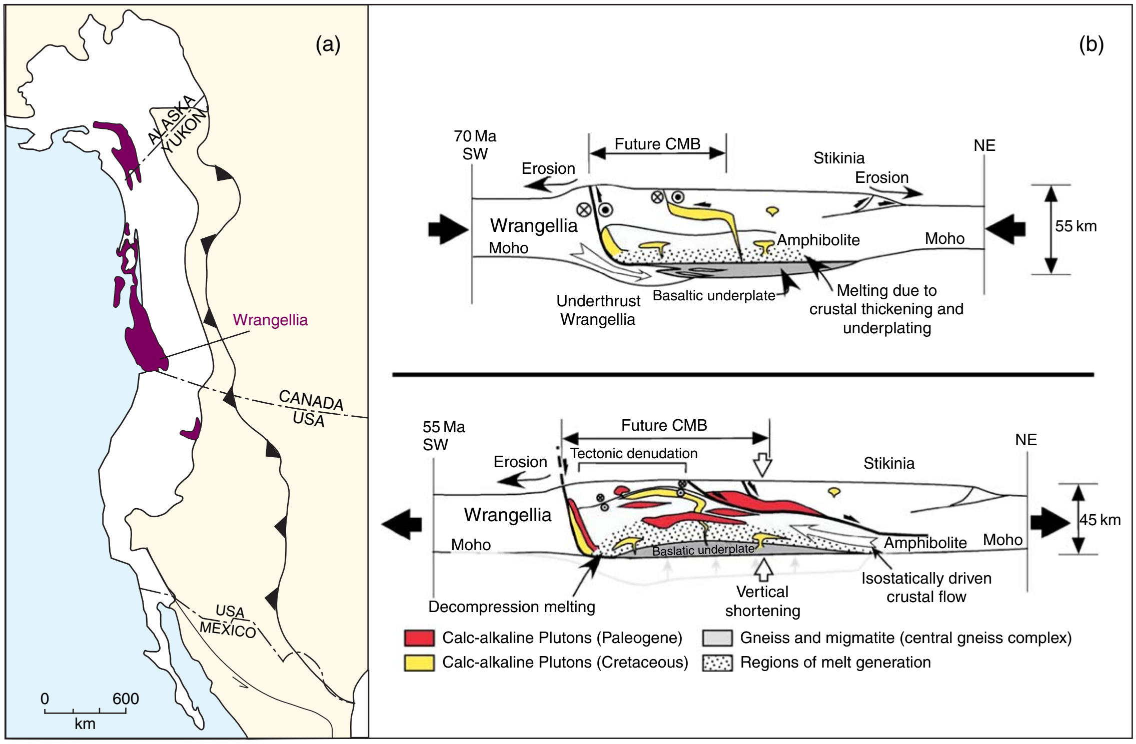

Figure 8 (a) Accretion plateaus along the western margins of North America and Alaska. The location of the Wrangallia plateau is marked in brown (after Ben-Avraham *et al*., 1981). (b) Model of plateau accretion and transformation of the oceanic crust to new continental crust and lithosphere (after Hollister and Andronicos, 2006). The figure shows processes of accretion, shortening, and melting. Note the stratigraphic organization of the Cretaceous and the Paleogene calc-alkaline granites and location of zones of melting due to crustal thickening and basaltic underplating. Amphibole appears above the Moho and its production could control Nb/Th(REE) fractionation (Stein *et al*., 1997).

chemical and geological evidence. The oldest preserved oceanic plateau sequences have been dated to 3.5 Ga in the Kaapvaal Shield, South Africa (Smith and Erlank, 1982; De Wit *et al.*, 1987) and the Pilbara craton in Australia (Green *et al.*, 2000). Greenstone belts of the Canadian Superior province, ranging in age from 3.0 to 2.7 Ga also contain lava groups that have been interpreted to be remnants of accreted oceanic plateaus (see recent review by Kerr (2003)). The evidence for an oceanic plateau origin is based on the occurrence of pillow basalts and komatiites without terrestrial sedimentary intercalations or sheeted like swarms, possessing the characteristics of Cretaceous oceanic plateaus. Below, we elaborate on the examples of crustal growth in the ANS and the Baltic Shield where evidence was found for formation of oceanic plateau and accretion of juvenile terrains to the existing older continents.

9.07.4.2 The Transformation of Oceanic Plateau to Continental Crust

The accretion of oceanic plateaus with a clear oceanic affinity to continental margins is of fundamental importance in the mechanism of crustal growth. The crustal structure of these plateaus, which was originally composed of one layer made of basalt, is modified with time and a typical structure of upper continental crust is developed. The most dramatic example of this process can be seen in British Columbia, northwest Canada. Here many accreted (or 'exotic') terranes were added to the continent during lithospheric plate convergence from the Archean to the present. The tectonic evolution of northwestern Canada involved a series of accretion events, alternating with periods of continental extension. Thus, it is tempting to suggest that the continent has 'grown' westward by the surface area of the accreted terranes (Cook and Erdmer, 2005). Recently, a very large-scale geophysical and geological transect, Slave-Northern Cordillera Lithospheric Evolution (SNORCLE), was carried out in northwestern Canada (Cook and Erdmer, 2005). The results show quite clearly that several of the large accreted terranes, which have traveled long distances embedded within oceanic plates before colliding with the continent, now have a continental crustal structure. For example, Wrangelia and Stikinia terranes, which were accreted during the Phanerozoic, and Hottah terrane, which was accreted during the Proterozoic, have continental crustal structure with an upper crustal layer having granitic Vp velocities (Clowes *et al.*, 2005).

In a recent paper Hollister and Andronicos (2006) have proposed that crustal growth in the Coast Mountains, along the leading edge of the Canadian Cordillera, was the result of processes associated with horizontal flow of material during transpression and subsequent transtension, and the vertical accretion of mantle-derived melts. Their model has two distinct tectonic phases. The first occurred during a period of transpression when the continental crust was thickened to about 55 km, and mafic lower crust of the oceanic plateau Wrangellia was pushed under the thickened crust where basalt from the mantle heated it to temperatures hot enough for melting. The basalt and the melts mixed and mingled and rose into the arc along transpressional shear zones, forming calc-alkaline plutons. The second phase occurred as the arc collapsed when a change of relative plate motions resulted in transtension that accompanied intrusion of voluminous calc-alkaline plutons and the exhumation of the core of the Coast Mountains Batholith (CMB) (**Figure 8(b)**).

The calculation made by Schubert and Sandwell (1989) suggests that accretion of all oceanic plateaus to the continents on a timescale of 100 My, would result in a high rate of continental growth of 3.7 km^3 yr^{-1}. This rate is much higher than continental growth rate based only on accretion of island arcs, which is 1.1 $km^3\,yr^{-1}$ (Reymer and Schubert, 1986), thus providing an explanation for the above-mentioned crustal growth rate dilemma.

9.07.4.3 Mantle Overturn and Crust Formation Episodes

In an attempt to link between various characteristics of mantle geochemistry to the dynamics and the apparent episodic mode of enhanced crustal growth (the Reymer and Schubert dilemma), Stein and Hofmann (1994) proposed that the geological history of the mantle-crust system has alternated between two modes of mantle convection and dynamic evolution, one approximating a two-layer convective style when 'normal mode' of plate tectonics prevails (Wilsonian tectonics) and the other characterized by significant exchange between the lower and upper mantle, when large plume heads form oceanic plateaus. The exchange periods apparently occurred over short time intervals – a few tens of million years (as estimated from the chronology of magmatism that is involved in the production of the oceanic plateaus), whereas the Wilsonian periods extend over several hundreds of millions of years (e.g., ~800 My in the case of the ANS, see below). The idea is that during mantle overturn episodes (termed by Stein and Hofmann as MOMO episodes) substantial amount of lower mantle material arrives to the Earth's surface (mainly via plume activity), replenishes the upper mantle in trace elements, and forms a new basaltic crust which eventually contributes juvenile material to the continental crust (via both subduction and accretion processes, Figure 9). Shortly after mantle upwelling and the production of thick oceanic crust, new subduction zones can be developed along the oceanic plateau margins as happened in the Mesozoic Pacific province. This mechanism views the formation of new subduction zones as a consequence of the production of thick oceanic crust and it's cooling.

