

Geochemical Perspectives



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NICHOLAS T. ARNDT

Formation and Evolution of the Continental Crust



Each issue of *Geochemical Perspectives* presents a single article with an in-depth view on the past, present and future of a field of geochemistry, seen through the eyes of highly respected members of our community. The articles combine research and history of the field's development and the scientist's opinions about future directions. We welcome personal glimpses into the author's scientific life, how ideas were generated and pitfalls along the way. *Perspectives* articles are intended to appeal to the entire geochemical community, not only to experts. They are not reviews or monographs; they go beyond the current state of the art, providing opinions about future directions and impact in the field.

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ABOUT THE COVER

The cover shows the Poulabrone dolmen, portal tomb, near Burren in Ireland. It is composed of granitic gneiss from the continental crust.

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PREFACE

In May of this year I asked myself what I would do during the summer. I had finished several major projects, the ICDP drilling program in the Barberton belt was at the stage where the core was being worked on by groups throughout the world and my yearly trip to South Africa was not needed. I was at a loose end. I had been becoming irritated by numerous references in the literature to intracrustal models for the formation of Archean granitoids, and perplexed by the recent swing to “preservationist” models that linked peaks in zircon ages to the supercontinent cycle. So when Tim Elliott suggested that I write a *Geochemical Perspectives* article on the subject I grabbed the chance.

The text that follows was written quickly from late June to the end of July, a pace that left me too little time to explore in detail many of the subjects that I discuss. I was not able to develop in a rigorous manner some of the ideas that came up during the writing – these are discussed at the end of the article. Nor was I able to produce new figures or tables and was obliged instead to borrow extensively from published sources. But what I was able to do was to outline a broad picture of the formation of the continental crust and the many intriguing and stimulating models, interpretations, arguments and debates that surround the subject.

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THE FORMATION AND EVOLUTION OF THE CONTINENTAL CRUST

ABSTRACT

In the Archean, like now, the granitoids that constitute the core of the continental crust formed in subduction zones. Hydrous basaltic magmas from the mantle wedge rose to the base of the crust where they fractionally crystallised or remelted underplated rocks to yield more evolved granitic magmas. Alternative models to explain Archean granitoids, which call on melting in intraplate settings such as the bases of oceanic plateaus, are implausible because such settings lack the water that is essential to form voluminous granitic melt.

From the end of the Archean to the late Proterozoic, the continental crust grew in a series of major pulses, each triggered by accelerated mantle convection. The arrival of large mantle plumes displaced material from the upper mantle, accelerating the rate of subduction and causing a pulse of crustal growth.

The Hadean crust was mafic and it underwent internal partial melting to produce the granitic melts that crystallised the Jack Hills zircons. This crust was disrupted by the Late Heavy Bombardment and from then on, since about 3.9 Ga, plate tectonics has operated.



We were drinking Bitburger in a pub in the Mainz Altstadt (Fig. 1.1) when *Jon Patchett* said “the formation of the continental crust was the most interesting thing that ever happened to the Earth”. At first we were dubious, but after some argument, and a few more beers, he convinced us of his point of view. The segregation of the core had a greater effect on the overall structure of the planet, but this happened very early on and has had little subsequent impact on the evolution of the planet. Oxygenation of the atmosphere and the emergence of life were clearly very important, but at that time, in the mid 1980’s, we really understood very little of these processes. *Jon* argued that without the continental crust, we would not have been sitting in that pub: the Earth would have been a drab waterworld enveloped by a stagnant basaltic crust and covered by a global ocean. Then the discussion turned to the *Armstrong* model.

I first heard of the polemic surrounding the formation of continental crust in 1973 when a friend told me that *Richard Armstrong* had quit Yale to take up a position at the University of British Columbia. The move apparently was triggered by Yale’s dismissal of *Armstrong*’s idea that most of the Earth’s continental crust had formed very early in Earth history. According to *Armstrong*, after an initial exuberant pulse of crustal growth, segregation of new crust was balanced by cycling of older crust back into the mantle. These ideas ran counter to prevailing opinion, which held that very little continental crust had formed before about 3.8 Ga – the age of the oldest rocks known at that time – and that the crust had grown progressively through Earth history. In our pub in the Mainz Altstadt, the *Armstrong* model received only lukewarm support from the geochemists who made up most of our group from the Max Planck Institut für Geochimie; they preferred the continuous-growth model, which holds that the continental crust has continued to form through all of earth history.

Next we discussed *Steve Moorbath*’s CADS. We knew that the ages of Archean rocks are not distributed uniformly through time but grouped in a small number of distinct peaks, at 2.7, 2.1 and 1.85 Ga, for example. In the 1980’s, enough reliable radiometric ages had appeared to show that these peaks were not an artefact of incomplete or biased sampling. To describe them, *Steve Moorbath*, a geochemist from Oxford, had coined the expression Crustal Accretion-Differentiation Super-event (which fortunately never stuck). At that time we all thought that the peaks marked periods of accelerated growth of the continental crust, an idea that was central to a paper published a few years later by *Moti Stein* and *Al Hofmann* (1994). In the past 3-4 years, *Chris Hawkesworth* and colleagues (e.g., *Hawkesworth et al.*, 2009; *Hawkesworth et al.*, 2010; *Condie et al.*, 2011) have come up with another interpretation. They link *Moorbath*’s CADS to the assembly of supercontinents and interpret them as the times of enhanced preservation of



the continental crust, and not to periods of accelerated generation of crust from its mantle source. The issue is still hotly debated and it will become one of the principal subjects of the present *Perspectives*.



Photo courtesy of Oliver Barchewitz.

Figure 1.1 A pub in the Mainz Altstadt.

Underpinning our entire discussion was the question of how the continental crust formed. Then, as now, we did not know if plate tectonics operated during the earliest part of Earth history, and, then as now, opinion was split as to whether the granites that make up the bulk of the crust formed in a subduction setting like modern granitoids, or through a process peculiar to the Archean and the early Proterozoic.

These questions form the basis of my contribution. I will start with an introductory section summarising current thinking about how granitic magma is generated, this being the crux of the process that builds the continental crust. Then, following a brief historical review, I discuss the experimental constraints on the conditions in which granitic melts are produced before discussing some of the broader issues related to the rate and mechanism of crustal growth. I review and then dismiss models that hold that Archean granitoids are generated in an intraplate setting and develop a model in which these rocks form in a subduction setting, much as those of the present day. In the final sections I address the question of episodic growth of the continental crust before summarising the information pertaining to the nature and origin of the first felsic crust of our planet.



2. GRANITOIDS-THEIR MINERALOGY, GEOCHEMISTRY AND OCCURRENCE

The continental crust is commonly described as having a granitic composition. This emphasises the contrast with the mafic rocks of the oceanic crust and ultra-mafic nature of the mantle, from which both the continental crust ultimately derives. However, there are important subtleties in the mineralogy and composition of silica rich rocks that have potential implications for their derivation and I will therefore start with the classification of silica-rich, 'granitic' (*sensu lato*) rocks, as more properly termed, granitoids.

2.1 Classification

2.1.1 Mineralogy and major elements

The term granitoid applies to the suite of felsic plutonic rocks whose compositions are summarised in Figure 2.1. The dominant minerals are quartz and various types of feldspar, the proportions of which define the rock type. Plagioclase predominates in *tonalite*, alkali feldspar in *granite* (*sensu stricto*) and both types are present in *granodiorite*. Trondjemite is a variety of tonalite that is unusually quartz rich and contains sodic, not calcic plagioclase. The proportion of mafic minerals varies from 20-40% in granodiorite and tonalite to only a few percent in granite. Pyroxene is rare, the most common mafic minerals being hydrated species such as hornblende and biotite.

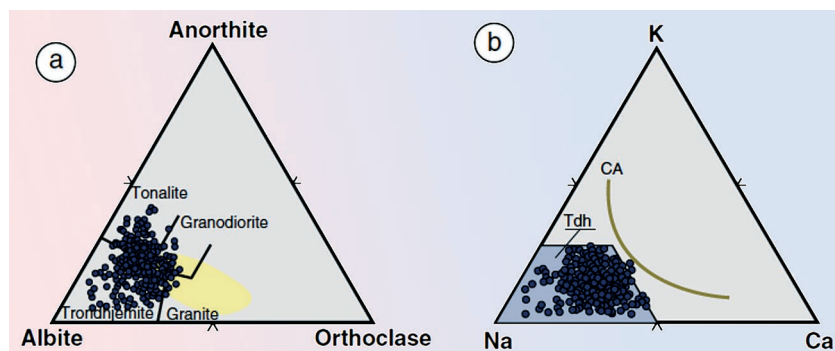


Figure 2.1 Mineralogical and chemical classification of granitoids. The dark blue dots represent the compositions of Archean TTG which have high Na and low K contents. CA – calc-alkaline trend; Tdh – trondjemite (from Moyen and Martin, 2012, with permission from Elsevier).



The mineralogy reflects the major-element compositions. Silica contents are high (60 to 75%), as are K_2O contents, while elements such as TiO_2 , FeO , MgO and CaO , which are removed during the crystallisation of mafic minerals or calcic plagioclase, are low. The relative proportions of Al_2O_3 vs. CaO , Na_2O and K_2O form the basis of a classification of granite types. In peralkaline granites, the molecular proportions of Al_2O_3 are greater than $(Na_2O + K_2O)$; in metaluminous granites $Al_2O_3 > (CaO + Na_2O + K_2O)$ but $Al_2O_3 > (Na_2O + K_2O)$; and in peraluminous granites, $Al_2O_3 > (CaO + Na_2O + K_2O)$. These differences in element ratios impart distinctive mineralogies such as alkali pyroxene and amphibole in peralkaline granites and aluminous minerals such as muscovite, cordierite, garnet and rarely corundum in peraluminous granites.

2.1.2 I and S type granites

Table 2.1 and Figure 2.2 summarise the well-known classification of granitoids based on their field and geochemical characteristics and the information these characteristics provide as to their origin. More complete discussions of the classification of granitoids and alternative classification schemes are given by Barbarin (1999), Frost *et al.* (2001), and Moyen and Martin (2012). I concentrate here on the Chappell and White scheme in part because it emphasises the contrasting nature of the sources of granitoids but mainly, I confess, because of my association with both authors during my years at the Australian National University. This alphabetical scheme has similarities with the mineralogically defined classification (above) but includes the use of isotopes and infers an origin for the granites.

Initially Chappell and White (1992, 2001) identified just two types of granite; S-types, where the “S” stands for “sedimentary”, and I-types, where the “I” stands for “igneous”. The former are strongly peraluminous, relatively potassic and have moderately high silica contents (64–77 wt % SiO_2). They contain low CaO , Na_2O and Sr contents, elements that are lost when feldspar alters to clay minerals. Their sedimentary source imparts a distinctive mineralogy – the presence of aluminous phases such as muscovite, cordierite and garnet. Isotope compositions can be extreme (high $^{87}Sr/^{86}Sr$, low $^{143}Nd/^{144}Nd$) when the source is Precambrian paragneiss. S-type granitoids normally intrude in relatively small volumes, particularly in active margins and continental collision zones.

The Granites of SE Australia – During my first research project in the final year of my bachelor’s degree at the Australian National University in Canberra, I mapped a series of granitoids and country rocks in hilly country near the NSW-Victoria border. Very often I collapsed exhausted at the summits of even the smallest hills – too much for a town boy whose sporting activities to that time had been billiards and guitar playing (see Canberra Blues <http://www.youtube.com/watch?v=ykzSQeCfsjY>). I remember the first nights in the field, alone in my tent, asking what the heck was I doing there. The same happened during PhD fieldwork in the Abitibi belt in Canada, alone in a tent with only a Yamaha trail bike for company. Communing with nature, then as now, is not my thing.



During my mapping and subsequent petrographic and geochemical analysis of the granites in my field area, I realised there were two types, a two-mica granite full of metasedimentary enclaves, and a hornblende granite containing enclaves of diorite and amphibolite. Later that year, 1969, I learnt that *Allan White*, the co-supervisor of my project, and his colleague *Bruce Chappell*, had recognised the same types throughout the Lachlan Fold belt and were developing what has now come to be known as the S- and I-type classification.

I-type granites are metaluminous to weakly peraluminous, relatively sodic, and have a wide range of silica contents, from 56 to 77 % SiO₂. They typically contain hornblende and biotite. In the original definition of *Chappell* and *White*, enclaves or xenoliths of diorite and gabbro or amphibolite are emphasised. I-type granites are emplaced in large volumes in mature island arcs and convergent margins, as well as within Precambrian granite-greenstone terrains. These rocks are the building blocks of the continental crust and most hypotheses for the origin of granitic magma apply specifically to them.

Table 2.1 Classification of granite types

	I-type	S-type	M-type	A-type
	Igneous	Sedimentary	Mantle	Anorogenic or anhydrous
Mineralogy and field characteristics	Amphibole and biotite; enclaves of diorite and gabbro	Biotite and muscovite, sometimes with cordierite and garnet; metasedimentary enclaves	Biotite and plagioclase, little to no alkali feldspar	Alkali pyroxene and amphibole
Geochemistry	Metaluminous to weakly peraluminous, relatively sodic, wide range of silica contents. Moderate ⁸⁷ Sr/ ⁸⁶ Sr, ¹⁴³ Nd/ ¹⁴⁴ Nd, d ¹⁸ O.	Peraluminous, potassic, high silica, low CaO, Na ₂ O and Sr. High ⁸⁷ Sr/ ⁸⁶ Sr, low ¹⁴³ Nd/ ¹⁴⁴ Nd, high d ¹⁸ O. Relatively oxidised.	Metaluminous. Moderate ⁸⁷ Sr/ ⁸⁶ Sr, ¹⁴³ Nd/ ¹⁴⁴ Nd, mantle-like d ¹⁸ O.	Peralkaline or calc-alkaline. High alkalis, moderate to high silica. ⁸⁷ Sr/ ⁸⁶ Sr, ¹⁴³ Nd/ ¹⁴⁴ Nd, d ¹⁸ O.
Origin	From (meta)igneous source rocks, typically basaltic	From metasedimentary source rocks	From the mantle, or from crystallisation of basaltic magma	Intruded in intraplate setting after orogenesis



A-type granitoids are not associated with orogenesis – the “A” stands for “anorogenic”, “alkaline” or “anhydrous”. They are rarely deformed and are thought to intrude after the main orogenic event. They are peralkaline and commonly contain pyroxene in the place of hydrous amphibole and biotite. M-type granites, where the “M” stands for “mantle”, are said to be produced by partial melting of mantle peridotite or of “juvenile” basaltic crust. A- and M-types are not common and do not play a major role in the formation of the continental crust. They are discussed no further in this contribution.

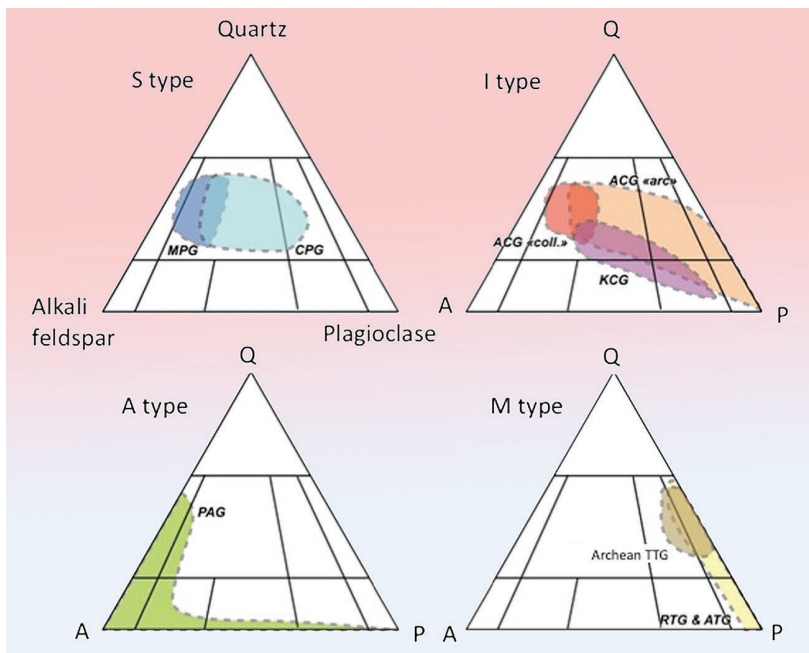


Figure 2.2 Mineralogy basis of the classification into S, I, A, and M type granites. MPG is muscovite peraluminous granite, CPG is cordierite peraluminous granite, ACG is amphibole calc-alkaline granite, KCG is K-rich calc-alkaline granite, PAG is peralkaline and alkaline granite, RTG is (midocean) ridge tholeiitic granite and ATG is arc tholeiitic granite.

2.1.3 Archean granitoids

The term TTG, for tonalite-trondhjemite-granodiorite, is commonly applied to the granitic rocks that are the dominant component of most Archean cratons, giving a certain mystique to this assemblage of rock types. The origin of the term TTG is variously attributed to Jahn *et al.* (1981) or to Barker and Arth (1976, 1979). These rocks have relatively sodic compositions, and normally lack the more

potassic varieties such as granite *sensu stricto*. Trondjhemite, the sodic, silica-rich granitoid, is an important member of the series that plots in a field close to the Na apex in Figure 2.1b and off the normal calc-alkaline trend.

2.2 Trace Elements

As is to be expected in evolved, felsic rocks, the concentrations of incompatible elements (those that are not accommodated in mafic minerals) are high. This is best seen in mantle-normalised trace element diagrams like Figure 2.3, which shows that the abundances of Rb, Ba and the light rare earth elements (LREE) are 50 to 500 times higher than in the primitive mantle whereas the heavy rare earth elements (HREE) are far less enriched. Distinctive positive or negative anomalies are present in all rocks from the continental crust; Pb is almost always present in excess, compared with neighbouring elements, and Nb-Ta and Ti are deficient. These features are commonly ascribed to the presence in rutile at the site of melting (*e.g.*, McCulloch and Gamble, 1991). Deficits in other elements

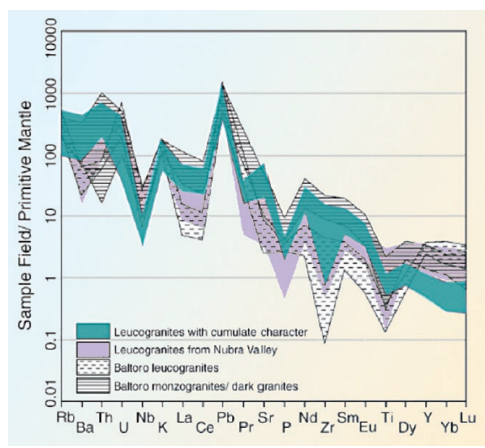


Figure 2.3

Trace elements, normalised to primitive mantle, for granitoids from the Himalayas (modified from Reichardt *et al.*, 2010). This figure, one of many that I could have chosen, illustrates the characteristic features of most granitoids – strongly enriched highly incompatible elements and low concentrations of the more compatible elements; a marked positive Pb anomaly and negative anomalies of Nb, P, Zr and Ti.

can be related to segregation during fractional crystallisation of specific minerals: Sr in feldspar, P in apatite, and Ti (in part) in Fe oxides. A low ratio of Nb to Ta is the signature of amphibole fractionation (Foley *et al.*, 2002). Or alternatively, these features could have been inherited, in whole or in part, from the source of the granitic magmas.

The HREE are of particular importance because their concentrations are unusually low in both Archean TTG and in modern adakites. The latter term is applied to modern volcanic rocks and granitoids of various ages that share distinctive geochemical features such as a low Yb content or high La/Yb, and relatively high Sr contents. These features point to the presence of garnet in the residue of melting and an absence of plagioclase, features



that point to melting at relatively great depths. The distinctive geochemical features of adakites are commonly explained by partial melting of basalt in subducting oceanic crust (Defant and Drummond, 1990).

The depletion of the HREE in Archean TTG is shown in a diagram made popular by *Hervé Martin* (Fig. 2.4) in which $(La/Yb)_N$ (where the “N” signals that the ratio is normalised to primitive mantle concentrations) is plotted against Yb_N : the HREE depletion is manifested as high La/Yb and low Yb . The HREE are compatible in garnet and the presence of this phase in the residue of melting provides a ready explanation for this distinctive geochemical feature. The Sr-Y ratio and Y concentration are commonly used in conjunction with La/Yb and Yb : modern granitoids generally do not show the depletion of the HREE and have relatively low Sr/Y and high Y, features that are ascribed to the presence of plagioclase (in which Sr is compatible) in the solid residue of melting.

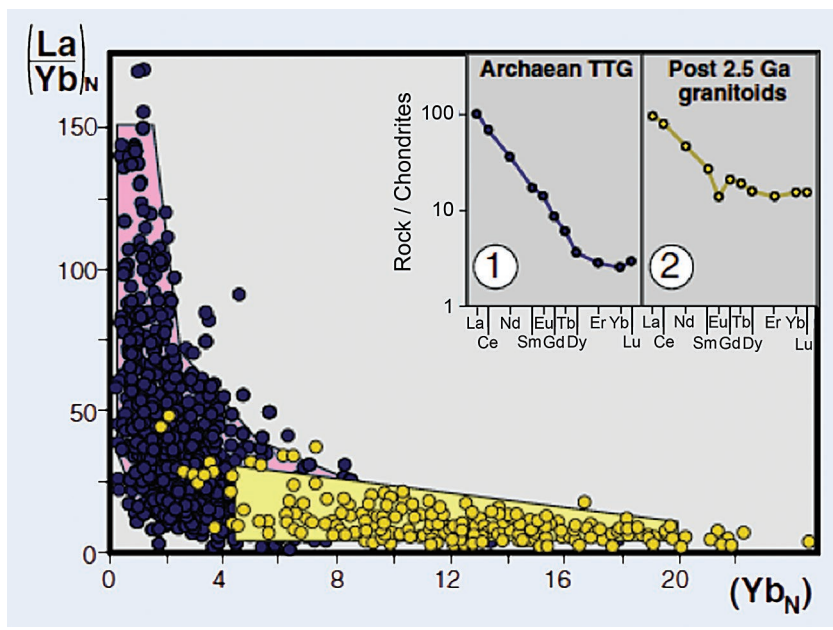


Figure 2.4

The La/Yb vs Yb diagram used by *Hervé Martin* (Martin, 1987, 1994a; Martin and Moyen, 2002; Moyen and Martin, 2012) to distinguish Archean TTG from modern granitoids. The subscript N indicates that the ratio or concentration is normalised to primitive mantle. The insets 1 and 2 illustrate the contrasting REE patterns of the two types of granitoid (from Moyen and Martin, 2012, with permission from Elsevier).



2.3 Isotopic Compositions

The isotopic compositions of granitoids span a wide range of values, from close to those of the mantle to values like those expected in old continental crust, as illustrated in the $\epsilon_{\text{Nd}(T)}$ vs $^{87}\text{Sr}/^{86}\text{Sr}$ diagram (Fig. 2.5). Magmas from the depleted mantle (DM) have high Rb/Sr and low Sm/Nd and they plot in the top left quadrant; magmas from a source in the continental crust have low Rb/Sr and high Sm/Nd and, if this crust is old, these magmas plot in the lower right quadrant. Figure 2.5 illustrates these differences: the I-type granites (“hornblende granites” in the figure) have $\epsilon_{\text{Nd}(T)}$ values ranging from +4 to about -9 and correspondingly low to moderate $^{87}\text{Sr}/^{86}\text{Sr}$ values. These values indicate a mixed source comprising juvenile material from the mantle and a small proportion of recycled crustal component. The composition identified as “DM” for depleted mantle, taken from McCulloch and Chappell (1982), has an $\epsilon_{\text{Nd}(T)}$ of about +6 and a $^{87}\text{Sr}/^{86}\text{Sr}$ of about 0.703, values that are significantly lower or higher, respectively, than those of the upper mantle source of mid-ocean ridge basalts. As shown in Figure 2.6, these values plot in the upper left part of the field occupied by lavas from island arcs, whose $\epsilon_{\text{Nd}(T)}$ extends from +8 to -10 and whose $^{87}\text{Sr}/^{86}\text{Sr}$ extend to 0.710. Figure 2.7 is a compilation of Nd and Hf isotopic data and illustrates the wide range in both isotopic systems. Although the assimilation of older crust may have influenced the composition of the more extreme samples, they can be used to infer the compositional range of the mantle source of these rocks. This source lies in the mantle wedge and contains a component derived from subducted sediment or oceanic basalt, but it nonetheless represents the composition of the source of magmas parental to the continental crust. As will be seen in later sections, this range is far larger than those normally assumed for the mantle source of crustal rocks.

The S-type granites plot in the crustal field in the bottom right of Figure 2.5, as is to be expected of magmas from a sedimentary source. The presence of the crustal component is manifested more clearly by zircon xenocrysts, which are common in these rocks (e.g., Kemp *et al.*, 2006). The mixing line in the figure shows the calculated proportions of juvenile and crustal components in the samples. Those plotting near DM have relatively low concentrations of older crust; these values would be still lower if values for the mantle source like those of island arc lavas were adopted.

Most Archean granitoids have been subjected to metamorphism and deformation after their crystallisation and because Rb and Sr are relatively mobile during alteration, the Rb-Sr isotopic method only rarely gives reliable results for rocks of this age. The Sm-Nd and Lu-Hf systems are more robust and give more useful data. A further advantage of these systems, which comprise refractory lithophile elements, is that the bulk Earth composition is well defined from a meteorite reference. Thus, unlike the Sr isotope system, Nd and Hf isotope ratios can be expressed in the epsilon notation and the composition of granitoids



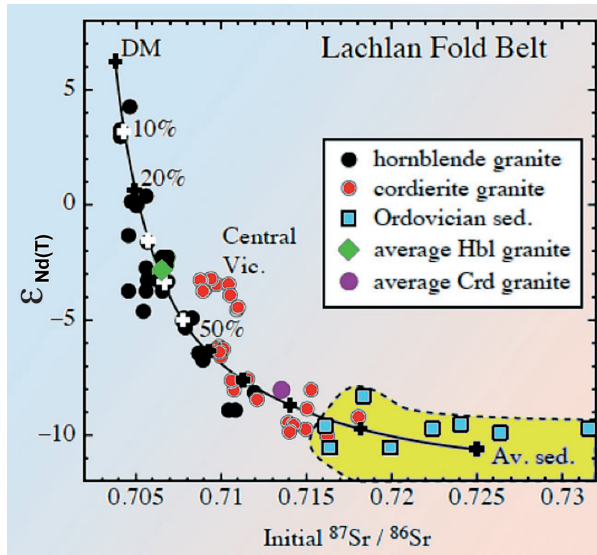


Figure 2.5 Isotope data for granites from the Lachlan Belt. DM stands for depleted mantle (modified from Kemp and Hawkesworth, 2004).

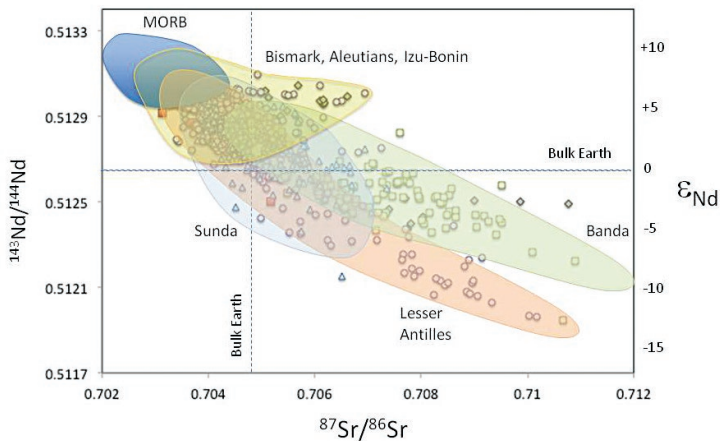


Figure 2.6 Nd and Sr isotopic compositions of lavas from island arcs. These provide an indication of the composition of the part of the mantle that melts to produce the magmas parental to the continental crust. The $\epsilon_{Nd(T)}$ ranges from about +10 to highly negative values and are very different from that of the depleted mantle source of mid-ocean ridge basalts (MORB) (figure redrawn based on data from Georoc).



relative to a bulk Earth composition is readily seen. Figure 2.8 shows how the initial Nd isotopic composition of granitoids, expressed as $\epsilon_{\text{Nd}(T)}$, varies through time. The compositions are distributed on both sides of the bulk Earth reference ($\epsilon_{\text{Nd}(T)} = 0$) with the dispersion increasing with decreasing age.

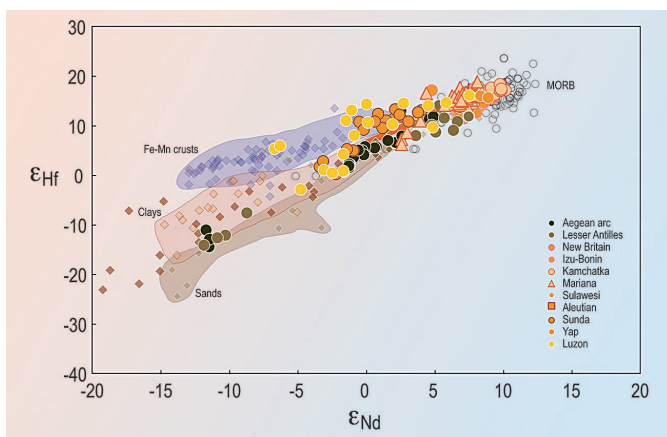


Figure 2.7 Nd and Hf isotopic compositions of lavas from some arcs (modified from Chauvel *et al.*, 2009).

The Hf isotopic compositions of zircons provide a powerful tool for discerning the origin and source of granitic magmas. Not only are zircons resistant to weathering and resetting, but they can also be dated, thus providing a reliable and well-dated record of the isotopic evolution of the continental crust. *In-situ* analyses using modern ICP-MS instruments allows the rapid measurement of both Hf and O isotopic compositions, yielding results like those shown in Figure 2.8. In this type of figure, which will be used extensively in the following sections, the line labelled DM represents the composition of the convecting upper mantle, the source of mid-ocean ridge basalts. In many papers in which the isotopic composition of zircons and granites are interpreted, this line is taken to represent that of the mantle source. In later sections I will show that this assumption is rarely valid. The black lines with positive slopes represent the change in the composition of crustal material. The line on the left in the $\epsilon_{\text{Hf}(T)}$ vs time diagram corresponds to the evolution of 4 Ga mafic crust with $^{176}\text{Lu}/^{177}\text{Hf} = 0.02$. Granitic crust has a lower $^{176}\text{Lu}/^{177}\text{Hf}$ of about 0.01 and evolves along steeper lines. Most of the Hf isotope values scatter about the zero value, between the DM line and down to highly negative values. All of these data can be interpreted as indicating generation from a mixed source containing one component from the mantle and another from older continental crust.



In many recent studies, the oxygen isotopic compositions of zircons are measured together with Pb and Hf isotopes. Zircons that crystallised from juvenile magma should have an oxygen isotopic composition like that of the mantle source, *i.e.* a $\delta^{18}\text{O}$ between +5 and +5.6‰ (Eiler, 2001). In contrast, zircons from recycled continental crust have higher values and can thereby be distinguished from their juvenile counterparts.

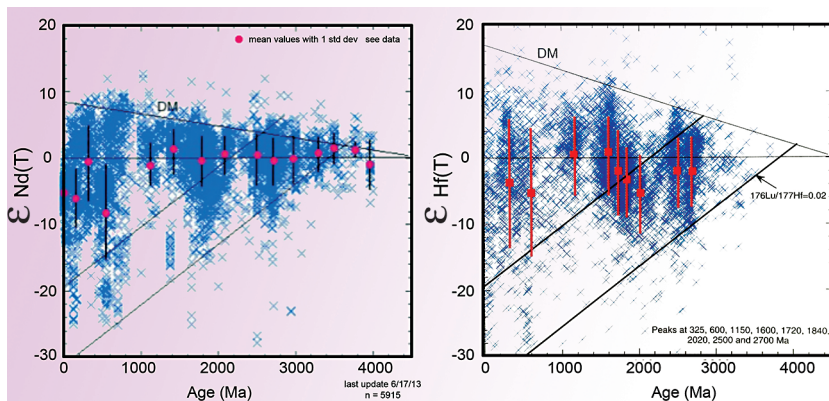


Figure 2.8

(left) Nd isotopic compositions of felsic igneous rocks, sedimentary rocks and their metamorphic equivalents, plotted against their age (from Condie, pers. comm. 2013). (right) Hf isotopic compositions of zircons from modern rivers, plotted against their age (modified from Condie *et al.*, 2011).



Were we to understand more about the origin of both basalt and granite, we would have a much better appreciation of how the solid earth functions. The origin of basalt is the easier part. We can see basaltic lava erupting from volcanoes, and its magmatic origin has not seriously been disputed since the later 18th century, when *Sir James Hall* slowly cooled molten basalt and reproduced the texture of the volcanic rock (Watt, 1804). Most geologists now accept that basalt forms by decompression melting of mantle peridotite beneath mid-ocean ridges or within mantle plumes, or in a subduction setting when the mantle wedge is fluxed by water from a subducting slab.

Petrologists were entertained by a famous debate on the origin of granite during the first part of the twentieth century. One school advocated granitisation (the conversion to rock of granitic character without passing through a magmatic stage; Read, 1957); the other school favoured a magmatic origin. The issue was effectively settled by the experiments of Tuttle and Bowen (1958), who showed that the composition of common granites coincided with a thermal minimum in the system SiO_2 - KAlSi_3O_8 - $\text{NaAlSi}_3\text{O}_8$ - H_2O . In their textbook written a few years later, Carmichael *et al.* (1974) drew on observations from regions of high-grade metamorphism and migmatitisation and concluded “*some granitic magmas form by fusion of mixed sedimentary rocks at the climax of regional metamorphism*” (page 580, Carmichael *et al.*, 1974). Basaltic magma “*derived from the mantle or from fusion of mafic rocks in the deep crust may contribute – through magmatic mixing ...*”. Earlier in their book, they speak of Precambrian granites being “*drawn directly from a primitive sialic crust, if such exists*”. Judging from their discussion of the origin of granitic magmas, they, like many other petrologists of that period, viewed granite magmatism as a process that was largely confined to a continental crust which persisted as a perennial layer at the surface of the Earth and was periodically reworked during orogenic events.

