TECTONIC MODELS FOR THE GEOLOGICAL EVOLUTION OF CRUST, CRATONS AND CONTINENTS IN THE ARCHAEOAN

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Abstract The earliest sialic crust was probably destroyed by meteorite impacts, leaving only detrital 4.2-3.9 Ga zircons in later sediments. The oldest rocks are 3.9-3.8 Ga high-grade gneisses commonly associated with lavas and sediments which were transported into the deep crust. They indicate that mantle recycling was extensive, and that large volumes of surprisingly mature continental crust had formed by 3.7-3.8 Ga. Decreasing radiogenic heat production of the Earth is consistent with increasing indication of appreciable Archaean oceanic lithosphere and plume-generated oceanic plateaux, which in turn require extensive subduction, which accounts for the current evidence for many island arcs in the Archaean. Accretion of collages of arcs led to formation of the first protocratons bordered by the first identifiable active continental margins. Sections of upper Archaean crust, seen in many greenstone belts, and of deep Archaean crust, represented by granulite-gneiss belts, indicate massive thrust-generated imbrication of diverse rock units and tectonic belts presumably in collisional events. The Archaean was a period of high crustal growth and the eventual formation of stable cratons of mature continental crust with thick keels of sub-continental lithosphere. Sedimentary successions in greenstone belts are comparable to those in Phanerozoic depositional basins, and preserve a record of sedimentation adjacent to oceanic islands, in calc-alkaline island arcs, in syn- to post-rift stable shelves, in foreland basins, and strike-slip pull-apart basins. The presence of several late Archaean sedimentary basins many kilometres deep indicates the local stability of continental crust. Towards the end of the Archaean impingement of mantle plumes beneath mature continental lithosphere in the centre of some cratons led to reheating, diapirism, and generation of diverse magma types. This heralded the beginning of the Proterozoic, by which time large continents or even supercontinents had formed.

Introduction The Archaean eon lasted from the age of the oldest rocks at about 4.0 Ga to 2.5 Ga, about a third of geological time (Condie, 1994; Windley, 1995). It is imperative to understand how the earth's crust evolved through this critical period of time, because the subsequent development of the continents was strongly influenced by the nature of that pre-2.5 Ga evolution. But primarily the Archaean was the period when the earliest crust segregated from the mantle, when the first cratons and terranes amalgamated by accretionary processes, and when the first orogens, continents and even supercontinents evolved (Rogers, 1996). The aim of this paper is to review current perspectives on how these early growth processes took place in relation to current ideas of crust-lithosphere tectonics.

The Earliest Crust Evidence for the existence of very early crustal material includes:

- Detrital zircons, quartzite, Australia: 4.2-3.9 Ga. (Maas *et al.* 1992)
- Detrital zircons, quartzite, Montana: 3.96 Ga. (Mueller *et al.* 1992)
- Detrital zircons, gneiss, Bavaria: 3.843 Ga (Gebauer *et al.* 1989)
- Detrital zircons, quartzite, N.E. China: 3.85 Ga (Liu *et al.* 1992)
- Granite in gneiss, N.E. China: 3.81 Ga (Song *et al.* 1996)
- Acasta gneiss, NW Canada : 3.962. (Bowring *et al.* 1989)
- Gneiss, Enderby Land, Antarctica: 3.927 Ga (Black *et al.* 1986)
- Uivaaq Gneiss, Labrador: 3.863 Ga (Schiotte *et al.* 1989)
- Itsaq Gneiss, Greenland: 3.9 Ga (Nutman *et al.* 1996)

Three important discussion points arise from consideration of this information:

1. The sources of the detrital zircons have never been found. Maybe they were destroyed by meteorite impacts, calculated to have been intense before 4.0 Ga (Taylor 1993).
2. The oldest rocks on Earth are deep crustal gneisses or granites in granulite-gneiss belts. The oldest upper crustal rocks in any greenstone belt are 3.5-3.7 Ga at Barberton,

South Africa (Kroner *et al.* 1996). Thus the key to understanding the earliest stages of our planet's evolution lies in unravelling the very complicated granulite-gneiss regions.

3. Segregation of crustal or sialic material from the earth's mantle was already advanced by 4.0-3.8 Ga, but it may have been sporadic. The existence of extremely positive initial ϵ_{Nd} values in 3.7-3.8 Ga ultramafic rocks from Greenland and Labrador demonstrates the presence of a highly LREE-depleted and fractionated mantle reservoir prior to 3.8 Ga (Collerson *et al.* 1991). The presence of these Nd isotope ratios higher than that of the bulk Earth suggests that by 3.8 Ga the volume of the crust was as large as 40% of the present value. The fact that only a minute fraction (<1 %) of that crust is preserved may be an indication of removal by mantle recycling, possibly therefore by subduction processes. According to Chase and Patchett (1988) the temporary storage of that subducted mafic-ultramafic oceanic crust was responsible for the high ϵ_{Nd} values.