The final product of the plume-thick oceanic-crust subduction system is a mass of continental crust that can be generated over a relatively short period of geological time. The sequence of events involving formation of an anomalously thick, oceanic lithosphere, which is modified by subduction at its margin explains both the formation of major continental segments during discrete episodes of time and their calc-alkaline affinities.

In an alternative view of the Reymer and Schubert dilemma, Patchett and Chase (2002) and Patchett and Samson (2003) suggested that the evidence for enhanced crust production during discrete episodes can reflect the tectonics of accretion processes alone, arguing for the role of transform faulting, which serves to pile up various terrains of juvenile crust in one restricted region (e.g., the Canadian–Alaskan cordillera, where slices of juvenile accreted crustal material were piled up following northward along-margin transport associated with transform faulting; see Figure 8). The transport and accretion of juvenile mantle terranes is a central part of the plume-accretion model of crustal growth (Figure 9). Yet, the additional important component in this model that is not required by the arc-accretion models is the important role of oceanic plateaus as the nucleus of the juvenile continental crust growth and possibly the locus of newly developed subduction zones. Moreover, the crustal terranes that were formed during the episodes of 'major orogeny' are not limited to geographically restricted regions but rather are distributed globally. The best example is that of the late Proterzoic Pan-African orogeny, when juvenile crust terranes were produced in Arabia, Nubia (the ANS discussed below), as well as in North Africa, New Zealand, and Patagonia. All these geographical-crustal segments were parts of the early Phanerozoic Gondwana. Considering a reconstruction of this old continent (Figure 10) it appears that continental segments within the Gondawana that can be

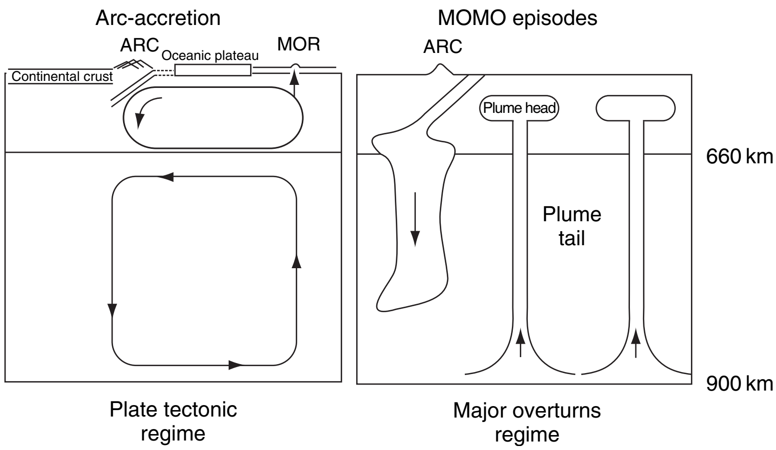

Figure 9 Mantle overturn and crust addition model (modified after the original MOMO model by Stein and Hofmann (1994)). The original MOMO model assumes that the mantle–crust system evolves through two modes of convection and dynamic evolution, one (MOMO overturn episodes) is characterized by significant exchange between lower and upper mantle and crust and the other by two-layer convective style when the typical plate tectonic Wilsonian cycle operates. The question of layered vs whole mantle convection has been continuously discussed in the literature and it is beyond the scope of this review. Here we present a modified representation of the model where we stress the accretion of the oceanic plateau to the existing continent and development of subduction zones on the margins of the plateau (similar to the scenario shown in Figure 8). Most of the existing subduction zones in the world are associated with plateaus though a few subduction zones are not clearly associated with plateaus.

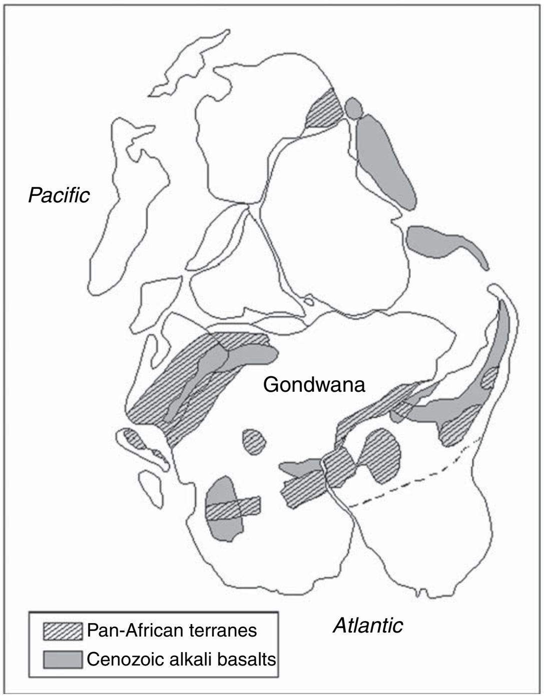

Figure 10 Schematic representation of regions that expose terranes of Late Proterozoic Pan-African juvenile magmatic sequences and Cenozoic alkali basalts over a reconstructed map of Gondwana. The juvenile magmatic sequences are characterized by Nd and Sr isotope compositions (e.g., εNd $= +5 \pm 1$ and $^{87}Sr/^{86}Sr = 0.7028 \pm 2$; Stein and Hofmann (1994); Stein (2003)). The Pb isotopic compositions of the Gondwana basalts are marked in **Figure 11**. These basalts show the plume-lithosphere connection) that were derived from the juvenile Pan-African lithospheric mantle. The areas that are best documented are those of the Arabian–Nubian Shield on both sides of the current Red Sea (closed in the reconstruction).The figure suggests that during the Late Proterozoic Pan-African orogeny juvenile terranes were accreted to an existing continental nucleus in a similar manner to the growth that was described for the North America continent. The distribution of the juvenile Pan-Africa terranes may mark the loci of the subduction zones that were developed at that time along the margins of the Pan-African oceanic plateaus.

identified as juvenile Pan-African crustal or lithospheric mantle terranes (see Stein (2003) and caption of **Figure 10**) form a 'belt' along the possible margins of a mid-or-late Proterozoic continent. All of these segments are characterized by juvenile Pan African Nd and Sr isotope compositions that in turn are consistent with plume (MOMO) type mantle (see MOMO evolution line in Stein and Hofmann (1994)).

9.07.5 Rapid Growth of Major Continental Segments

In this section we describe the magmatic histories of several prominent continental crust segments (termed as major orogens). We first elaborate on the evolution of continental crust comprising the ANS, which is one of the best-exposed crustal terranes (due to the Cenozoic uplift of the Red Sea shoulders). Then, we describe in chronological order the other major orogens. Overall, we show the general similarity in the evolutionary pattern of all these orogenic episodes.

9.07.5.1 The ANS Orogeny (~0.9–0.6 Ga)

The excellent preserved exposures of the Late Proterozoic crystalline basement of the ANS and the sequences of Phanerozoic alkali basalts provide an opportunity to monitor the magmatic history of the ANS over the past ~900 My. The low initial $^{87}Sr/^{86}Sr$ ratios of the Late Proterozoic magmas of Sinai, Jordan, northern Saudi Arabia, and the Eastern Desert of Egypt preclude significant involvement of an old pre-Pan-African felsic continental precursor in this part of the northern ANS, indicating that basement magmas were derived from either a pre-existing mafic crust or directly from the mantle. The absence of older continental crust is further indicated by mantle-type oxygen isotope ratios (e.g., $\delta^{18}O = +5.5$; Bielski (1982)).

The Late Proterozoic magmatic history of the ANS can be divided into four evolutionary phases.