The notion that the source of granitic magma is to be found in the mantle was clearly expressed by T.H. Green and Ringwood (1968). *Trevor* (T.H.) is the brother of *David* (D.H.) *Green*, who, together with *Ted Ringwood*, undertook a series of experimental petrological studies at the Australian National University in Canberra in the 1960's and 1970's. These experiments form the basis of much of our present understanding of the genesis of basaltic magmas.

Life on the Other Side of the ANU Campus – At the time *Green* and *Ringwood* were conducting their experimental studies at the Research School of Earth Sciences, I was studying for my BSc in geology, also at the Australian National University. All our courses were given in the undergraduate teaching section, which was located in another part of the campus. Only in the final honours year did we hear anything of the research being done in the Research School and never, during that period, did I meet any of the scientists involved. This had to wait until 1976, after my PhD in the University of Toronto and during a post-doc at the Geophysical Laboratory in Washington DC, when I returned to Australia to give my first scientific talk at the



25th International Geological Conference in Sydney. By that time *David Green* had left to take up the position of director of the Geology Department in the University of Tasmania, but I met and talked with *Ted Ringwood*, *Bill Compston*, *Malcolm McCulloch* and the other scientists who were undertaking fundamental research at the Research School of Earth Sciences.

In their 1968 paper, *Trevor Green* and *Ted Ringwood* described the results of their experimental studies of the calc-alkaline rocks, the suite that includes both the volcanic rocks of subduction zones and most of the granitic rocks of the continental crust. As shown in Table 3.1, they reviewed five hypotheses for the origin of these magmas: fractional crystallisation of basaltic magma; melting of pre-existing sial; contamination of basaltic magma with sialic crust; partial melting of quartz eclogite; and partial melting of basalt under wet conditions. They dismissed the first three hypotheses, thereby moving the science beyond the older ideas that involved continuous reprocessing of an ancient sialic layer, and proposed a hypothesis that underpins most modern discussion of the origin of granite. According to them, *“In the first stage ... large piles of basalt are extruded on the earth’s surface. Subsequently this pile of basalt may, under dry conditions, transform to quartz eclogite, sink into the mantle and finally undergo partial melting at 100–150 km depth. ... Alternatively, if wet conditions prevail in the basalt pile and the geotherms remain high, partial melting of the basalt may take place near the base of the pile, at about 10 kb pressure”* (Green and Ringwood, 1968, page 151).

Table 3.1 Hypotheses for the formation of granitic magma

Green and Ringwood (1968)	Carmichael et al. (1974)	Moyen and Martin (2012)
1) Fractional crystallisation of basaltic magma	1) Fusion of mixed sedimentary rocks at the climax of regional metamorphism	1) Fractional crystallisation of wet basaltic magma
2) Melting of pre-existing sial	2) Melting of primitive sialic crust	2) Direct melting of (metasomatised) mantle
3) Contamination of basaltic magma with sialic crust	3) Derivation from ultramafic mantle	3) Partial melting of greywacke
4) Partial melting of dry quartz eclogite	4) Partial melting of dry quartz eclogite	4) Partial melting of a hydrated basalt metamorphosed at high pressure and transformed to eclogite or garnet bearing amphibolite
5) Partial melting of basalt under wet conditions	5) Contamination (mixing) of basaltic magma with crustal rock	



As we shall see in later sections of this *Perspectives*, the idea that the parental magmas of most granites are generated by re-melting of hydrated basalts and/or fractional crystallisation of such magma at the base of the crust is found in most current models for the formation of the continental crust. It was suggested on the basis of field evidence from the old gneisses of Greenland by McGregor (1979) who did the fundamental mapping that established the antiquity of these rocks. The idea was also promoted by other authors working on the nature and origin of Archean granitoids (Barker and Arth, 1976) and also provides an explanation for the origin of granites in modern island arcs (e.g., Tatsumi, 1986; Arculus, 1999; Jagoutz, 2010), convergent margins, and some collisional orogens (Thompson *et al.*, 2002). Finally, it also forms the basis of models in which Archean granites are said to form in settings unrelated to plate margins, such as at the base of an oceanic plateau or below thick Archean oceanic crust (e.g., Zegers and van Keken, 2001; Bédard, 2006; Rollinson, 2010).

An alternative to the basaltic underplate model is the hypothesis that Archean granitoids result from melting of oceanic crust that forms part of a subducting slab (Martin *et al.*, 2005). Many papers published in the past two decades focus on the geochemical similarities between Archean TTG and an intermediate to felsic volcanic rock known as adakite (Defant and Drummond, 1990). These rocks have unusual geochemical features such as strong relative depletion of the heavy rare earth elements (HREE) and relatively high Sr contents that some authors attribute to partial melting of subducting oceanic crust at large depths where plagioclase has been replaced by garnet in the melting residue. Archean granitoids share these geochemical features, and this has led to the hypothesis that they too formed by melting of ocean crust in subducting slabs. Because Archean heat production was 2 to 4 times greater than today (Brown, 1986 and as discussed in a later section) both the Archean mantle and subducting slabs would have been hotter, and the normal Archean thermal regime may have been rather like that of modern zones of 'hot' subduction where adakite is produced. High temperatures in the Archean oceanic crust and mantle favoured melting of the subducting oceanic crust, rather than fluid-fluxed melting in the mantle wedge, the process that operates in modern subduction zones (Grove *et al.*, 2006).

Since the 1970's numerous experimental studies have shown convincingly that partial melting of basalt or its altered or metamorphosed equivalents – amphibolite, mafic granulite or eclogite – produces magma of granitic composition. Moyen and Stevens (2006) summarised the experimental work that had been carried out in this domain between about 1991 and 2007. Although most of the experimental work summarised in that paper focused on Archean TTGs, the conclusions concerning the compositions of partial melts and the conditions required to stabilise garnet in the residue apply to the formation of most granitic magmas.

Moyen and Martin (2012) recognised four possible granite-forming processes, each of which may be valid in a different tectonic setting: (1) fractional crystallisation of wet basaltic magma; (2) direct melting of (metasomatised) mantle; (3) partial melting of greywackes; (4) partial melting of a hydrated basalt that had been transformed to eclogite or garnet-bearing amphibolite during



high-pressure metamorphism. Reference to Table 3.1 shows that these hypotheses are not very different from those proposed by Green and Ringwood (1968) and like them, Moyen and Martin (2012) opted for partial melting of garnet-bearing amphibolite or eclogite as the most probable origin of Archean granitoids. The influence of the composition and type of basalt and the specific conditions of melting are discussed in more detail in the following section.

In current models, granitic magma is generated in tectonic settings ranging from subduction zones, oceanic plateaus and oceanic crust, within continental crust during orogenesis and in intraplate settings. In this *Perspectives*, we are not so interested in within-crust melting of pre-existing rocks of the continental crust (although the contamination of mantle derived magma with crustal rock will be shown to have an important influence on the composition of many granites); instead we will focus on the extraction of felsic magma from the mantle, usually by way of the two-stage process outlined by Green and Ringwood (1968).

The basaltic material that constitutes the source of granitic magma may be found in the following tectonic contexts:

- (1) in oceanic crust within subducting slabs which, when young and hot, may melt directly to form adakitic magmas (Martin, 1987; Defant and Drummond, 1990)
- (2) at the bases of piles of basalt in settings that range from thick Archean oceanic crust (de Wit and Hart, 1993; Rollinson, 2010), oceanic plateaus (Bédard, 2006) to the lower parts of island arcs or convergent margins (Kay and Mahlburg-Kay, 1991; Thompson *et al.*, 2002; Jagoutz *et al.*, 2013).

A notable feature of Archean TTGs, particularly those that formed before about 3.2 Ga, is a low magnesium number ($Mg\# = Mg/(Mg+Fe)$) and low contents of Ni and Cr, compared with younger TTG and modern granitoids (Smithies, 2000; Smithies and Champion, 2000b; Martin and Moyen, 2002). The higher values of the younger rocks have been attributed to interaction with the surrounding mantle peridotite during ascent of the magma from its source in the subducting slab. The lower values in the older rocks are explained by flat subduction (Smithies *et al.*, 2007) in which case the magma would have traversed little to no mantle peridotite before reaching the crust, or they are cited as evidence that the granitoids formed by a process unrelated to plate tectonics (Smithies and Champion, 2000a; Condie and Benn, 2006; Van Kranendonk, 2010).

In other papers, the emphasis is on processes that operated in the lower part of the crust, either beneath island arcs or in convergent margins. Getsinger *et al.* (2009) describe experiments designed to investigate how melt interacts with the rock matrix as it segregates from its metabasaltic source. This study is particularly pertinent to the geodynamic setting of granitic magma formation. Steven Moorbath (Hildreth and Moorbath, 1988) invented an acronym for these processes – MASH – which stands for mixing, assimilation, storage and homogenisation. According to Hildreth and Moorbath (1988) “*basaltic magmas that ascend from the mantle wedge become neutrally buoyant, induce local melting, assimilate and mix extensively, and either crystallise completely or fractionate to the degree necessary*”



to re-establish buoyant ascent." The magmas that emerge from MASH zones have acquired intermediate to felsic compositions and, if trapped in the crust, solidify as diorites or granitoids. The chemical compositions of these rocks reflect the mixture of juvenile and reworked crustal components in their source. In a later section I will call on this type of model, particularly as developed by Annen *et al.* (2006), to explain the formation of Archean granitoids and the generation of continental crust in general.

My Career as a Mining Magnate – During my PhD at the University of Toronto I worked on komatiites, the emblematic Archean rock type but one that will receive little attention in this *Perspectives*. In 1969, the year preceding my thesis, I worked for Project Mining, a small Australian mineral exploration company, at the height of the nickel exploration boom. We enjoyed ourselves but found no ore deposits of any importance. The only ore that I ever found was amethyst. Some years earlier, when working as a summer student with an Australian exploration company that was prospecting in the Broken Hill region in outback NSW, I found some outcropping veins of this mineral. The claims of the exploration company were for base metals, not quartz, so I staked my own claim and during the next two years extracted about 200 kg of mineral specimens that I sold to dealers in Adelaide and Melbourne. I earned several hundred dollars from these sales which made me far better off than some of my fellow students. My enterprise was even tainted by a whiff of scandal when my method for cleaning the samples was questioned – see the letter on page 423. In fact, I just used weak hydrochloric acid to remove the secondary carbonate on the surface of the samples and in a follow-up letter was able to convince the dealer of my innocence (Fig. 3.1).

REGULATIONS—MINING ACT, 1906—AS AMENDED

DIVISION 11.
SCHEDULE 20.
(See Regulation 82.)

**APPLICATION FOR AUTHORITY TO PROSPECT
AND/OR CONSTRUCT RACES, &c. (SEC. 17).**

To the Honourable
The Secretary for Mines,
Sydney.

(Address) 16 Hopton Ave, Broken Hill
(Date) 12th February 1968

Sir,

By virtue of my miner's right, number 26, issued at BROKEN HILL dated 5th February, 1968, I hereby make application for authority, under the 17th section of the Mining Act, 1906-as amended, to prospect for and/or ~~construct and use~~ rock quartz on that portion of Crown land, situated in the parish of BARRERIGARRAN, county of TORRAWANGEE, containing 40 acres, hereby described, of which I took possession on the 11th day of February, 1968, at the hour of 11 o'clock in the forenoon, in accordance with Regulation 6, by erecting posts and cutting trenches at each angle thereof. ^{on 20 acres a piece of} ~~area~~ ^{area were not up to 5ft long.}

The posts were 4 inches in diameter and projected 4 feet above the ^{12 inch} surface of the ground. Each arm of each trench was 5 feet in length and 12 inches in depth. The datum post is at the NW angle of the land, and the notice prescribed in Regulation 6 was affixed thereto.

The area applied for is more particularly defined in the sketch attached hereto. I deposit herewith the sum of \$100 as provided for by Regulation 82.

Yours faithfully,

(Signature of Applicant) M. T. Corbett



KOVAC'S GEMS & MINERALS

CK/cb

12th March, 1968.

Dear Mr. Arndt,

Thank you kindly for the box of specimens which we received a few days ago, and we admit that we are very pleased with the promptness of your dispatch. Please find enclosed our cheque to the value of \$61 as per your letter.

We would like to say that we are quite satisfied with your classification into different grades, with the exception of first grade amethyst, for which we were expecting a deeper coloured material, however we will let it pass at this stage. The only other complaint we have is that the agreement was for specimens to be cleaned by you, and therefore we would appreciate it if with future orders you adhere to this point.

Kindly send us some more smoky quartz crystals and green quartz crystals at your earliest convenience as we are advertising these in the next issue of "The Gemhunter", and we have practically sold out of the material you sent us. In The Gemhunter we are advertising Green Quartz specimens at \$5, \$8 and \$10 per piece because at the time of placing the ad, we did not know that you had smaller specimens available, and we based the price on the larger chunks you sold us when you were in the shop.

In your letter you state that the specimens were cleaned only with a nylon brush and water, but this does not appear to be the case as there was a strong smell of sulphuric acid present and some of the paper was burnt, which indicates that for some reason sulphuric acid was applied to the specimens.

We would like to point out that if the green material has been in one way or the other artificially changed from amethyst in colour either by heating or by any other chemical means, we are not interested in further purchases. We have a good reputation for quality specimens and we cannot afford to sell any article that is not genuine. However we do take it in good faith that all the specimens are as found, and we hope to do a lot more business with you in the near future.

Hoping to hear from you soon.

Yours sincerely,
KOVAC'S GEMS & MINERALS

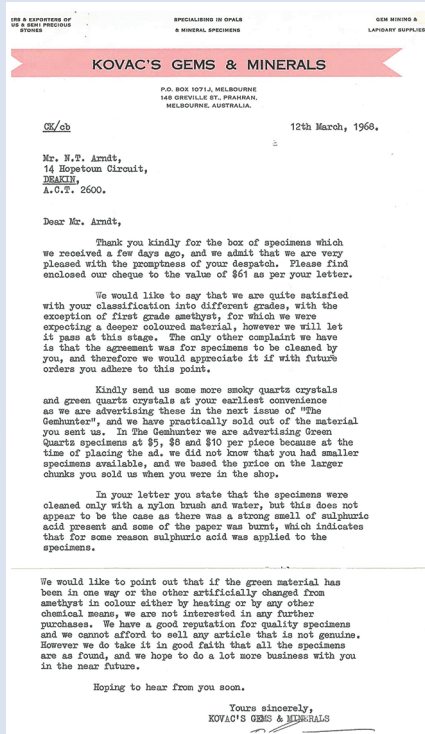


Figure 3.1

(left) My claim for the amethyst deposit in the Broken Hill region of NSW. (above) A letter from a mineral dealer in Melbourne who questioned the way I had prepared the amethyst samples.





Figure 4.1 A simple recipe for granite that should present no problems for even a starting chef.

As shown in Figure 4.1, the recipe for “granite ordinaire” is simple; all that is needed is a basaltic source, a generous dose of water, some “condiments” to add flavour, and a source of heat. But the simplicity of the recipe should not mislead – each ingredient is essential.

4.1 The Basaltic Source

It is now widely accepted that basalt is the dominant component in the source of high-volume granitic magmas (*e.g.*, Green and Ringwood, 1968; Barker and Arth, 1976; Kay and Mahlburg-Kay, 1991; Moyen and Martin, 2012). The material that melts may be solid, as in the basaltic underplate at the base of an island arc or oceanic plateau; or it may be liquid, when basaltic magma fractionally crystallises to produce more evolved felsic liquids. Some granitic magmas are produced by remelting of metasediment or orthogneiss, and others, the so-called sanukitoids, may come directly from mantle peridotite (Shirey and Hanson, 1986). But Archean TTG and most granitoids that intrude in large volumes in modern orogenic zones must ultimately have come from material that is produced in abundance in the mantle. And the normal product of mantle melting is basalt.



Moyen and Stevens (2006) summarised the large volume of experimental work that has shown conclusively that partial melts of basalt or its metamorphosed equivalent have granitic compositions (Fig. 4.2). Although the starting materials in these experiments include some more siliceous rock types such as basaltic andesite and andesites, the majority of experiments were carried out on basalts of tholeiitic, komatiitic and high-Al compositions, these representing the most common partial melts of mantle peridotite. More recently, experiments have been carried out on compositions corresponding to mafic rocks in the oldest supracrustal belts (Nagel *et al.*, 2011). In a few cases fresh (unaltered) basalt or andesite were used, but more commonly the starting material consists metamorphic rocks like amphibolite, mafic granulite, or eclogite. The chemical compositions of these materials vary significantly: SiO₂ from 47 to 60%, K₂O from 0.1 to 1.8%, Na₂O from 1 to 4.3% and Mg# from 38 and 71. Experimental conditions span a range corresponding to that at the surface to deep within subduction zones: pressures are from 1 to 35 kbar and temperatures are from 750 to 1200 °C.

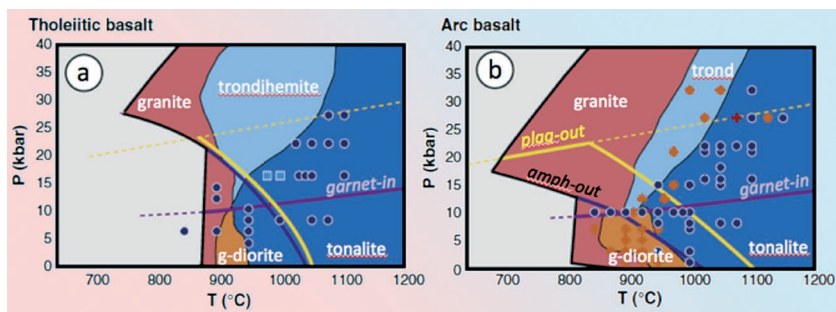


Figure 4.2 Summary of experiments designed to investigate the formation of granitic magmas by partial melting of mafic sources (modified from Moyen and Martin, 2012). These experiments were carried out under hydrous but fluid-absent conditions (water was present in hydrous minerals but not as a separate aqueous fluid) on two starting materials – a tholeiitic basalt and an arc basalt. The different fields are labelled with the compositions of the common types of granitoids.

4.2 The Need for Water

Two issues must be discussed here: is water necessary for the formation of granite, and is water present as a separate fluid phase or only in hydrous minerals? There is ample evidence that most TTG and orogenic granitoids formed from a water-bearing source. The most obvious evidence is the presence of hydrous minerals such as amphibole or biotite, which are ubiquitous in both TTGs and calc-alkaline granites. This is not the case for some felsic plutonic rocks in intraplate or convergent margin settings: in granophyres from Iceland, for example, the dominant



ferromagnesian minerals are pyroxenes, fayalitic olivine and only minor high-T amphibole (arfvedsonite, richterite edenitic hornblende) (Jørgensen, 1987). Likewise, in A-type granitoids, biotite and amphibole are uncommon and where present they contain high contents of Cl and F (Eby, 1990). The source of these types of granites is thought to be wholly or in part in water-poor portions of the lower continental crust: the low water content of this source is reflected in the anhydrous nature of their mafic minerals. The mineralogy of these near-anhydrous types of granitoid stands in sharp contrast with that of TTG and orogenic granitoids, which almost always contain amphibole or mica and evidently were derived from a hydrous source. The relatively low volumes of granite in intraplate or convergent margin settings may be linked to the low productivity of their water-poor sources.

To stabilise hornblende in intermediate to felsic liquids requires the presence of at least 5% water (Allen and Boettcher, 1983; Moyen and Stevens, 2006; Maksimov, 2009). If we assume that 20% partial melting produced the parental magma of the granitic suite and that the melt was water-saturated, from mass balance we can calculate that the source would have contained at least 0.5% water. An amphibolite comprising 60% hornblende contains more than 1% H₂O.

Davidson *et al.* (2013) used geochemical criteria to trace amphibole fractionation in arc magmas. They noted that amphibole phenocrysts are uncommon in erupted lavas and proposed that the amphibole accumulated in mid-crustal magma chambers where the magma solidified. An abundance of water in the crust above subduction zones is to be expected because magmas produced by fluid-fluxed melting are hydrous and these magmas transport the water into the crust. Arc magmas contain up to 6% water (Wallace, 2005) and much of this water will be released when the magmas enter the MASH zones in the lower crust.

Most of the experiments summarised by Moyen and Stevens (2006) were undertaken in the presence of water, either within hydrous minerals such as amphibole or biotite, or as a separate fluid phase. Moyen and Stevens (2006) and most other authorities conclude that those experiments carried out in the presence of water best reproduce the chemical compositions and mineralogy of TTG. They opted for what they call “fluid-absent” or “dehydration” melting, the process during which melt is produced during the breakdown of the hydrous minerals. In this process the water that is released from the hydrous mineral is immediately dissolved in the newly formed silicate melt and a separate fluid phase is absent. Fluid-present melting, *i.e.* melting in the presence of a separate aqueous fluid, is thought not to be important by Moyen and Martin (2012) according to whom “*the general consensus is that fluid-present melting at >10 kbar (i.e. garnet stability field) is unlikely in nature*”. In their opinion, and that of many other authors (*e.g.*, Rapp and Watson, 1995; Foley *et al.*, 2002), the quantity of water at the deep sites where granitic melts are produced is always relatively low and what water is present is contained in hydrous minerals. I am not totally convinced by this argument. Given the high water contents of subduction-zone magmas (6-8%; Wallace, 2005)



it seems possible that remelting in the lower parts of island arcs or the base of the crust in convergent margin settings could proceed in the presence of a separate fluid phase. This idea will be developed in a later section.

The initial source of water in terrestrial oceans and subduction zones was in the volatile-bearing meteorites and/or comets that impacted the Earth during its first 500 million years. The oceans probably formed at that stage (Valley *et al.*, 2002), and ever since that time water has circulated from the mantle into the oceans and then back into the mantle. The most efficient, and probably the only important method of returning water to the mantle is by subduction of hydrated basalts and sediments. The oceanic crust is hydrated at two stages, first at mid-ocean ridges and subsequently as the slab bends and fractures as it starts to subduct. Most of the water of the altered crust is released to the overlying mantle wedge as the hydrous minerals in the crust break down. The process thereby introduces water directly into a hot part of the mantle where partial melting can take place. Taking the argument further, the crust above a subduction zone is probably the only setting in the Earth (or in any other planet) where the three main ingredients of granite – basalt, water and heat – are found together (Fig 4.3).

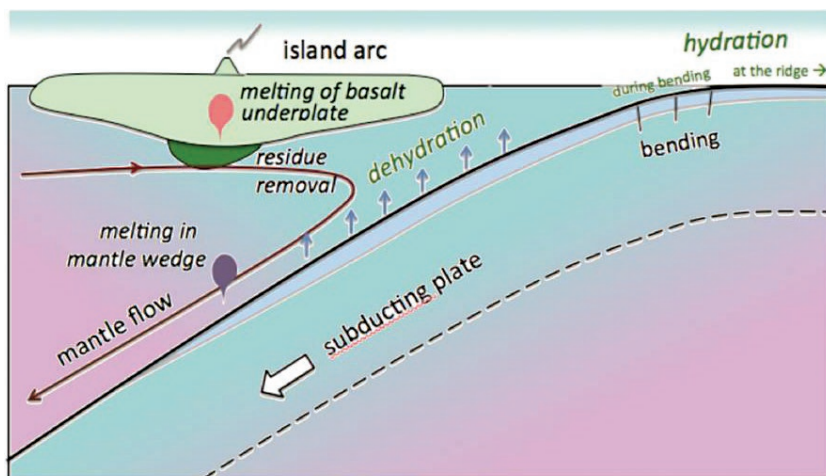


Figure 4.3 A sketch of a subduction zone showing how it efficiently transports water from the oceans into the site of melting.

4.3 The Condiments

I use this term to refer to those components of newly formed granitic magmas that do not come from the mantle but from older continental crust. Some of this material is derived from older mafic rocks or mafic-ultramafic cumulates whose



compositions are not very different from that of the basaltic source, but other materials, particularly old metasediments or felsic gneisses, have distinctive geochemical and isotopic compositions that differ radically from those of the mantle-derived component. Their presence in the source influences, or dominates, the trace-element and isotopic composition of the granitic magma.

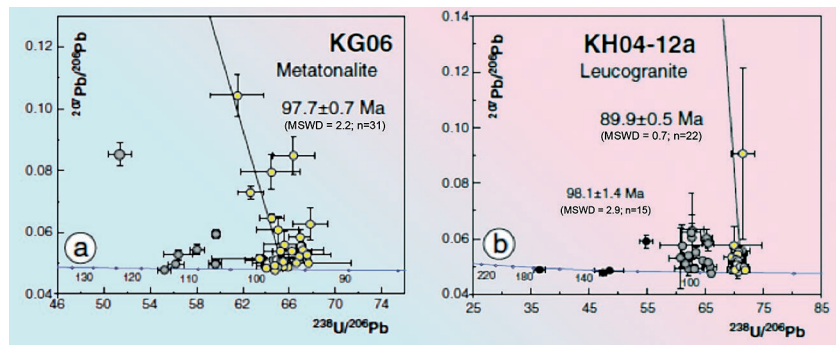


Figure 4.4 Pb isotopic compositions of zircons in granitoids from the Kohistan palaeo-island arc in Pakistan (modified from Bosch *et al.*, 2011). The crystallisation age of the rocks is about 65 Ma but each granitoid contains older zircon xenocrysts with ages up to 170 Ma.

Virtually every granite contains one or more components derived from older continental crust. The most obvious record of these components is the presence in many granitoids of zircons whose age exceeds that of the granite, as in the examples shown in Figure 4.4. In granites from modern and ancient island arcs (Bahlburg *et al.*, 2011; Bosch *et al.*, 2011) convergent margins (Folkes *et al.*, 2011) and collisional orogens (Belousova *et al.*, 2006), as well as from Archean granite-greenstone terrains, the zircons with the youngest ages commonly display magmatic features such as pronounced oscillatory zoning and are interpreted to have crystallised directly from the granitic magma (*e.g.*, Belousova *et al.*, 2006). The older zircons may lack such zoning and instead display rounded morphologies, corrosive borders and other features that indicate they are xenocrysts that most probably were derived from older crustal rocks.

I emphasise that the zircon xenocrysts are not sprinkled individually into the newly formed granitic magma but are added within fragments of country rock that had been assimilated by the granitic magma. Whenever present, they provide *prima facie* evidence of an older crustal component in the source of the granite. In some examples, the proportion of assimilated material can be seen to be high from the large proportion of xenocrysts (*e.g.*, KH04-12a in Fig. 3.4) but the exact amount cannot be determined using the zircons alone. Instead, the amount that was added can usually be quantified using Nd or Hf isotopic data.



Field Work in Canada – Laurie Curtis and I arrived at University of Toronto in 1970, to start our PhDs after working for a year with Project Mining in Western Australia and a six-month tour of South America. We remembered well our undergraduate studies with *Chappell* and *White* at ANU, and during a wonderful two-week field trip through the Canadian Shield organised by *Allan Goodwin* of the University of Toronto, we expected to see granites. Nothing doing – all we were shown were volcanic and sedimentary rocks.



Figure 4.5 (left) A photo taken at the end of a hard day in the field with Tony Naldrett, my thesis advisor. This was in my “seedy Mexican” days, the only post-adolescent period when I had no beard. (right) The first field trip on the Canadian Shield with Tony Naldrett and students on the left and a be-ponchoed Laurie Curtis on the right.

From 1980 to 1990, I was a Wissenschaftlicher Mitarbeiter in the Max Planck Institute in Mainz, working, as ever, on komatiites. In 1982 *Jon Patchett* joined the group and he told us all about the work he had done on granitic rocks in Scandinavia. He explained how, by using a combination of geochemical and isotope data, it was possible to determine the origin of the rocks – whether they came directly from the mantle or from recycling of older continental crust. We thought this was all very exciting and decided to organise our own field excursion to the Canadian Shield. This was the first Canadian field work for *Catherine Chauvel*, one of the members of the field team. She arrived fresh from France where she had worked on basalts in the bucolic countryside of the Massif Central; her surprise at having to drive kilometres on unpaved gravel roads was followed by alarm when she saw the boat that would take us around enormous Reindeer Lake in northern Saskatchewan – a 14ft runabout charged to the gunnels with three weeks of provisions. A highlight of that trip, organised very capably by *Mel Stauffer* of the University of Saskatchewan, was a night in a tent on a small island where she felt safe until told that bears could swim. The product of that expedition was about 100 kg of Proterozoic granitic, metavolcanic and sedimentary rocks that we subsequently analysed for major and trace elements and Sr, Nd and Pb isotopic compositions.





Left: Photo courtesy of Kaye Patchett.



Figure 4.6 (left) Jon Patchett and Dalila Ben Othman in the lab of the Geochemistry Division of the Max Planck Institut in Mainz. (right) Sampling granites in the Trans-Hudson belt. Pat Bickford and John Lewry are taking notes.

Using data obtained after our Canadian field trip, Chauvel *et al.* (1987) calculated the fraction of old Archean crust in the granitoids that had intruded during the Trans-Hudson orogeny (Fig. 4.7). By making some assumptions about the Nd contents and isotopic compositions of the juvenile mantle component and of the Archean contaminant, they were able to show that the proportion of older crust ranged from close to zero in the island arc that had formed remote from the Archean craton, to about 50% in the batholith and Peter Lake terrain that formed as an convergent margin adjacent to the craton. In a later paper with *Wolfgang Todt*, the specialist of Pb isotopic analysis in the Mainz Geochemistry Division, we produced additional constraints on the sources of the contaminants. For the belts far from the craton, the material came from the upper continental crust but closer to the craton, the source was dominantly in the lower crust (Arndt and Todt, 1994).

Zircon xenocrysts, and Sr, Nd, Pb or Hf isotopic evidence of an older crustal component, are found in granites of all ages. Inheritance from older crust is of course greater in S-type granitoids – those with a sedimentary protolith (Chappell and White, 2001) – but is also common in I-type granitoids from a mainly igneous sources. Granitoids from the Andes (Hildreth and Moorbath, 1988), the Himalayas (Goss *et al.*, 2011), modern island arcs (Bosch *et al.*, 2011), Proterozoic mobile belts like the Limpopo, and still older granite-greenstone terrains like the Yilgarn in Australia (Hill *et al.*, 1989, 1992) all contain zircon xenocrysts from older crustal rocks. The only notable exceptions are granites in certain Archean and Proterozoic greenstone belts whose Sr, Nd or Hf isotopic compositions point to their formation in a setting remote from older continental crust. Examples include



the granitoids of the Abitibi belt in the southern Superior Province of Canada (Ayer *et al.*, 2002; Chown *et al.*, 2002) and portion of the Birimian terranes in West Africa (Abouchami *et al.*, 1990; Boher *et al.*, 1991). But even in these regions, isotopic evidence for the assimilation of older crust is present in granitoids that intruded adjacent to Archean Man craton, as is the case for the Superior Province (Davis *et al.*, 2005) and Trans-Hudson granitoids studied by Chauvel *et al.* (1987).

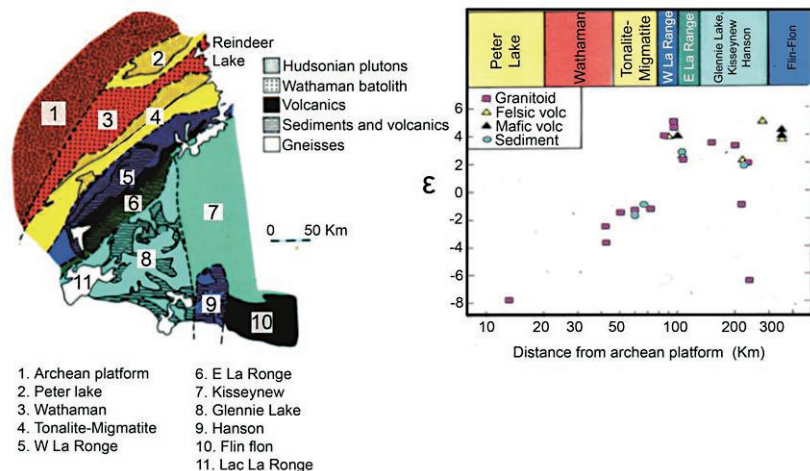


Figure 4.7 (left) Sketch map showing part of the Trans-Hudson orogeny in northern Saskatchewan. A series of Mesoproterozoic litho-tectonic belts are accreted to the Archean platform in the northwest. (right) The Nd isotopic composition ($\epsilon_{Nd(T)}$) plotted against distance from the Archean platform. The $\epsilon_{Nd(T)}$ value decreases steadily (except within the W La Ronge island arc) reflecting an increasing component of Archean crust in the source (modified from Chauvel *et al.*, 1987).

There are two reasons for my insistence on the presence of recycled material in most granitoids and on the distinction between juvenile and recycled components. The first will become apparent in a later section when I discuss some of the broader questions related to the rate of growth of the continental crust when it is particularly important to distinguish real growth (*i.e.* the transfer of material from mantle to crust) from recycling of material within the crust. The second is related to a conclusion reached in many studies of the geochemistry of granitoids. Moyen and Martin (2012), for example, conclude on the basis of their trace-element modelling that “the TTG source was basaltic ... and relatively enriched”. Smithies *et al.* (2009) reached a similar conclusion following their study of TTG from the Pilbara. By “enriched”, these authors mean that the concentrations of incompatible trace elements are higher than in normal mid-ocean ridge basalt. Figure 4.8 shows Moyen *et al.*'s (2007) estimates of the compositions of

the source of granitoids from the Barberton granitoid-greenstone terrain of South Africa. Conspicuous in the mantle-normalised trace element patterns is a strong enrichment in the more incompatible elements, combined with distinct negative Nb anomalies and positive Pb and Sr anomalies. Although these estimates are model dependent, it is notable that the calculated trace-element spectra display many of the features that characterise both magmas from subduction zones and most materials – granitoids, felsic gneisses and sedimentary rocks – of the continental crust.