The reason for only localised development of very early crust may have been because mid-oceanic ridges before 4.0 Ga stood above sea-level, preventing interaction between the mantle and hydrosphere, and efficient production of continental crust. Not until 4.0-3.6 Ga were ridges drowned, allowing appreciable continental crust to evolve (de Wit *et al.* 1992; de Wit and Hynes 1995). However, de Wit *et al.* (1992) suggested that the period 4.0-3.6 Ga was one of intraoceanic obduction, and that modern-style processes of subduction did not start until 3.0 Ga when the volume of low-density serpentinite had decreased, because of the decline of the Mg content of oceanic crust. This last point seems unlikely in view of geochemical and isotopic evidence from West Greenland for the production of widespread tonalite-trondhjemite-granodiorite (TTG) suites by 3.7 Ga interpreted as the result of melting of subducted mafic oceanic crust (Nutman *et al.* 1993).

Plumes and Plateaux Not long ago the following argument prevailed. The total heat flux of the Earth and the heat production from the breakdown of radiogenic isotopes were two to three times greater in the Archaean than today (Richter, 1985). Some sixty-five percent of the heat loss from the Earth today is used in oceanic crust creation and cooling (Sclater *et al.* 1981). Therefore, in the absence of any other efficient method of dissipating heat, the production of Archaean oceanic crust was higher than today. Then came the realisation that oceanic plateaux, generated from hot mantle plumes, could have provided an important additional heat-mantle contribution to early growth (Reymer and Schubert 1986), or indeed that plumes may have been the primary means of cooling of the Earth (Vlaar *et al.* 1994). This can be put into perspective when one considers that the rate of production of a single, albeit the largest, extant oceanic plateau, the 5.0 x 10⁷ km³ Aptian Ontong Java plateau, was 8-22 km³ yr⁻¹, which probably exceeded the contemporaneous global production rate of the entire mid-oceanic ridge system (Coffin and Eldholm 1993). As Stein and Hoffmann (1994) proposed, the evolution of the earth may have been punctuated by periods of increased plume activity, of which the Archaean was the first and most prominent.

The discovery of 3.2-3.3 Ga diamonds derived from over 150 km depth demonstrates that a stable continental root existed by that time below the Kaapvaal craton (Richardson *et al.* 1984). From their Re-Os isotope studies Pearson *et al.* (1995) found that shallow, spinel-facies and deep, diamond-facies Kaapvaal peridotites have similar ages of 3.3-3.5 Ga and concluded that 150 km of mantle lithosphere had accumulated quickly and that the stabilisation of cratonic lithosphere occurred by, at least, 3.5 Ga, when the lithosphere was over 200 km thick. Only hot mantle plumes would have been capable of generating rapidly such thick lithospheric keels by a process of harzburgite crystallisation from high-degree (50 %) mantle melts.

The tholeiitic lavas in greenstone belts, which have high Fe, Ni and Cr, low Al, and depleted incompatible trace elements, are comparable geochemically to modern flood basalts (Arndt 1994) which are widely recognised as having been generated by plume-lithosphere interaction (Tarduno *et al.* 1991). These geochemical similarities suggest that magmas derived from mantle plumes may have been important in building Archaean crust (Hill 1993, Choukroune *et al.* 1997).

The bimodal volcanic rocks in Archaean greenstone belts could have been produced from mantle plumes, the abundant tholeiitic basalts from melting in the cool head of the plume, and the relatively rare komatiites by melting in the hot conduit axial jet (Campbell *et al.* 1989, Arndt *et al.* 1997). Modern hot spots and MORE share a common depleted source and thus Archaean komatiites appear to have come from the same deep depleted reservoir as modern picrites and MORE (Anderson 1994). Kroner (1991) proposed that in the early pre-plate tectonic Earth islands of Iceland-type, thick, plume-induced crust gave rise to the basaltic-komatiitic volcanism. In contrast, MORB-like greenstone lavas underlain by tonalitic-trondhjemitic plutons could have formed as a consequence of subducting a ridge segment within 500 km of a hotspot located beneath a continent, as happened 3-6 Ma ago in the Taitao Peninsula in Chile (Nelson and Forsythe 1989, Abbott 1996).