9.07.5.1.1 Phase I

Production of thick sequences of tholeiitic during a very short interval of time (~900 ± 30 My). The (meta) tholeiites are characterized by chondritic REE patterns and 'mantle-type' Th/Nb ratios and were interpreted as oceanic basalts similar to the plateaus basalts (Stein and Goldstein (1996); Stein (2003) and references therein). It is important to emphasize that along with the production of the plateau magmas, thinner sequences of meta-tholeiites were interpreted as 'normal' mid-oceanic tholeiites erupting along spreading ridges (the Old Meta Volcanics in Egypt, described by Stern (1981)). The two ANS meta-tholeiite suites resemble the magmatic–tectonic association of the above-described Pacific Ontong Java plateau–Nauru Basin basalts.

9.07.5.1.2 Phase II

Upon reaching a continental margin, the thick oceanic plateau lithosphere resisted subduction and plate convergence occurred on its margins, forming the ANS calc-alkaline magmas. The subduction activity lasted for ~200 My (between ~870 and 650 Ma), similar to the duration of the activity of Phanerozoic arc magmatism in Japan. The arcs and oceanic plateaus were accreted to the old Gondwanaland producing a thick and mainly mafic crust and lithospheric mantle, which served later as sources for the calc-alkaline granitic batholiths, alkali granites, and Phanerozoic alkali basalts.

The enriched plume-type mantle of the early ANS was exhausted rapidly by the subduction processes and the Gabel Gerf ophiolitic magmas (~800–750 Ma), which are assumed to represent the arc environment, show isotope mixtures between the enriched ('plume lithosphere mantle') and depleted ('MORB-type mantle') components (see evolution lines in **Figure 11**).

9.07.5.1.3 Phase III

Rapid production (~640–600 Ma) of vast amounts of calc-alkaline granitic and dioritic magmas. They form batholiths that cover large areas, mainly in the northern ANS and their mechanism of production is not entirely clear. Melting of a metamorphosed mafic (e.g., amphibolitic) crust probably led to the production of the calc-alkaline granites.

However, the production of the calc-alkaline grantic magmas certainly required the supply of 'heat and water', a problem that is not clearly understood. One possibility is a relation between asthenospheric rise associated with delamination of a thick lower crust and lithosphere that were produced in the earlier evolutionary stage of the island arc magmatism.

9.07.5.1.4 Phase IV

The orogenic cycle terminated by production of alkali granites and related rhyolites (~600–530 Ma; Bielski (1982); Mushkin *et al.* (2003)). The parental magmas of the alkaline magmas were probably derived from the lithospheric mantle and underwent extensive differentiation at shallow crustal levels (Mushkin *et al.*, 2003).

The transition from calc-alkaline magmatism to alkaline magmatism associated with the lithospheric mantle melting after 600 Ma reflects the ending of plate convergence. In the ANS this transition is associated with a major uplift (Black and Liegeois, 1993). Similar rapid transitions in the stress field associated with continental collisions were described in other localities such as the Variscan province in Europe, the Basin and Range province in the USA, and the Tibetan plateau (Klemperer *et al.*, 1986; Rey,

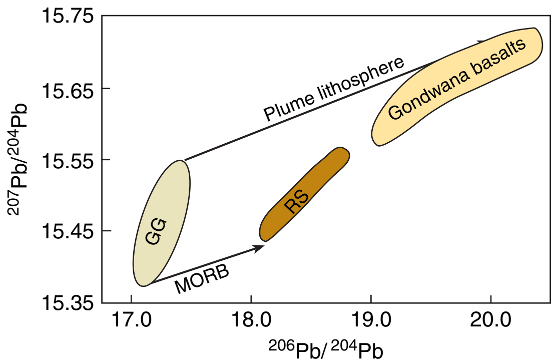

Figure 11 The distribution of Late Proterozoic oceanic basalts (the Gabel Gerf (GG) ophiolites, Red Sea basalts (RS), and Gondwana basalts in the Pb–Pb diagram (after Stein and Goldstein, 1996; Stein, 2003). Stein and Goldstein proposed that the GG ophiolites lie in the diagram on a mixing trend between the asthenopsheric MORB mantle and asthenopsheric plume-type mantle (in similar tectonic-magmatic settings to the configuration of the Nauru Basin and East Pacific Rise in the Pacific Ocean). The Gondwana basalts (location marked in **Figure 10**) lie on an evolution trend together with the Late Proterozoic enriched ophiolites and other magmas (e.g., galenas and K-feldspars not shown here, see Stein and Goldstein, (1996)) and suggest that the lithospheric mantle beneath the Arabian continent was generated during the Late Proterozoic crust-formation events. It can be regarded as a 'frozen' mantle wedge, which previously accommodated the production of calc-alkaline magmas and fractionation–transportation processes of trace elements en route to shallower levels of the crust (the subject is elaborated below). The Red Sea basalts lie on a mixing trend between the lithospheric derived basalts and the MORB-type asthenosphere. Stein and Hofmann (1992) noted that enriched-type Red Sea basalts erupt on the southern and northern edges of the Red Sea Basin while depleted MORB erupt only in the central trough where 'real' spreading occurs. They regarded this configuration as a strong evidence for the existence of the plume-type mantle above the asthenospheric MORB. The latter is rising into the rifting lithosphere and continental crust. This also provides strong evidence for the MORB-type composition of the upper mantle. The Gondwana basalts show a mixing relation with the Red Sea asthenosphere suggesting that MORB is rising into the rifting lithosphere (e.g., in the Ross Sea rift, Ahaggar, and the Dead Sea rift). This upwelling has important implications for the heating of the lithosphere and the lower crust, as well as for production of basaltic melts from the lithosphere and provide evidence for asthenospheric underplating (Stein, 2006).

1993; Costa and Rey, 1995). In the Himalayas, major uplift and the appearance of alkaline volcanism has been related to the thinning and melting of the lithosphere after the ending of plate convergence.

9.07.5.2 The Sumozero–Kenozero and Kostomuksha Orogeny (~2.8–2.9 Ga)

The magmatic history of the ANS appears to represent a complete MOMO cycle illustrating a primary means of continental crust formation and cratonization throughout Earth history. The model may be particularly relevant in cases of rapid continental growth as demanded by the Reymer and Schubert dilemma. Evidence for the existence of magmatic provinces that resemble oceanic plateaus was found in other older continental crust provinces (Figure 12). The 2.9 Ga Sumozero–Kenozero greenstone belt in the SE Baltic Shield comprises a several kilometer-thick lower unit that is made of submarine mafic–ultramafic units (with plume-type Nb/U ratios of 43 $\pm$ 6) that were interpreted as representing oceanic plateau magmas and the upper unit consisting of arc-type magmas (Puchtel *et al.*, 1999). The authors suggested that the over-thickened plume-related oceanic crust and the overlying arc-type magmas were later accreted to the exisiting TTG-type older crust of the Vodla block.

The ~2.8 ka Kostomuksha greenstone belt in the NW Baltic Shield consists of komatiitic–basaltic