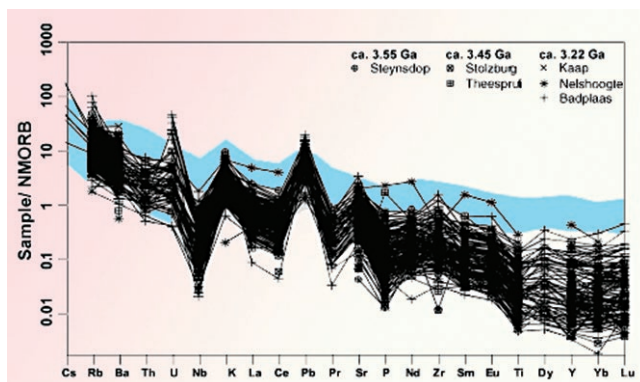


Figure 4.8 Moyén *et al.*'s estimated compositions of the sources of TTG of the Barberton belt in South Africa (modified from Moyén *et al.*, 2007).

The emphasis on the inferred composition of the source arises in part from the hypothesis that certain Archean granitoids, like modern adakites, are the product of partial melting of subducting oceanic crust (*e.g.*, Martin, 1987; Martin *et al.*, 2005). In such a case, the source should have a depleted character, like normal mid-ocean ridge basalt (MORB). The convincing evidence of enrichment requires instead that the source was either another type of basalt (*e.g.*, “enriched” MORB), or that it contained a contaminant of one type or another. In some papers, the origin of the enrichment is attributed, without further explanation, to “mantle metasomatism”. Another possibility is that the enriched character is inherited from subducted sediments that contaminated the source of the basalts in the mantle wedge of a subduction zone. A third alternative is that the compositions presented in Figure 4.7 record a mixed signal, a mixture between basalt with only moderate enrichment of the incompatible elements and with subdued Nb-Ta, Ti and Pb anomalies, and material from the continental crust. A distinction between the two interpretations might be made using isotope data, ideally a combination of radiogenic isotopes (Sr, Pb, Nd and Hf) and stable (O) isotopes, which should be able to identify the presence of old continental crust, though not recycled material with a short crustal residence time.



4.4 The Source of Heat: Temperatures in Modern and Archean Mantle

Granitic magma forms at temperatures that are low compared with those of other types of magma. As shown in Figure 4.2, the wet liquidus is at 850 °C at atmospheric pressure, declines to a minimum of 700-750 °C at 2.0-2.5 GPa as water becomes more soluble in the silicate liquid, then increases to >850 °C as the hydrous phases become unstable. Collision-related granitoids such as those in the Himalayas are produced at particularly low temperatures (<800 °C) by low-degree partial melting of mainly metasedimentary precursors at mid crustal depths (ca. 15 km) (Vigneresse and Burg, 2003). The melting is linked to tectonic deformation, thrusting and shear heating, and both the amount of granitic melt and the rate of melt production is low (Harrison *et al.*, 2012).

Some within-crust melting takes place in subduction settings due to a combination of deformation (tectonic inversion during thrusting, shear heating, etc.) and advection of heat within magmas derived from deeper levels. Most high-volume granitic magmas, however, incorporate mafic melts from the underlying subduction zone. The temperatures required to produce these melts are met in almost all parts of the mantle, including the asthenosphere beneath thin oceanic crust and within the mantle wedge above subducting slabs. The limitation to producing granitic magma in the mantle is therefore neither temperature, nor a source of basalt – what is lacking in most parts of the mantle is a source of water.

In the Archean, the mantle as a whole was hotter than at present (Herzberg *et al.*, 2010). The high temperatures in the mantle beneath Archean ocean ridges produced oceanic crust that was thicker than modern oceanic crust and on average had a more magnesian composition (Sleep and Windley, 1982). Like its modern counterpart, Archean oceanic crust was differentiated into lower mafic-ultramafic cumulates, an intermediate gabbroic layer, and upper layers of basaltic lavas. Although the volcanic layers of the Archean crust would have been intruded by sills of olivine-rich rocks, even some with komatiitic compositions, most of these magnesian and dense magmas would have been trapped by the crustal filter before reaching the surface, just as modern picrites rarely erupt on the floor of today's oceans (Rhodes, 1982). As in modern oceanic crust, the uppermost part of Archean oceanic crust had a basaltic composition.

Oceanic plateaus as well are differentiated into lower mafic-ultramafic cumulates, intermediate gabbros and an upper basaltic layer (Farnetani *et al.*, 1996; Kerr, 2013). Although some oceanic plateaus are thought to have been emplaced at or near oceanic spreading centres (Neal *et al.*, 1997), the rate of eruption was probably greater than that of ocean-ridge basalts. More importantly, the scale and geometry of circulation of seawater through the crust was unlike that at a mid-ocean ridge. An organised, deep-rooted system of hydrothermal circulation comprising diffuse downwelling of seawater and focused upward flow of hydrothermal fluid was probably absent. Beneath an oceanic plateau, the sources of magma, and therefore of heat, are dispersed over a broad area, not confined



to a narrow planar zone as at an oceanic spreading centre. The consequence is that the extent of hydrothermal alteration of the upper portion of the oceanic plateau was less than in normal oceanic crust and most probably restricted to the uppermost layer of volcanic rocks (Banerjee *et al.*, 2004). The lower portion of the plateau consists of cumulates that were derived from plume-derived, near-anhydrous magmas. This lower portion contained very little water. The relevance of this discussion will become clearer in a later section when I evaluate models in which Archean TTG are said to result from partial melting in the lower parts of oceanic plateaus.

4.5 Conditions of Melting during the Production of Granitic Magma

In Figures 4.2 and 4.9, I reproduce several diagrams from papers by *Jean-François Moyen* and colleagues (Moyen and Stevens, 2006; Moyen and Martin, 2012) that summarise the results of experiments designed to understand the formation of TTGs. The diagrams show the range of conditions in which the partial melting is thought to take place: at pressures ranging from about 10 to 30 kbar and temperatures from as low as 700 °C to about 1000 °C.

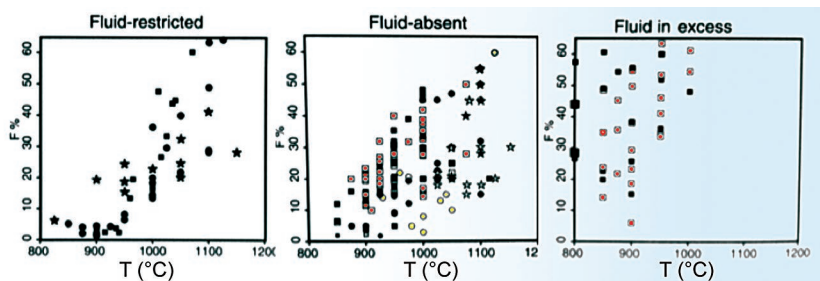


Figure 4.9 The relationships between temperature (T °C) and the degree of melting (F %) during experiments summarised by *Moyen et al.* (modified from *Moyen et al.*, 2007).

To understand the relationship between the composition of the melts and the experimental conditions, consider first the effects of temperature, the degree of melting (F) and the presence or absence of fluid, as shown in Figure 4.9. Three different melting regimes are shown in the figure: “fluid-absent” and “fluid-restricted” refer to melting of a source containing hydrous phases that break down to release the fluid that triggers the melting but without a separate aqueous fluid. During “fluid-absent” melting, the proportion of hydrous minerals is low and the fluid that they release dissolves quickly in the melt. During “fluid-restricted” melting, the proportion of hydrous minerals and fluid is greater and, at a given temperature, the amount of melt increases with the availability of fluid.



Some first-order conclusions can be drawn from Figures 4.2 and 4.9:

- Granitic magmas are produced by low degrees of melting (0-20%), at relatively low temperatures (700-900 °C) and at a wide range of pressures (<5-40 kbar)
- Primary magmas of granodioritic composition form under a restricted range of conditions, from only 900-950 °C in the tholeiitic basalt (though from 900-1000 °C in arc basalt) and at pressures less than 10 kbar
- Tonalites form at higher degrees of melting (40-60%) at high temperatures (>950 °C) and low to high pressures
- Trondjemite forms only at pressures greater than 15 kbar and moderate to high temperatures (900-1100 °C) and moderate degrees of melting (30-50%).

The compositions of the two basaltic starting materials shown in Figure 4.2 are not very different. "Arc basalt" contains slightly more quartz and more amphibole, which explains the lower solidus and the earlier on-set of dehydration melting. It also contains higher K₂O and Na₂O and lower CaO, which explains the broader field in which granitic magmas are produced. Another difference between the two basalts is their trace element contents. The tholeiite has a flat REE pattern and lacks the subduction signature of high La/Yb, negative Nb and Ti anomalies and a positive Sr anomaly that is present in the arc basalt. Only the latter has the "enriched" characteristic inferred for the source of TTG (Smithies *et al.*, 2009; Moyen *et al.*, 2007).

The limit of garnet stability is clearly shown in both diagrams; this mineral becomes stable at a pressure about 10 kbar and this limit is relatively independent of temperature. Archean TTG and other granitoids that display the distinctive trace element pattern of low HREE and high Gd/Yb that records the presence of garnet at some stage during the production or evolution of the magma must have formed or evolved at pressures at least 10 kbar. This pressure corresponds to depths of 27-30 km, as in the lower part of an island arc or continental margin, or in the upper portion of subducting oceanic lithosphere. This is not a major constraint. All that it imposes is that garnet must have been left in the residue of melting – of the subducting slab, of peridotite in the mantle wedge, or of basalt at the base of the crust. Alternatively, garnet could have been removed during fractional crystallisation of magmas derived by any of these processes. To produce the greater extent of depletion observed in early Archean TTG requires, however, that a large amount of garnet remained in the residue of melting (Moyen and Stevens, 2006) and this points to higher depths of melting.

A final point concerns the role of water, which destabilises plagioclase and stabilises garnet (Müntener *et al.*, 2001). The presence of excess aqueous fluid at the site of melting may contribute to the development of the garnet-subtraction signature (depletion of HREE or a high Gd/Yb ratio) of Archean TTGs.



This section focuses on the process that generates the continental crust. I will argue that throughout most of geological history, the rocks of the continental crust were generated in subduction zones by processes very much like those that operate today. I will develop the idea in detail at the end of the section, but before doing this I consider some alternative models and explore the reasons why many authors reject subduction-related processes, at least during the earliest Archean crust.

Foremost of these arguments is the notion that through much or all of the Archean, plate tectonics did not operate because Archean oceanic crust was too buoyant to subduct. I do not accept this and instead will argue that the oceanic crust subducted through most of Earth history, except perhaps during the first 500 million years. On this basis it would seem unnecessary to invoke mechanisms for the generation of continents that do not involve subduction, but hypotheses of this type are common in the literature. I therefore will discuss models that hold that Archean granitic magma was produced through partial melting in the basal portions of oceanic plateaus or even in the lower part of normal oceanic crust. In my opinion such models are not credible, principally because the lower parts of thick piles of basalt are essentially anhydrous. Unlike in subduction settings where a constant flux of aqueous fluid enters the zone of melting, no mechanism exists to transport large amounts of water to the base of an oceanic plateau or to the base of the oceanic crust. Without water, felsic magma is produced only in small volumes and the composition of this magma is unlike that of Archean TTGs.

5.1 Archean Geodynamics: Plate Tectonics or Something Else?

Figure 5.1 is the image taken from course material of *Robert Barminski* of Hartnell College in California which is published on the following website: <http://www.hartnell.edu/faculty/rbarminski/geology%20ppt/chapter8%20hadean%20and%20archean.ppt>. The figure shows how new felsic crust is created in island arcs above subduction zones and then accretes to form the continents. I submit that throughout Earth history, crust formation was as simple as this.

Figure 5.2 is a sketch of a subduction zone beneath an island arc. It shows that the oceanic crust interacts with seawater and becomes hydrated, initially as a result of hydrothermal circulation at the mid-ocean ridge and later as seawater penetrates along fractures as the crust bends before it subducts. Water is stored in altered oceanic basalt, in sediments and in peridotite altered when water penetrates along deep fractures. At depths between 20 and almost 160 km, the subducting slab releases aqueous fluids, first by dewatering of sediments, then by progressive destabilisation of hydrous minerals in the altered basalts and finally



by breakdown of serpentine in the peridotite (Poli and Schmidt, 2002). As shown in the diagram, the fluid flux is greatest at depths from 20 to about 80 km, where clay minerals, chlorite and amphibole break down.

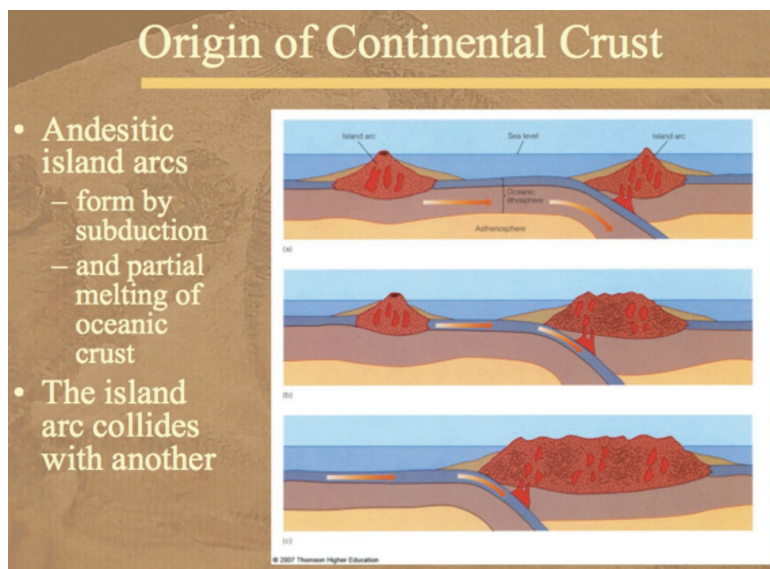


Figure 5.1 Robert Barminski's view of the formation of the continental crust.

In most island arcs, the volcanic front is fixed at a position where the depth to the underlying subduction zone is about 100 km (Tatsumi, 1986). The melting that produces the volcanic rocks is triggered by influx of aqueous fluids into the mantle wedge, which mainly occurs at depths up to about 80 km (Fig. 5.2) but the melting that produces arc magmas takes place at greater depths. To explain this apparent discrepancy, it is commonly accepted (*e.g.*, Poli and Schmidt, 2002; Grove *et al.*, 2006; Kelley *et al.*, 2010) that peridotite in the mantle wedge overlying the subducting slab is first hydrated by influx of fluid from the subducting plate, then is transported to greater depths by convection within the mantle wedge. Viscous coupling with the subducting slab induces circulation within the wedge that draws in hot asthenosphere from behind the island arc. As newly hydrated peridotite of the mantle wedge is dragged down, it heats up and eventually melts.

The product of flux melting in the wedge is hydrous magma of magnesian basaltic composition. These magmas rise from their source and ascend to the base of the crust where they undergo the complex series of processes (mixing, assimilation, storage and homogenisation) that leads to their transformation into more felsic magmas. Finally the evolved magmas ascend into the middle to upper crust where they solidify as granitic plutons.



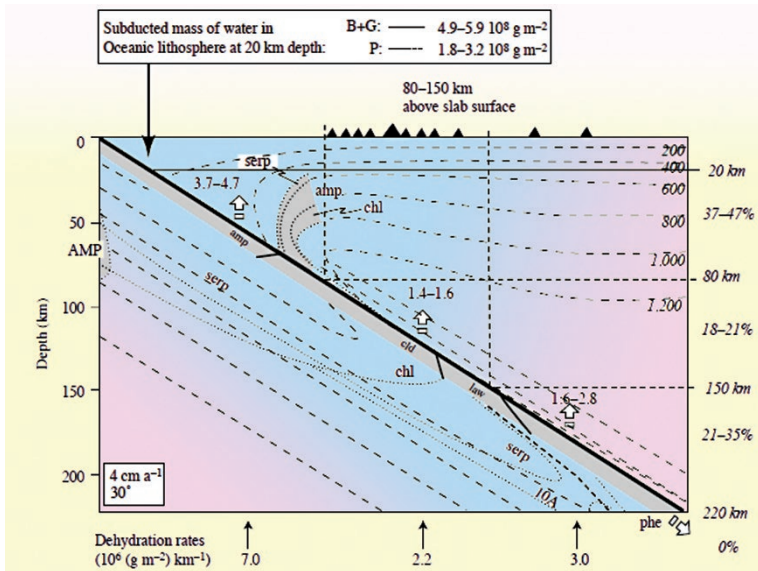


Figure 5.2 Schematic cross section of a subduction zone beneath an island arc. Dehydration of the subducting slab triggers hydration and melting of the mantle wedge (the part of the mantle above the subducting slab). Stabilities of hydrous phases and masses of H_2O in a unit section (in 10^8 g m^{-2}) are given for an intermediate convergence rate (4 cm yr^{-1}). Dark grey: oceanic crust, light gray: lithosphere of the overriding plate. The dotted lines denote maximum stability fields of serpentine, chlorite and amphibole (modified from Schmidt and Poli, 2004).

This brief and no doubt oversimplified account illustrates why a subduction zone is such an efficient granite factory. A subduction zone is the only place in present-day Earth where the two basic ingredients of granite – basalt and water – are transported together in large volumes to a site where temperatures are high enough to induce partial melting. The conveyor belt driven by forced convection in the mantle wedge continually draws hot, hydrated mantle into the site of melting, a process that produces large volumes of granitic magma, as can be observed in any mature island arc or active continental margin. Given that we have a mechanism that seems perfectly capable of producing copious amounts of granitic melt, why seek an alternative? The answer comes from the long-standing debate about the tectonic regime that operated through the Archean.

EGU Debates – Each year for about 8 years I have convened or co-convened the “Great Debates in the Geosciences” at the General Assembly of the European Geosciences Union (EGU). The idea for these sessions emerged during a planning meeting organised by John Ludden, then president of EGU, who had asked for suggestions for new



types of sessions. I had just received a letter from my father who was a professor of economics at ANU in which he wrote that he had been asked to be a “responder” at an economics conference. His role was to follow a keynote lecture delivered at the conference with his critical appraisal of the talk. I liked this idea and proposed it to the participants of our meeting, but we agreed after some discussion that it would be difficult to find geoscientists willing to play by these rules. Then we had another idea – to organise an “Oxford Union” style debate at EGU.

This we did, with varying degrees of success, over much of the next decade. At first the debate were strictly scientific, such as one on “Flood volcanism is the major cause of mass extinctions”. Right from the start I had planned to organise one on the motion “Armstrong was right – the continental crust formed in the Hadean”; but rapidly we drifted to more political or societal themes such as “In 30 years petroleum will have become a little-used energy source” or “The carbon footprint of EGU is bigger than necessary”. Most of the debates were not very successful: it turned out that very few European scientists understood, or accepted, the notion of a debate – a rhetorical exercise in which the members of each team are required not to present their latest scientific results or to convince the audience of their heart-felt conviction that “Geoscientists could do more for society”, but to defend one side or the other of the motion. The only really successful debate was one when a staunch advocate of geoengineering was told upon his arrival in Vienna that a member of the opposing team had pulled out and he argued, with vigour and enthusiasm, that “The risks of geoengineering outweigh its benefits”.

I still hope that, some time in the future, we will return to the original format and debate “Plate tectonic started in the Phanerozoic” or “Archean granitoids do not form in subduction zones”.

5.2 The Onset of Plate Tectonics – a Great Debate in the Geosciences

A quick survey of the literature reveals an immense volume of papers, books, special issues, blogs and informal communications that explore the question of when plate tectonics began. I have attempted to summarise these papers in Table 5.1. Estimates of the beginning of plate tectonics vary from about 800 Ma to close to 4.2 Ga, with a clear preference for the end of the Archean. Indeed, most of the more recent papers place the start even earlier, around or before 3.0 Ga. If we accept this interpretation, there would seem no need to explore mechanisms of large-volume granite formation that are not linked to subduction; nonetheless, a significant number of well-respected scientists continue to deny that modern plate tectonics operated in the Archean. What is the basis of their arguments?

There are two reasons. The first is theoretical – it is argued that the Archean mantle was too hot (or too wet, or too dry) to produce subductable oceanic lithosphere; instead an alternative, specifically Archean, geodynamic setting is said to have existed where granitic magmas could be generated. The other reason is based on geological observation – the identification of a set of “petrotectonic



signatures” that record the operation of modern plate tectonics. Several authors have concluded that these signatures are absent in Archean or even Proterozoic terranes, and that plate tectonics had not operated during these periods.

Table 5.1 Timing of the start of plate tectonics

Time	Authors
ca. 800 Ma	Stern (2008, 2013), Hamilton (2008, 2011)
1.8-2.7 Ga	Bédard (2006), Rollinson (2010), Brown (2007), Keller and Schoene (2012)
by 2.7 Ga	Condie and Kröner (2008), O’Neill <i>et al.</i> (2007), Korenaga (2006), Davies (2007), van Hunen and Moyen (2012)
3.0 Ga	Condie and Benn (2006), Cawood <i>et al.</i> (2006), Pease <i>et al.</i> (2008), Richardson and Shirey (2008), Polat (2012), Dhuime <i>et al.</i> (2012), Naeraa <i>et al.</i> (2012)
3.3-3.5 Ga	Zegers and van Keken (2001), Moyen <i>et al.</i> (2006), Smithies <i>et al.</i> (2007), Van Kranendonk (2007), Griffin <i>et al.</i> (2013)
4.3 Ga	Harrison <i>et al.</i> (2008)

Stern (2008, 2013) has adopted what is perhaps the most extreme position. He argues in a series of papers (and an excellent blog; Stern, 2013) that there was “*a progression of tectonic styles from active Archean tectonics and magmatism to something similar to plate tectonics at ~1.9 Ga to sustained, modern style plate tectonics ... beginning in Neoproterozoic time*” (Stern, 2008, page 265). In his opinion, petrotectonic indicators of modern plate tectonics such as ophiolites, blueschists, and ultra-high pressure metamorphic terranes are missing in Proterozoic and Archean sequences. In his estimation, a record of plate tectonics appeared first in the Neoproterozoic. He also argues that the thick oceanic oceanic crust generated from hot Archean mantle (Sleep and Windley, 1982) would have been too buoyant to subduct, and on this basis questioned whether plate tectonics could have operated during the first three quarters of Earth history.

Warren Hamilton is another active opponent of Archean plate tectonics. In a recent article (Hamilton, 2011) he argues against both this process and mantle plumes (another issue where we do not see eye-to-eye; see Arndt, 2012). His arguments are rather like those of Stern. As he puts it (Hamilton, 2011, page 4), “*indicators of seafloor spreading and subduction, ... including voluminous accretionary wedge mélanges, blueschists, ophiolites, and oceanic and continental magmatic arcs are sparse and incomplete in the Neoproterozoic and wholly lacking in older Precambrian terrains. When I began looking at Archean geology about 20 years ago, I expected to find evidence for modified plate tectonics, but instead found unfamiliar rock types,*



associations, and structures, and none of the indicators of Phanerozoic-type convergent and divergent plate interactions. I next assumed that plate tectonics began in early Proterozoic time, but abandoned that notion also when I looked more closely.

He raises some very valid questions about the nature of igneous rocks in modern island arcs, which differ in several important ways from those in Archean greenstone belts. He discusses the structure of Archean granite-greenstone belts and questions the validity of various sightings of Archean ophiolites. In his opinion, through most of Earth history, global geodynamics has been dominated by a mafic global protocrust that had segregated very early from the mantle and “contained most of the material recycled and polycycled into both continental and oceanic crust of all subsequent ages” (page 8). This view is not far removed from the older notion that granites are the products of reworking of the “primitive sialic layer” which I discuss in Section 3.

Some other authors (McLaren *et al.*, 2005; Bickford and Hill, 2007) have questioned whether plate tectonics operated during the formation of the Proterozoic and Archean orogens that they studied, but do not exclude it for other regions. Jean Bédard, who will figure prominently in the coming discussion, rejects plate tectonics as a major force in the construction of Archean continental crust. Jean, a geologist with the Geological Survey of Canada, developed what he calls a “catalytic delamination-driven model for coupled genesis of Archean crust and sub-continental lithospheric mantle” during his work in the Archean Minto block in northern Quebec. The model has much in common with those of other authors (*e.g.*, Zegers and van Keken, 2001; van Thienen *et al.*, 2004; Van Kranendonk, 2010) but he goes further by attempting to explain not only the generation of the continental crust, but also that of the subcontinental lithosphere and even ultramafic magmas like komatiites, all in a setting far removed from normal subduction. He rejects the slab-melting hypothesis of Hervé Martin and others (Martin, 1987, 1994b) on the basis that the physical conditions within the slab were inappropriate – the subducting crust was either too old or moved too fast to melt, or underwent shallow dehydration. He speaks of “melt productivity deficits” by which he refers to differences between his estimated rate of formation of the TTG and other lithologies of the Minto block and those that might be predicted from plate tectonics. According to his calculations, the rate of crust formation during the 300 Ma history of the block was 67 km³/Ma/km of subduction zone, a value that exceeds by a factor of about 3 commonly accepted estimates for Phanerozoic subduction zones. The calculation depends strongly, however, on his estimation of the length of past subduction zones, the volume of the Minto block and his concept of its structural evolution: in particular, he assumes that the current 40 km thickness of this part of the craton was that of the initial crust and was not generated by tectonic thickening. For modern active margins such as the Andes, the competing contributions of magmatic addition and deformation to the current >80 km thickness are subject to considerable discussion. The estimated magmatic contribution may be as high as 50%, but most authors (*e.g.*, Jaillard *et al.*, 2002; Francis and Hawkesworth, 1994; Haschke and Günther, 2004) prefer a figure closer to 20%. In the tectonic-thickening interpretation, the high Andean



plateau results mainly from thrust stacking and deformation of a once-much-wider swath of continental crust. To take another example, a large fraction of the North American Cordillera consists of a collection of island arcs, microcontinents and active margin belts that amalgamated during its 300 Ma evolution. As *Bédard* points out, the Minto block fits comfortably into a fraction of the Cordillera. If we accept that the present thickness and structure of this region results, at least in part, from the accretion of once dispersed arcs and other terranes, combined with tectonic thickening, then the rate of generation of this part of the Archean continental crust was probably no greater than in many modern orogens.

Bédard bolsters his arguments against Archean plate tectonics by citing the absence of many of the petrotectonic indicators identified by *Stern* and *Hamilton*, and refers as well to the intriguing alternation of arc-like volcanics with komatiites in several Archean greenstone belts, an association that has led to discussions about the co-habitation of subduction zones and mantle plumes (Wyman, 1999; Sproule *et al.*, 2002). This provides a convenient introduction to the crux of *Bédard's* model, which holds that continental crust forms at the base of a basaltic pile overlying one or more large Archean plumes. But before going into this, I need to discuss in more detail the issue of whether Archean oceanic crust could ever have subducted.

5.3 Subduction of Archean Oceanic Crust

Stern's (2008, 2013) conjecture that Precambrian oceanic crust was not readily subductable is crucial to any discussion of Archean geodynamics. The basis of the argument is shown in Figure 5.3. This figure, taken directly from *Stern's* blog, shows that modern oceanic lithosphere remains buoyant until cooling and downward growth of the mantle portion has made it denser than the underlying asthenosphere. At this stage, subduction is possible even in the absence of other tectonic driving forces. In practice, whether or not a slab subducts depends on the balance between the buoyancy forces that drive subduction and the mechanical and buoyancy forces that resist subduction (Billen, 2008). Most on-going modern subduction is driven in part by the drag from already-descending parts of the plate within which the basaltic portion of the crust has transformed to denser eclogite. Local compression and other tectonic forces may help to overcome buoyancy while opposing forces such as resistance to bending of the lithosphere, frictional plate-coupling and viscous shear in the mantle resist the process (Schutt and Leshner, 2006; van Hunen and van den Berg, 2008; van Hunen and Moyen, 2012). Whether or not subduction or some other process cycled back into the mantle the crust that was produced by melting of hot mantle is a matter of conjecture. Johnson *et al.* (2013) report a recent analysis of the density of Archean oceanic crust and underlying residual mantle and conclude that the base of even fully hydrated >30 km thick crust would have been denser than underlying mantle.



However, they do not advocate subduction as such but invoke foundering of the base of overthickened crust. Sivoza *et al.* (2010) describe a range of subduction styles that may have operated during the Archean.

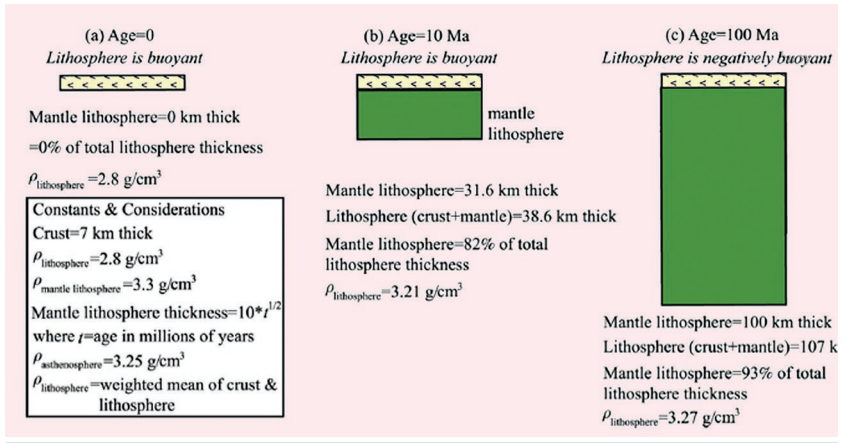


Figure 5.3 Stern's estimations of the structure and density of modern oceanic lithosphere (modified from Stern, 1998).

Figure 5.4 shows my conception of the composition and structure of Archean oceanic lithosphere, based on Herzberg and Rudnick's (2012) model for the generation of Archean crust. Herzberg and Rudnick propose that the dominant magma in the crust had a picritic composition and contained about 18.5% MgO. It was formed by a high degree of partial melting and erupted in large volumes to produce oceanic crust that was 20-40 km thick. If we assume 30 km and adopt Herzberg and Rudnick's estimate of 30% partial melting, this crust would be underlain by a 70 km thick interval of mantle that had become depleted in fusible components by the extraction of the magma. Downwards through this zone of depleted mantle, the composition of olivine, the dominant component, changes from Fo92 at the Moho (where the degree of melting was greatest) to about Fo89, the ambient mantle value at the base of the melting zone. Using Leonid Danyushevsky's Petrolog program (Danyushevsky *et al.*, 2002), I calculate that a picritic magma containing 18.5% MgO would crystallise 33% olivine, 10% cpx, and 7% of plagioclase (I ignore the minor minerals) before arriving at a basaltic composition containing only 8% MgO. The latter composition corresponds to that of the basalts that make up the upper portions of oceanic plateaus like Ontong Java (Fitton *et al.*, 2004) as well as normal mid-ocean ridge basalt. The dense mafic minerals that fractionated from the parental picrite will remain as cumulates in the lower part of the crust while the evolved liquids migrate towards the top where they intrude as gabbro or erupt as basalt. As a result, the crust acquires the layered structure shown in Figure 5.4.



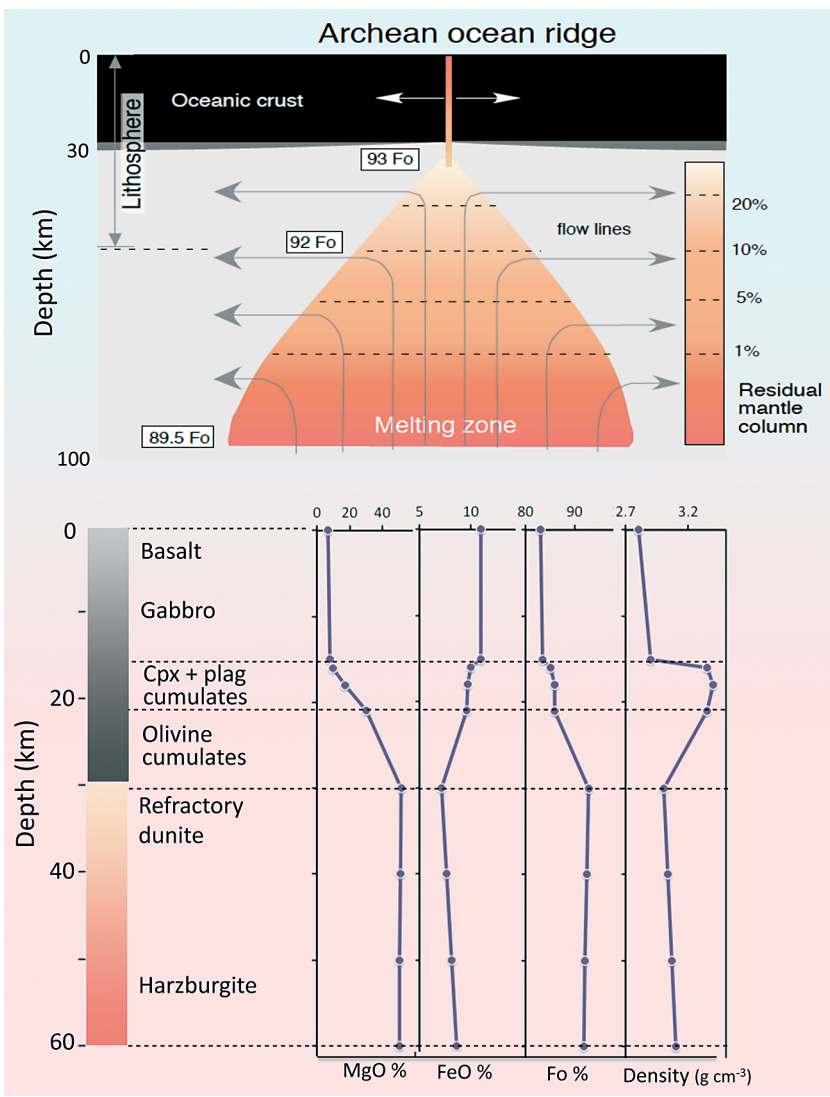


Figure 5.4

My model of the structure of Archean oceanic lithosphere. The upper diagram shows the situation at the mantle ridge, as hot mantle wells up and melts in the interval from 135 to about 30 km (the base of the crust) before flowing sideways and out of the melting region. The degree of partial melting and the amount of extracted melt increases upward through the melting zone and changes the composition of the depleted upper layer of the mantle. The lower diagram shows 60 km thick lithosphere some distance from the



spreading centre. The crust is 30 km thick and differentiated into lower olivine then clinopyroxene (Cpx) and clinopyroxene plus plagioclase (plag) cumulates overlain by evolved gabbros and basalts. The lithosphere extends only part way through the melt-depleted portion of the underlying mantle. The graphs to the right illustrate variations in composition and density of the various layers. The densities were estimated assuming a conductive gradient across the lithosphere, from 1400 °C at the base to 0 °C at the surface.