Most debate about modern analogues of ancient plateaux has concerned the voluminous upper lavas, such as the tholeiitic-komatiitic suites of the Malarctic-Val d'Or area in the Canadian Superior Province, which are comparable geochemically to Tertiary lavas of the oceanic plateau on Gorgona island (Storey *et al.* 1991, Kimura *et al.* 1993). But what about the deeper parts of plateaux? Based on the account of Nivia (1996) of the Bolivar mafic-ultramafic complex in Columbia, as the obducted lower crust of the Caribbean-Columbian

oceanic plateau, Kent *et al.* (1996) suggested that the deeper portions of an Archaean plateau consisted of norite underlain by Iherzolite, orthopyroxenite, gabbro-norite and dunite. As yet, descriptions of possible deep sections of Archaean plateaux are rare. One example may be in the Superior Province where the basal part of the Vicien belt consists of serpentinite schists with gabbroic pods overlain by gabbros and pillow basalts (Skulski and Percival 1996). Not surprisingly controversy has arisen about the interpretation of some Archaean sequences. For example, Kusky and Kidd (1992) interpreted a thrust-based 6.5 km thick succession of 2.7 Ga basaltic and peridotitic komatiites in the Belingwe greenstone belt in Zimbabwe as a fragment of an accreted and fragmented oceanic plateau. However, Bickle *et al.* (1994) reaffirmed the existence of an unconformity between the lavas and underlying granites and gneisses, negating the plateau model. But what about the ridge subduction plume model of the Taitao Peninsula referred to above?

Oceanic Lithosphere Modern-type ophiolites have not yet been reliably documented from Archaean terranes. Nevertheless, some interesting, partly controversial, examples have been proposed including: the Barberton ultramafic-mafic complex (de Wit *et al.* 1987); an 11 km thick mafic intrusive-extrusive complex in the Slave Province in Canada, where sheeted mafic dykes occur (Helmstaedt *et al.* 1986) with pillowed basalts which have MORB-type chemistry (Cunningham and Lambert 1989); also in the Slave Province mafic-ultramafic assemblages with an underlying high-temperature dynamothermal aureole (Kusky 1990); dismembered ophiolitic fragments in Wyoming (Harper 1985); and a succession of pillow basalts, mafic dykes, gabbros and serpentinites on a terrane boundary in the Minnesota River Valley (Southwick and Chandler 1996).

More recently, 3.1-3.3 Ga low-K tholeiites with MORB-type chemistry were described in Australia by Ohta *et al.* (1996), who proposed that the potential mantle temperature was about 1200°C higher than today and that the oceanic crust would have been 2-3 times thicker than today; crustal thicknesses of about 20-22 km thick were suggested by Sleep and Windley (1982), Hoffman and Ranalli (1988), and Bickle *et al.* (1994). Delamination of volcanic units in a greenstone belt in the Superior Province led to preservation of just pillow basalts with modern type oceanic crust chemistry (Tomlinson *et al.* 1996).

Many oceanic-looking basalts from Archaean greenstone belts have geochemical affinities more like modern supra-subduction back-arc basalts, and thus an oceanic back-arc model has been proposed. Examples are in the Superior Province (Tomlinson *et al.* 1996) and the Zimbabwe craton (Jelsma *et al.* 1996). However, basalts dredged from the modern southern Chile Ridge close to the advancing continental margin have hybrid MORB-arc geochemical characteristics, comparable to those of some Archaean greenstone basaltic lavas, leading support to the idea that ridge subduction may have been an important mechanism in the Archaean (Karsten *et al.* 1996).

Arcs and Subduction Zones The creation of extensive oceanic lithosphere in the Archaean would necessitate a high degree of subduction to maintain a non-expansive Earth. Evidence for the existence in Archaean greenstone belts and granulite-gneiss belts of rocks with chemical affinities comparable to modern island arcs and continental arcs is abundant. For example there is a huge quantity of high quality structural-geochemical-isotopic data that indicate that a vast segment of the Superior Province of Canada consists of arc-derived crust that formed in the period 3.1-2.65 Ga (Card 1990, Williams *et al.* 1992, Kimura *et al.* 1993, Sutcliffe *et al.* 1993). Several

process-oriented problems are worth considering for Archaean arcs:

1. Many U-Pb age determinations indicate that the locus of subduction and arc accretion of the Superior Province migrated southwards (Thurston *et al.* 1991). The two principal tectonic models to account for this accretion are:
 - a. Arc-arc collision. This model, widely accepted by early workers (e.g. Hoffman 1989, Williams 1990, Thurston and Chivers 1990), involved the progressive southward accretion of new arcs, each on their own subduction zones, the arcs or collections of arcs being separated by sedimentary prisms.
 - b. Migrating or prograding arc-trench model (Hoffman 1991, Kimura *et al.* 1993, Jackson and Cruden 1995). Many oceanic or continental fragments were swept northwards and accreted on one subduction zone to create an extensive accretionary package. This mechanism requires that new arcs were developed on each newly accreted accretionary prism, that the single subduction zone and trench backstepped or migrated oceanwards, this being caused by choking of the subduction zone, and that both the initiation and cessation of arc magmatism show an oceanward migration (Kimura *et al.* 1993). It is interesting that Sengor *et al.* (1993) employed a similar model to explain the accretionary collage of Central Asia through the Palaeozoic.
2. Modern oceanic crust is typically 8-10 km thick, and oceanic plateaux are up to 32 km thick (Gladchenko *et al.* in press). Archaean oceanic crust has been commonly calculated to have been about 20-22 km thick (references given above), and Archaean oceanic plateaux would have been even thicker (50 km?). Therefore, accretion of Archaean oceanic crust would have been close in tectonic style to that of modern oceanic plateaux (Kimura *et al.* 1993). Because Archaean oceanic lithosphere was very chemically depleted (Chase and Patchett 1988), it is likely that oceanic plateaux would have been buoyant and thus would have accreted rather than subducted, so increasing the possibility of choking of subduction zones.
3. In modern or ancient accretionary prisms we never find complete sections of either oceanic crust or plateaux, but most commonly only thin slices of basalt and pelagic sediment. The permeability contrast between low temperature-altered, more buoyant upper oceanic crust from the remainder of the downgoing slab causes delamination, the former being obducted and accreted, and the denser, mantle-dominated lower part of the plate being subducted (Kimura and Ludden 1995). To overcome the buoyancy problem, Hoffman and Ranalli (1988) suggested similar subduction-related delamination of Archaean oceanic crust. A comparable process may affect the upper and lower parts of accreting island arcs.
4. Once a substantial collage of Archaean accreted arcs, plateaux, continental fragments and accretionary prisms had been built up, they would act as an incipient microcontinent or proto-craton, the leading edge of which would effectively become an active continental margin, so giving rise to Andean-type magmas. Percival *et al.* (1994) described such an active 2.72 Ga continental margin magmatic arc in the Vizion greenstone belt in Canada, which has been

imbricated with a sliver of 2.78 Ga plume-related oceanic plateau crust, and a 2.72 Ga volcanic sequence representing continental rift deposits (Skulski and Percival 1996). Likewise in the Superior Province, in the Golden Pond sequence of the Abitibi belt intraoceanic arc rocks have been imbricated with mantle-derived oceanic rift rocks (La Flache and Camire, 1996), and in the Beardmore-Geraldton greenstone belt rocks from oceanic, arc and back-arc crusts have been delaminated and juxtaposed (Tomlinson *et al.* 1996).

5. The paper by Martin (1986) has had profound influence on ideas of Archaean subduction tectonics. He proposed that the location of calc-alkaline magma genesis in subduction zone environments has changed with time from more slab melting without dehydration in the Archaean to more mantle wedge melting as a result of slab dehydration in post-Archaean time. This process would have facilitated the formation of abundant Andean-type tonalites, which indeed we see today in many Archaean granulite-gneiss belts such as West Greenland where melting of subducted mafic oceanic crust produced (3.70 Ga microcontinents consisting of tonalite-trondhjemite-granodiorite (TTG) suites (Nutman *et al.* 1993). Similar high Al TTG suites continue to form in modern arcs where hot oceanic crust less than 25 Ma old has been subducted (Defant and Drummond 1990). The 3.0 Ga TTG suite in the Lewisian of NW Scotland were also probably the product of partial melting of subducted mafic crust (Rollinson 1996).

Archaean Tectonics The previous sections have concerned data and models which relate to the origin of Archaean rocks. Here we consider the style and interpretation of structures which deformed the rocks, and which were responsible for their tectonic emplacement and transformation to their present state. Ideas about Archaean tectonics have changed considerably over the last few decades; most recent interpretations of Archaean structures relate them to plate tectonic processes (Myers and Kroner 1994).

Many Archaean terranes, generally termed greenstone-granite belts formed in the upper crust and contain rocks that come from oceanic crust, oceanic plateaux, island arcs, fore-arcs, back-arcs, continental arcs, and accretionary wedges (e.g. the Superior Province, Williams *et al.* 1992). In contrast, many terranes, generally termed granulite-gneiss belts, contain rocks that have been deformed and metamorphosed in the deep continental crust. However, the presence of meta-supracrustal rocks shows that upper crustal material has been transported to the deeper parts of continental terranes, where they were tectonically intercalated with rocks that had formed in the deep crust.