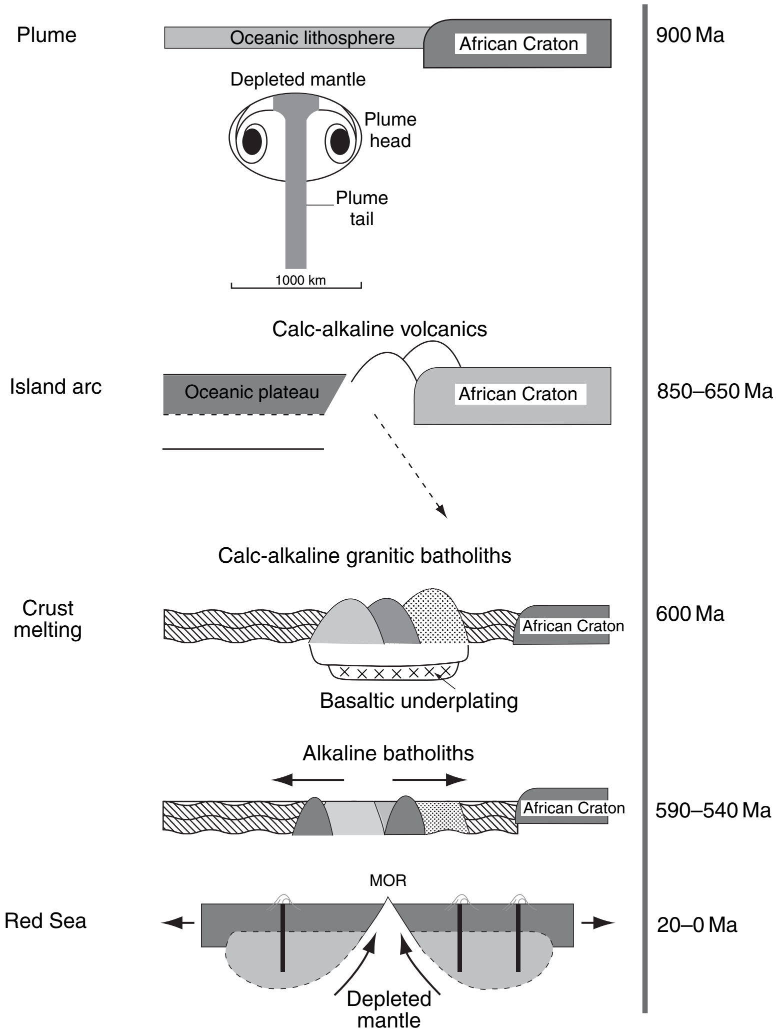

Figure 12 Schematic model illustrating the growth and fate of the continental crust and lithospheric mantle in the Arabian–Nubian Shield. Similar (but clearly not identical) structure of magmatic histories can be ascribed to other major crustal segments (e.g., Canadian Shield, Birimian orogen, the Mesozoic Pacific).

submarine lavas and volcano-clastics whose formation was attributed by Puchtel *et al.* (1997) to plume activity and the production of a thick oceanic crust–oceanic plateau. The plateau reached the Baltic continental margin but was too buoyant to subduct and became a new part of the continent. Using the example of the Cretaceous Caribbean plateau (Kerr *et al.*, 1997), Puchtel and colleagues suggested that only the upper part of the thick komatiitic–basaltic crust of the proposed Baltic oceanic crust was imbricated and obducted on the ancient continent, whereas the deeper zones were subducted to the mantle. The subduction of this lower crust gave rise to subduction-related volcanism that erupted within the accreted sequences. This mechanism distinguished the thick plateau type oceanic crust from the thinner crust that is represented in the ophiolite sections. Moreover, it provides a plausible explanation for the sequence of events that occurred within a 'cycle' of plateau formation–accretion–delamination–subduction–calc-alkaline arc-type magmatism.

9.07.5.3 The Superior Province (~2.7 Ga)

The ~2.7 Ga Superior province, the largest Archean craton on Earth comprises several continental and oceanic mega-terranes that were formed and accreted rapidly (the ages of the various superprovinces lie in the range of 3.1–2.67 Ga and acceretionary tectonics occurred between 2.74 and 2.65 Ga (cf. Percival and Williams, 1989). The geological history and magmatic stratigraphy of the Superior province was divided by Arth and Hanson (1975) into four stages: (1) production of tholeiitic basalts by shallow melting of mantle peridotites, followed by fractional crystallization that produces basaltic-andesites; (2) formation of trondhjemite, tonalite, and dacites by melting of eclogites in mantle depth; (3) intrusion of quartz monzonites; and (4) intrusion of post-kinematic syenites.

The general pattern of this history is strikingly similar to the four growth phases and magmatic evolution of the above-described late Proterozoic ANS or the Proterozoic Birimian orogeny. In the framework of the MOMO model this means: production of thick sequences of tholeiitic magmas in the plume-related intraplate (oceanic) environment, followed by development of subduction zones and their related magmatism, followed by intracrustal and intralithospheric melting that recycled the material formed in the early two phases. The main difference between these orogenic suites lies in the appearance of the Archean TTG assemblage.

Polat and Kerrich (2001a, 2001b) and Polat *et al.* (1998) described the formation of the 2.7 Ga Wawa and Abitibi greenstone subprovinces of the Superior province as an example of episodic continental growth, by lateral spread of subduction-accretion complexes. They recognized two principal volcanic associations: (1) tholeiitic basalt-komatiite; and (2) tholeiitic to calc-alkaline bimodal basalt-rhyolite. Tholeiitic basalts of the former association are characterized by near-flat REE patterns and Th/U, Nb/Th, Nb/La, and La/Sm-pm ratios that span the primitive mantle values. Transitional to alkaline basalts and Al-depleted komatiites have fractionated REE patterns and OIB-like trace-element signatures. Polat and his co-authors interpreted these assemblages as representing a oceanic plateau derived from a heterogeneous mantle plume. The bimodal association has fractionated REE patterns and negative Nb, Ta, P, and Ti anomalies typical of arc magmas. Tonalite plutons derived from partial melting of subducted oceanic slabs intrude the subduction-accretion complex as the magmatic arc axis migrated toward the trench. The eruption of voluminous ocean plateau and island arc volcanic sequences and the following deposition of kilometer-thick siliciclastic trench turbidites, and intrusion of the supracrustal units by syn- to post-kinematic TTG plutons occurred in the relatively short time period of 2.75–2.65 Ga (a classic representation of the Reymer and Schubert super-rapid crustal formation events). The above-mentioned events were accompanied by contemporaneous intense poly-phase deformation, regional greenschist to amphibolite facies metamorphism, and terrain accretion along a north-northwest-dipping subduction zone. The description of the history of the Superior province is consistent with the plume upwelling – terrane accretion model where the growth of the new continental crust involved formation and accretion of oceanic plateaus, production of magmatic arcs, and accretion of older continental fragments.

9.07.5.4 The Birimian Orogeny (~2.2 Ga)

Another well-studied case of episodic crustal formation is that of the Birimian orogeny that lasted between ~2.2 and ~2.1 Ga (Abouchami *et al.*, 1990; Boher *et al.*, 1992). The early stage in the growth of the Birimian juvenile crust was marked by eruption of thick sequence of basalts in oceanic basin that were interpreted as oceanic plateau (formed in an environment remote from any African-Archean crust). Shortly after, calc-alkaline magmas were produced, probably in island arcs that were developed on the margins of the oceanic plateau, which then collided with the Man Archean Craton. Overall, the magmatic-geodynamic history of the Birimian orogeny strikingly resembles that of the ANS and to some extent that of the Superior province.

9.07.6 Summary

The continents come from the mantle. Large submarine oceanic plateaus whose production was associated with mantle plumes may have formed the basis for the growth of continental crust, at least since the late Archean time. The chronological evidence suggests that plateau formation was rapid and episodic and that it was followed by enhanced activity of arcs and calc-alkaline magmatism. Thick oceanic plateaus are buoyant and resist subduction. Yet, some plateaus were subducted, reflecting probably internal differentiation within the plateaus and the geodynamic conditions (e.g., thermal state of the mantle). Subduction of segments of plateaus can in turn enhance arc magmatism, because its altered parts can increase the water flux at depth of magma generation. It should be also emphasized that oceanic plateaus may well be an important component of current arc magmatism, as some of the currently active arcs are located close to anomalously thick oceanic crust (e.g., the central Aleutians and Lesser Antilles).

Along with the production of new crust, the lithospheric mantle is evolved . It probably represents a frozen mantle wedge where production of magmas and migration of fluids and incompatible trace elements occurs, thus both the continental crust and the lithospheric mantle preserved the complicated history of continent growth. Besides the plateau and arc accretion and calc-alkaline magmatism, delamination of lithospheric mantle and lower crust and underplating by rising asthenospheric magmas, as well as lithospheric melting are all important processes that lead to the magmatic and thermal maturation of the continents. The fate of the continent relates eventually to the geodynamics and thermal conditions in the mantle.