The fate of this crust as it migrates away from the spreading centre can be predicted if we compare the inherent density of the individual layers with that of the underlying mantle. At the ridge, the upper basaltic/gabbroic parts of the crust and the underlying melt-extraction zone are less dense than the underlying mantle. Most of the ultramafic cumulates contain more Fe than the underlying ambient mantle and are therefore denser.

As the crust migrates away from the ridge, it cools and the lithosphere thickens. The 1350 °C isotherm that marks the base of the lithosphere probably descends more slowly in the Archean oceanic lithosphere than in the modern lithosphere because, although the cooling rate was higher than now, the crust started hotter. Nonetheless, after a certain time, the base of the lithosphere will lie well within the melt-extraction layer, as shown in the figure. Heat is extracted from the upper part of the crust by conductive cooling, augmented by hydrothermal circulation, and may eventually cool the lithosphere to the extent that it becomes denser than the underlying mantle.

Before calculating the density of the whole lithosphere, however, we must first consider the extent of hydration of the uppermost layers. Contrary to prevailing opinion, the proportion of hydrothermally altered Archean crust was probably greater than that in modern oceanic crust. The reason is shown in Figure 5.5: in modern crust, the conductive thermal gradient is from about 1200 °C at the Moho at 9 km depth to 0 °C at the surface, to produce an average gradient of about 130 °C per km. In Archean crust, the gradient is from about 1400 °C at 30 km to 0 °C at the surface, an average gradient of only 47 °C per km. The breakdown of chlorite to amphibole that marks the transition from water-rich greenschist facies rocks with 10-12% H₂O, to water-poor amphibolite-facies rocks with 1-2% H₂O, occurs at a temperature of about

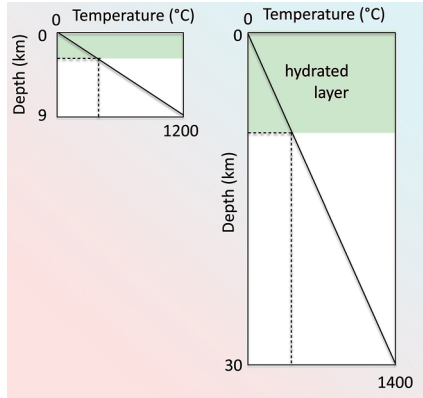


Figure 5.5 Sections through 9 km thick modern oceanic crust and 30 km thick Archean oceanic crust with the temperature gradients from surface to the Moho. The gradient is shallower in the Archean crust and therefore the layer of hydrated crust where the temperature is less than 500 °C, is thicker.

400 °C (Fig. 5.6). In modern crust this isotherm lies at about 3 km depth, whereas in the Archean crust illustrated in Figure 5.5, it is at about 10 km. Of course this analysis is oversimplified. It does not take into account the advection of heat into the upper portions of the crust within mafic intrusions, which would have perturbed the temperature gradient. It is also probable that the Archean oceans were warmer than those of today (Robert and Chaussidon, 2006) which would have (very slightly) steepened the gradient. However, hot (30-50 °C) Archean seawater would also have more efficiently hydrated the rocks it encounters, and once it penetrates more than a few hundred metres, circulating 30-50 °C seawater cools the crust. Even when these factors are taken into account it remains probable that a significant thickness of the Archean crust was hydrated. I emphasise this conclusion for two reasons. First, hydrated basalt is less dense than fresh basalt, and its presence must be considered when calculating the overall density of the oceanic lithosphere. Second, the subduction of a thick layer of hydrated crust provides a mechanism of transferring large amounts of water into the mantle and into the site where the magmas of the continental crust are generated.

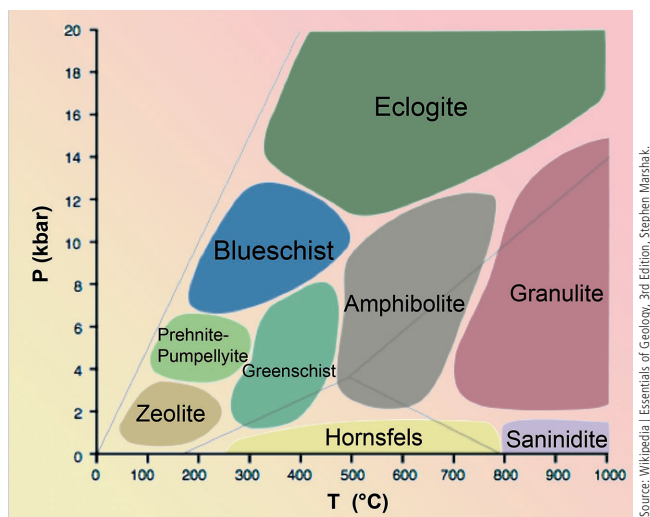


Figure 5.6 Summary of metamorphic facies showing the limits of the amphibolite facies that marks the transformation of hydrous to anhydrous assemblages.

Another factor that must be reckoned with is the rate of plate movement. In most papers on Archean geodynamics (*e.g.*, Bickle, 1978; de Wit and Hart, 1993; Van Kranendonk, 2007), it is argued that because the mantle was hotter, it would have been less viscous, and if less viscous it would have convected more rapidly. Rapid convection and rapid formation and destruction of the oceanic lithosphere were necessary to evacuate heat from the mantle. If plate movement is coupled to mantle convection, then the Archean plates would have moved quickly. A



popular image of the Archean ocean basins is one of numerous spreading centres, small, rapidly migrating plates, and numerous sites where the plates sink back into the mantle (de Wit and Hart, 1993). In such a situation, the oceanic lithosphere would rarely, if ever, reach the age required for it to become denser than underlying mantle.

But is this view of the lifetime of Archean oceanic crust correct? If continental crust was sparse in the early part of Earth history (as accepted by most authors and discussed in a later section), then the Archean oceans must have been large. If Archean oceanic lithosphere were initially more buoyant than modern oceanic lithosphere, it would not easily subduct; instead, the plates may have migrated for considerable distances across the broad Archean oceans, reaching considerable ages. Small and numerous plates are not a unique solution to the problem of how to remove heat from the Archean mantle. If thick refractory, viscous depleted mantle occupied the upper part of the mantle, as argued by Davies (2007) and Korenaga (2006), then plate movement would not have been rapid. In Figure 5.4, I show schematically the structure of 60 Ma old Archean oceanic lithosphere. In this diagram, the base of the lithosphere lies at a depth of about 60 km, and the overall density of the lithosphere is equivalent to that of the underlying mantle. Given some drag from downgoing eclogitised parts of the plates, recycling of this crust to the mantle is possible.

It might also be argued that the recycling that took place in the Archean was different from modern subduction – that only the base of the crust was involved (Johnson *et al.*, 2013) or that the crust foundered as irregular diapirs (van Hunen and van den Berg, 2008). This leads to consideration of the strength of the oceanic lithosphere, which may have been reduced by higher temperatures in the crust and mantle, or the influx of mantle melts (Sivoza *et al.*, 2010). On the other hand the presence of a thick interval of refractory, dry, viscous residual dunite in the mantle portion of the lithosphere would have added rigidity.

Another variant on this theme is the question of whether the temperature in the upper mantle remained constant or decreased in a regular manner through the Archean (Herzberg *et al.*, 2010). In Section 7, I discuss various interpretations of the uneven distribution the U-Pb zircon ages of crustal rocks and develop a model that invokes irregular mantle convection and episodic rise of major mantle plumes. In this model, the arrival of the plumes in the upper mantle accelerates the rate of subduction, leading to pulses of crustal growth. In the present context, what is particularly relevant is the variation in the temperature of the upper mantle – it is high after the arrival of the plumes because heat is transferred from the plumes but relatively low before their arrival. This means that the oceanic crust at those times was thin and the plates rigid and subductable.

What emerges from this discussion is that opinion remains divided on whether subduction was possible through the Archean. Modelling of the density and dynamics of ancient oceanic crust produces equivocal results, as does the search for “petrogenetic indicators” of plate tectonics. The absence of blueschists and ophiolites may simply be a consequence of moderately high mantle and



crustal temperatures and may provide little direct information about the tectonic style. The presence of rocks with the geochemical compositions of boninites and other subduction-related lavas in >3.7 Ga regions (Polat *et al.*, 2002) counters the apparent absence of the other petrogenetic indicators of plate tectonics. Nonetheless, in spite of the uncertainty surrounding this crucial question, many authors reject the plate tectonic interpretation and have developed alternative models for the formation of granitic melts. What are these models, and how credible are they?

5.4 Sagduction and Similar Models

Bédard (2006) proposes that the granitic rocks of the continental crust are produced by partial melting of the lower portions of a thick pile of basalts. In many respects the model is not new, resembling as it does the ideas expressed by *Green* and *Ringwood* some forty years ago. Where it, and other more recent models (e.g., *Abbott* and *Mooney*, 1995; *Zegers* and *van Keken*, 2001; *van Thienen et al.*, 2004; *Van Kranendonk*, 2010; *Condie*, 2011), differ is that the basalts are said to have erupted as part of an oceanic plateau, the product of melting in a mantle plume. In *Bédard's* model (Fig. 5.7), the oceanic plateau first forms from picritic to basaltic magmas generated from the plume. Then heat is transported into the plateau within magmas (from the same plume?) and this causes partial melting of earlier basalts and the formation of tonalitic magma. The ascent of buoyant tonalitic magma into the upper crust triggers convective overturn of the crust and produces the dome-and-basin structure of Archean cratons. Large bodies of dense eclogitic residue penetrate deep into the mantle where they may guide the ascent of new mantle plumes; smaller bodies of dense residue worm their way into a newly arrived plume, causing additional melting and the generation of more mafic and even ultramafic magmas.

The model is marvellously flexible, being able to produce not only the granitoids of the continental crust but also the sub-continental lithospheric mantle (SCLM) and a large range of other rock types. *“Physical addition of delaminated crustal restites would refertilise the refractory mantle, allowing extraction of additional melt increments, and might explain the ultra-depleted and orthopyroxene-rich nature of the SCLM. A hybrid source composed of 10% eclogitic restite of ... tonalite generation, mixed with harzburgitic residues from 25% melting of primitive mantle, yields model melts with trace element signatures resembling typical Munro komatiites. Variations in the mineralogy and geochemistry of the delaminated component might account for the diversity of komatiite types. Degassing of hornblende-rich delaminated restites would transfer large ion lithophile elements (LILE) to surrounding depleted mantle and could generate boninites. Fusion of undepleted metabasalt sandwiched among denser restites could generate sanukitoids. Mantle melt pulses generated by catastrophic delamination events would underplate nascent TTG crust and trigger renewed crustal melting, followed by delamination of newly formed eclogitic restites, triggering additional mantle melting, and so on.”* (Bédard, 2006, page 1188).



But, to repeat the question posed above, is the model credible? In my opinion it has several minor inconsistencies and one major flaw. Without water, basalt is a miserly source of partial melt. As summarised by Moyen and Stevens (2006) and discussed in Section 4.1, an anhydrous basaltic source yields only small amounts of melt and the composition of this melt differs from that of Archean TTG. More significantly, the mineralogy of common granites, most notably the near ubiquitous presence of hydrous minerals such as amphibole and biotite, provides indisputable evidence that their source contained water. Indeed, in *Bédard's* modelling of the generation of tonalitic magmas at the base of the oceanic plateau, he assumes that the source was hydrous and that amphibole fractionally crystallised.

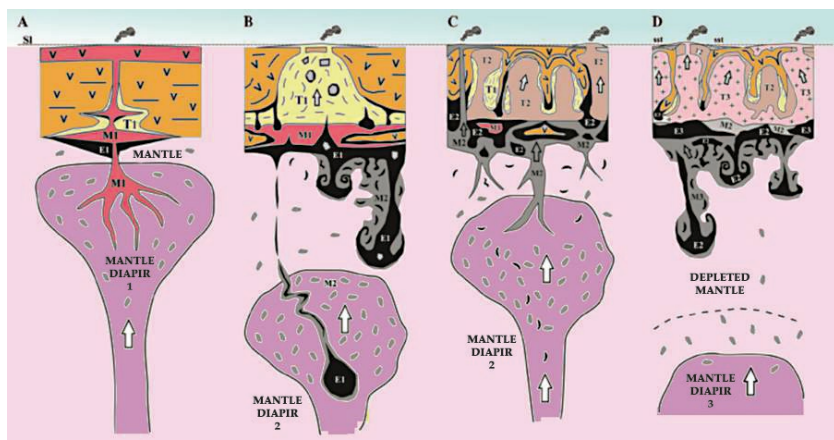


Figure 5.7 Jean Bédard's model for the formation of granitic magmas at the base of an oceanic plateau (modified from Bédard, 2006).

Ample theoretical and observational arguments support the notion that the lower parts of an oceanic plateau are dry. The first argument depends on the relationship between temperature and mineralogy in mafic crustal assemblages. To meet the requirements of producing the characteristic HREE depletion of TTG, the source basalt must melt in the field of stability of garnet, at a pressure of at least 10 kbar or a depth of about 30 km. Assuming a temperature gradient of 30 °C/km (a modest value for an oceanic plateau above a mantle plume) yields a temperature of 900 °C, which, as seen in Figure 5.6, lies well above the limit of the amphibole stability and within either the granulite or the eclogite facies. At this temperature, amphibole and biotite have broken down to anhydrous pyroxene and feldspar and/or garnet. This interpretation is likewise supported by various geological and geophysical observations. Spear (1993) predicts a transition from amphibolite to granulite facies metamorphism at the base of relatively thin (6-9 km) modern oceanic crust and IODP drilling through the oceanic crust has revealed mineralogical transformations that correspond to those predicted

from the thermal gradient. Rupke *et al.* (2004) report a transition from 1.5-3% H₂O in upper 1 km of lavas, to near zero at 3 km depth. Studies of exposed sections through mafic-ultramafic complexes in the basal parts of the ancient Kohistan and the Talkeetna island arcs in Pakistan similarly reveal anhydrous mineral assemblages that record temperatures of 800– 1000 °C at pressures of 10-15 kbar (Jagoutz, 2010). In addition, even if water were able to penetrate deep into the lower part of an oceanic plateau or thick ancient oceanic crust, the refractory, cumulus, mafic-ultramafic nature of the rocks that constitute the lower portions of such crust make them extremely infertile sources of evolved magmas.

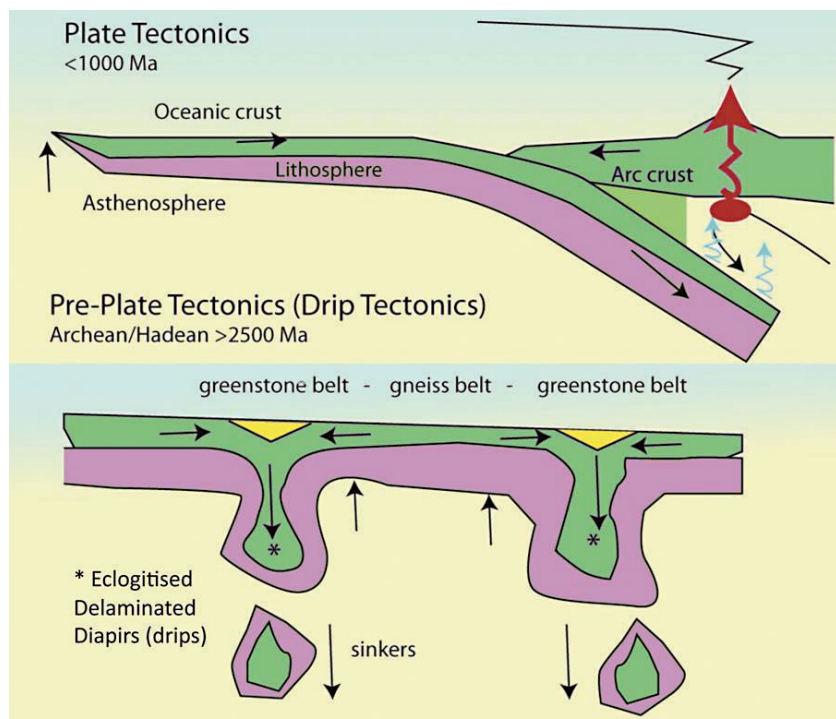


Figure 5.8 A comparison of plate and drip tectonics (from Stern, 2013, with permission from the Geological Society of America).

Advocates of the oceanic plateau model propose that submarine basalts make up most of the plateau and that “*this affords the opportunity for uptake of H₂O*” (Bédard, 2006, page 1195). This is incorrect. There are sound theoretical and observational arguments for proposing that oceanic plateaus are layered, and that basaltic lavas are restricted to the uppermost portions. The simplest argument depends on the difference in composition between that of the erupted basalts (as sampled during drilling of the Ontong Java plateau, Fitton *et al.* (2004)) and



the compositions of primary melts generated by partial melting of the mantle deep below the plateau. The latter are picrites – highly magnesian magmas that crystallise a large amount of olivine, pyroxene and plagioclase before evolving to the erupted basalt composition. The crystallised phases are retained in the lower part of the plateau while the less-dense evolved basaltic liquids migrate to the surface (Cox, 1980). Farnetani *et al.* (1996) predict such evolution for the Ontong Java plateau and demonstrate that seismic velocities measured in the lower two-thirds correspond to gabbros and mafic-ultramafic cumulates. Kerr *et al.* (1997) find evidence for similar stratification of an oceanic plateau in obducted portions of the Caribbean large igneous province.

I emphasise the layered structure of oceanic plateaus because in such a structure the lavas that erupted under submarine conditions and may have interacted directly with seawater are limited to the uppermost part of the plateau. These relatively porous and fractured rocks can indeed contain high water contents, but they are found only in the uppermost 1-5 km of the 20-30 km thick plateau. Circulation of seawater by diffuse downward flow may penetrate into the upper portions of the gabbroic layer, but these more compact, glass-free rocks are less susceptible to alteration and the depth of such circulation again will be no more than a few kilometres.

Robin and Bailey (2009) discuss the distribution of water in volcanic piles during their numerical modelling of the dome-and-basin structure of old Archean cratons. They predict rapid generation of negatively buoyant diapirs of dense volcanic rocks that descend from near the surface to the base of the crust in about 10-100 Ma, and they propose that these descent rates will allow the water to be retained in hydrated volcanic rocks as they descend onwards into the mantle. Although certain aspects of the model can be questioned (it requires rapid eruption and serpentinisation of a thick layer of komatiite above a layer of felsic crust, which is not an obvious disposition of these rock types), it would seem initially to provide a possible explanation for downward transport of hydrated volcanic rocks to the levels where temperatures are high enough to cause partial melting. But is the descent rate sufficiently high? A modern oceanic slab subducting at 10 cm yr^{-1} reaches a depth of 100 km in about 1 million years; during this time, the hydrous minerals in altered crust break down and release their water well before the temperatures required for partial melting are reached. In a diapir that requires 10 to 100 Ma to reach mantle depths, the minerals will have completely dehydrated before arriving at the base of the crust.

Returning to *Bédard's* model, let us consider what happens to the basalt/eclogite pods (Fig. 5.7) as they descend through the crust and founder beneath a basaltic plateau. Let us suppose that, contrary to the arguments developed in the previous paragraph, some hydrous minerals remain within the pods as they descend into the mantle. The fluids released would ascend upwards into the overlying parts of the crust or diapir where they encounter only rock that is *cooler* than at the source. In such a situation, no water-fluxed melting will take place.



Compare this to what takes place in a subduction zone (Figs. 5.8, 5.9). As explained more fully in the following section, the fluids released from dehydrating oceanic crust move upwards into the mantle wedge where they encounter a continuous flux of hot mantle being drawn in by forced convection. A copious source of aqueous fluid meets a continuous supply of hot peridotite – just what we need to produce abundant hydrous magma!

The partial melting in the wedge does not produce felsic magma. This takes place at the bases of island arcs or active continental margins, where basalt is reprocessed into more felsic magma. During this process, the water needed for the process is supplied by the hydrous magmas generated within the subduction zone (Figs. 5.8, 5.9). Such is not the case at the base of an oceanic plateau, which are built up from magmas from the mantle plume. Plume-derived magmas have low to negligible water contents.

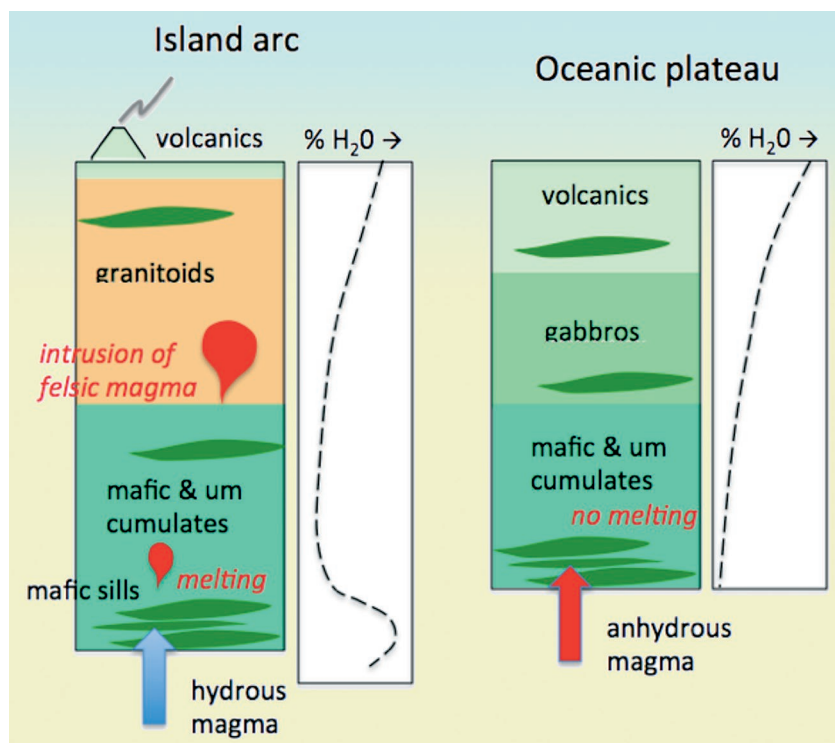


Figure 5.9

Comparison of a section through an island arc (or convergent margin) and an oceanic plateau (or thick Archean oceanic crust). The graphs on the right of each section schematically show the variation of water content, which decreases downwards through the crust. Water is supplied to the base of the crust of the island arc in hydrous magmas but not to the base of the oceanic plateau where



plume-derived magmas are anhydrous. The input of hydrous magma at the base of the arc allows the partial melting of previously intruded basalts and the generation of felsic magmas. These ascend into the crust and solidify as granitoid intrusions. Little to no melting takes place at the base of the oceanic plateau which consists of refractory, dry mafic and ultramafic (um) cumulates.

Before leaving this subject I will touch on two other questionable aspects of the oceanic plateau model: the composition of partial melts of plume-derived basalt, and the timing of plume impact at the base of the plateau.

An argument used in defence of the oceanic plateau model is that felsic magmas with many of the geochemical characteristics of continental granitoids are found in Iceland. Willbold *et al.* (2009) report a continental geochemical signature (a calc-alkaline composition, strong enrichment in Na relative to K, high Pb/Ce, La/Nb, and Ta/Nb ratios) in dacitic volcanic rocks from Iceland and propose that the signature results from high-pressure partial melting of basaltic lower crust, followed by fractional crystallisation of amphibole, plagioclase and ilmenite. Although this model is entirely feasible for the rocks that Willbold *et al.* (2009) studied, it cannot be applied more generally to the generation of the granitoids of the continental crust. Iceland is a special case, an emergent island that formed above a mid-ocean ridge; a situation where meteoric water and seawater can penetrate several kilometres into the basaltic crust to trigger hydrous partial melting. However, as Martin *et al.* (2010) point out, even in these rather exceptional circumstances, felsic rocks make up only a very small fraction of the largely mafic Icelandic crust and these rocks do not show the strong HREE depletion that characterises Archean TTG. To generate such magma requires that garnet remains in the residue of melting, and this is possible only at depths greater than about 20-25 km – depths well beyond the limit of penetration of surface waters. Thus, although limited amounts of dacite and granophyre are indeed generated within the basaltic Icelandic crust, wholesale transformation of such crust into granitic continental crust seems highly unlikely.

An additional difficulty is that the oceanic plateau model requires either a single, long-lived plume or the arrival beneath the same portion of the crust of a series of plumes. A continual supply of magma, or successive pulses of magma, is needed to supply the heat needed to re-melt older underplated basalt. It is true that some parts of the Earth have been the target of multiple plumes – 3 or 4 plumes impinged at the base of southern African crust, for example (White, 1997) – but this occurred over a 3 billion year time interval. The Ontong Java plateau records two pulses of magma, one at 120 Ma and the other at 90 Ma, but the second was small and occurred some 30 Ma after the first (Fitton *et al.*, 2004). It would be remarkable if the plumes that build large igneous provinces, which are generated deep in the lower mantle, would repeatedly find their way to the same part of the lithosphere. In addition, the flood volcanism that constructs Phanerozoic large igneous provinces normally is remarkably short-lived – most large continental basaltic plateaus were emplaced in less than 1 million years (Courtillot and Renne, 2003). Required for the oceanic plateau model for the



generation of Archean TTG is a totally different type of behaviour; either the arrival of a series of plumes beneath the same restricted portion of the crust (perhaps guided by sinking slugs of eclogite, as suggested by Bédard, 2006) or abnormally protracted flood volcanism.

One can never dismiss a model out of hand, but I now ask the reader to compare the models in which granitic magmas form in an intraplate setting, at the base of either an oceanic plateau or thick oceanic crust, with its uniformitarian alternative; *i.e.* generation of granitoids in a subduction setting.

5.5 Generation of Granitic Magmas in Subduction Zones

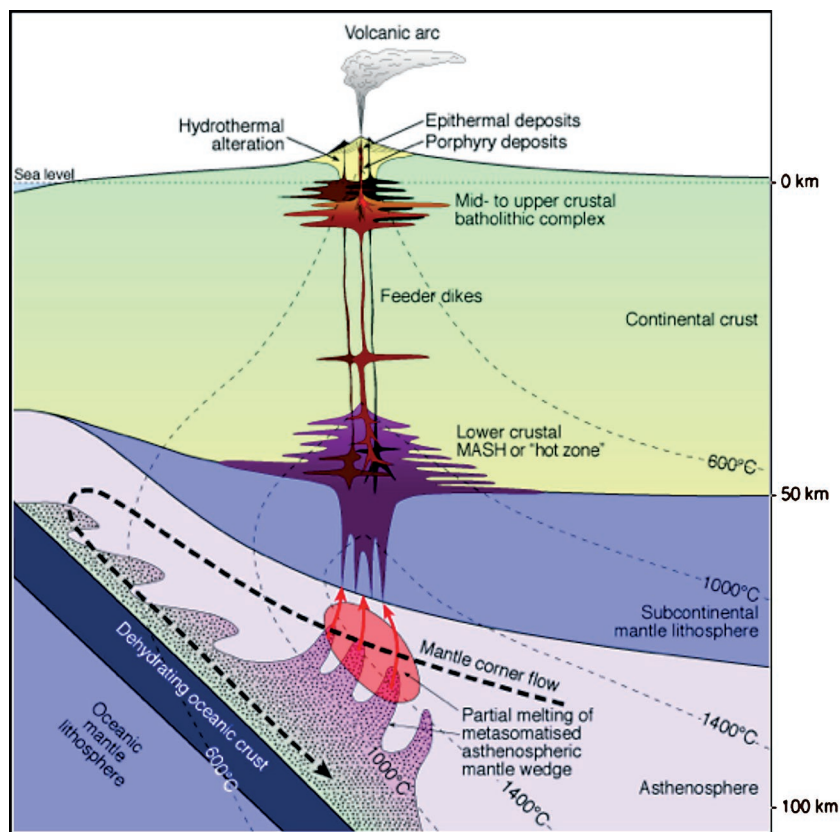


Figure 5.10 A sketch of the processes involved in the formation of granitic magmas (from Richards, 2011, with permission from Elsevier).



Figures 4.3, 5.2 and 5.10 show why a subduction zone is such an efficient granite factory. Oceanic crust becomes hydrated first at the mid-ocean ridge and then later as the plate bends and fractures before entering the subduction zone. If the lithosphere is young and hot, fluids retained within the subducting slab may trigger melting of the basaltic oceanic crust to produce adakitic melts (Defant and Drummond, 1990). More commonly, the fluids released by destabilisation of hydrous minerals ascend upwards into the mantle wedge where they trigger partial melting in the mantle wedge (Grove *et al.*, 2006). Forced convection driven by viscous coupling between slab and overlying mantle drags mantle material down to the site of melting in the subduction zone. A conveyor belt is thereby set up that continually transports a supply of hot peridotite to an ascending stream of fluid – an ideal set-up to produce abundant hydrous basaltic magma. This magma migrates to the base of the crust, *i.e.* to the base of an island arc or an active continental margin, and there it undergoes the complex processing that eventually yields granitic magma. A combination of fractional crystallisation of the incoming basalt, partial melting of older solidified basalt, and assimilation or mixing of crustal material with the new magmas produces tonalitic magmas. These magmas segregate from their crustal source and move up into the crust and evolve to more siliceous granitic melts (Fig. 5.10). Variations in the composition and mineralogy of the source (including the proportion of juvenile magma and reworked crustal rocks), the composition of incoming magmas, the degree of partial melting, and the extent of fractionation crystallisation in crustal magma chambers all contribute to the diversity of composition of the granitoids.

Research on the Continental Crust in Mainz – During the mid 1980's a large proportion of the members of the Geochemistry Division of the Max Planck Institut in Mainz were working on subjects related more or less directly with continental crust formation. One group, led by *Alfred Kröner* of the University of Mainz was working on the lower continental crust of India and Sri Lanka; *Wolfgang Todt* used mainly the U-Pb zircon method to date granites from various locations in Europe and elsewhere, and *Al Hofmann* and *Hans Voshage* were conducting joint research with *Georgio Rivalenti*, *Silvano Sinigoi* and *Maurizio Mazzucchelli* on the Ivrea Zone in Italy. There they obtained intriguing trace element and isotopic data from the mafic complex at the base of the lower crustal section and interpreted these data to indicate that almost the entire complex was massively contaminated with metasedimentary country rocks (Voshage *et al.*, 1990). At the same time I was working with *George Jenner* on komatiites from the Kambalda region of Western Australia and we found that these rocks shared much of the same geochemical characteristics as the Ivrea rocks. In the first manuscript that *George* and I submitted, we followed the line of argument strongly promoted by the British school of mantle geochemistry and proposed that the komatiites had come from "an enriched source in metasomatised sub-continental lithospheric mantle". On a field trip to the Ivrea zone, *Al*, *Georgio* and *Maurizio* showed us migmatitic melting of metasediments surrounding the mafic intrusion and abundant enclaves of melt-depleted metasediment within the intrusion. In combination with their geochemical



data, this convinced us of the validity of their interpretation and this led *George* and I to radically change our interpretation of the komatiites. We subsequently published a paper called “Crustally contaminated komatiites and basalts from Kambalda” (Arndt and Jenner, 1986), and since then I have remained extremely sceptical about the entire notion that “metasomatised sub-continental lithospheric mantle” is the source of high-volume mafic magmas (see Arndt and Christensen, 1992; Arndt *et al.*, 1993, 1998; Arndt, 2013).

A little later *Steve Goldstein* and I started thinking about how mafic magmas would interact with the rocks at the base of the continental crust, and how these processes might be linked with broader questions related to the evolution of the crust and the mantle. One issue that intrigued us had been raised a few years earlier in an insightful paper by *Claude Herzberg* (1983). *Claude* was interested in the densities of the rocks produced by melting and differentiation at the base of the crust and he pointed out that the Fe-rich residues of crustal melting or fractional crystallisation should be denser than the underlying mantle and should not remain in the crust. We took the argument further and explored the relative volumes and densities of the differentiation products of basaltic magmas at the base of the crust. The bottom line was that the extraction of a tonalitic melt from a basaltic source leaves a residue that has four to five times the volume of felsic melt. In other words, the generation of the continental crust has to contend with a massive garbage disposal problem. We wrote this up in a manuscript called “An open boundary between lower continental crust and mantle: its role in crust formation and crustal recycling” (Arndt and Goldstein, 1989). When we first showed the manuscript to *Al Hofmann*, who normally was very supportive, he remarked that it was not particularly innovative but thought that if we added a veneer of numerical modelling to make it respectable, we should be able to get it published. Google Scholar tells me that the paper has since been cited 204 times.