If the Superior Province represents lateral accretion of juvenile crust at the margins of a protocraton, then the Dharwar craton of India may reflect reheating of lower and middle crust by plume impact in the centre of a craton (Choukroune *et al.* 1997). The Dharwar craton is characterised by dome-basin structures, granitic diapirs and intervening triple junctions of mafic material, which Chardon *et al.* (1996) ascribed to the impact of a plume at 2.5 Ga, which resulted in diapirism and vertical growth of crust. Under such mature continental regions, where the subcontinental lithosphere is thicker, the interaction of a hot mantle plume would yield a higher proportion of basalts relative to komatiites, as seen, for example, in the Aravalli lavas of NW India (Ahmad and Tarney 1994).

One of the key challenges in Archaean tectonics is to work out the geological environments in which different rock groups formed and how and why they were tectonically

juxtaposed. In this respect a prime problem concerns the tectonic relationship between upper and lower crustal belts. Only rarely do we find evidence of a complete crustal section from the upper to the lower Archaean crust, as seen in the Kapuskasing Uplift in Ontario and confirmed by the LITHO-PROBE seismic reflection profile (Ludden *et al.* 1993, Percival and West, 1994). However, there is controversy concerning the interpretation of this section. High-precision U-Pb ages of Krogh (1993) demonstrate a progressive downward younging of crustal growth. According to Percival *et al.* (1997) the greenstone lavas formed in an island arc, and the deeper crustal gneisses were generated underneath the lavas in a slightly younger Andean-type continental margin. In contrast, Choukroune *et al.* (1997) assert that the mafic lavas of the upper crust are geochemically primitive and probably formed in an oceanic environment and not on mature crust. They bear no geochemical relationship with the underlying mature tonalitic gneisses and plutons and therefore the upper supracrustal rocks are regarded as allochthonous with respect to the mid-lower crust.

The development of extensive sedimentary basins in the late Archaean indicates that the continental crust had locally attained sufficient rigidity to sustain the load of sedimentary piles many kilometres thick. Examples include: the 11 km-thick, 3.1-2.7 Ga Witwatersrand Supergroup, the 8 km-thick, 2.7 Ga Ventersdorp Supergroup, and the 15 km-thick, 2.56 Ga Transvaal Supergroup all in southern Africa, the 3.5-5 km-thick, 2.11-2.1 Ga Fortescue Group in NW Australia, and the 3 km-thick, 2.79 Ga Oraniemi Group in Finland (refs. in Windley, 1995).

From their exhaustive survey Eriksson *et al.* (1994) concluded that Archaean greenstone belts contain six lithological associations that can be matched with the following Phanerozoic depositional environments: 1. barred lagoons and bays

around oceanic volcanic islands and sediment-starved platforms adjacent to coalesced volcanoes in inter-arc, intra-arc and back-arc basins. 2. forearc trenches and marine volcano-plutonic arcs. 3. cratonic extensional basins in arc-continent and intracontinental rifts. 4. continent-adjacent syn- to post-rift stable shelves and arc-adjacent post-rift stable shelves. The proportion of these sedimentary successions increased from 4.0 to 2.5 Ga in response to the progressive growth of the continents (Eriksson and Fedo 1994). 5. compressional foreland basins of arc-continent collisional and compressional-arc tectonic basins. 6. strike-slip collisional graben in hinterland tectonic-escape and terrane-accretion orogens.

Many recent studies show that both greenstone-granite belts and granulite-gneiss belts are composite terranes that formed by the tectonic amalgamation of contrasting types of rock groups of different age and origin (e.g. the composite Slave Province, Kusky 1990). Geochronological data show that both types of belts have a long history (Thurston *et al.* 1991, Friend and Nutman 1991), and structural studies demonstrate that both types of belts comprise an imbricate stack of inter-thrust rock groups of different provenance (e.g. Skulski and Percival 1996, Nutman *et al.* 1991). In these respects it is important to recognise that many greenstone-granite belts and granulite-gneiss belts have several features in common; they both consist of an imbricated amalgamation of rocks of diverse origin. Much mature continental crust with thick sub-continental lithosphere had developed to form orogens or small continents by the end of the Archaean. Impingement of mantle plumes beneath such lithosphere would give rise to crustal reheating and diapirism, more basalts relative to komatiites, and more diverse magmas including alkaline types (Ahmad and Tarney 1994). This heralded the beginning of the Proterozoic at 2.5 Ga by which time large continents or even supercontinents had formed.

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