Many of the abovementioned topics require further study. In particular, we emphasize the importance of understanding the origin of the 'crustal type' Vp velocities that indicate a continental structure for some of the oceanic plateaus. Are these velocities associated with significant magmatic differentiation before accretion of the plateau to the existing continents? The mechanism of production of the calc-alkaline granitic batholiths is not as well established as is the composition of 'mid-crust'. The processes of asthenospheric underplating, crustal foundering, and crustal recycling and their relation to the plate tectonic cycles and mantle dynamics through geological time require more attention.

Finally, among the numerous works and thorough reviews on continental crust composition and evolution we consider and 'rephrase' the Taylor and Campbell statement: no water, no oceans, no plate tectonics, no granites, no continents, as a major characteristic of the environment of production of the Earth continental crust.

References

Abbott D and Mooney W (1995) The structural and geochemical evolution of the continental-crust – Support for the oceanic plateau model of continental growth. *Reviews of Geophysics* 33: 231–242.

Abouchami W, Boher M, Michard A, and Albarede F (1990) A major 2.1 Ga event of mafic magmatism in west Africa – An early stage of crustal accretion. *Journal of Geophysical Research-Solid Earth and Planets* 95: 17605–17629.

Albarède F (1998) The growth of continental crust. *Tectonophysics* 296: 1–14.

Anderson AT (1982) Parental melts in subduction zones: Implications for crustal evolution. *Journal of Geophysical Research-Solid Earth and Planets* 87: 7047–7070.

Anderson DL (1994) Superplumes or supercontinents? *Geology* 22: 39–42.

Armstrong RL (1968) A model for the evolution of strontium and lead isotopes in a dynamic Earth. *Reviews of Geophysics* 6: 175–199.

Armstrong RL (1981a) Radiogenic Isotopes – The case for crustal recycling on a near-steady-state no-continental-growth Earth. *Philosophical Transactions of the Royal Society of London Series A-Mathematical Physical and Engineering Sciences* 301: 443–472.

Armstrong RL (1981b) Comment on 'Crustal growth and mantle evolution: Inferences from models of element transport and Nd and Sr isotopes'. *Geochimica et Cosmochimica Acta* 45: 1251.

Armstrong RL (1988) Mesozoic and early Cenozoic magmatic evolution of the Canadian Cordillera. *Geological Society of America Special Paper* 218: 55–91.

Armstrong RL (1991) The persistent myth of crustal growth. *Australian Journal of Earth Sciences* 38: 613–640.

Armstrong RL and Ward PL (1993) Late Triassic to earliest Eocene magmatism in the North American Cordillera: Implications for the Western Interior Basin. In: Caldwell WGE and Kauffman EG (eds.) *Geololgical Association Canada Special Paper 39: Evolution of the Western Interior Basin*, pp. 49–72. Boulder, CO: Geological Society of America.

Arth JG and Hanson GN (1975) Geochemistry and origin of Early Precambrian crust of Northeastern Minnesota. *Geochimica et Cosmochimica Acta* 39: 325–362.

Barth MG, McDonough WF, and Rudnick RL (2000) Tracking the budget of Nb and Ta in the continental crust. *Chemical Geology* 165: 197–213.

Ben-Avraham Z, Nur A, Jones D, and Cox A (1981) Continental accretion and orogeny: From oceanic plateaus to allochthonous terranes. *Science* 213: 47–54.

Bennett VC and Depaolo DJ (1987) Proterozoic crustal history of the Western United-States as determined by neodymium isotopic mapping. *Geological Society of America Bulletin* 99: 674–685.

Bielski M (1982) *Stages in the Evolution of the Arabian–Nubian Massif in Sinai*. PhD Thesis, 155pp, The Hebrew University of Jerusalem.

Black R and Liegeois J (1993) Cratons, mobile belts, alkaline rocks and continental lithospheric mantle: The Pan-African testimony. *Journal of the Geological Society* 150: 89–98.

Boher M, Abouchami W, Michard A, Albarede F, and Arndt NT (1992) Crustal growth in west Africa at 2.1 Ga. *Journal of Geophysical Research-Solid Earth* 97: 345–369.

Bohlen SR and Mezger K (1989) Origin of granulite terranes and the formation of the lowermost continental-crust. *Science* 244: 326–329.

Bowring SA and Housh T (1995) The Earth's early evolution. *Science* 269: 1535–1540.

Bowring SA, Housh TB, and Isachsen CE (1990) The Acasta Gneisses: Remnants of Earth earliest crust. In: Newsom HE and Jones JH (eds.) *Origin of the Earth*, pp. 319–343. New York: Oxford University Press.

Brown GC (1979) The changing pattern of batholith emplacement during Earth history. In: Atherton MP and Tarney U (eds.) *Origin of Granite Batholiths: Geochemical Evidence*, pp. 106–115. Orpington: Shiva.

Campbell IH and Taylor SR (1983) No water, no granites – No oceans, no continents. *Geophysical Research Letters* 10: 1061–1064.

Carlson RL, Christensen NI, and Moore RP (1980) Anomalous crustal structures on ocean basins: Continental fragments and oceanic plateaus. *Earth and Planetary Science Letters* 51: 171–180.

Clarke FW (1889) The relative abundance of the chemical elements. *Philosophical Society of Washington Bulletin* XI: 131–142.

Clarke FW and Washington HS (1924) The composition of the Earth's crust. *USGS Professional Paper* 127: 117.

Cloos M (1993) Lithospheric buoyancy and collisional orogenesis – Subduction of oceanic plateaus, continental margins, island arcs, spreading ridges, and seamounts. *Geological Society of America Bulletin* 105: 715–737.

Clowes RMC, Hammer P T, Fernández-Viejo G, and Welford JK (2005) Lithospheric structure in northwestern Canada from Lithoprobe seismic refraction and related studies: A synthesis. *Canadian Journal of Earth Sciences* 42: 1277–1293.

Collerson KD and Kamber BS (1999) Evolution of the continents and the atmosphere inferred from Th-U-Nb systematics of the depleted mantle. *Science* 283: 1519–1522.

Condie KC (1990) Growth and accretion of continental crust: Inferences based on Laurentia. *Chemical Geology* 83: 83–194.

Condie KC (1993) Chemical-composition and evolution of the upper continental-crust – Contrasting results from surface samples and shales. *Chemical Geology* 104: 1–37.

Condie KC (1997) *Plate Tectonics and Crustal Evolution*. Oxford, UK: Butterworth-Heinemann.

Condie KC (1998) Episodic continental growth and supercontinents: A mantle avalanche connection? *Earth and Planetary Science Letters* 163: 97–108.

Condie KC (2000) Episodic continental growth models: Afterthoughts and extensions. *Tectonophysics* 322: 153–162.

Condie KC (2002) Continental growth during a 1.9-Ga superplume event. *Journal of Geodynamics* 34: 249–264.

Coney PJ, Jones DL, and Monger JWH (1980) Cordilleran suspect terranes. *Nature* 288: 329–333.

Cook FA and Erdmer P (2005) An 1800 km cross section of the lithosphere through the northwestern North American plate: Lessons from 4.0 billion years of Earth's history. *Canadian Journal of Earth Sciences* 42: 1295–1311.

Costa S and Rey P (1995) Lower crustal rejuvenation and growth during post-thickening collapse – Insights from a crustal cross-section through a variscan metamorphic core complex. *Geology* 23: 905–908.

Cross TA and Pilger RH (1978) Constraints on absolute motion and plate interaction inferred from cenozoic igneous activity in Western United States. *American Journal of Science* 278: 865–902.

De Wit MJ, Hart RA, and Hart RJ (1987) The Jamestown ophiolite complex Barberton mountain belt: A section through 3.5 Ga oceanic crust. *Journal of Africa Earth Sciences* 6: 681–730.

DePaolo DJ (1981) Neodymium isotopes in the Colorado Front Range and crust–mantle evolution in the Proterozoic. *Nature* 291: 193–196.