If we assume that the tonalitic melt that remains in the continental crust results from 20% partial melting of a solid basaltic source, or 80% crystallisation of a basaltic liquid, the mass of the residue is 4 times greater than that of the resultant felsic magma. The generation of a 40 km thick layer of continental crust leaves behind a 160 km thick layer of residue. This residue is richer in Fe (and, depending on the pressure and Al content, in garnet) than normal mantle peridotite and most of it (depending on Fe content and mineralogy) is denser than the underlying harzburgitic mantle (Herzberg *et al.*, 1983; Arndt and Goldstein, 1989; Kay and Mahlburg-Kay, 1991). Given enough time and appropriate physical properties of the underlying peridotite, the residue may founder into the underlying lithospheric mantle. At the time we published the “Open boundary” paper (Arndt and Goldstein, 1989), we explored whether the dense residue would become gravitationally unstable and descend as negative diapirs into the mantle, but simple calculations assuming normal lithospheric mantle rheology indicated that the times required were far too long. This conclusion remains valid for all tectonic settings like those beneath stable continental crust where the lithospheric mantle contains very little water and has a very high viscosity.



In an active subduction zone, however, the situation is different, for two reasons. First, as emphasised repeatedly in this paper, the mantle above a subduction zone has a high water content that considerably reduces its viscosity. Second, forced convection driven by the subducting slab continually drags mantle material past the region where the residue is being formed. This convection will remove the residue from beneath the mixing, assimilation, storage and homogenisation (MASH) zone and transport it first sideways then downwards where it eventually may be reprocessed during melting triggered by upwelling fluids from the subducting slab. A subduction zone has a built-in, very efficient, garbage disposal system!

Things are different beneath an oceanic plateau – there the mantle is dry. Because of its relatively high temperature, the viscosity of the root of a mantle plume will not be as low as that in stable subcontinental lithospheric mantle, and the type of residue foundering inherent to sagduction models for the formation of continental crust may not be totally unreasonable. However, a setting beneath an oceanic plateau lacks the efficient means of removing the residue that is inherent to the subduction setting.

There is now general acceptance that granitic magmas are produced in subduction zones at active continental margins and for modern and ancient island arcs. For example, it is the standard model for granitic magmatism in the Andes (*e.g.*, Hildreth and Moorbath, 1988; Kay and Mahlburg-Kay, 1991; Petford *et al.*, 2000; Otamendi *et al.*, 2009; Goss *et al.*, 2011), as well as for ancient island arcs such as those exposed in crustal sections in Pakistan (Jagoutz, 2010; Jagoutz *et al.*, 2011, 2013). There is also growing support for the generation of Archean TTG in such a setting (*e.g.*, Polat *et al.*, 2002; Nagel *et al.*, 2012; Polat, 2012). Yet other authors do not accept that this applies to the formation of the oldest TTG.

Several authors (*e.g.*, Bergman, 1987; Zegers and van Keken, 2001; Bédard, 2006; Smithies *et al.*, 2009; Van Kranendonk, 2010) have proposed that the geological setting, petrology or compositions of some or all Archean TTG preclude their formation in a subduction setting. This opinion is in part conditioned by the conviction that plate tectonics did not operate in the Archean (as discussed in the previous section), but it also stems from differences between the average compositions of granitoids in modern orogens and those of their Archean counterparts. As shown in Figure 2.1, the modern granitoids tend to be more potassic, on average, than Archean TTG, and they plot on a calc-alkaline trend that extends from tonalite through granodiorite to granite. Archean TTG, in contrast, plot mainly on a sodic, low-potassium trend from tonalite to trondjemite. A possible explanation for at least part of the difference in compositions is that the modern granites form in settings where their sources contain a higher proportion of evolved crustal rocks such as granite or detrital metasediment. The end stage of the evolution of the continental crust involves massive transfer of material from base to the top of the crust, combined with loss of refractory material to the mantle (Rudnick and Fountain, 1995). This transfer results in



an enrichment of potassium and fluid-mobile trace elements in the upper crust and corresponding depletion of these elements in the lower crust. The process generally follows major periods of crustal growth such as the one at the end of the Archean, a process that was repeated during subsequent orogenic events. The consequence is that from the late Archean to the present day, the crust as a whole has evolved and, crucially in this context, potassium-rich felsic gneisses and metasedimentary rocks have become more abundant in the sources of newly formed granitic magmas.

It should also be emphasised that granitoids with sodic, siliceous compositions like those of Archean TTG are well documented in modern orogenic belts and some of these even display the depletion of HREE that is said to distinguish the Archean granitoids. Atherton and Petford (1993) identified such granitoids in the Andes some twenty years ago and Jagoutz *et al.* (2013) recently did the same in the Kohistan island arc.

The Birimian Field Trip – In the late 1980's I went on a three-week field trip to the early Proterozoic, 2.1 Ga Birimian belt in Ivory Coast, West Africa. The idea was to study this belt, which can be seen as a transition between Archean terrains like the granite-greenstone belts of Canada and Australia, and the Proterozoic Trans-Hudson orogeny, which seemed to have formed by process very like modern plate tectonics. *Francis Albarède* had organised the trip and he, *Don Lowe* (from Stanford University and an expert on Precambrian sedimentary rocks) and I spent a lot of time debating the origin of the rocks and structures that we saw. For my point of view, the trip was not scientifically very fruitful: all we saw were pillow lavas, usually altered, some calc-alkaline volcanics, various types of sedimentary rocks, and a few granites; nothing extra-ordinary, I thought, so I took very few samples.

Francis was smarter and collected two comprehensive suites, one of the volcanic rocks that became the basis of the thesis project and excellent paper of *Wafa Abouchami* (Abouchami *et al.*, 1990); the second of granitoids, which likewise gave rise to the thesis and paper of *Muriel Boher* (Boher *et al.*, 1991). In these papers *Wafa*, *Muriel* and *Francis* developed the idea that oceanic plateaus were directly implicated in the creation of Precambrian continental crust. *Francis* introduced me then, in about 1987, to the notion of oceanic plateaus – before that I, like most of the geological community, had never heard of them. Figure 5.11, replete with its Viking ship fleeing flood volcanism, shows the model developed in *Muriel's* paper: an oceanic plateau forms above a mantle plume and subsequently focuses subduction at its margins to produce the granitoids of the continental crust. In its essence, this is the model that I defend in the present paper.



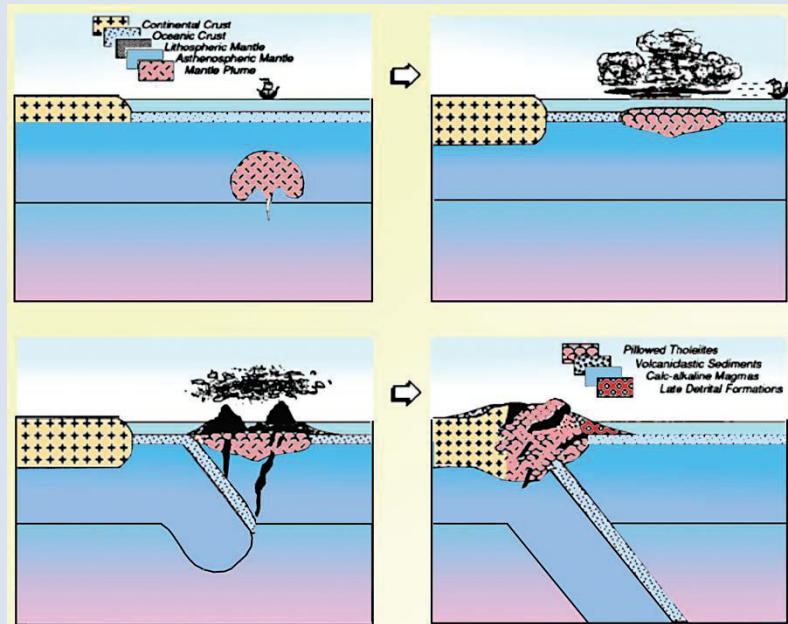


Figure 5.11 A model for the formation of continental crust at the margins of an oceanic plateau (modified from Boher *et al.*, 1991).

Some specific aspects of models for the formation of Archean continental crust deserve further comment:

5.5.1 Melting of Subducting Oceanic Crust

From what I can establish, the hypothesis that Archean TTG result from partial melting of subducting oceanic crust was originally proposed by *Bor-ming Jahn* from the University of Rennes in France in a paper describing the origin of granitoids from Finland (Jahn *et al.*, 1980). *Hervé Martin*, now at the University of Clermont-Ferrand, has since developed and has strongly defended the hypothesis over the past three decades (Martin, 1987, 1994a,b; Martin *et al.*, 2005). In this model, Archean oceanic crust, being hotter than most modern crust, reaches its solidus before the breakdown of the major hydrous minerals, and it melts directly to produce felsic magmas (Fig. 5.12). A comparison can be made with modern adakites, which are generated when unusually young or hot oceanic crust subducts (Defant and Drummond, 1990b). The model has many attractive features; in particular, it explains the characteristic depletion of HREE of Archean

TTG in terms of melting of garnet-bearing eclogitic crust and attributes the transition to normal granitoids that lack the HREE depletion to gradual cooling of subducted crust and a transfer of the site of melting to the mantle wedge. It has been criticised, however, because it fails to explain the unusually low Ni, Cr and Mg# of some early Archean TTG, as discussed in the following section. My opinion is that slab melting probably did occur in the Archean and was no doubt more common than in modern tectonic settings, but, then as now, the main source of granitic melt was in the lower part of the island arc or continental margin, and not within subducting oceanic crust.

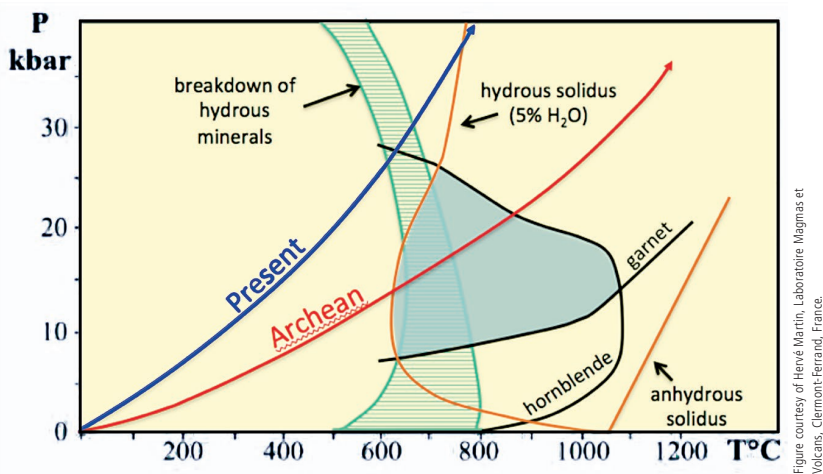


Figure 5.12 Hervé Martin's model for the formation of TTG in the Archean shows the wet solidus (in orange) and the limits of stability of hydrous minerals in the subducting slab (in green). The P-T paths of present-day and Archean subducting slabs are shown as blue and red arrows. Hot Archean oceanic crust reaches its solidus before the main hydrous phases break down, and the basaltic component (converted to garnet amphibole or eclogite) partially melts. In modern subduction zones, the oceanic crust dehydrates before the solidus is reached. The fluids escape into the mantle wedge where they produce hydrous basaltic magmas.

5.5.2 Low Ni, Cr and Mg# and Magma-Mantle Interaction

Smithies and co-workers (Smithies and Champion, 2000; Martin *et al.*, 2005) have noted that the oldest, pre 3.4 Ga granitoids from the Pilbara and Barberton regions contain lower concentrations of Ni and Cr and lower Mg# ($(\text{Mg}/(\text{Mg}+\text{Fe}))$) than the younger granitoids from these and other regions. The higher values in the younger granitoids are generally attributed to interaction of siliceous melts from the subducting slab with peridotite of the mantle wedge. Initially it was



proposed that the Archean TTG were derived from a slab that subducted at such a shallow angle that negligible mantle lithosphere intervened between the source and the base of the crust (Smithies *et al.*, 2005, 2007). This model was subsequently discarded and instead the low Ni and Cr and lower Mg# are cited as evidence for the “oceanic plateau” model. The younger Pilbara granitoids are said to have formed in a setting that resembled modern subduction while the older ones formed by melting at the base of a plume-generated oceanic plateau (Smithies *et al.* 2009; Van Kranendonk, 2010).

In my opinion, the change to higher Ni and Cr and Mg# of the younger TTGs can be linked to an evolution in the composition of the mantle wedge and the lithosphere beneath the continents. If we accept that the continental crust, and the root of mantle peridotite that underlies it, formed in large part during the Archean, as argued in Section 7, then the refractory nature of the lithosphere would not have fully developed at the time of formation of the oldest TTG. More specifically, the oldest TTG of the Pilbara – those with low Mg#, Ni and Cr – would have formed as part of the earliest phase of crust formation in that region, and before the development of a refractory lithospheric mantle root. The high Mg# and high Ni and Cr contents of the cratonic lithosphere are attributed to the extraction of partial melt which is thought to have accompanied the formation and stabilisation of the continental crust (Boyd and Mertzman, 1987; Pearson *et al.*, 2004). The magmas parental to the oldest Archean TTG would have formed from, or interacted with, mantle that was more primitive and more fertile; *i.e.* with lower Mg#, Ni and Cr, than that implicated in the formation of younger granitoids.

5.5.3 Depletion of HREE in Archean TTG

One final, outstanding, question remains to be answered – how can we explain the depletion of HREE and high Sr contents of Archean TTGs? It is generally accepted that these features result from melting at greater-than-normal depths where plagioclase is unstable and garnet is stable. From the experimental work summarised in Section 2, this depth is not very great, probably around 20-25 km. These depths correspond to those in the intermediate to lower part of the crust beneath an island arc or an active convergent margin. Any magma produced at these levels in the crust, either by partial melting of a basaltic source or fractional crystallisation of a basaltic liquid, could have left garnet in the residue and could have acquired the characteristic depletion of the HREE. To produce the more extreme depletion observed in early Archean TTG requires that a large amount of garnet was left in the residue. The critical question is why should this have occurred more commonly in the Archean than at later times.

I suggest two possible explanations. First, because of higher temperatures in the Archean mantle and subducting oceanic lithosphere, the degree of partial melting in the mantle wedge would have been higher and the volume of magma produced would have been greater. This resulted in a thicker pile of basalt at the



base of the arc or convergent margin and therefore higher pressures at the site of melting. The second explanation depends on the experimental work of Müntener *et al.* (2001) who showed that a high water content stabilises garnet in mafic or ultramafic materials. If, as I argued in Section 5.3, the Archean oceanic crust were more hydrated than modern crust, it would have released more fluids to the mantle wedge and would have generated abundant water-rich magmas. These magmas transported water to the base of the crust where they stabilised garnet at the site of melting. With the decline in mantle temperatures, both the volume of basaltic melt and the quantity of aqueous fluid released from subducting oceanic crust declined, decreasing the depth and the amount of water in the MASH zones that generate granitic magmas.



6.1 Introduction

Figure 6.1 (left) is a geological map of what I see when I look out of my window here in Grenoble, a view of the stacked thrust sheets of limestones and marls of the exterior zone of the French Alps. The satellite image on the right shows the distinctive “dome-and-basin” structure that characterises old (>3 Ga) granite-greenstone terrains. The photo is of the old yellowing cover of my copy of a book called *Archean Geology* by Glover and Groves (1981). I checked on Google Earth for more recent images of the region but none illustrates the pattern as well. In the image, we see rounded, light-coloured, granitoid batholiths surrounded by cusp-shaped belts of dark-coloured metavolcanic and metasedimentary rocks. The pattern differs radically from that of the portion of the Alps in the left-hand image; there we see linear belts of mainly flat-lying volcanic and sedimentary rocks, intruded by sparse granitoids and cut by low-angle thrusts.

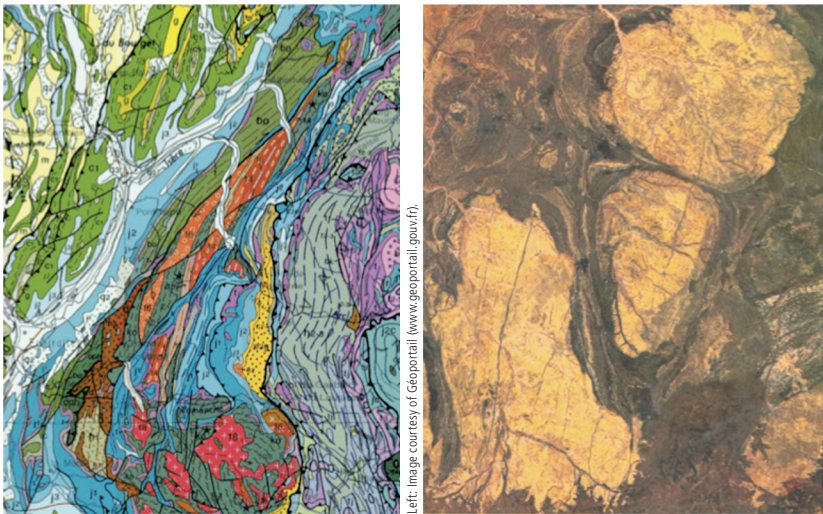


Figure 6.1

Two images of portions of the continental crust. On the left we see part of the geological map of France. Grenoble is in the middle of the Y-shaped valley in the top left of the image. On the right I show an old satellite image of the granitoid domes and cusp-shaped greenstone belts of the Pilbara region of Western Australia (from the cover of *Archean Geology* by Glover and Groves, 1981).

The difference between the two structural styles is the focus of yet another debate in the general theme of formation of the continental crust. Once again there are two schools of thought. One of these holds that the style of deformation in Archean granite-greenstone belts was not significantly different from those of the younger orogenic belts (Bickle *et al.*, 1980; de Wit, 1982; Myers and Watkins, 1985; White *et al.*, 2003; van der Velden *et al.*, 2006). In this interpretation, the domes and basins are interpreted as a fold-interference pattern. In this essentially uniformitarian view of the structure of the early crust, plate tectonics is assumed to operate, and the deformation is attributed to lateral plate movement. The compressive regime during deformation of the crust resulted in crustal thickening facilitated by numerous low-angle thrusts, and multiple episodes of deformation triggering several periods of folding.

The other school believes that the “dome-and-basin” structure is peculiar to the Archean, particularly the older parts, and reflects unusual conditions in the continental crust of that time (Bouhallier *et al.*, 1983; Choukroune *et al.*, 1995, 1997; Chardon *et al.*, 2002; Van Kranendonk *et al.*, 2004; Robin and Bailey, 2009). Vertical rather than horizontal movements are thought to dominate the deformation. The granitic domes are interpreted as rising diapirs of low-density felsic material and the surrounding supracrustal sequences as downward-moving portions of denser metavolcanic and metasedimentary rock. For such movements to occur, the lithosphere must have had a lower viscosity than that of the present crust.

Research on Archean structures and Tectonics by the Team from Geoscience Rennes

– When I arrived at the University of Rennes in 1990 after my stint at the Max-Planck-Institut in Mainz, *Pierre Choukroune* was just becoming interested in the Archean. *Pierre* is a highly experienced structural geologist who had previously worked in younger mountain belts, particularly the Pyrenees and the Alps. A few years earlier, *Raymond Capdevilla* had led a group of petrologists and geochemists from Rennes – *Sylvain Blais*, *Bernard Auvrey* and *Bor-ming Jahn* – to study Archean and Proterozoic granitoids and greenstone belts in Finland, a project that led to the publication of a series of important early papers on Archean petrology, geochemistry and tectonics (*e.g.*, Jahn and Shih, 1974; Blais *et al.*, 1978; Martin *et al.*, 1983). Soon after that, an exchange program was set up between *Jean-Jacques Peucat* and *Hervé Martin* in Rennes and Professors *Mahabalashwar* and *Jayananda* in the University of Bangalore in India with the aim of investigating the geochronology and petrology of the granitic rocks of the Indian Shield (Jayananda *et al.*, 1995).

Pierre Choukroune led a number of trips to the Indian shield where he and his students investigated the structural style of the granitoids and their enclosing supracrustal sequences. In 1992, I went on one of these trips with the intention of following up the petrological-geochemical studies of granitoids that we had started in Mainz. To my disappointment, I saw very few rocks that I was happy to interpret as the products of crystallisation of granitic magmas. We spent most of the time working on rocks that had intruded at mid-crustal levels and had been through several stages of high-degree metamorphism and deformation. The rocks were foliated or lineated, and almost all



were invaded by migmatitic patches and veins that indicated partial melting. *Pierre* used these observations to propose that the granitic domes had been emplaced not as granitic magmas, as had been suggested in some earlier studies, but as diapirs of only slightly molten granitoids and metamorphic rocks. With his PhD students, *Hughes Bouhallier* and *Dominique Chardon*, he published a series of papers (*Bouhallier et al.*, 1983; *Choukroune et al.*, 1995, 1997; *Chardon et al.*, 2002) that developed the diapiric model for the dome-and-basin structure. The diagrams copied below (Figs. 6.3-6.5) are from sketches in *Pierre's* notes.

At the same time, the debate between proponents of the two interpretations of the structure of Archean cratons was heating up. Those who favoured the fold-interference school were at that time developing structural models for greenstone belts that incorporated abundant low-angle thrusting and structural duplication of the volcano-sedimentary sections. This debate also continues to the present day, as reflected by the ongoing allochthonous vs. autochthonous interpretations of greenstone belts in Canada and Zimbabwe (*Kusky and Kidd*, 1992; *Bickle and Nisbet*, 1993; *Bickle et al.*, 1994). *Maarten de Wit* of the University of Cape Town had published an interpretation of the Barberton Belt in South Africa (*de Wit*, 1982) that incorporated many of these ideas, much to *Pierre's* dismay. He was convinced that the structures he had studied in the Indian craton, and in other Archean regions in China and West Africa, were fundamentally different from those that he had mapped in the Pyrenees and the Alps. In 1986 we both went on another field trip, led by *Steve McCourt* of the University of KwaZulu-Natal who showed us the structures and geology of the Limpopo and Barberton Belts. What we saw only reinforced *Pierre's* idea that the deformation and structure in the Barberton sequence was very different from that in the younger mountain belts and provided ideas in his subsequent papers on Archean structures and tectonics.

Pierre had previously worked in the northern and central parts of the continent, and was amazed by the excellent roads, public buildings and infrastructure of the Barberton region. Throughout the entire trip he repeated "ça, ce n'est pas l'Afrique!"



Figure 6.2 Two photos from the Barberton region.

6.2 Archean Domes and Basins

The main structural features of Archean domes and basins, as envisaged by the Rennes group, are illustrated in Figures 6.3 and 6.4. What is most striking in these images are the similarities between the structures mapped in the field in India and those observed in experiments and theoretical studies of diapirism. Most notable is the distribution of strain within the two elements of the structures. According to Dixon and Summers (1983), Brun (1983), Bouhaillier *et al.* (1983), Choukroune *et al.* (1995) and Cagnard *et al.* (2011), a diapir produced by solid-state deformation is distinguished by the distinctive patterns of lineations illustrated in the figures. In triple junctions in the downward-moving basins, both foliations and lineations are near vertical.

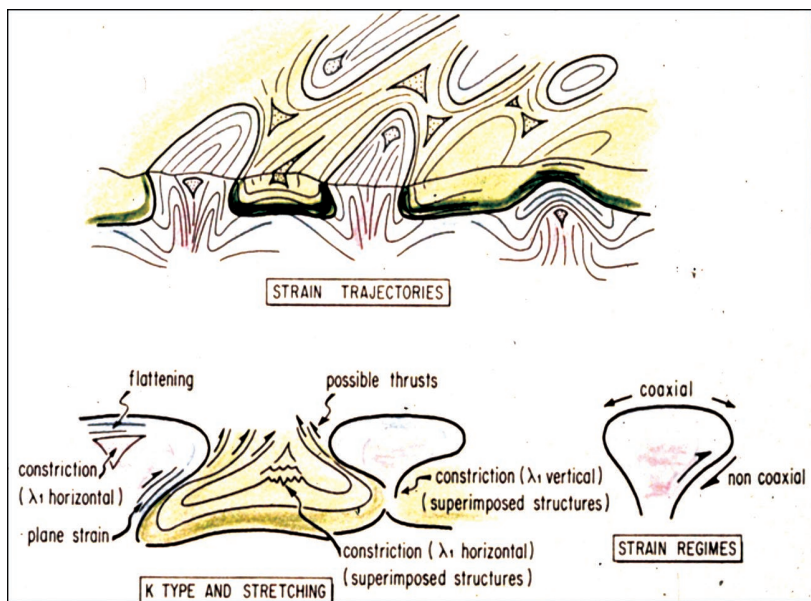


Figure 6.3 Diagrams of strain regions in a medium affected by diapirism, as inferred by Pierre Choukroune from the experimental work of Dixon and Summers (1983) and his field observations (modified from Choukroune *et al.*, 1995).

To explain these structures, the Rennes group referred to studies such as those of Dixon and Summers (1983) who had shown in centrifuge experiments that when a layer of high-density material overlies a less dense layer, if the viscosity of the two layers is sufficiently low and given enough time, diapirs of the low-density material from the lower layer penetrate upwards into the upper layer. The process has been modelled numerically by Mareschal and West (1980), de Bremond d’Ars *et al.* (1999) and Robin and Bailey (2009).



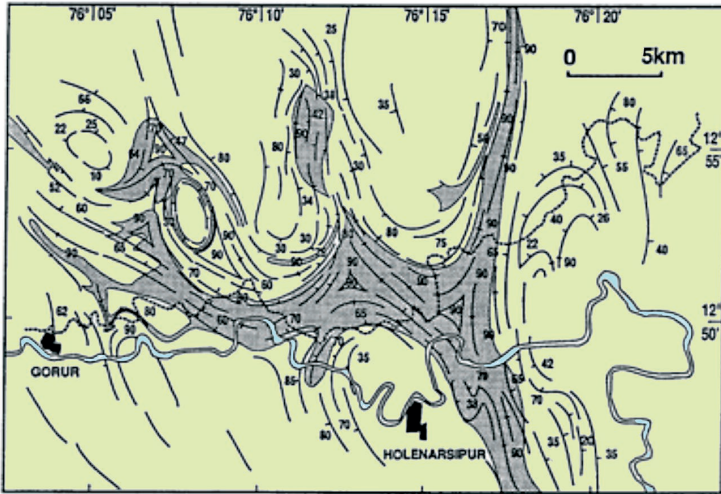


Figure 6.4

(top) Bouhallier *et al.*'s (1983) structural map of the Holenasipur belt in India showing felsic domes (colourless) and cusp-shaped supracrustal belts (grey) (modified from Bouhallier *et al.*, 1983). (bottom) Strong vertical lineation in felsic volcanic rocks in a cusp between granite domes; from south of Tjkakastad in the Barberton belt.

Another aspect of Archean regimes that intrigued the Rennes group was the large shear zones that cut through many granite-greenstone belts. These are shown diagrammatically in Figure 6.5, which shows a transition from the

dome-and-basin structure of the Pilbara to a structure created by strong horizontal shortening in the Hebei region in China. To *Pierre Choukroune*, this type of deformation provided further evidence that the rheology of the Archean crust was very different from that of the modern crust.

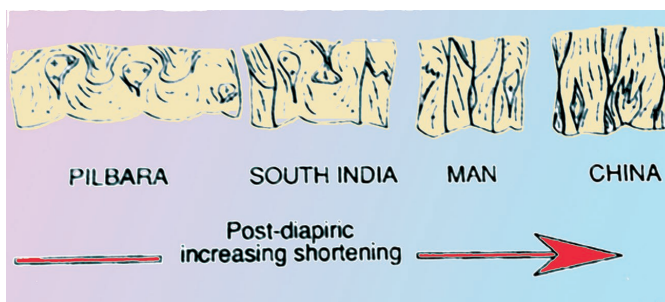


Figure 6.5 The transition from a well-developed dome-and-basin structure in the Pilbara craton, through transitional situations in the South Indian and Man (West Africa) cratons, to one of strong horizontal shortening in the Hebei region of China (modified from Choukroune *et al.*, 1995).

Alternative Models for the Formation of Archean Structures – The reader might be asking at this stage why I make so little mention of the alternative interpretation – that the structures observed in Archean cratons are the result of fold interference in a tectonic setting dominated by lateral movements. The answer is simple – I’ve never been particularly interested in structural geology. I recognise that the discipline is important but I look upon structural geologists in the same way as I look upon university administrators: what they do is essential if the system is to function and I am very glad that some people are prepared to do the necessary work. Since this is my personal perspective of crust-forming processes, I refer the reader to the large literature on the subject to get the other side of the story.



Figure 6.6 (left) A remarkable fold in flood basalts of the Ethiopian plateau. (right) Arnaud Pêcher enjoying his noodles during a trip to the Panzhihua region in China.



Evidence that I do work with structural geologists comes from my more recent collaboration with *Arnaud Pêcher* of the University of Grenoble. On one occasion I took him with me to study the Ethiopian flood basalts. *Arnaud* is mountain geologist with considerable experience in the Alps and the Himalayas, and he did not look forward to spending three weeks in a volcanic plateau where the maximum dip was expected to be no more than 3 degrees. Imagine his surprise and delight when we came across the magnificent structure (shown above) that had developed during syn-volcanic deformation of the lava pile. On another occasion we went together to China where, during the first day of his visit to the Panzhihua intrusion, he realised that the previous interpretation of its structure was quite wrong (*Pêcher et al.*, 2013).

An issue we do need to discuss is the reason why the Archean crust deformed as it did. If we now accept, as discussed in Section 5, that plate tectonics operated through most of the Archean, why is the structure of granite-greenstone belts so different? In many models it is concluded that to explain the dome-and-basin structure requires that all components of the Archean crust had low viscosities. In some models, the low viscosity is ascribed to lithology or mineralogy of the rocks of the Archean crust. De Bremond d'Arc *et al.* (1999) and Robin and Bailey (2009), for example, appeal to complete serpentinisation of abundant komatiite in a upper dense layer that erupted onto less dense granitic rocks. This type of explanation does not get us very far because komatiite is a relatively rare rock in Archean greenstone belts (de Wit and Ashwal, 1997; Arndt *et al.*, 2008) and where present the komatiitic lavas are only partially serpentinised. In addition, the starting situation adopted for these analogues or numeral experiments – a thick mafic-ultramafic layer overlying a felsic layer – is unrealistic. All Archean cratons consist of a complex assemblage of different lithologies: the volcanic package is dominated by basalt but also includes felsic volcanics and sedimentary sequences; granitic plutons intrude the greenstones and mafic bodies intrude the gneissic rocks deeper in the crust. The clue to the deformation is not to be found in the lithology or crustal structure; rather it requires that the viscosity of the Archean crust as a whole was lower than that of modern continental crust.

The mineralogy and composition of Archean granitoids and metasedimentary or metavolcanic rocks provides little indication that these rocks were less viscous than their modern counterparts. As discussed in Section 2, the mineralogical differences are limited to a little more quartz and a little less potassium feldspar, not enough to account for a major difference in their rheologies. A more promising line of investigation stems from their temperature – most probably the rocks of the Archean continental crust were less viscous because they were hotter. To explain these higher temperatures, Choukroune *et al.* (1995), in a paper in which I was a co-author, appealed to the conduction or advection of heat from plumes at the base of the lithosphere to account for higher crustal temperatures. I no longer believe this: there is ample evidence that a thick layer of depleted peridotite accumulated to form the lithospheric mantle at more or less the same time as the continental crust was generated. This layer would have insulated the crust from the source of heat. A more probable explanation is that internal heating



caused the high temperatures. As argued by Mareschal and Jaupart (2006) and Sandiford and McLaren (2006), the Archean crust and mantle were far hotter than their modern counterparts because the rate of heat production in all parts of the earth has declined exponentially as radioactive elements decayed. Although the concentrations of K, U and Th in Archean TTGs are low, much less than those of modern granitoids, the proportions of short-lived and highly active radionuclides such as ^{235}U would have been far higher. The heat produced by these nuclides in the felsic portions of the Hadean crust was sufficient to maintain the granitic material at its solidus (see Section 8). By 3.5 Ga, the age of the Kaapvaal and Pilbara Cratons, heat production had declined but remained high enough to decrease the viscosity of these rocks and allow for considerable vertical movement and the development of domes and basins. At 2.7 Ga, the crust as a whole had cooled to the extent that lateral tectonics became more important. At this stage, the granite-greenstone belts were able to acquire the linear features that led *John Ludden* and colleagues to develop plate tectonic models for the Superior Province in Canada (Kimura *et al.*, 1993; Choukroune *et al.*, 1997).



7.

EPISODIC GROWTH OF THE CONTINENTAL CRUST

In the previous section I proposed that Archean granitoids, like those of the present day, formed mainly in subduction settings. Two questions can now be posed. First, was the tectonic process that generated the continental crust invariant through all of geological time and second, was the rate of growth of the crustal constant or did it vary?

7.1 The Rate and Mechanism of Crust Formation

Figure 7.1 is a compilation of U-Pb zircon ages of rocks and minerals of the continental crust, reproduced from Condie and Aster (2010). It shows these ages are not distributed uniformly through geological time but are grouped in a number of prominent peaks, particularly in the period from 2.7 Ga to about 1.1 Ga.