DePaolo DJ, Linn AM, and Schubert G (1991) The continental crustal age distribution – Methods of determining mantle separation ages from Sm-Nd isotopic data and application to the Southwestern United States. *Journal of Geophysical Research-Solid Earth and Planets* 96: 2071–2088.

Dewey JF and Windley BF (1981) Growth and differentiation of the continental-crust. *Philosophical Transactions of the Royal Society of London: Series A-Mathematical Physical and Engineering Sciences* 301: 189–206.

Drummond MS, Defant MJ, and Kepezhinskas PK (1996) The petrogenesis of slab derived tronhejmite – tonalite – dacite/adakite magmas. *Transactions of the Royal Society of Edinborough.*

Eldholm O and Coffin MF (2000) Large igneous provinces and plate tectonics. *AGU Monograph* 121: 309–326.

Galer SJG (1991) Interrelationships between continental freeboard, tectonics and mantle temperature. *Earth and Planetary Science Letters* 105: 214–228.

Gallet S, Jahn BM, Lanoe BV, Dia A, and Rossello E (1998) Loess geochemistry and its implications for particle origin and composition of the upper continental crust. *Earth and Planetary Science Letters* 156: 157–172.

Gao S, Luo TC, Zhang BR, *et al.* (1998) Chemical composition of the continental crust as revealed by studies in East China. *Geochimica et Cosmochimica Acta* 62: 1959–1975.

Goldschmidt VM (1933) Grundlagen der quantitativen Geochemie. *Fortschr. Mienral. Kirst. Petrogr* 17: 112.

Goldschmidt VM (1958) *Geochemistry*. Oxford: Oxford University Press.

Goodwin AM (1991) *Precambrian Geology*. London: Academic Press.

Green M (2001) Early Archaean crustal evolution: Evidence from $\sim$3.5 billion year old greenstone successions in the Pilgangoora Belt, Pilbara Craton, Australia, The University of Sydney.

Green MG, Sylvester PJ, and Buick R (2000) Growth and recycling of early Archaean continental crust: Geochemical evidence from the Coonterunah and Warrawoona Groups, Pilbara Craton, Australia. *Tectonophysics* 322: 69–88.

Griffin WL and O'reilly SY (1987) Is the Continental Moho the crust–mantle boundary. *Geology* 15: 241–244.

Haschke M and Ben-Avraham Z (2005) Adakites along oceanic transforms? *Geophysical Research Letters* 32: L15302 (10: 1029/2005GLO23468).

Hattori Y, Suzuki K, Honda M, and Shimizu H (2003) Re-Os isotope systematics of the Taklimakan Desert sands, moraines and river sediments around the Taklimakan Desert, and of Tibetan soils. *Geochimica et Cosmochimica Acta* 67: 1195–1205.

Hawkesworth CJ, Gallagher K, Hergt JM, and Mcdermott F (1993) Mantle and slab contributions in arc magmas. *Annual Review of Earth and Planetary Sciences* 21: 175–204.

Hofmann AW (1988) Chemical differentiation of the Earth: The relationship between mantle, continental crust, and oceanic crust. *Earth and Planetary Science Letters* 90: 297–314.

Hofmann AW, Jochum KP, Seufert M, and White WM (1986) Nb and Pb in oceanic basalts: New constraints on mantle evolution. *Earth and Planetary Science Letters* 79: 33–45.

Hollister LS and Andronicos CL (2006) Formation of new continental crust in western British Columbia during transpression and transtension. *Earth and Planetary Science Letters* 249: 29–38.

Hurley PM and Rand JR (1969) Pre-drift continental nuclei. *Science* 164: 1229–1242.

Ishikawa A, Shigenori M, and Komiya T (2004) layered lithospheric mantle beneath the Ontong Java plateau: Implications from xenoliths in Alonite, Malaita, Solomon Islands. *Journal of Petrology* 47: 2011–2044.

Kay RW and Kay SM (1981) The nature of the lower continental-crust – Inferences from geophysics, surface geology, and crustal xenoliths. *Reviews of Geophysics* 19: 271–297.

Kay RW and Mahlburgkay S (1991) Creation and destruction of lower continental-crust. *Geologische Rundschau* 80: 259–278.

Kelemen PB (1995) Genesis of high Mg-number Andesites and the continental-crust. *Contributions to Mineralogy and Petrology* 120: 1–19.

Kerr AC (2003) Oceanic plateaus. In: Rudnick R (ed.) *Treatise on Geochemistry, vol. 10: The Crust*, pp. 537–565. Amsterdam: Elsevier.

Kerr AC, Marriner GF, Arndt NT, *et al.* (1996) The petrogenesis of Gorgona komatiites, picrites and basalts: New field, petrographic and geochemical constraints. *Lithos* 37: 245–260.

Kerr AC, Tarney J, Marriner GF, Klaver GT, Saunders AD, and Thirlwall MF (1996) The geochemistry and petrogenesis of the late-Cretaceous picrites and basalts of Curacao, Netherlands Antilles: A remnant of an oceanic plateau. *Contributions to Mineralogy and Petrology* 124: 29–43.

Kerr AC, Tarney J, Marriner GF, Nivia A, and Saunders AD (1997) The Caribbean–Colombian Cretaceous igneous province: The internal anatomy of an oceanic plateau. In: John JM and Millard FC (eds.) *Geophysical Union Monograph: Large Igneous Provinces; Continental, Oceanic and Planetary Flood Volcanism*, pp. 45–93. Washington, DC: American Geophysical Union.

Kessel R, Stein M, and Navon O (1998) Petrogenesis of late Neoproterozoic dikes in the northern Arabian–Nubian shield – Implications for the origin of A-type granites. *Precambrian Research* 92: 195–213.

Kimura G and Ludden J (1995) Peeling oceanic-crust in subduction zones. *Geology* 23: 217–220.

Klemperer SL, Hauge TA, Hauser EC, Oliver JE, and Potter CJ (1986) The Moho in the northern Basin and Range province along the COCORP 40°N seismic reflection transect. *Geological Society America Bulletin* 97: 603–618.

Kroenke LW (1974) Origin of continents through development and Coalescence of oceanic flood basalt plateaus. *Transactions-American Geophysical Union* 55: 443–443.

Kroner A, Compston W, and Williams IS (1989) Growth of Early Archean crust in the Ancient Gneiss Complex of Swaziland as revealed by Single Zircon Dating. *Tectonophysics* 161: 271–298.

Langmuir CH, Klein EM, and Plank T (1992) Petrological constraints on mid-ocean ridge basalts. In: Morgan JP and Blackman DK (eds.) *Geophysical Monograph: Mantle Flow and Melt Generation at Mid-Ocean Ridges*, pp. 182–280. Washington, DC: American Geophysical Union.

Larson RL (1991) Latest pulse of Earth – Evidence for a midcretaceous superplume. *Geology* 19: 547–550.

Lassiter JC, Depaolo DJ, and Mahoney JJ (1995) Geochemistry of the Wrangellia Flood-Basalt Province – Implications for the role of continental and oceanic lithosphere in flood-basalt genesis. *Journal of Petrology* 36: 983–1009.

Mahoney JJ, Storey M, Duncan RA, Spencer KJ, and Pringle M (1993a) Geochemistry and age of the Ontong Java Plateau. In: Pringle MS, Sager WW, Sliter WV, and Stein S (eds.) *American Geophysical Union Monograph: The Mesozoic Pacific: Geology, Tectonics, and Volcanism*, pp. 233–261. Washington, DC: American Geophysical Union.

Mahoney JJ, Storey M, Duncan RA, Spencer KJ, and Pringle M (1993b) Geochemistry and geochronology of Leg 130 basement lavas; nature and origin of the Ontong Java Plateau. *Proceedings of the Ocean Drilling Program, Scientific Results* 130: 3–22.

McCulloch MT and Bennett VC (1993) Evolution of the early Earth – Constraints from Nd-143-Nd-142 isotopic systematics. *Lithos* 30: 237–255.

McCulloch MT and Bennett VC (1994) Progressive growth of the Earths continental-crust and depleted mantle – Geochemical constraints. *Geochimica et Cosmochimica Acta* 58: 4717–4738.

McCulloch MT and Wasserburg GJ (1978) Sm-Nd and Rb-Sr chronology of continental crust formation. *Science* 200: 1003–1011.

McGeary S, Nur A, and Benavraham Z (1985) Spatial gaps in arc volcanism – The effect of collision or subduction of oceanic plateaus. *Tectonophysics* 119: 195–221.

McKenzie D and Bickle MJ (1988) The volume and composition of melt generated by extension of the lithosphere. *Journal of Petrology* 29: 625–679.

McKenzie D and O'Nions RK (1983) Mantle reservoirs and ocean island basalts. *Nature* 301: 229–231.

McLennan SM (1988) Recycling of the continental-crust. *Pure and Applied Geophysics* 128: 683–724.

Mezger K (1992) Temporal evolution of regional granulite terranes: Implications for the evolution of the lowermost crust. In: Fountain DM, Arculus R, and Kay RV (eds.) *Continental Lower Crust*, pp. 447–478. New York: Elsevier Science.

Monger JWH (1993) Canadian cordilleran tectonics: From geosynclines to crustal collage. *Canadian Journal Earth Science* 30: 209–231.

Moorbath S (1975) Evolution of Precambrian crust from strontium isotopic evidence. *Nature* 254: 395–398.

Moorbath S (1978) Age and isotope evidence for evolution of continental crust. *Philosophical Transactions of the Royal Society of London Series A-Mathematical Physical and Engineering Sciences* 288: 401–112.

Moorbath S, Gale NH, Pankhurs RJ, Mcgregor VR, and Onions RK (1972) Further rubidium–strontium age determinations on very Early Precambrian rocks of Godthaab District, west Greenland. *Nature-Physical Science* 240: 78–82.

Moorbath S, Whitehouse MJ, and Kamber BS (1997) Extreme Nd-isotope heterogeneity in the early Archaean – Fact or fiction? Case histories from northern Canada and west Greenland. *Chemical Geology* 135: 213–231.

Mushkin A, Navon O, Halicz L, Hartmann G, and Stein M (2003) The petrogenesis of A-type magmas from the Amram Massif, southern Israel. *Journal of Petrology* 44: 815–832.

Neal CR, Mahoney JJ, Kroenke LW, Duncan RA, and Petterson MG (1997) The Ontong Java Plateau. In: Mahoney JJ and Coffin M (eds.) *American Geophysical Union Monograph : On Large Igneous Provinces*, pp. 183–216. Washington, DC: American Geophysical Union.

Nelson BK and DePaolo DJ (1984) 1,700-Myr Greenstone volcanic successions in Southwestern North-America and isotopic evolution of proterozoic mantle. *Nature* 312: 143–146.

Nelson BK and DePaolo DJ (1985) Rapid production of continental-crust 1.7 to 1.9B.Y. Ago – Nd isotopic evidence from the basement of the North-American mid-continent. *Geological Society of America Bulletin* 96: 746–754.

Nur A and Ben-Avraham Z (1977) Lost Pacifica continent. *Nature* 270: 41–43.

Nur A and Ben-Avraham Z (1978) Speculations on mountain building and the lost Pacifica continent. *Journal of Physucs of the Earth Supplement* 26: S21–S37.

Nur A and Ben-Avraham Z (1981) Volcanic gaps and the consumption of aseismic ridges in South America. *Memoirs of the Geological Society of America* 154: 729–740.

Nur A and Ben-Avraham Z (1982) Displaced terranes and mountain building. In: Hsu KJ (ed.) *Mountain Building Processes*, pp. 73–874. London: Academic Press.

Nye CJ and Reid MR (1986) Geochemistry of primary and least fractionated lavas from Okmok Volcano, central Aleutians – Implications for arc magmagenesis. *Journal of Geophysical Research-Solid Earth and Planets* 91: 271–287.

O'Nions RK, Evensen NM, and Hamilton PJ (1979) Geochemical modeling of mantle differentiation and crustal growth. *Journal of Geophysical Research* 84: 6091–610.

Oxburgh ER and Parmentier EM (1977) Compositional density stratification in the oceanic lithosphere-causes and consequences. *Journal of the Geological Society of London* 133: 343–355.

Patchett PJ (1996) Crustal growth – Scum of the Earth after all. *Nature* 382: 758–759.

Patchett JP and Arndt NT (1986) Nd isotopes and tectonics of 1.9-1.7 Ga crustal genesis. *Earth and Planetary Science Letters* 78: 329–338.

Patchett PJ and Bridgwater D (1984) Origin of continental-crust of 1.9-1.7 Ga age defined by Nd isotopes in the Ketilidian Terrain of south Greenland. *Contributions to Mineralogy and Petrology* 87: 311–318.

Patchett PJ and Chase CG (2002) Role of transform continental margins in major crustal growth episodes. *Geology* 30: 39–4.

Patchett PJ and Samson SD (2003) *Treatise on Geochemistry: Ages and Growth of the Continental Crust from Radiogenic Isotopes*. Amsterdam: Elsevier.

Patchett PJ, White WM, Feldmann H, Kielinczuk S, and Hofmann AW (1984) Hafnium rare-Earth element fractionation in the sedimentary system and crustal recycling into the Earth's mantle. *Earth and Planetary Science Letters* 69: 365–78.

Pearcy LG, Debari SM, and Sleep NH (1990) Mass balance calculations for two sections of island arc crust and implications for the formation of continents. *Earth and Planetary Science Letters* 96: 427–442.

Percival JA and Williams HR (1989) Late Archean Quetico accretionary complex, Superior Province. *Canada Geology* 17: 23–25.

Pilger RH (1978) Method for finite plate reconstructions, with applications to Pacific-Nazca plate evolution. *Geophysical Research Letters* 5: 469–472.

Plank T (2005) Constraints from thorium/lanthanum on sediment recycling at subduction zones and the evolution of the continents. *Journal of Petrology* 46: 921–944.

Plank T and Langmuir CH (1998) The chemical composition of subducting sediment and its consequences for the crust and mantle. *Chemical Geology* 145: 325–394.

Polat A and Kerrich R (2001a) Geodynamic processes, continental growth, and mantle evolution recorded in late Archean greenstone belts of the southern Superior Province, Canada. *Precambrian Research* 112: 5–25.

Polat A and Kerrich R (2001b) Magnesian andesites, Nb-enriched basalt-andesites, and adakites from late-Archean 2.7 Ga Wawa greenstone belts, Superior Province, Canada: Implications for late Archean subduction zone petrogenetic processes. *Contributions to Mineralogy and Petrology* 141: 36–52.

Polat A, Kerrich R, and Wyman DA (1998) The late Archean Schreiber-Hemlo and White River Dayohessarah greenstone belts, Superior Province: Collages of oceanic plateaus, oceanic arcs, and subduction-accretion complexes. *Tectonophysics* 289: 295–326.

Puchtel IS, Arndt NT, Hofmann AW, *et al.* (1998) Petrology of mafic lavas within the Onega plateau, central Karelia: Evidence for 2.0 Ga plume-related continental crustal growth in the Baltic Shield. *Contributions to Mineralogy and Petrology* 130: 134–153.

Puchtel IS, Haase KM, Hofmann AW, *et al.* (1997) Petrology and geochemistry of crustally contaminated komatiitic basalts from the Vetreny Belt, southeastern Baltic Shield: Evidence for an early Proterozoic mantle plume beneath rifted Archean continental lithosphere. *Geochimica et Cosmochimica Acta* 61: 1205–1222.

Puchtel IS, Hofmann AW, Amelin YV, Garbe-Schonberg CD, Samsonov AV, and Schipansky AA (1999) Combined mantle plume-island arc model for the formation of the 2.9 Ga Sumozero-Kenozero greenstone belt, SE Baltic Shield: Isotope and trace element constraints. *Geochimica et Cosmochimica Acta* 63: 3579–3595.