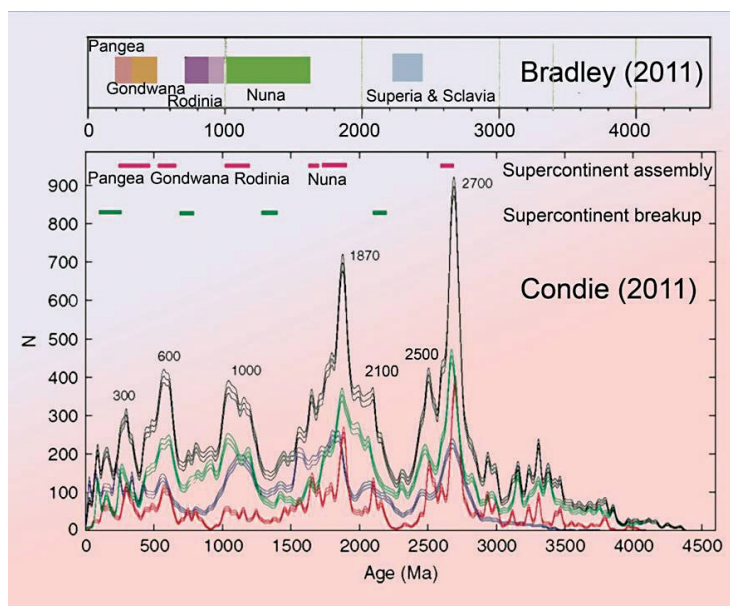


Figure 7.1

(bottom) Distribution over the past 4.5 Ga of U/Pb zircon ages in detrital zircons (blue, black and green spectra) and orogenic granitoids (red spectrum). Shown in the upper part of the diagram are the estimated times of assembly and breakup of supercontinents (modified from Condie, 2011). The peaks at 2100 and 2500 Ma are not labelled in Condie's (2011) diagram but are identified here and by Condie *et al.* (2011). **(top)** An alternative interpretation of the life cycles of supercontinents (from Bradley, 2011, with permission from Elsevier).



These peaks have been known since at least 1960 when Gastil (1960) observed that radiometric ages were distributed in a similar manner: he identified peaks at 2710-2490, 2220-2060, 1860-1650, 1480-1300, 1100-930, 620-280 and 120 to the present. As mentioned in the first section, there was some initial doubt as to whether the peaks were not an artefact of incomplete sampling of the crust, or a problem inherited from the analytical methods used at that time, but in the intervening 50 years, the peaks have been reproduced using data from all continents and by multiple precise and accurate dating methods. In addition, the peaks that are seen in ages obtained from whole-rock samples of granitic rocks are also evident in compilations of the ages of detrital zircons in major rivers, which can be assumed to sample large parts of the continental crust in a relatively unbiased manner.

Condie *et al.*'s (2009, 2011) compilation of the U-Pb zircon ages of orogenic granitoids and detrital zircons (Fig. 7.1) displays 5-7 prominent peaks separated by "troughs" or intervals within which there are far fewer data. The peaks are seen, though not with exactly the same magnitude and position, in the compilations of both the ages of granitoids and of detrital zircons. Condie and Aster (2010) identified peaks at about 2.7, 1.85, 1.0, 0.6 and 0.3 Ga, and more or less the same peaks have been identified by other authors (*e.g.*, Rino *et al.*, 2004; Wang *et al.*, 2009; Belousova *et al.*, 2010; Hawkesworth *et al.*, 2010; Voice *et al.*, 2011; Griffin *et al.*, 2013). In the other compilations, including that of Condie *et al.* (2011), two additional peaks are identified at about 2.5 and 2.1 Ga.

The peak at 2.7 Ga is the largest and in many different ways the most important. (The compilation of Griffin *et al.* (2013) is an exception in that its 2.5 Ga peak is larger, perhaps because this compilation contains abundant data from China and Australia, where zircons of this age are abundant). The 2.7 Ga peak registers a global event that affected rocks on all of the present-day continents (McCulloch and Wasserburg, 1978; Reymer and Schubert, 1986; Stein and Hofmann, 1994; Condie, 1998). Rocks formed at 2.7 Ga include a large amount of granitoid, much of which is said to be "juvenile". This term refers to granitic material that was extracted rapidly from the mantle, in most cases via a basaltic intermediary, but without a long residence time in the crust. Granitoids are identified as juvenile by their Nd, Hf or Pb isotopic compositions, which are similar to those of their mantle source. Another way to put it is to say that their model age is the same as their crystallisation age. They are distinguished from granitoids that are derived wholly or partially from older crustal rocks which have more "evolved" isotopic compositions; *i.e.* higher $^{87}\text{Sr}/^{86}\text{Sr}$ and lower $^{143}\text{Nd}/^{144}\text{Nd}$ or $^{176}\text{Hf}/^{177}\text{Hf}$ than the mantle. Granites from a crustal source usually have $\delta^{18}\text{O}$ values outside the mantle range of about 4.8 to 5.6‰.

Examples of juvenile 2.7 Ga old granitoids include those of the southern Superior province of Canada (Card, 1990), some parts of the Zimbabwe craton (Bickle and Nisbet, 1993), and in several smaller belts in Finland (Martin *et al.*, 1983), Brazil (Arndt *et al.*, 1989) and Siberia (Puchtel *et al.*, 1993). Other 2.7 Ga granitoids contain a component of older continental crust, as in the eastern Yilgarn of Australia (Hill *et al.*, 1992; Myers and Swagers, 1997) and other parts of Zimbabwe (Blenkinsop *et al.*, 1993; Nisbet *et al.*, 1993).



Large volumes of volcanic rock erupted during the 2.7 Ga event. This is the age of komatiites – *Steven Moorbath* remarked in the 1980's that all isotopic analyses of komatiites available at that time plotted within error of a 2.7 Ga isochron. These ultramafic lavas, together with abundant tholeiitic basalts, probably were derived from mantle plumes (Storey *et al.*, 1991; Herzberg, 1995; Arndt *et al.*, 1997; Barnes *et al.*, 2012): they provide clear evidence that enhanced mantle activity was part of the 2.7 Ga event. The continental flood basalts of the Fortescue (Australia) and Ventersdorp (South Africa), another manifestation of plume activity, also erupted during this period (Nelson *et al.*, 1992).

The 2.7 Ga greenstone belts also contain mafic to felsic calc-alkaline volcanic rocks that most authors are happy to assign to a subduction setting (*e.g.*, Lisher *et al.*, 1986; Kimura *et al.*, 1993). The association of the two types of volcanic activity has led to considerable debate about the apparent co-existence at that time of plumes and subduction zones (*e.g.*, Wyman, 1999; Sproule *et al.*, 2002). In regions of older continental crust, the 2.7 Ga peak is manifested as a thermal event. In the Hearne and Superior Provinces of Canada, the Pilbara and western Yilgarn in Australia, the Man Craton in West Africa, and in parts of Greenland, Antarctica and Siberia, metamorphic events of variable intensity are dated at 2.7 Ga (Brown, 2010).

In contrast to the 2.7 Ga peak, the younger events in Figure 7.1 are not global in extent but are restricted to specific parts of the present-day continents, as illustrated by the ages of detrital zircons in Figure 7.2 from *Iizuka et al.* (2010). The 2.5 Ga event is well documented in China (*Diwu et al.*, 2011) and India (*Mojzsis et al.*,

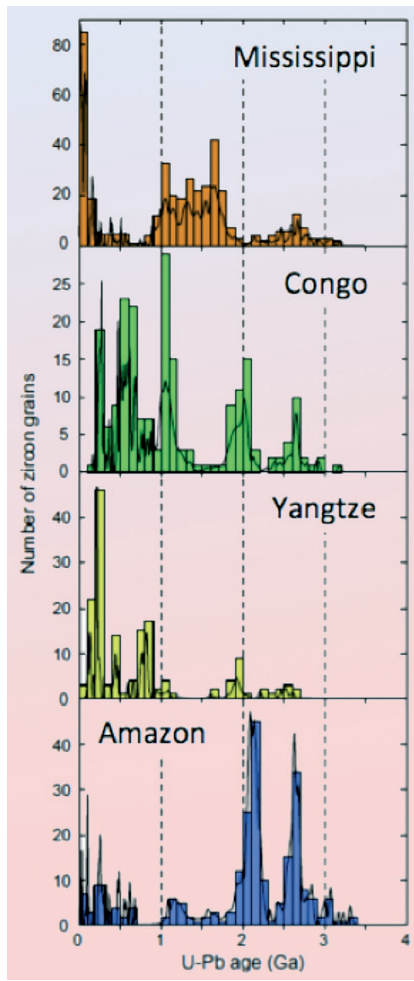


Figure 7.2 The ages of detrital zircons in four major rivers (from *Iizuka et al.*, 2010, with permission from Elsevier). Note that the 2.7 Ga peak is present in all compilations (though small in the Yangtze) but the other peaks are more sporadically distributed.

2003) as well as in the Gawler Craton of Australia (Fraser and Neumann, 2010). The 2.1 Ga peak is registered in South America (Goldstein *et al.*, 1997), north-western North America (Holm *et al.*, 2005) and much of West Africa (Boher *et al.*, 1991). Juvenile volcanic and plutonic rocks make up a major portion of the latter region (Abouchami *et al.*, 1990).

The peak at 1.85 Ga is largely absent in South America (Goldstein *et al.*, 1997) but is conspicuous in North America, Scandinavia and northern Australia (DePaolo, 1981; Patchett and Bridgwater, 1984). In many of these regions, rocks with ages corresponding to the 2.5 Ga and 2.1 Ga peaks are missing, and the 1.85 Ga crust formed adjacent to or upon 2.7 Ga crust. Under such circumstances, it is very straightforward to use isotopic data to distinguish between juvenile and recycled components. A key region is the Trans-Hudson Orogen in Canada where a variety of 1.9–1.8 Ga rocks accreted onto two separate Archean cratons. The Circum-Superior belt wraps around Superior Craton, from Labrador in the east to Saskatchewan in the west where it disappears under younger cover (Baragar and Scoates, 1981). The northern and eastern parts of the belt, which consists primarily of volcanic and sedimentary rocks, formed as a passive margin; in the west, a series of island arcs and exotic terranes accreted against the older craton. Rifting of the continent was associated with, and perhaps triggered by, the arrival of one or more mantle plumes at the base of the continental lithosphere, as along the margins of the present-day Atlantic Ocean (Storey, 1977; Courtillot *et al.*, 1999). Plume-derived magmas are readily recognised in the komatiitic sequences of the Cape Smith belt (Arndt, 1982) and the flood basalts of the Belcher Islands (Baragar and Lamontagne, 1980). What is arguably the oldest well-documented ophiolite, a relict of oceanic crust, has been found in the Purtunig region in the eastern part of the belt (Scott *et al.*, 1992).

Field Work in the Cape Smith Belt of Northern Canada – Probably the most enjoyable, and arguably the most hazardous field trips I ever made were in the volcanic belts of northern Quebec. On a trip with *Don Francis*, *Hubert Staudigel* and *Alan Zindler* we spent 3 weeks in the remote Cape Smith belt. We were dropped by helicopter at the start, without radio or any other contact and trusted that the same helicopter would return to pick us up at the end of our stay. It was a few days late and during that time all of us started to imagine hearing the chop-chop of rotor blades at all times of the day. I learnt to dislike freeze-dried food and enjoy Rusty Nails.

This trip introduced me to the excellent work being done by *Don Francis* and *Andrew Hynes* on the Cape Smith Belt, a series of Proterozoic (≈ 1.9 Ga) volcanic and sedimentary rocks that developed on the margin of the Archean Superior craton, much in the same way as the Tertiary North Atlantic Large Igneous Province formed during rifting of Greenland from Europe. The belt provides a convincing example of plate tectonic processes in the mid Proterozoic.

With *Bob Baragar* and his field crew from the Canadian Geological Survey I spent three weeks on Gilmour Island in the east part of Hudson Bay. The island is too far from the coast to be reached by a single-engine floatplane but no twins were in the region at the time we went there. The only way to get to the island was in an old Inuit fishing boat. The geology is spectacular – glacially polished outcrops of



komatiites – and it is without doubt the best place in the world to study how these lavas erupted. It was there, in 1981, that we realised that these flows were inflated from the interior by newly incoming lava, an interpretation that anticipated by about 15 years the now widely accepted inflation model for the formation of flood basalts and other types of sheet flow.

These komatiites erupted at 1.9 Ga and constitute convincing evidence that mantle plume activity immediately preceded the massive emplacement of granitoids that marks the 1.9 Ga crustal growth peak.

Hazards of this trip included polar bears sniffing around our tents and precarious navigation. One day we set out in bright sunshine towards another island that could easily be seen on the horizon only about 5 kilometres away, only for a thick blanket of fog to descend on us at mid-passage. Thirty minutes later we should have reached the island, but only saw it, well to our stern, when the fog briefly lifted. Our course would have taken us to eventually to Baffin Island, some 1000 km farther to the north.

In the Trans-Hudson belt in northern Saskatchewan and Manitoba (Lewry and Stauffer, 1990), a series of Mesoproterozoic island arcs, microcontinents and an active continental margin accreted to the older Archean platform to the northwest (Fig. 4.7). The volcanic rocks of these belts are tholeiitic and calc-alkaline and they closely resemble rocks in modern subduction settings (Stauffer *et al.*, 1975). Likewise, the felsic plutonic rocks, which comprise tonalites, granodiorites and some granites, resemble those in modern island arcs or convergent margins. Chauvel *et al.* (1987) used Sm-Nd isotopic data to demonstrate that the components in the source of the volcanic and plutonic rocks changed from almost 100% juvenile in the belts farthest from the Archean platform to about 50% recycled in the belt closest to the Archean platform (Fig. 4.7).

In the segment of the age spectrum (Fig. 7.1) younger than 1.8 Ga, the peak-and-trough distribution is less pronounced, and only the 1.1 Ga peak, which marks the well-known Grenville orogeny, is present in all compilations. In the segment older than 2.7 Ga, the spectrum is quasi continuous. A small peak around 3.3 Ga appears in data from Australia but is largely absent in data from other areas (Kemp *et al.*, 2006; Condie and Aster, 2010).

A deep trough separates the 2.5 to 2.1 Ga peaks. This trough is present in compilations of the ages of detrital zircons (DePaolo, 1981; Patchett and Bridgewater, 1984; Rino *et al.*, 2004; Campbell *et al.*, 2005; Wang *et al.*, 2009; Iizuka *et al.*, 2010) and of zircons from granites (DePaolo, 1981; Patchett and Bridgewater, 1984; Rino *et al.*, 2004; Campbell *et al.*, 2005; Condie *et al.*, 2009; Wang *et al.*, 2009; Condie and Aster, 2010; Iizuka *et al.*, 2010). These troughs are perhaps the most puzzling aspect of the age spectrum because they imply either that the granitic magmas ceased to be generated in these periods, or that the zircons produced during this interval were not preserved.

What emerges from this summary is a conclusion I will emphasise repeatedly in this section: the peaks in the U-Pb age spectra register the times of massive input of magmas from two separate sources. Mafic-ultramafic lavas



were derived from mantle plumes while granitoids and calc-alkaline volcanic rocks probably formed in a subduction setting. In each region and among the samples that define each peak, a variable – minor to major – proportion of the granitoids are juvenile whereas others contain a high proportion of recycled older continental crust.

7.2 Interpretation of U-Pb Zircon Age Peaks

The peaks in Figure 7.1 have been interpreted in two different ways. One school relates them to episodic convection of the mantle which led to short spurts of accelerated crustal growth, separated by periods of geodynamic inactivity (Moorbath and Taylor, 1981; Reymer and Schubert, 1986; Stein and Hofmann, 1994; Condie, 1998). In this interpretation, the pulses of crustal growth are attributed to (1) superplumes from the deep mantle (e.g., Condie, 1998; Isley and Abbott, 1999), (2) return mantle flow triggered by slab avalanches (e.g., Stein and Hofmann, 1994; Davies, 1995; Condie, 2004) or (3) periods of accelerated plate motion and subduction (e.g., O'Neill *et al.*, 2007).

The second school links the age peaks to the assembly of supercontinents, and thus to the global plate-tectonic cycle (Gurnis, 1988; Kemp *et al.*, 2006; Maruyama *et al.*, 2006, Campbell and Allen, 2008; Hawkesworth *et al.*, 2009; Condie and Aster, 2010; Hawkesworth *et al.*, 2010). In this interpretation, the peaks do not coincide with pulses of true crustal *growth* but to periods of enhanced *preservation* of continental crust.

7.2.1 Age peaks and the supercontinent cycle

Chris Hawkesworth and co-workers (Hawkesworth and Kemp, 2006; Kemp *et al.*, 2006; Hawkesworth *et al.*, 2009, 2010), the main advocates of the preservation hypothesis, explain it with reference to Figures 7.3 and 7.4: *“The volume of continental crust added through time via juvenile magma addition appears to be compensated by the return of continental and island arc crust to the mantle. ... if plate tectonics has been operating since around 3.0 Ga (see Cawood *et al.*, 2006) then a volume equal to the total current volume of continental crust would have been recycled into the mantle (Scholl and von Huene, 2007). The implication is that the net growth of continental crust at the present day is effectively nil, and convergent plate margins are sites of crustal recycling and reworking, as well as continental addition.”* (Hawkesworth *et al.*, 2009, page 241). Further, *“the record of magmatic ages is likely to be dominated by periods when supercontinents assembled, not because this is a major phase of crust generation but because it provides a setting for the selective preservation of crust. The preservation potential, particularly for crystallisation ages of zircons, is greater for late-stage collisional events as the supercontinents come together, rather than for subduction- and extension-related magmatism.”*



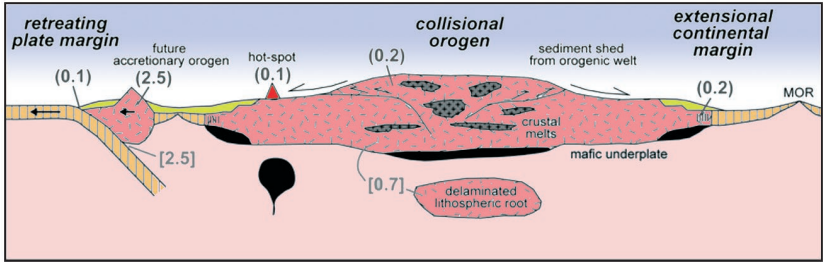


Figure 7.3 A sketch of a subduction zone and a collisional orogeny (modified from Hawkesworth *et al.*, 2010). The numbers in parentheses represent the rates of crust formation (transfer of felsic material from the mantle to the surface, in $\text{km}^3 \text{yr}^{-1}$) and those in brackets are those of crust removal.

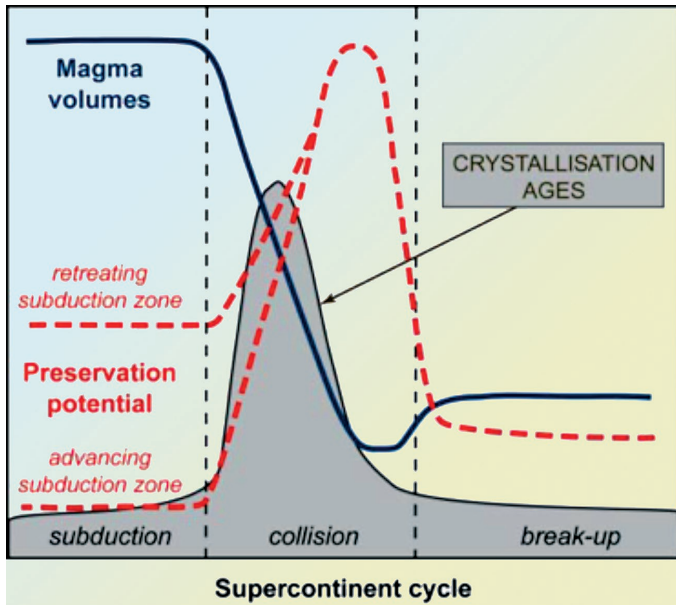


Figure 7.4 Hawkesworth *et al.*'s (2010) diagram demonstrating how crust might be preferentially preserved in collisional orogens. In a subduction zone (to the left of the diagram) the flux of magma to the surface is high, but so is the rate of crust removal and the proportion that is preserved is small. During collision, the magmatic flux declines, but not as rapidly as that of crust removal, and the preservation potential increases.

According to this hypothesis, during continent-continent collisions, granulite-facies metamorphism at the bases of collisional mountain chains stabilises new crustal material that becomes an integral part of the continent, isolated from the margins of the continents where crustal material is cycled back into the mantle (Brown, 2007; Hawkesworth *et al.*, 2010). In this interpretation, the peaks in U-Pb zircon ages do indeed register the times of crystallisation of granites or metamorphism associated with collision, but these do not necessarily coincide with periods of net crustal growth.

Kent Condie's Talk at AGU – A few years ago I attended a talk given by *Kent Condie*, probably at AGU. The talk, as usual, was clear and well argued, but its content astonished me. At that time I was not following closely the debate about the rate of crust formation, and I had not heard that *Kent*, who for many years had persuasively defended the model that related the age peaks to periods continent growth, had become a preservationist. At the end of the talk I asked him why he had changed his mind and he replied that he had recently become convinced of the validity of the preservation model. He has since expressed these views in a series of papers (*e.g.*, Condie and Aster, 2010; Condie, 2011; Condie *et al.*, 2011) and I look forward to debating the issue with him at several upcoming meetings.

7.2.2 What's wrong with the preservation model?

The following aspects of the preservation model warrant discussion:

- (1) The lack of a good correlation between the age peaks and the estimated times of supercontinent assemblage
- (2) Some questionable aspects of the capture hypothesis
- (3) The protracted duration of supercontinent assembly, which should not produce sharp age peaks
- (4) The lack of a convincing explanation for the high proportion of juvenile crust generated during the age peaks
- (5) Alternative explanations of isotopic evolution in the mantle and crust.

The argument most commonly advanced to support the preservation model is an apparent coincidence between the times of assembly of supercontinents and the zircon age peaks. In my opinion, none of the attempts to demonstrate a correlation is convincing: our knowledge of the timing of the assembly and dispersion of particularly the older (>1.9 Ga) supercontinents is so poorly known that any claim for a correlation must be treated with extreme caution. Figure 7.1 compares the distribution of U-Pb age peaks with two estimates of the times of the assembly and breakup of supercontinents. In the lower part of the diagram, taken from Condie (2011), the coincidence between the two types of event is striking. However, in Bradley's (2011) more recent estimates of the timing of the supercontinent cycles, shown in the upper part of the diagram, a correlation



between the two types of event is weak to absent. In the latter compilation, none of the major Precambrian U-Pb zircon peaks coincides with the assembly of a supercontinent: the first supercontinent assembled well after the 2.7 Ga peak and the Nuna supercontinent formed well after the 1.9 Ga peak. In Bleeker's (2003) compilation (Fig. 7.5), the U-Pb zircon peaks at 2.7 and 1.9 Ga do coincide with the assembly of a supercontinent, but the peaks at 2.5 and 2.1 Ga do not.

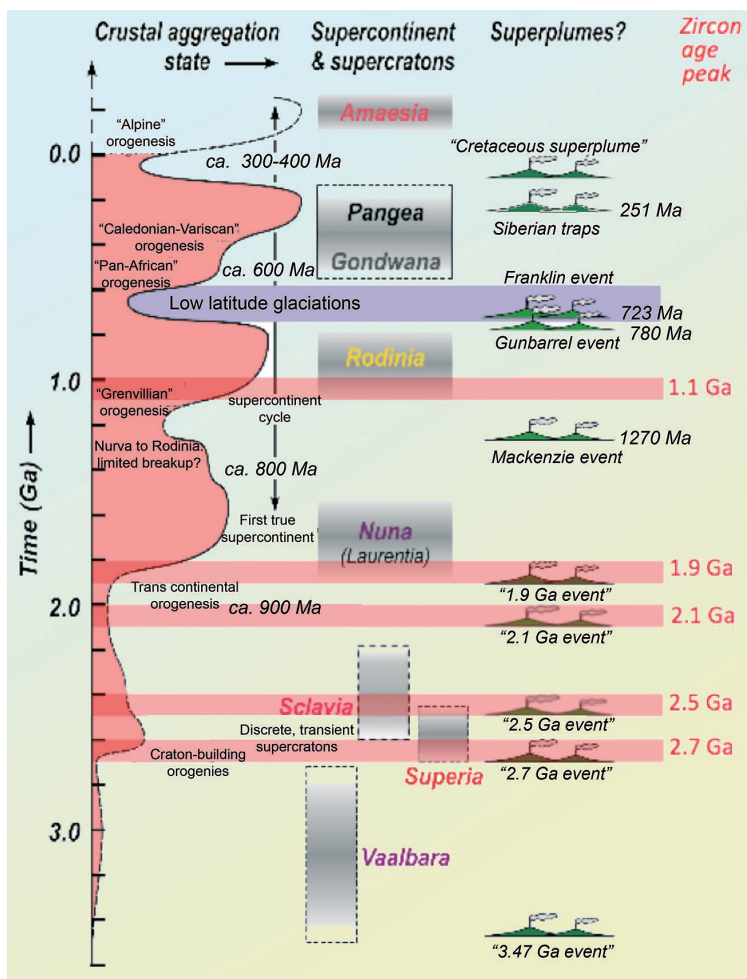


Figure 7.5 Bleeker's compilation of the crustal aggregation state, the timing of the assembly and breakup of supercontinents, and the timing of superplumes. The pink bands represent the U-Pb age peaks (from Fig. 7.1 in Bleeker, 2003, with permission from Elsevier).



Another troubling aspect of the data reported in Figure 7.1 is the relationship between the amplitude of U-Pb age peaks and the amount of continental crust that can be inferred to have existed at the time. Unless we accept an extreme version of the *Armstrong* model (an issue I will discuss a little later in this section), relatively little continental crust had grown before 2.7 Ga. Rapid crustal growth at the end of the Archean, or a pulse of enhanced stabilisation of continental crust at this time, is implicit in all models that deal with this issue. Yet what we see in the age spectrum is an enormous peak at 2.7 Ga, a time when relatively little crust should have been available to assemble into a supercontinent. At this time, relatively few zircons were available for recycling or preferential preservation in a newly accreted supercontinent. The post-2.7 Ga supercontinents were no doubt bigger, being composed of more abundant continental crust that had grown by the end of the Archean, *yet all* of the younger U-Pb zircon peaks are far smaller than the 2.7 Ga peak.

A far more convincing correlation can be made between the U-Pb age peaks and another type of geodynamic event. Shown in Bleeker's (2003) diagram is the timing of what he calls "superplumes" – the massive mantle upwellings that give rise to large igneous provinces (LIPs). In his diagram, a superplume arrives immediately before each of the major U-Pb age peaks in the interval 2.7 to 1.9 Ga. The association between LIPs and the age peaks is well documented, particularly by *Dallas Abbott* and *Richard Ernst* (Isley and Abbott, 1999; Ernst and Buchan, 2001; Condie *et al.*, 2002), and it also forms the basis of Stein and Hofmann's (1994) MOMO model.

What emerges from this discussion that during the period from 2.7 to 1.0 Ga, vast areas of continental crust were generated in a series of pulses whose timing is not directly related to the supercontinent cycle. Instead the peaks coincide with the emplacement of large igneous provinces. In the following sections I will develop a model that relates the age peaks to periods of accelerated true crustal growth driven by episodes of more vigorous mantle convection.

7.2.3 Interpretation of U-Pb zircon age peaks as periods of accelerated crustal growth

Figure 7.6 shows the distribution of ages of "juvenile" granitoids. This diagram, from Condie (1998, 2011), is a compilation of U-Pb zircon data from rocks whose Nd isotopic compositions are within 2 epsilon units of a source in the convecting upper mantle or depleted mantle. Another expression for juvenile is that these rocks have short "crustal residence times", which means that they formed from material that had been extracted from the mantle immediately before the formation of the granitic magma, without spending any significant time as part of the continental crust. Juvenile granites are thereby distinguished from granites that contain a significant component derived from older continental crust. In a later section I will make the case that it is not appropriate to use the depleted mantle as a reference because the initial source of most granitic magmas is in the mantle



wedge above a subduction zone, which evolves along a line with a significantly lower ϵ_{Hf} . When such a reference is used, the proportion of “juvenile” granites is far higher than calculated by Condie (2011), or more recently by Griffin *et al.* (2013).

What emerges nonetheless is that the pattern of age peaks of juvenile granitoids (Fig. 7.6) is very similar to that of the global spectrum of U-Pb zircon ages (Fig. 7.1), which includes both juvenile and recycled granitoids. In particular, the positions and the relative sizes of the Precambrian peaks are similar in both diagrams. Given that a juvenile granitoid is extracted rapidly from the mantle, the pattern of age peaks in the interval from 2.7 to 1.8 Ga provides strong evidence that these peaks register periods of rapid crustal growth.

There is ample additional evidence that the U-Pb zircon peaks coincide with periods of enhanced mantle activity. O’Neil *et al.* (2007) associated the peaks with periods of accelerated plate motion that they link with episodic subduction during the Precambrian. Osmium isotopic data suggest that mantle depletion occurred in pulses corresponding to some of these peaks (Pearson *et al.*, 2007). The isotopic compositions of iron formations record increased input of mantle-derived material into the oceans (Rasmussen *et al.*, 2012). Some 70% of all known komatiites erupted in the interval 2.73–2.70 Ga (Arndt *et al.*, 2008); others erupted at about 1.9 Ga, during a later peak of crustal growth (Arndt, 1982; Arndt *et al.*, 2008). Komatiites form through melting at depths well below subduction zones, most probably in unusually hot mantle plumes (Herzberg and Ohtani, 1988; Arndt, 2003). The association between the times of komatiite eruption and the time of granite emplacement thereby provides a direct link between plume activity and the U-Pb age peaks. The ages of late Archean and early Proterozoic greenstone belts (Condie, 1994) also coincide with the peaks. The geochemistry of the tholeiites that dominate the volcanic sequences of these greenstone belts is very similar to that of oceanic plateau basalts (Arndt *et al.*, 1997; de Wit and Ashwal, 1997), providing another link between plumes and crustal growth.

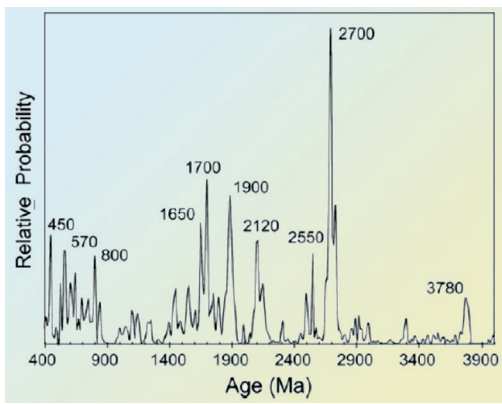


Figure 7.6

Condie *et al.*'s compilation of the ages of “juvenile” granitoids; *i.e.* samples whose $\epsilon_{\text{Nd}}(T)$ values fall within 2 epsilon units of the depleted mantle growth curve (from Condie *et al.*, 2009, with permission from Elsevier).

7.3 Hf Isotopic Compositions and Model Ages

Figure 7.7 is a compilation of Hf isotopic compositions, expressed as $\epsilon_{\text{Hf}(t)}$ values and plotted against their U-Pb age, of zircons from granitoid intrusions (from Guitreau *et al.*, 2012) and detrital grains in rivers (from numerous sources). Four reference lines are included to help with the interpretation of the data: the horizontal line at $\epsilon_{\text{Hf}(t)} = 0$ represents the primitive mantle; the line labelled DM is that of the depleted upper mantle, taken from Workman and Hart (2005) as adopted by Dhuime *et al.* (2011). The line labelled “Arc mantle” represents Dhuime *et al.*'s (2011) estimate of the composition of the source of granitic magmas in subduction zones. The fourth line, labelled “continental crust” shows the evolution of the isotopic composition of a portion of continental crust that was emplaced at 2.7 Ga ($^{176}\text{Lu}/^{177}\text{Hf} = 0.012$).

The tendency for the ages to cluster at the peaks is not immediately apparent because the abundant data within these peaks plot one upon another: Condie and Aster's (2010) histograms (Fig. 7.1) show this far better. The $\epsilon_{\text{Hf}(t)}$ values vary widely, from positive values that extend up to the two mantle reference lines to highly negative values. Much has been made in the literature (Hawkesworth *et al.*, 2009; Pietranik *et al.*, 2010) of the fact that most data fall below the line representing the evolution of depleted mantle. The rarity of zircons with what are described, on this basis, as “juvenile” has been used by many authors (Hawkesworth *et al.*, 2009; Wang *et al.*, 2009; Hawkesworth *et al.*, 2010; Iizuka *et al.*, 2010; Condie, 2011; Iizuka *et al.*, 2013) to argue that little new continental crust had formed during the time-intervals represented by the U-Pb zircon age peaks and that recycled older continental crust was the dominant component in the source of the granitoids. This argument is then used to support models in which re-fusion of older rocks, or metamorphism associated with continent-continent collisions, is the main source of the zircon.

Another way to make the same point is to compare the Hf model ages of detrital zircons with their U-Pb zircon ages, as is done in Figure 7.8 (Voice *et al.*, 2011). For each of the five age brackets, which correspond broadly to the peaks in Figure 7.1, the Hf model age is much older than the crystallisation age. This difference is also taken as evidence that the zircons crystallised from granitic magmas that contained a large fraction of older crustal rock, with little direct input from the mantle (*e.g.*, Hawkesworth *et al.*, 2009; Wang *et al.*, 2009; Hawkesworth *et al.*, 2010; Iizuka *et al.*, 2010; Pietranik *et al.*; 2010, Condie, 2011; Voice *et al.*, 2011; Iizuka *et al.*, 2013).

A flaw in this type of argument was noted by Dhuime *et al.* (2011) who asserted that the source of crustal granitoids is best represented not by depleted upper mantle but by the mantle wedge beneath subduction zones. When a line with slightly lower $\epsilon_{\text{Hf}(t)}$ values is used to represent the mantle source (“Arc mantle” in Fig. 7.7), a large number of zircons and granites plot upon or close



to mantle values and can be considered juvenile. When the Dhuime *et al.*'s "Arc mantle" is used as the reference, the offset between the crystallisation and model age is smaller; about 200 Ma less at 1 Ga and 100 Ma less at 2.7 Ga.

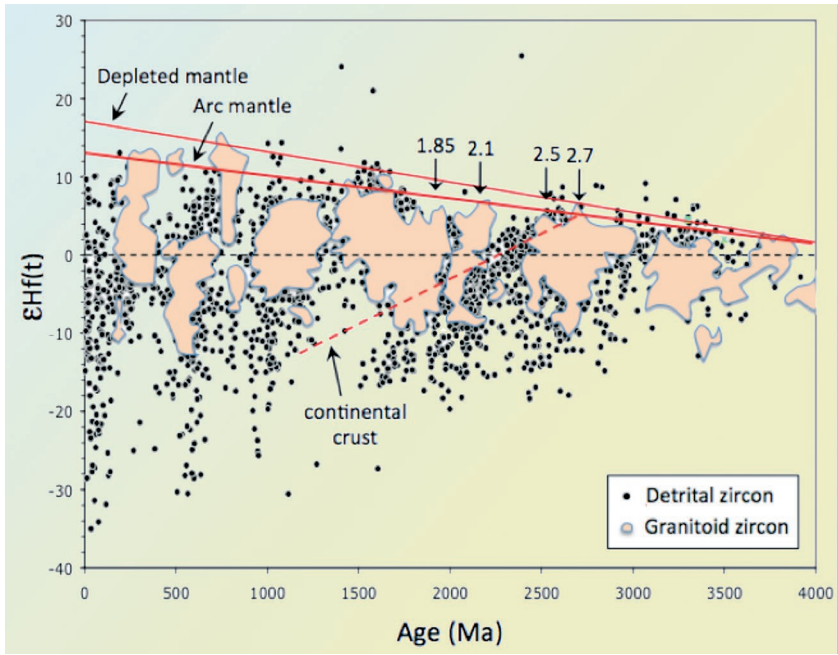


Figure 7.7 Compilation of Hf isotopic compositions (expressed as $\epsilon_{\text{Hf}}(t)$ values) of detrital zircons (black circles) and zircons from granitoids (pink fields), plotted against U-Pb crystallisation age. The principal sources of data for detrital zircons are Davis *et al.* (2005), Wang *et al.* (2009), Iizuka *et al.* (2010), Voice *et al.* (2011), Yao *et al.* (2011), Ying *et al.* (2011), Sarma *et al.* (2012) and Iizuka *et al.* (2013); the data for granitoids are from Guitreau *et al.* (2012). The line labelled "depleted mantle" represents the evolution of the upper mantle (from Condie, 2011); the line labelled "arc mantle" represents the source of granitic magmas in subduction zones (from Dhuime *et al.*, 2011).

This argument can be taken further. As I showed in Figures 2.6 and 2.7, the isotopic compositions of mafic volcanic rocks from modern island arcs span a wide range, from $\epsilon_{\text{Hf}}(t) = +12$ to 0 (ignoring the more extreme values). This range reflects a mixture between a depleted peridotite of the mantle wedge and a mainly sedimentary component from subducted oceanic crust. The latter component was once part of the continental crust and in a sense its incorporation into newly formed melts represents a form of recycling. However, the dominant component of all samples with positive $\epsilon_{\text{Hf}}(t)$ values is from the mantle. The principle involved is illustrated in Figure 7.9. The content of Nd, and to a lesser extent



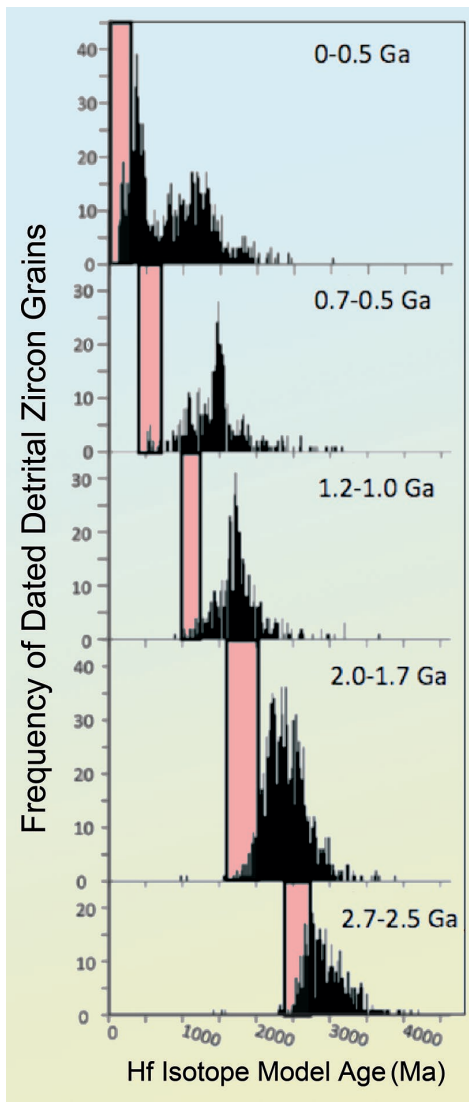


Figure 7.8 Comparison between the distribution of U-Pb ages and Hf model ages of detrital zircons (modified from Voice *et al.*, 2011).

Hf, is so much higher in material from the continental crust that in juvenile magma only a small amount of contaminant is sufficient to change drastically the $\epsilon_{\text{Nd}(t)}$. Consider, for example, the 1.87 Ga Trans-Hudson granitoids of the Reindeer Lake zone, where the depleted mantle composition is well defined from the compositions of komatiites and tholeiitic basalts and the contaminant is represented by Archean granitoid. As seen in Figure 7.9, only 10% of the latter is needed to reduce the $\epsilon_{\text{Nd}(t)}$ from +5 to zero. In this example, 90% of the flux from mantle is juvenile. Although Nd isotopes were used in this example, the same principles apply to Hf and to mixing of a component from recycled sediment into the mantle wedge. With this perspective, all the samples in Figure 7.7 with positive $\epsilon_{\text{Hf}(t)}$ can be considered to record the transfer of magma from their source in the mantle wedge into the crust.

How should we interpret the zircons in Figure 7.7 with negative $\epsilon_{\text{Hf}(t)}$ values? A dominant theme throughout this *Perspectives* is that almost every granitoid of the continental crust contains recycled older crustal material. This point was made persuasively over two decades ago by Hildreth and Moorbath (1988) who used the MASH model (mixing-assimilation-storage-homogenisation) to explain the compositions of



Andean granitoids (see Section 3). The red dashed line in Figure 7.7 shows the evolution of material from 2.7 Ga upper continental crust whose $\epsilon_{\text{Hf}(t)}$ becomes negative after a few hundred million years. Patchett *et al.* (1984, 1986) and Chauvel *et al.* (1987) demonstrated clearly using Nd isotopic data that the range of isotopic compositions of Mesoproterozoic granitoids reflect variable contributions from juvenile and recycled components (Fig. 7.9). The same reasoning applies to the Hf isotopic data plotted in Figure 7.7.

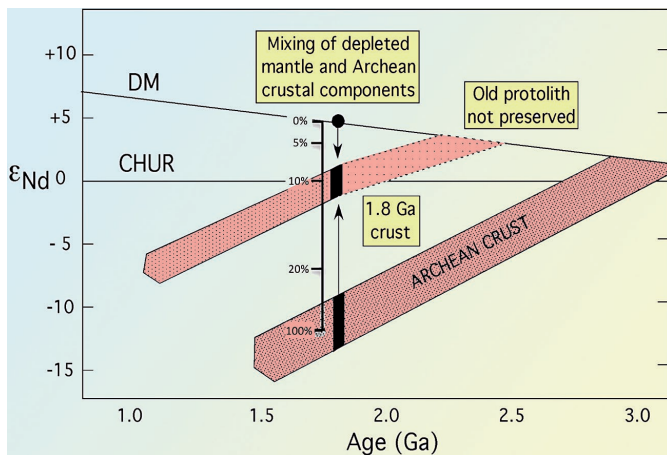


Figure 7.9

Patchett and Arndt's diagram illustrating how the Nd isotopic composition of Proterozoic granitoids can be interpreted as a mixture of a juvenile component and recycled material from Archean crust. Because of the far higher content of Nd in old crust compared with magma from the mantle, the hybrid granitoids with an $\epsilon_{\text{Nd}(t)}$ value of zero consist dominantly ($\approx 90\%$) of juvenile material (modified from Patchett and Arndt, 1986).

The proportion of samples with highly positive $\epsilon_{\text{Hf}(t)}$ values is greater within the age peaks, particularly those at 2.7 and 2.5 Ga, indicating a greater flux from the mantle at these times. The time interval between about 2.4 and 1.7 Ga is exceptional because very few zircons have strongly positive $\epsilon_{\text{Hf}(t)}$ values and plot close to the mantle reference lines. This interval includes two important age peaks, those at 2.1 and 1.85 Ga. Of the zircons within the 2.1 Ga peak only a small number of granitoids from the Birimian belt have dominantly juvenile compositions (Blichert-Toft *et al.*, 1999; Guitreau *et al.*, 2012) and at 1.8-1.9 Ga, neither granitoids nor detrital zircons, are juvenile. In a later section I develop the idea that a vast amount of new granitic crust formed during the major pulse of crustal growth at 2.7 Ga and this provided enough material to dominate or strongly influence the compositions of all magmas that formed in the succeeding 1 billion years.



A final point must be made before leaving Figure 7.7. Many authors, when arguing in favour of the crustal preservation model, disregard the juvenile component that is present in most crustal granitoids. Guitreau *et al.* (2012) do just the opposite; they appear to ignore the recycled component in the granitoids that they analysed. They propose that the data plotted in Figure 7.7 “*demonstrate that the time-integrated Lu/Hf of the mantle source of TTGs has not significantly changed over the last 4 Gy. Continents therefore most likely grew from nearly primordial unfractionated material extracted from the deep mantle via rising plumes that left a depleted melt residue in the upper mantle.*” In my view, as expressed repeatedly in preceding sections, the $\epsilon_{\text{Hf}(t)}$ values of TTGs reflect mixtures of juvenile and recycled components. As mentioned above, at 1.8 Ga, only a small proportion of recycled crust is needed to bring the $\epsilon_{\text{Hf}(t)}$ value from +5 to zero. The data straddling the bulk earth line ($\epsilon_{\text{Hf}(t)} = 0$) in Figure 7.7 correspond to mixtures containing between 10 and 30% recycled crust. The small but relatively constant proportion is controlled by the heat budget during interaction between mantle-derived basaltic magma and older continental crust. As explained by Huppert and Sparks (1988), Bohron and Spera (2001) and Thompson *et al.* (2002), the heat needed to warm up and assimilate crustal rocks is provided by the incoming magma and latent heat of crystallisation. Under normal circumstances, this heat is sufficient to assimilate no more than a third of the mass of the magma. When the data are considered in this way, there appears to be little justification for proposing that the source was unfractionated material from the deep mantle.

7.4 Plumes and Subduction

In Section 4 we reached the conclusion that granitic magmas are generated in subduction settings and in the preceding paragraphs I draw a link between crustal growth and mantle plumes. What is the connection between the two processes?

In many granite-greenstone terrains a cycle of magmatic activity opens with mafic-ultramafic volcanism, reaches a climax about 30 Ma later with the massive intrusion of TTGs associated with calc-alkaline mafic-to-felsic volcanism, and terminates with the emplacement of more evolved granites. Figure 7.10 shows two examples, one from the Yilgarn province in Western Australia and the other from the Abitibi Belt in Canada (Rey *et al.*, 2003). The initial mafic-ultramafic volcanics are attributed to melting in a mantle plume whereas the later felsic plutonism to magmatism in a subduction setting. The pattern is repeated in the Proterozoic Birimian belt of West Africa where initial mafic volcanism at 2130 Ma was followed by the intrusion of granitoids at about 2100 Ma (Abouchami *et al.*, 1990; Boher *et al.*, 1991).



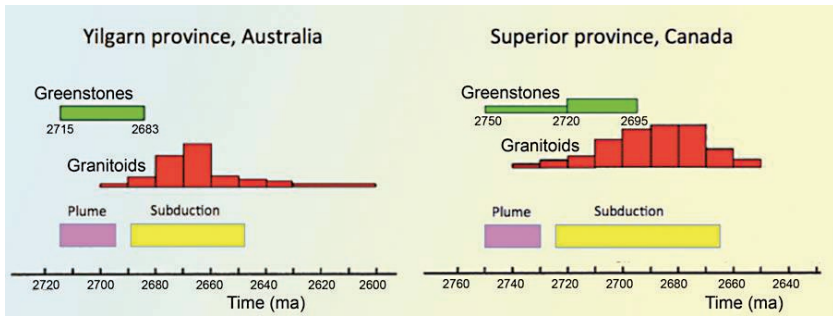


Figure 7.10 A schematic view of magmatic evolution in two Archean granite-greenstone belts. In each case the magmatism opens with mafic-ultramafic magmatism which is attributed to melting in a mantle plume followed by the intrusion of granitoids related to a subduction setting (modified from Rey *et al.*, 2003).

The association of mafic-ultramafic volcanic rocks and felsic plutonism has led to speculation about simultaneous operation of a mantle plume or plumes and one or more subduction zones (Wyman, 1999; Sproule *et al.*, 2002). More likely is an alternation of the two types of activity: as proposed initially by Boher *et al.* (1991), basaltic magmas from a mantle plume could have erupted as an oceanic plateau that focused the locations of subduction zones. Granitic magmas were then generated in the subduction setting via the process illustrated in Figure 5.10.

The oceanic plateaus of the Pacific Ocean provide a modern analogue of this process. About 120 Ma ago, an enormous volume of mainly basaltic magma erupted simultaneously at three sites to form what we know as the Ontong Java, Manihiki and Hikuranga oceanic plateaus (Mahoney and Coffin, 1997). This event coincided with the end of the Cretaceous magnetically quiet period during which the polarity of Earth's magnetic field in the core ceased to flip, an association that supports the hypothesis that the magmas were formed in a plume from a deep mantle. There is some evidence, not universally accepted, that the emplacement of these oceanic plateaus coincided with a period of unusual rapid seafloor spreading (Fig. 7.11; Larson, 1991; Seton *et al.*, 2009). In addition, in Japan and other islands of the western Pacific, a pulse of granitic magmatism coincided with the period of rapid convergence that immediately followed the emplacement of the oceanic plateaus (Takagi, 2004; Kimura, 2006; Fig. 7.11).

In the next section, I describe a model that I developed in collaboration with *Anne Davaille* of the Université Paris-Sud, which attempts to link these processes with the episodic generation of the continental crust.

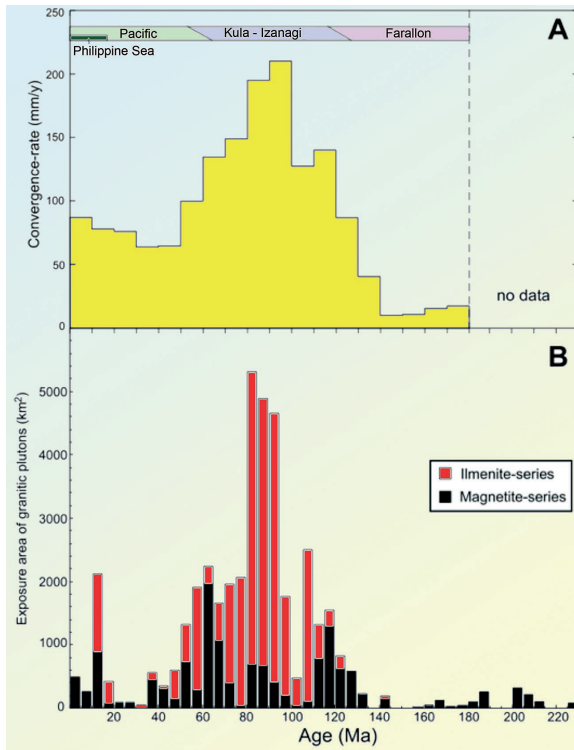


Figure 7.11 Variation of the convergent rate of the Pacific Plate and the area of granitic plutons in Japan over the past 220 Ma. After a period of slow plate movement and minimal generation of granite, both increased dramatically around 120 Ma, just after the emplacement of the Ontong Java oceanic plateau (modified from Takagi, 2004).

7.5 A Model that Links Mantle Convection to Episodic Growth of the Continental Crust

Any successful model for the generation of continental crust during the period from 2.7 to 1.1 Ga must explain the following features :

- The distribution and the relative size of the five major age peaks, at 2.7, 2.5, 2.1, 1.85 and 1.1 Ga (Fig. 7.1)
- The world-wide distribution of the 2.7 Ga peak and the more restricted distribution of the younger peaks



- The abundant juvenile granite that crystallised at the time of each peak
- The association of two types of geodynamic activity; mantle plumes followed by subduction.

The foundation of the model I develop here comes from experiments conducted by *Anne Davaille* and described in a paper that we recently published (Arndt and Davaille, 2013). I give only a brief summary here and for full details the reader is referred to the publication. The experiments were carried out in a large tank filled with viscous fluids (typically sugar syrup). A layer of slightly denser fluid at the base of the tank was heated from below, and after a certain time lapse during which the liquid in the upper layer convected actively, this layer destabilised to form a series of large “domes” or plumes that rose to the top of the tank. As they did so, they displaced cooler liquid from the upper layer which descended into the interior of the tank. At the same time, transfer of heat from the upwelling plumes increased the temperature of the upper layer (Fig. 7.12). The first period of plume activity was succeeded by a quiet interval before the basal layer once again destabilised to produce a second and subsequent generation of plumes. The first plume event was the largest and occupied the whole tank; subsequent generations were smaller and sporadically distributed. Once the lower layer was exhausted, the liquid in the entire tank convected as before.

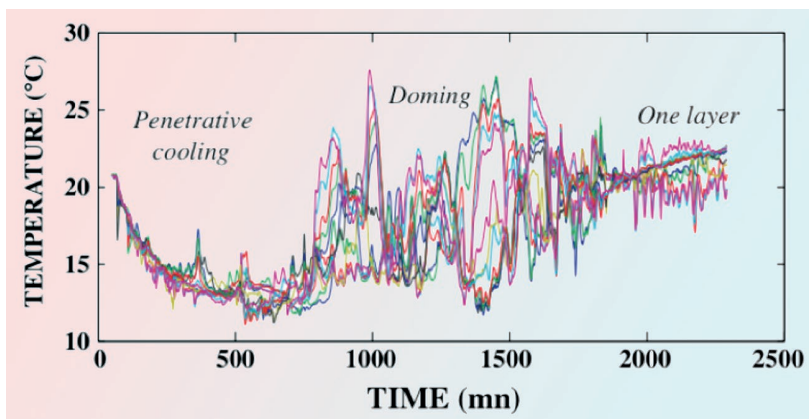


Figure courtesy of Nick Arndt and Anne Davaille.

Figure 7.12

Variation in temperature in one of Anne Davaille’s experiments. The pattern defines three stages. In the first and last, penetrative cooling driven by convection maintained near constant temperatures in the tank. In the intervening stage the ascent of major “domes” or plumes periodically heated the top of the tank.



We use these experiments to explain the pattern of crust formation described in earlier paragraphs. The first and last stages in the experiments (Fig. 7.13), when the fluid convected normally, correspond to periods when normal plate tectonics operated on the Earth; *i.e.* the period from the end of the Hadean to 2.7 Ga and the period after 1.1 Ga. During these periods, seafloor spreading and subduction operated quasi-continuously and as Stern and Scholl (2010) calculate, there was little to no net growth of the continental crust. Mantle plumes generated large igneous provinces like the 120 Ma Ontong Java plateau or the 3.0 Ga Pongolo basalts, but this activity had little influence on global tectonics.

During the period from 2.7 to 1.1 Ga, enormous mantle plumes were generated in the lower mantle and rose into the upper mantle. As they approached the surface they underwent partial melting to yield the komatiites and basalts that erupted as oceanic (or in some cases continental) plateaus. The arrival of an enormous volume of hot plume displaced material from the upper mantle – this material had to return to the lower mantle and we propose it did so in subduction zones.

This is a key element of our model. We do not think that the continental crust was derived directly from the mantle plumes, nor from melting at the bases of oceanic plateaus; rather we propose that the arrival of the mantle plumes in the upper mantle accelerated the rate of plate movement and subduction, which in turn boosted the rate of magma production in the subduction zones. The eruption at the surface of large volumes of magma produced a thick basaltic pile and allowed the generation at its base – at high pressures and in the presence of abundant hydrous fluid – of the TTGs that are the foundation of the continental crust.

An important element of the model is periodic change in the temperature of the upper mantle engendered by the arrival of the megaplumes. In the periods preceding plume activity, the temperature in the upper mantle is relatively low and as a result, the oceanic crust produced by melting at mid-ocean ridges is relatively thin and relatively rigid. When the plume bulldozers its way into the upper part of the mantle, this ocean lithosphere is readily subductable and penetrates rapidly into the mantle, releasing its fluids and generating abundant granitic magma. The transfer of heat from the mantle plumes then increases the temperature of the upper mantle which leads to the production of abnormally thick oceanic crust. This crust, underlain in large areas by the hot, refractory, low-density residues of the mantle plumes, resists subduction and imposes a period of sluggish tectonic activity until the arrival of the next set of mantle plumes. The activity continued in an episodic manner, with short periods of rapid subduction alternating with sluggish periods, until about 1.1 Ga when the source of the major plumes was exhausted and geodynamic processes evolved to their present state.



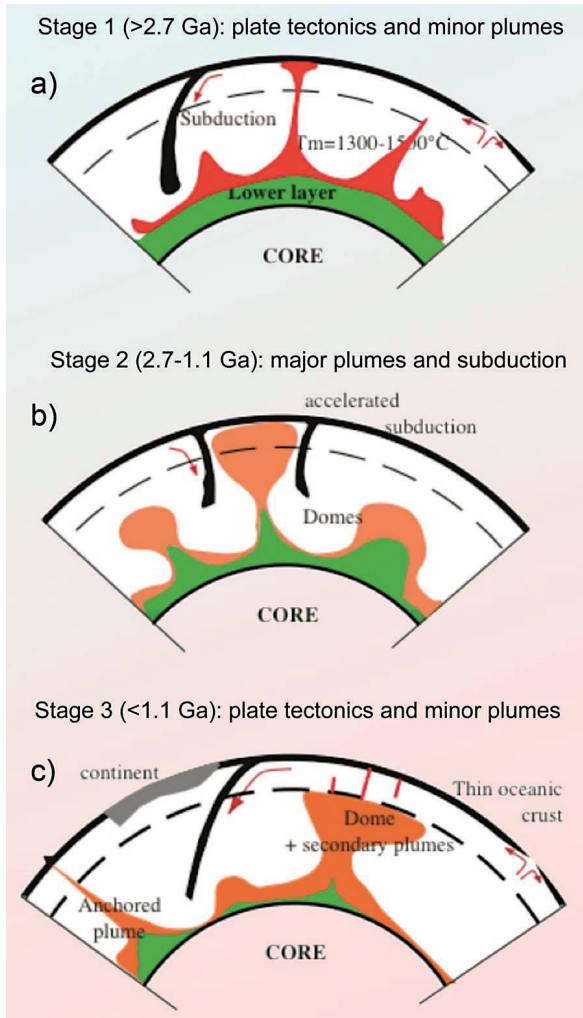


Figure courtesy of Nick Arndt and Anne Davaille.

Figure 7.13

Cartoon illustrating three stages of evolution of the Earth.



8.1 Introduction

In this section I discuss the nature and origin of the first crust of planet Earth. This crust formed in the first 500 Ma of the history of the planet, a period from which almost no geological record remains: all we have are zircons, the majority from two localities in Western Australia, complemented by isolated grains from other regions. Our knowledge of the first crust, and the basis for our speculations as to its size, distribution and origin, come entirely from geochemical analyses of minute crystals – or even portions of these crystals – a type of research that became possible only during the past decade thanks to the development of new analytical techniques. In the past few years, very valuable information has been produced by new high-performance inductively coupled plasma mass spectrometers (ICP-MS) but the first clues to the existence of Hadean felsic crust came from the ion microprobe developed at the Australian National University.

8.2 The SHRIMP

In the late 1980's, geochemists of the Australian National University in Canberra made one of the most important discoveries of the Earth Sciences – the analysis of a few zircons from Western Australia, revealed these zircons to be older than 4 billion years.

Birth of the SHRIMP – On a visit to the Research School of Earth Sciences, some time in the 1980's, my host took me to a large room on the ground floor and showed me a very long steel tube coupled to an enormous magnet. He told me it was to become a new mass spectrometer designed expressly to determine the ages of zircons, by *in-situ* analysis using the U-Pb method. The project had started several years earlier, was well behind schedule, and well over budget. My host thought that *Bill Compston* (Fig. 8.1), who was leading the project, had embarked on a wild goose chase and predicted that the instrument would never work. During another visit a few years later, he honestly admitted he had been quite wrong. In the meantime *Bill Compston* and his team had dated the Jack Hills zircons and the discovery was being debated and/or acclaimed throughout the world.





Photo courtesy of the Research School of Earth Sciences, Australian National University.

Figure 8.1 A severe-looking Bill Compston glaring by his SHRIMP. Ian Williams tells me that the photo was taken at the end of a day-long tour of the lab by journalists. His expression reflects his irritation of having to pose, zircon mount in hand, through several hours of photography.

As explained on the ANU website “*The first ion microprobes were able to resolve individual unit masses, but could not deal with the complex molecular interferences produced in the sputtering process. Professor William Compston of ANU saw the solution to this problem in the design of a large ion microprobe capable of high mass resolution while maintaining high sensitivity. And so SHRIMP was born. Dr Steve Clement, an RSES graduate student who designed his own mass spectrometer for his PhD, was retained to design SHRIMP I based on ion optical parameters of Professor Hisashi Matsuda of Osaka University.*”

8.3 The Ages of Jack Hills Zircons and their Significance

The first paper describing SHRIMP analyses of very old zircons was published by *Derek Froude*, a PhD student from New Zealand who had been asked by *Bill Compston* to determine the ages of metasedimentary rocks from the western part of the Yilgarn craton in Western Australia (Froude *et al.*, 1983). An interview of *Bill Compston* at <http://science.org.au/scientists/interviews/c/bc.html#9> provides a nice account of the initial discovery of old zircons from Mt. Narryer, and from the more prolific site subsequently found at the Jack Hills.



Figure 8.2 is a concordia diagram showing the ages of zircons from the Jack Hills, including the world-record 4.4 Ga result. The total range of ages is given in *Aaron Cavosie's* compilation of all the ages that had been measured before about 2007 (Fig. 8.3). Here a peak is seen at about 4.1 Ga with declining numbers at both older and younger ages. Blichert-Toft and Albarède (2007) have proposed that 4.1 Ga was the time of crystallisation of the majority of Jack Hills and that the older ages were artefacts resulting from analytical difficulties. Most authors, however, accept that even though certain analyses may be unreliable, the existence of pre-4 Ga zircons is not in question. Figure 8.4 shows images of some of these ancient zircons.

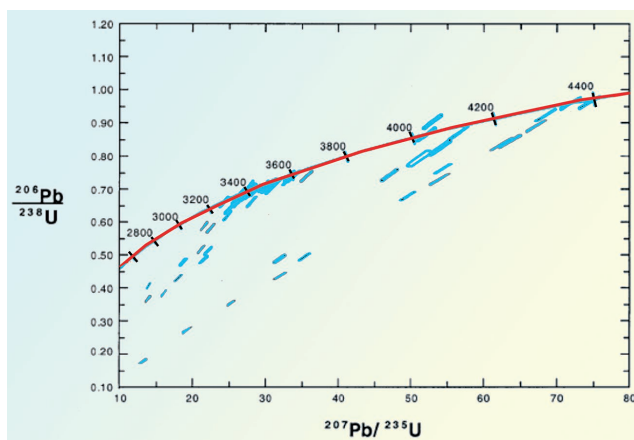


Figure 8.2 Concordia diagrams with the ages of Jack Hills zircons (modified from Peck *et al.*, 2001). Other studies of zircons from this locality have produced abundant ages in the 4.0 to 4.4 Ga interval.

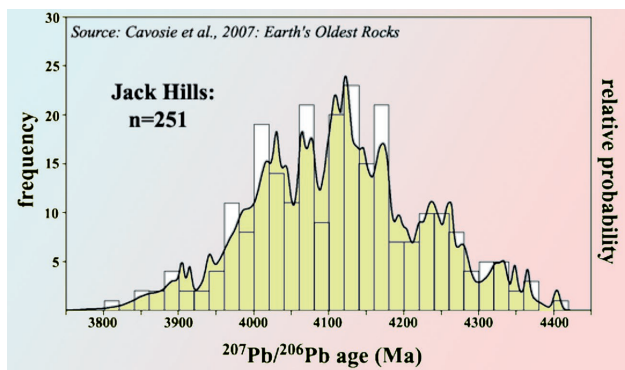


Figure 8.3 Aaron Cavosie's compilation of ages of Jack Hills zircons (modified from Cavosie *et al.*, 2007).



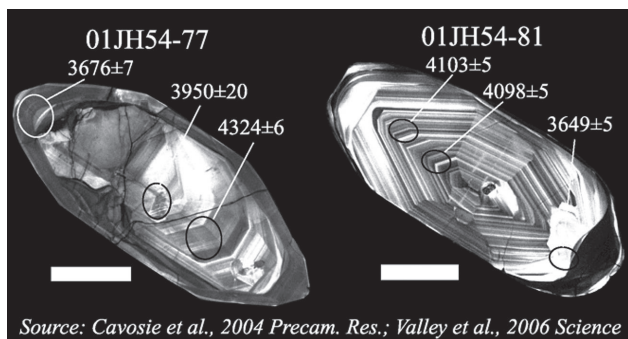


Figure 8.4 Cathode luminescence images of two Jack Hills zircons illustrating magmatic zoning and differing ages of parts of each grain. The portion with an age greater than 4.3 Ga is seen in the grain to the left (from Cavosie *et al.*, 2004, with permission from Elsevier).

The significance of the very old zircon ages was immediately obvious. If, as generally accepted, zircon crystallises from magma of granitic composition, then the discovery of 4.2 Ga zircon provides evidence for 4.2 Ga granite. A discussion about whether the Jack Hills zircons could not have come from a mantle source, like zircons in kimberlites, was largely resolved when Maas *et al.* (1992) published the trace element compositions of the zircons that indicated a granitic source (but see Coogan and Hinton, 2006).

From here it was only a small step to conclude that if granite existed, then continents must also have formed. Ian Campbell and Ross Taylor (1983) defended this idea when, in a paper entitled “*No water, no granite; no oceans, no continents*” (Ian is good at catchy titles), they explained that “*water is essential for the formation of granite, and granite, in turn, is essential for the formation of continents.*”

A major point of contention was whether the existence of a small number of zircons from just two localities in Western Australia could be used to support Armstrong’s (1981, 1991) model that invoked massive early generation of the continental crust. Steve Moorbath (1983) famously remarked “*One swallow does not necessarily make a summer, and four zircon grains from a single quartzite sample do not necessarily make a continent*”. This was written in a News and Views article published in *Nature* to provide background to the Froude *et al.* (1983) paper; it was therefore published before the discovery of more abundant pre-4 Ga zircons in the Jack Hills. Since then, although there have been many attempts to find old zircons in Eoarchean metasediments the world over, these first met with very little success. No zircons with ages much older than their host sedimentary rocks, and none with ages greater than 4 Ga, were initially found in areas like Isua in Greenland (Nutman *et al.*, 2009), the Beartooth Mountains in the USA (Mueller *et al.*, 1992), the Superior and Rae provinces of Canada (Davis *et al.*, 2005; Hartlaub *et al.*, 2006), and the Limpopo Belt of South Africa (Zeh *et al.*, 2008). The apparent absence of Hadean zircons in these regions led Stevenson and Patchett (1990) to conclude “*The data strongly suggest inheritance of pre-3.0 Ga zircons only in areas*

where pre-3.0 Ga old crust exists today, and imply that the quantity of continental crust prior to 3.0 Ga ago was not much greater in extent than the pre-3.0 Ga crust exposed today. Small amounts of continental crust prior to 3.0 Ga ago and rapid addition of continental crust between 2.5 and 3.0 Ga ago are consistent with the gradual growth of continental crust, and argue against no-growth histories."

I'm not sure if *Stevenson* and *Patchett* still think this way. In the 23 years that have since elapsed, Hadean zircons have been found in scattered locations the world over: in the Acasta gneiss (*Iizuka et al.*, 2006), in the Beartooth Mountains of the USA (*Maier et al.*, 2012) and in various locations in northern China (*Geng et al.*, 2012, *Cui et al.*, 2013). It can still be argued that these zircons are few and far between, and that if there had been as much old crust as *Armstrong* had proposed, we should see more of it. Mark Harrison (2009) put it this way in his review paper on the early continental crust "supporters of the slow growth paradigm point to the distribution of age provinces and the absence of >4 Ga crust, Archean-Proterozoic sediment REE patterns, the lack of fractionation of the Nd/Hf isotopic systems, the uniformity of Ce/Pb in basalts throughout time, Nb-U-Th systematics in mantle-derived rocks, and the implausibility of making early felsic crust." An alternative interpretation, to which I subscribe, is that it is quite remarkable that any relict whatsoever of the Earth's crust survived from Hadean times, a period when the mantle was hotter and convected vigorously and when the surface of the planet was subjected to intense meteorite bombardment.

I argued this way in a paper published in 1991 in the *Bulletin of the Geological Society of Denmark*; a fine journal but not one to guarantee wide dissemination of scientific results. In this paper (*Arndt and Chauvel*, 1991) we wrote, referring to the Jack Hills zircons, "The fact that such old zircons constitute 2% of the zircon population in a sediment deposited some 1 Ga after they formed is astonishing: either there was a remarkable preservation mechanism that enabled these zircons to survive for a billion years at the surface of an unstable early Earth, or the source of the old zircons was abundant and voluminous. Since the most likely platform that could have preserved the zircons is buoyant, low density, felsic crust, the very existence of old zircons is strong evidence that voluminous felsic material in one form or another existed on the Hadean Earth." (*Arndt and Chauvel*, 1991, page 146).