Puchtel IS, Hofmann AW, Mezger K, Jochum KP, Shchipansky AA, and Samsonov AV (1998) Oceanic plateau model for continental crustal growth in the archaean, a case study from the Kostomuksha greenstone belt, NW Baltic Shield. *Earth and Planetary Science Letters* 155: 57–74.

Reischmann T, Kroner A, and Basahel A (1983) Petrography, geochemistry and tectonic setting of metavolcanic sequences from the Al Lith area, southwestern Arabian Shield. *Bulletin of Faculty of Earth Sciences, King Abdulaziz University* 6: 365–378.

Rey P (1993) Seismic and tectonometamorphic characters of the lower continental-crust in Phanerozoic areas – A consequence of post-thickening extension. *Tectonics* 12: 580–590.

Reymer A and Schubert G (1984) Phanerozoic addition rates to the continental-crust and crustal growth. *Tectonics* 3: 63–77.

Reymer A and Schubert G (1986) Rapid growth of some major segemtns of continental crust. *Geology* 14: 299–302.

Richards MA, Duncan RA, and Courtillot VE (1989) Flood basalts and hot-spot tracks – Plume heads and tails. *Science* 246: 103–107.

Richardson WP, Emile A, and Van der Lee S (2000) Rayleigh – wave tomography of the Ontong – Java Plateau. *Physics of the Earth and Planetary Interiors* 118: 29–51.

Rocholl A, Stein M, Molzahn M, Hart SR, and Worner G (1995) Geochemical evolution of rift magmas by progressive tapping of a stratified mantle source beneath the Ross Sea Rift, Northern Victoria Land, Antarctica. *Earth and Planetary Science Letters* 131(3): 207–224.

Rudnick RL (1992) Xenoliths—samples of the lower continental crust. In: Fountain DM, Arculus R, and Kay RW (eds.) *Continental Lower Crust*, pp. 269–316. Amsterdam: Elsevier.

Rudnick RL (1995) Making continental-crust. *Nature* 378: 571–578.
Rudnick RL and Taylor R (1987) The composition and petrogenesis of the lower continental crust: A xenolith study. *Journal of Geopysical Research* 92: 13981–14005.
Rudnick RL and Fountain DM (1995) Nature and composition of the continental crust: A lower crustal prespective. *Reviews of Geophysics* 33: 267–.310.
Rudnick RL and Gao S (2003) *Treatise on Geochemistry: Composition of the Continental Crust*. Amsterdam: Elsevier.
Samson SD and Patchett PJ (1991) The Canadian Cordillera as a modern analogue of Proterozoic crustal growth. *Australian Journal of Earth Sciences* 38: 595–611.
Samson SD, McClelland WC, Patchett PJ, Gehrels GE, and Anderson RG (1989) Evidence from neodymium isotopes for mantle contributions to Phanerozoic crustal genesis in the Canadian Cordillera. *Nature* 337: 705–709.
Samson SD, Patchett PJ, Gehrels GE, and Anderson RG (1990) Nd and Sr isotopic characterization of the Wrangellia terrane and implications for crustal growth of the Canadian cordillera. *Journal of Geology* 98: 749–762.
Samson SD, Patchett PJ, McClelland WC, and Gehrels GE (1991) Nd isotopic characterization of metamorphic rocks in the coast mountains, Alaskan and Canadian cordillera: Ancient crust bounded by juvenile terranes. *Tectonics* 10: 770–780.
Schubert G and Reymer APS (1985) Continental volume and freeboard through geological time. *Nature* 316: 336–339.
Schubert G and Sandwell D (1989) Crustal volumes of the continents and of oceanic and continental submarine plateaus. *Earth and Planetary Science Letters* 92: 234–246.
Shirey SB and Hanson GN (1984) Mantle-derived Archean Monozodiorites and Trachyandesites. *Nature* 310: 222–224.
Sleep NH and Windley BF (1982) Archean plate tectonics constraints and inferences. *Journal of Geology* 90: 363–379.
Smith HS and Erlank AJ (1982) Geochemistry and petrogenesis of komatiites from the Barberton greenstone belt. In: Arndt NT and Nisbet EG (eds.) *Komatiites*, pp. 347–398. London: Allen and Unwin.
Stein M (2003) Tracing the plume material in the Arabian–Nubian shield. *Precambrian Research* 123: 223–234.
Stein M (2006) The rise of the asthenopsheric mantle into the Arabian lithosphere and its cosnsequences: Heating , melting uplifting. IAVCEI meeting, Guangzhou.
Stein M and Hofmann AW (1992) Fossil plume head beneath the Arabian lithosphere? *Earth and Planetary Science Letters* 114: 193–209.
Stein M and Hofmann AW (1994) Mantle plumes and episodic crustal growth. *Nature* 372: 63–68.
Stein M and Goldstein SL (1996) From plume head to continental lithosphere in the Arabian–Nubian shield. *Nature* 382: 773–778.
Stein M, Navon O, and Kessel R (1997) Chromatographic metasomatism of the Arabian–Nubian lithosphere. *Earth and Planetary Science Letters* 152: 75–91.
Stern CR and Kilian R (1996) Role of the subducted slab, mantle wedge and continental crust in the generation of adakites from the Andean Austral Volcanic Zone. *Contributions to Mineralogy and Petrology* 123: 263–281.
Stern RA, Syme EC, and Lucas SB (1995) Geochemistry of 1.9 Ga morb-like and oib-like basalts from the Amisk Collage, Flin-Flon Belt, Canada – Evidence for an intraoceanic origin. *Geochimica et Cosmochimica Acta* 59: 3131–3154.
Stern RJ (1981) Petrogenesis and tectonic setting of late Precambrian ensimatic volcanic rocks, central eastern desert of Egypt. *Precambrian Research* 16: 195–230.
Sylvester PJ (2000) Continent formation, growth and recycling – Preface. *Tectonophysics* 322: Vii–Viii.
Sylvester PJ, Campbell IH, and Bowyer DA (1997) Niobium/uranium evidence for early formation of the continental crust. *Science* 275: 521–523.
Taylor SR (1967) The origin and growth of continents. *Tectonophysics* 4: 17–34.
Taylor SR and McLennan SM (1985) *The Continental Crust: Its Composition and Its Evolution*. Oxford: Blackwell.
Taylor SR and McLennan SM (1995) The geochemical evolution of the continental-crust. *Reviews of Geophysics* 33: 241–265.
Taylor SR and McLennan SM (1996) The evolution of continental crust. *Scientific American* 274: 76–81.
Tera F, Brown L, Morris J, Selwayn Sacks I, Klein J, and Middleton R (1986) Sediment incorporation in island-arc magmas: Inferences from 10Be. *Geochimica et Cosmochimica Acta* 50: 535–550.
Veizer J and Jansen Sl (1979) Basement and sedimentary recycling and continental evolution. *Journal of Geology* 87: 341–370.
Weinstein Y, Navon O, Alther R, and Stein M (2006) The role of fluids and of pyroxenitic veins in the generation of alkali-basaltic suites, northern Arabian plate. *Journal of Petrology* 47: 1017–1050.
Wilde SA, Valley JW, Peck WH, and Graham CM (2001) Evidence from detrital zircons for the existence of continental crust and oceans on the Earth 4.4 Gyr ago. *Nature* 409: 175–178.
Windley BF (1995) *The Evolving Continents*. Chichester: Wiley.
White WM and Patchett PJ (1984) Hf–Nd–Sr isotopes and incompatible-element abundances in island arcs; implications for magma origins and crust–mantle evolution. *Earth Planetary Science Letters* 67: 167–185.
Wise (1974) Continental margins, freeboard and volumes of continents and oceans through time. In: Burke CA and Drake CL (eds.) *The Geology of Continental Margins*, pp. 45–58. Berlin: Springer.
Wörner G, Zindler A, Staudigel H, and Schmincke HU (1986) Sr, Nd, and Pb isotope geochemistry of Tertiary and Quarternary alkaline volcanics from West Germany. *Earth Planetary Science Letters* 79: 107–119.