Two-author Papers – *Arndt and Chauvel* (1991) is the only two-author paper I ever published with my wife, *Catherine*. She arrived at the Max-Planck Institut in 1982, fresh from her PhD on basalts from the Massif Central supervised by *Bor-ming Jahn* in the University of Rennes. *Al Hofmann*, who directed the geochemistry department, suggested that she change subjects and work instead on the Nd isotopic compositions of Archean volcanic rocks. The idea was that we would select samples with near-chondritic Sm-Nd ratios, which would mean that their measured $^{143}\text{Nd}/^{144}\text{Nd}$ would evolve parallel to that of a primitive mantle source and any enrichment or depletion relative to that of the source would be immediately apparent. In the event it turned out that the komatiitic and tholeiitic basalts we selected had such complicated geological histories that the method had little value. *Catherine* instead published a nice paper demonstrating that assimilation by komatiite magma of continental crust produced mixing lines that looked like isochrons but gave apparent ages that were far too high (*Chauvel et al.*, 1985).



I had hoped that we would be able to set up a cosy working relationship – she would slave over a hot fume hood producing geochemical data and we would both co-sign any papers that emerged from this work. But this never happened – *Lina* got to her first! *Lina Echeverría* was with us in Mainz in the early 1980's and she gained a solid international reputation working on the dinkum komatiites from Gorgona (Echeverría, 1980, 1982) before rising to a management position with Corning Glass. *Lina* told *Catherine* that if she published papers with me as a co-author, many readers would conclude that it was me who supplied the science while she, eight years my junior, just supplied the analyses. Sound advice indeed, for a young postdoc. But *Catherine* kept this up for 25 years! Only during the last five years, a period in which she has been scientifically more productive than me (if we ignore the generous helping hand extended to me by my colleague *Alex Sobolev*), have things equilibrated.

8.4 Oxygen Isotopic Data

In the early 2000's the Jack Hill zircons sprung two more surprises on the geological community. The first was when *John Valley* and co-workers (Peck *et al.*, 2001; Valley *et al.*, 2002) reported oxygen isotopic analyses (Fig. 8.5) of the pre 4 Ga zircons some of which were shown to have $\delta^{18}\text{O}$ values well above the accepted mantle values of +5.0 to +5.6. Valley *et al.* (2002) pointed out that oxygen isotopes fractionate strongly only at low temperatures and not significantly at magmatic temperatures. The higher-than-mantle $\delta^{18}\text{O}$ values of the Jack Hills zircons indicated that their granitic source contained a component that had been altered at or near the surface of Earth, most probably during interaction with liquid water. The existence of this altered component in turn implied the presence of liquid water, most probably oceans at the surface of the planet; hence the notion of a “cool early Earth” (Fig. 8.6).

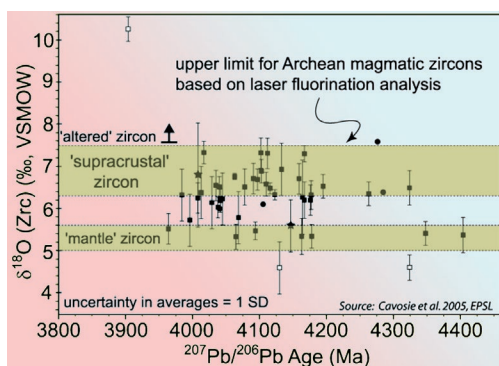


Figure 8.5

A compilation of the oxygen isotopic compositions of Hadean zircons from the Jack Hills (modified from Cavosie *et al.*, 2005).





Image credit: Steve Munsinger | Science Source.



Image credit: Don Dixon | cosmographica.com.

Figure 8.6

This figure combines two views of the Hadean landscape: the first is one of those wonderful artist's impressions of fiery volcanoes, frequent meteorite impacts and a looming close-by Moon. The second shows a more clement, cool, early Earth on which a shallow ocean laps over a rugged shoreline of volcanic and impact craters.



The notion of an altered component in the source of the granites received support from Bruce Watson's (2005) development of the Ti-in-zircon geothermometer, which produced low crystallisation temperatures for the Jack Hills zircons and their host granites. Studies combining oxygen isotopes, Ti-geothermometry and the trace-element contents of the zircons produced similar results (Trail *et al.*, 2007; Harrison, 2009). Additional evidence came from the discovery of inclusions of crustal minerals like quartz and muscovite in some of the zircons (Hopkins *et al.*, 2010) although the significance of these minerals has been questioned (Rasmussen *et al.*, 2011).

8.5 Hafnium Isotopic Data

The second surprise came with the accumulation of Hf isotope analyses of the Jack Hills zircons. Many different groups have worked on this project, using a variety of different analytical methods (*e.g.* Amelin, 1998; Harrison *et al.*, 2005; Blichert-Toft and Albarède, 2007; Harrison *et al.*, 2008). As mentioned in Sections 2 and 7, *in-situ* methods have allowed simultaneous or sequential analysis of lead, hafnium and oxygen isotopes in small portions of single zircon grains and this has provided invaluable information about their ages and the nature of their sources. In Figure 8.7, a recent compilation from Bell *et al.* (2011), we see the following features:

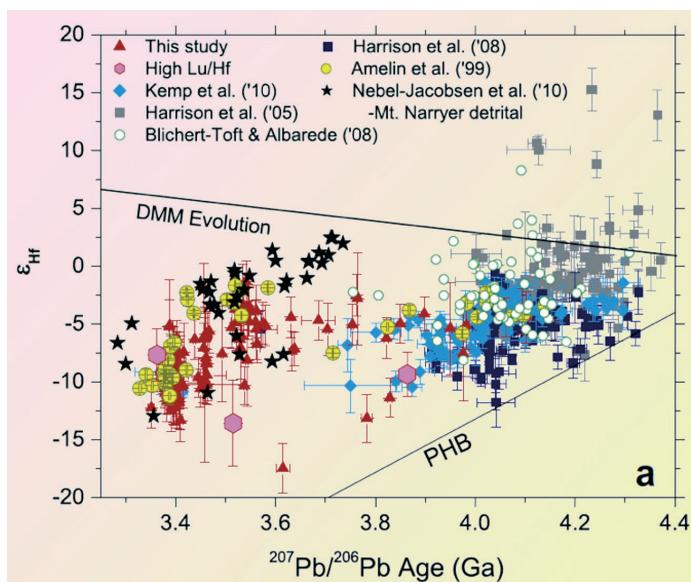


Figure 8.7

A compilation of Hf isotopic compositions and Pb-Pb ages of zircons from the Jack Hills and Mt Narryer regions (from Bell *et al.*, 2011, with permission from Elsevier).



- The data display a large range of isotopic compositions, from ϵ_{Hf} values close to the “forbidden zone” (the line labelled PHB shows the evolution of a Lu-free source that formed at 4.56 Ga) to some highly positive values.
- Almost all data plot below the line labelled “DMM evolution”, which represents that of depleted upper mantle.
- The data identified by blue diamonds are from Kemp *et al.* (2010). These data were selected using oxygen isotope analyses and other criteria and they define a more restricted linear trend than the bulk of the data. This trend extends from a primitive mantle value at 4.5 Ga and has a slope that corresponds to a $^{176}\text{Lu}/^{176}\text{Hf}$ value of 0.02, which corresponds to that of mafic crust.
- Zircons with ages less than about 3.8 Ga plot in a separate field; they also lie on an array with a positive slope but the data are displaced to higher ϵ_{Hf} values and the trend intersects the DMM curve at about 3.8 Ga.

How should these data be interpreted? Do they provide evidence for the existence of very old continental crust, as predicted by the *Armstrong* model? Before entering into a discussion of the possible geological implications, a word of caution about the significance of the calculated ϵ_{Hf} values of these very old zircons is warranted. As summarised by Griffin *et al.* (2013), the reliability of these data depends on the accuracy of the Hf isotope analysis and particularly of the age determination. The Jack Hills zircons are extremely complex and, as illustrated in Figure 8.4, contain zones of very different ages. The interpretation of these ages is not straightforward: some no doubt reflect the time of crystallisation of the zircon from the host magma but others may result from disturbance of the U-Pb system during metamorphism or perhaps alteration of detrital grains during the period preceding deposition of the host quartzite around 3.1 Ga. Most of the data reported in the figure were obtained by measuring separately the Lu-Hf and Pb isotopic compositions in selected parts of zircon grains, a procedure that leads to uncertainty as to whether the two measurements were made on the same portion of the zircon. This is crucial because metamorphism or alteration may perturb the Pb isotopes without changing the Hf isotopic composition. As argued by Vervoort *et al.* (2013) the assignment of an incorrect age may result in the calculation of ϵ_{Hf} values that are falsely positive or far more negative than the real values. One way to resolve this problem is to measure Pb and Hf isotopic data simultaneously either by peak jumping (*e.g.*, Kemp *et al.*, 2009) or using techniques to split the argon stream and conduct it to two separate mass spectrometers (*e.g.*, Xie *et al.*, 2008).

For the moment I will work with the data that exist, as plotted in Figure 8.7. The key observation is that, within the scatter and particularly for the filtered data of Kemp *et al.* (2010), the analyses of >3.8 Ga zircons plot on a broad trend in which ϵ_{Hf} decreases progressively with decreasing age. The decrease of ϵ_{Hf} indicates that the source was enriched (*i.e.* $^{176}\text{Lu}/^{177}\text{Hf}$ was low), which means that the concentration of Hf, the more incompatible element, was higher than Lu.



This is typical of the trace element patterns of TTG, as illustrated in Figure 2.4. The uniform positive slope indicates that, within limits of error, the Lu-Hf ratio of this source was constant, and remained so for over 300–400 Ma. This type of evolution is highly unusual. In modern geodynamic settings, ancient continental crust may undergo partial melting to yield magmas with low ϵ_{Hf} , but such melting normally lasts no longer than 10–50 Ma and is not commonly repeated over such a long time span. In addition, the continental crust, being heterogeneous, gives rise to melts of diverse compositions. Finally and most significantly, in island arcs and convergent margins, most granitic magmas come from a source that includes a large juvenile component from the underlying mantle. This input, combined with the incorporation of material from older continental crust, results in isotopic compositions that scatter around near-zero ϵ_{Hf} values, as illustrated in Figure 7.7.

The steadily declining trend defined by the ϵ_{Hf} values of Jack Hills zircons requires either a single source of constant, enriched composition (*i.e.* a uniform and low Lu/Hf), or a series of sources, all with the same composition. There are two opinions about the nature of this source. Harrison *et al.* (2005, 2008) and Blichert-Toft and Albarède (2007) opt for a granitoid source, perhaps TTG with the characteristic depletion of HREE and thus a very low Lu/Hf ($^{176}\text{Lu}/^{176}\text{Hf} \approx 0.01$). Kemp *et al.* (2010) note that the slope of the trend through their filtered data set ($^{176}\text{Lu}/^{176}\text{Hf} \approx 0.02$) corresponds to that of slightly enriched basalt. On this basis they propose that the Jack Hills granitoids resulted from “*protracted intra-crustal reworking of an enriched, dominantly mafic protolith that was extracted from primordial mantle at 4.4–4.5 Ga, perhaps during the solidification of a terrestrial magma ocean.*”

8.6 The Hadean Crust

Kemp *et al.*'s (2010) model is illustrated in Figure 8.8 and described in the caption of this figure. The key features are the formation of an enduring and widespread protocrust composed of enriched basalt that was extracted from the mantle following solidification of the terrestrial magma ocean. Parts of this crust became hydrated through interaction with the hydrosphere, and these portions repeatedly melted to produce the felsic magmas that crystallised the Jack Hills zircons.

The model has several attractive aspects. (1) It provides an explanation for the uniform, enriched composition of the source of the granitic magmas; *i.e.* the enriched basaltic protocrust. (2) Very high abundances of short-lived radioactive elements in all Hadean rocks provide a source of heat and a mechanism for periodically remelting this source. (3) The widespread protocrust may have constituted a stagnant lid beneath which the mantle convected; such a situation might explain the absence of juvenile input during the period that the protocrust persisted.

There remains, nonetheless, the problem that I raised when we discussed the “ocean-plateau” model for the formation of Archean TTG in Section 5. Two lines of evidence – the oxygen isotope data and the Ti-in-zircon geothermometry – suggest that the Jack Hills felsic magmas were hydrous; but the Kemp



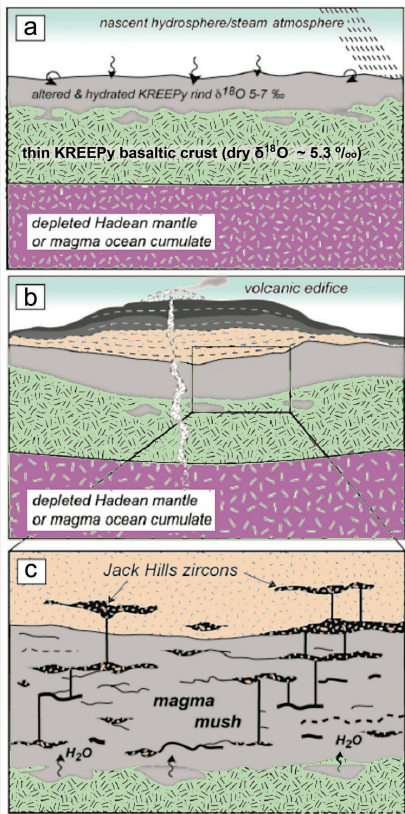


Figure 8.8

Kemp *et al.*'s model for the formation of the felsic magmas that crystallised the Jack Hills zircons. In the first stage (a), a crust is formed from basaltic magma derived from an enriched source left in the mantle after solidification of the terrestrial magma ocean (the term KREEPy refers to the enrichment of potassium and REE that characterises the source of certain lunar basalts). The upper rind of this crust is altered through interaction with the hydrosphere or atmosphere. In the second stage (b and c), the hydrous rind is buried by new basaltic eruptions and it undergoes repeated partial melting to produce the felsic magmas that crystallised the Jack Hills zircons (modified from Kemp *et al.*, 2010).

et al. (2010) model does not provide a mechanism for repeatedly supplying water into the site of melting. Just as in an oceanic plateau, the lower portions of the protocrust would have been anhydrous, and unlike magmatism in a subduction setting, there was no continuous supply of hydrous magmas into the zone of melting. Kemp *et al.* (2010) propose that the surface of the crust became hydrated through interaction with the hydrosphere and “speculate that foundering of hydrated basaltic shell to deeper crustal levels in locally thickened eruptive centres, possibly facilitated by volcanic resurfacing, led to the intermittent generation of melts”. Perhaps ... but such a mechanism would only supply water to the first generation of felsic melts. When these melts cooled and crystallised (to form the zircons on which all this story is based), they would have lost fluids to the surface and these fluids would have been unavailable for subsequent melting. What seems to be required is a process whereby the felsic magmas remained in contact with the oceans (hydrosphere if you prefer). Kemp *et al.* (2010) speak of volcanic resurfacing and it is tempting to call on this process to re-bury erupted felsic melts – probably pyroclastic deposits – to depths where they could heat up a re-melt. But what was the source of the volcanics that did the resurfacing? They cannot have come from the underlying mantle because the defining feature of the Jack Hills Hf isotopic array, particularly Kemp *et al.*'s filtered data set, is a lack of input of juvenile material.



What seems to be involved is the persistence of a layer of partially molten felsic material at the surface, in contact with the oceans. The zircons could have crystallised in the crust of melt sheets, which became hydrated through interaction with the oceans; then portions of the crust could have foundered into the underlying melt sheet to provide xenocrysts in subsequent portions of solidified crust. Assimilation of the surrounding basaltic protocrust buffered the evolution of the Hf isotopic composition and had it evolve along the relatively shallow slope of the Kemp *et al.* (2010) trend, rather than the steeper slope that would be generated by repeated melting or reprocessing of a purely granitic source.

Another issue is the amount of felsic rock that was present in the Hadean crust. Did the granites that crystallised the Jack Hills zircons make up only a small proportion of a mafic protocrust, as proposed by Kemp *et al.* (2010), or were they more voluminous, as argued by Harrison (2005, 2008)? Kamber *et al.* (2005, 2010) envisage that mafic protocrust containing a small felsic component mixed back into the mantle at the end of the Hadean, an event that may have been triggered by the late heavy bombardment. Kemp *et al.* (2010) argue along similar lines, but propose that the ancient felsic crust contributed to the source of Archean granitoids. An argument in favour of larger and more stable Hadean continents was made over 20 years ago in that two-author paper with Catherine Chauvel (1991). Somehow the zircons that crystallised during multiple granite-forming events through the Hadean survived at the Earth surface for up to one billion years, through the late heavy bombardment and periods of vigorous mantle convection. For this to have happened, the zircons must have been part of a stable platform and this platform most probably was continental crust.

8.7 The Onset of Plate Tectonics at 3.8 Ga

An intriguing aspect of the data for >3 Ga zircons plotted in Figure 8.7 is the re-appearance of zircons with positive ϵ_{Hf} values at about 3.8 Ga. The data from Jack Hills and the few >4.0 Ga grains that are being found in other regions plot on the trend of falling ϵ_{Hf} with decreasing age (the trend interpreted above as being due to continued reprocessing of the same enriched primitive crust); the jump to higher values at about 3.8 Ga signals a renewed input of juvenile material from the mantle. Somehow the geodynamic regime that had previously prevented input of this material was disrupted.

The period from about 4.0 to 3.9 Ga was a time when the Earth was bombarded by a heavy flux of meteorites. The earlier idea that this was the tail of the accretion of the planet (Hartmann *et al.*, 2007) has now been replaced by the “late heavy bombardment” hypothesis which holds that there was a renewed pulse of impacts at this time, perhaps due to perturbation of the asteroid belt triggered by a change in the orbits of the giant planets (Gomes *et al.*, 2005). This period of bombardment could have disrupted the mafic crust within which the



Jack Hill felsic rocks were being generated (Kamber, 2007, 2010; Kemp *et al.*, 2010), and could have resulted in a change in the global geodynamic regime. What did it change to?

The oldest terrestrial volcanic rocks are preserved in the Isua belt of Greenland (Rosing *et al.*, 1996) and the Nuvvuagittuq belt of Canada (O'Neil *et al.*, 2007). At Isua, mafic pillow lavas are recognised, some of which, according to Polat *et al.* (2002) have the geochemical signature of boninites; *i.e.* magmas that formed in subduction zones. Intruding both belts and recognised in several other parts of the world, most notably the Acasta area, are abundant granitoids. In Section 5, I made the case that granitic rocks most likely formed in a subduction setting. Subduction is a consequence of plate movement. For me, the change in geodynamic style at about 3.8 Ga corresponded to the onset of plate tectonics.



9.1 Formation of the Continental Crust

The subject of this *Geochemical Perspective* is the continental crust – how it formed and at what rate. Important constraints come from the mineralogical and chemical compositions of granitoids, particularly those of the tonalites, trondjemites and granodiorites (TTG) that are the dominant component of Archean continental crust. When used in conjunction with the results of experimental studies, these data help define the conditions of partial melting and the nature of the source of granitic magmas, as well as the geodynamic setting in which they formed. We learned that the dominant component in their source is mantle-derived basalt. This basalt is present either in the form of solid intrusions that partially melt to produce more silicic magmas, or as a liquid, which undergoes a combination of fractional crystallisation and contamination to produce more evolved magmas. Trace element and isotopic data record the presence, in quantities that vary from near zero to almost 100 percent, of rocks from pre-existing, often far older continental crust. The same data can also be used to show that other granitoids are juvenile; *i.e.* they were extracted rapidly from their source in the mantle.

Trace elements constrain the depth of melting: a depletion of heavy rare earth elements characterises most TTG and points to the presence of abundant garnet in the residue of melting, as is possible only at depths from about 20 to 60 km. Another important result of the experimental studies, confirmed by geochemical modelling, is that magmas with the compositions of Archean TTGs and modern orogenic granitoids form in large volumes only when water is present in the source. The water can be present as a separate fluid phase or within hydrous minerals, but without it abundant granitic magma does not form.

The requisite conditions are met in subduction zones, which constitute a remarkably efficient factory for the production of granite. The process proceeds in two steps. In the first, subducting hydrated oceanic crust loses its water into the overlying, convecting peridotite of the mantle wedge. Hot, fertile mantle material is dragged into an ascending stream of aqueous fluid which fluxes melts and produces copious amounts of hydrous mafic magma. In the second stage this magma ascends to the base of the crust – an island arc or convergent margin – where it undergoes the complex combination of fractional crystallisation, remelting of previously-intruded basalts, and contamination with older crustal rocks that yield melts of intermediate compositions. Finally these magmas ascend to higher in the crust where they fractionally crystallise into silicic granitic magmas. Other processes produce only minor amounts of granite. When the subducting oceanic crust is unusually hot it may partially melt to give adakitic magmas. Granitoids also form in anorogenic settings, but not in the same volumes as in active subduction zones.



If plate tectonics operated in the Archean, there is every reason to believe that the processes described above generated Archean TTG. The more pronounced HREE depletion of the Archean rocks can be attributed to a higher magma flux that produced thicker crust in the region of compression above the subducting plate, or strongly hydrated subducting Archean crust released more abundant aqueous fluids that stabilised garnet at the site of melting.

There are no compelling reasons to argue that plate tectonics did not function from the start of the Archean. However, it may have operated episodically. During the middle part of Earth history, from 2.7 Ga to about 1.8 Ga, the mantle convected in an irregular manner. During 4 or 5 major events, large plumes rose from deep in the mantle and as they approached the surface, they displaced material from the upper mantle; that is, they accelerated the rate of subduction. The increased subduction resulted in accelerated generation of granitic magma and to a pulse of growth of the continental crust. Heat transfer from the plumes increased the temperature of the upper mantle and this resulted in the generation of thick oceanic crust that resisted subduction. A period of sluggish geodynamics therefore followed each crustal growth peak. Subsequent cooling led to the production of progressively thinner oceanic crust that was readily subducted on the arrival of another plume.

Models that call for the formation of Archean granitoids in non-subduction settings are predicated on the notion that plate tectonics did not operate during the first part of Earth history. Most such models invoke melting in the lower portions of thick piles of basalt and are unconvincing because the basal parts of oceanic plateaus or thick oceanic crust consist of refractory, infertile ultramafic cumulates. These cumulates are dry – the water needed to produce granitic magma in large volumes is missing. In the subduction setting, water is supplied by dehydrating oceanic crust.

The Hadean crust probably consisted of slightly enriched basalt. Because of strong internal heat production from short-lived nuclides, the crust repeatedly partially melted to produce the felsic magmas that crystallised the Jack Hills zircons. This crust may have been widespread and the quantity of felsic material may have been large; but it was disrupted and dispersed during the late heavy bombardment. Relicts of this crust left an isotopic trace in all granitoids that formed in the first part of the Archean.

Current and Future Research – Since 1990 I have worked in Grenoble. My contacts with the geophysicists of my department, now called ISTerre (Institut des Sciences de la Terre) led to a fruitful line of research on the sub-continental lithospheric mantle. One project involved the origin of the continental lithospheric mantle (*e.g.*, Arndt *et al.*, 2002). My research with *Helle Pedersen*, my favourite geophysicist, has required that I learn something about seismology and she something about petrology (she can now talk about metasomatised harzburgite without the slightest hesitation). After the organisation of a workshop in Canada (Pedersen *et al.*, 2009) and the publication of several papers on the subject (Bruneton *et al.*, 2004), we are now speculating on the existence of a carbonate-rich layer in the upper part of the lithosphere. This work also



reinforced my scepticism about any model that invokes the derivation of high-volume magmas or magmatic ore deposits from metasomatised sub-continental lithospheric mantle (Arndt, 2013).



Images courtesy of Dominique Weis, Al Hofmann, Nick Arndt and the Fondation Ecologie d'Avenir (Institut de France).

Figure 9.1

People I have worked with on subjects related, more or less directly, with formation of the continental crust. They are, from the top right and in no particular order: Al Hofmann, Anne Davaille, Claude Herzberg, Tim Elliott, Catherine Chauvel, Francis Albarède (shucking oysters before a New Year's dinner), Steven Moorbath, Bill White, Chris Hawkesworth, Pierre Choukroune, Janne Blichert-Toft, Robo and Billy, Jean de Bremond d'Ars and Bernard Auvrey, Catherine (again) with Steve Goldstein, Jon Patchett.



Another long-term project that deals indirectly with questions related to the formation of the continental crust is a program of scientific drilling in the Barberton Greenstone Belt in South Africa (see Fig. 9.2). This project, subtitled “Peering into the Cradle of Life” aims to recover complete sections of core through the volcanic and sedimentary rocks of the belt and in so doing to provide samples that would help us understand the conditions that reigned at the surface and in the interior of Earth during the first part of her history. The project also allowed me to maintain an interest in my main passion, komatiites. Preliminary studies of the drill sites resulted in a new theory for the origin of these ultramafic lavas (Robin *et al.*, 2011) and the kilometre of komatiite core that we recovered will provide material for years of future studies.



Figure 9.2 The page from the ICDP website (http://www.icdp-online.org/front_content.php?idart=2709) describing the Barberton drilling project.

9.2 Perspectives

What must we continue to work on, if we are to understand better the formation of the continental crust? It is obvious that more fieldwork is needed in regions of old crust, and more analyses using the latest methods are required on old rocks and particularly on old zircons. Most recent advances in our knowledge of the generation and evolution of the continental crust have been driven by the development of new analytical techniques which have produced new types of analytical data. Multiple-isotope *in-situ* analysis of zircons is the best example of this. There is also an important role for analogue and numerical modelling of the geodynamics of a younger, hotter planet. In the following sections, I focus on several key geological issues which, in my opinion, must be explored in more detail.



- (1) *The role of water, and how it gets into the site of melting.* As I have emphasised repeatedly in this *Perspectives*, large volumes of granitic magma are only produced if water in one form or another is available. An abundant source of aqueous fluid is accessible in subduction zones (though not in intraplate settings) but it is not clear how this fluid, initially released by destabilisation of hydrous minerals in the subducting slab, finds its way to source of granitic magma in the lower part of the crust. We can surmise that an influx of water triggers melting in the mantle wedge and that the result is hydrous basaltic magmas that ascend into the crust where they will fractionally crystallise to more evolved intermediate and felsic magmas. But there is ample evidence that granitic production also involves remelting of solid basalt most probably within the lower crust of the island arc or convergent margin. How does water find its way into the melting region? A separate aqueous fluid could in theory permit “fluid-excess” melting of solidified wall rocks, but it is not clear how the conditions imposed in experiments, where rock and fluid are confined in the same capsule, are reproduced in nature. Fluid does not migrate up the temperature gradient surrounding a zone of melting; rather the opposite – it will be driven away from the region where granitic magma is generated. Very probably aqueous fluid is released as basaltic magma stagnates and crystallises and these fluids react with and transform surrounding rocks to assemblages of hydrous minerals. A likely process is that the dominantly ultramafic cumulates that constitute the basal zones of island arcs will transform, at least in part, to amphibole which, when reheated, will trigger dehydration melting. The same will happen to amphibole that crystallises from solidifying hydrous basalt. Through such a process, the products of solidification of incoming basaltic magmas will first be hydrated, then will re-melt to yield more felsic magmas.

A problem is that amphibole is not stable in the lower parts of thick crust; *i.e.* at the depths where garnet can remain as an abundant residue phase. The presence of garnet is required to explain the distinctive HREE-depleted trace-element pattern of Archean TTG. What is needed is closer consideration of the stability of two key minerals – amphibole and garnet – under conditions appropriate for the formation of granitic magma.

- (2) *Hydration of Archean oceanic crust.* The extent of hydration of the thick oceanic crust that formed in the Archean is important for several reasons. First, it affects the density of the crust and thus its subductability. If a significant thickness of the basaltic upper part of the crust is transformed to hydrous assemblages, as I argue in Section 5, then it will make the crust more buoyant and more difficult to subduct. The mechanism of interaction of hotter Archean seawater with thicker, on average more magnesian oceanic crust requires further investigation. The depth that seawater penetrates into oceanic crust, both at the ridge and during deformation preceding subduction, is poorly



known. A crucial aspect is whether aqueous fluid can reach the more olivine rich lithologies deeper in the crust and below the upper layer of evolved, olivine poor basalt. Hydration of these layers produces water-rich serpentine or chlorite rather than water-poor amphibole. More strongly hydrated Archean oceanic crust releases more water into the mantle wedge, but since both mantle and slab are hotter, the dehydration will take place at shallower depths. The influence of more abundant water on the stability of garnet – the mineral that so influences the composition of Archean TTG – also merits further study.

- (3) *Composition of the lower part of the crust.* In almost all experiments designed to investigate the origin of granitic magma, the starting material is basalt. The composition varies, from komatiitic through tholeiitic to calc-alkaline, and much is made of the degree to which the compositions of experimental melts match those of modern or ancient granitoids. Yet all such basalts are evolved magmas whose compositions represent neither those of the primary magmas produced in the mantle source, nor those of the rocks that make up the lower parts of an island arc (or any other type of crust). Primary magmas generated in the mantle have more magnesian, picritic compositions; when these magmas reach the Moho, they differentiate into evolved basalts, which ascend to the surface, and mafic-ultramafic cumulates, which remain at the base of the crust, or founder into the underlying mantle. The rocks susceptible to hydration and re-melting to generate granitic magmas may have ultramafic, not basaltic compositions. This should be taken into account when selecting material for experiments and interpreting the results of these experiments.

Further work is needed to understand the factors that influence whether mafic-ultramafic cumulates will remain in the crust or be cycled back into the mantle. Important controls are the densities and viscosities of the cumulates and the underlying mantle, which are influenced in part by their compositions but more strongly by their temperature. At magmatic temperatures, the mantle viscosity is low enough that refractory cumulates regularly founder, exposing less mafic more fertile rock at the base of the crust. This process, which needs to be understood more fully, is crucial to the broad-scale differentiation that builds the continental crust.

- (4) *Isotopic analyses of zircons.* As discussed in Section 7, the Pb ages and Hf and O isotopic compositions of zircons are our main source of information on the rate and mechanism of formation of the continental crust. The Jack Hills zircons have provided almost all the information that we have about the first crust of the Earth and the analysis of these samples, and the search for comparable occurrences, will certainly continue. The interpretation of the data – the trace-element compositions of the zircons, the nature of origin of their inclusions and above all the interpretation of their ages and Hf isotopic compositions – is



not straightforward. Some of the pitfalls have been discussed recently by Rasmussen *et al.* (2011), Vervoort *et al.* (2013), Griffin *et al.* (2013). Simultaneous analysis of Pb and Hf isotopic compositions in the same portions of zircon grains will help resolve many of the ambiguities.

More significant in the context of this *Perspectives* is the use of Hf isotopic compositions to infer the rate of crust formation. It is crucial to distinguish between juvenile input into the crust from the mantle and internal recycling within the crust. To do this requires that we know the Hf isotopic composition of the mantle source. Very commonly that of the depleted upper mantle is used as a reference. In two recent papers, for example, a juvenile granite is said to have an isotopic composition within two epsilon units (Condie, 2011), or $\pm 0.75\%$ (Griffin *et al.*, 2013), of the depleted mantle curve. This choice is questionable. It may be appropriate for basalts of the oceanic crust and for adakites and other magmas produced by melting of the basaltic portion of this crust, but it cannot be used for the more common magmas in subduction settings. If the goal is to monitor growth of the continental crust; *i.e.* the flux of magma from the mantle source into the crust, then the composition of the mantle wedge above a subduction zone, as determined from the compositions of island arc lavas, is a better choice. Dhiume *et al.* (2011) argued this way when they proposed their “new crust” reference curve. In my opinion they did not go far enough. Their reference has a present-day ϵ_{Hf} value of about +13 but the data in Figure 2.7 show that this value corresponds to the more positive portion of the island arc field: the compositions of arc lavas in fact extend continuously to $\epsilon_{\text{Hf}} = 0$ with extreme values to $\epsilon_{\text{Hf}} = -15$. To be sure, the lower values are due to the presence of a component from subducted sediment, which in large part represents recycled continental crust, but, if a “juvenile” magma is defined as magma transferred rapidly from its mantle source without a long residence in the crust, then the full range of ϵ_{Hf} plotted in Figure 2.7 could qualify as candidates for the mantle source composition.

This issue is pertinent to calculations of the proportion of “juvenile” vs “recycled” components in granites. As pointed out in Sections 2, 5 and 7, the proportion of recycled crust in a magma with an ϵ_{Hf} or ϵ_{Nd} close to zero is low: in the example used in Figure 7.9, the proportion of old crust is only around 10% and 90% of the magma added to the crust comes from the mantle. When more appropriate mantle reference values are used, the proportion of juvenile material becomes far greater than reported by, for example, Condie *et al.* (2009; their Fig. 7.6) or Griffin *et al.* (2013). Such issues should be considered when evaluating the proportion of new continental crust associated with the mid-Precambrian age peaks.



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LIST OF ACRONYMS

CAD	Crustal Accretion-Differentiation super-event
HREE	Heavy Rare Earth Elements
LILE	Large Ion Lithophile Elements
LIP	Large Igneous Province
LREE	Light Rare Earth Elements
MASH	Mixing, Assimilation, Storage and Homogenisation
MOMO	Mantle Overturn, Major Orogeny
REE	Rare Earth Elements
SCLM	Sub-Continental Lithospheric Mantle
SHRIMP	Sensitive High-Resolution Ion Microprobe
TTG	Tonalite Trondjemite Granodiorite



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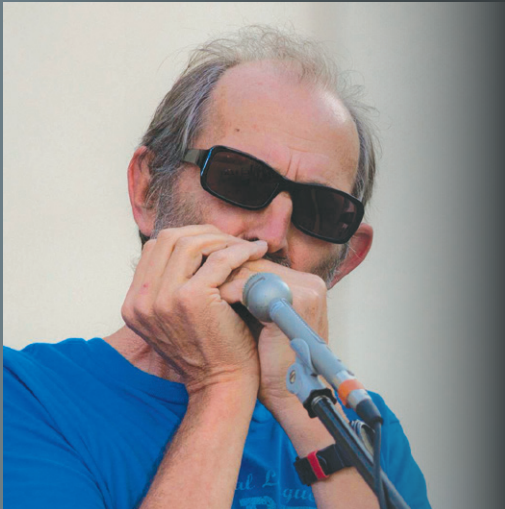
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