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Archean greenstone-tonalite duality: Thermochemical mantle convection models or plate tectonics in the early Earth global dynamics?

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Abstract

Mantle convection and plate tectonics are one system, because oceanic plates are cold upper thermal boundary layers of the convection cells. As a corollary, Phanerozoic-style of plate tectonics or more likely a different version of it (i.e. a larger number of slowly moving plates, or similar number of faster plates) is expected to have operated in the hotter, vigorously convecting early Earth. Despite the recent advances in understanding the origin of Archean greenstone–granitoid terranes, the question regarding the operation of plate tectonics in the early Earth remains still controversial. Numerical model outputs for the Archean Earth range from predominantly shallow to flat subduction between 4.0 and 2.5 Ga and well-established steep subduction since 2.5 Ga [Abbott, D., Drury, R., Smith, W.H.F., 1994. Flat to steep transition in subduction style. Geology 22, 937–940], to no plate tectonics but rather foundering of 1000 km sectors of basaltic crust, then “resurfaced” by upper asthenospheric mantle basaltic melts that generate the observed duality of basalts and tonalities [van Thienen, P., van den Berg, A.P., Vlaar, N.J., 2004a. Production and recycling of oceanic crust in the early earth. Tectonophysics 386, 41–65; van Thienen, P., Van den Berg, A.P., Vlaar, N.J., 2004b. On the formation of continental silicic melts in thermochemical mantle convection models: implications for early Earth. Tectonophysics 394, 111–124]. These model outputs can be tested against the geological record. Greenstone belt volcanics are composites of komatiite–basalt plateau sequences erupted from deep mantle plumes and bimodal basalt–dacite sequences having the geochemical signatures of convergent margins; i.e. horizontally imbricated plateau and island arc crust. Greenstone belts from 3.8 to 2.5 Ga include volcanic types reported from Cenozoic convergent margins including: boninites; arc picrites; and the association of adakites–Mg andesites- and Nb-enriched basalts.

Archean cratons were intruded by voluminous norites from the Neoarchean through Proterozoic; norites are accounted for by melting of subduction metasomatized Archean continental lithospheric mantle (CLM). Deep CLM defines Archean cratons; it extends to ~350 km, includes the diamond facies, and xenoliths signify a composition of the buoyant, refractory, residue of plume melting, a natural consequence of imbricated plateau-arc crust. Voluminous tonalites of Archean greenstone–granitoid terranes show a secular trend of increasing Mg#, Cr, Ni consistent with slab melts hybridizing with thicker mantle wedge as subduction angle steepens. Strike-slip faults of 1000 km scale; diachronous accretion of distinct tectonostratigraphic terranes; and broad Cordilleran-type orogens featuring multiple sutures, and oceanward migration of arcs, in the Archean Superior and Yilgarn cratons, are in common with the Altaid and Phanerozoic Cordilleran orogens. There is increasing geological evidence of the supercontinent cycle operating back to ~2.7 Ga: Kenorland or Ur ~2.7–2.4 Ga; Columbia ~1.6–1.4 Ga; Rodinia ~1100–750 Ma; and Pangea ~230 Ma. High-resolution seismic reflection profiling of Archean terranes reveals a prevalence of low angle structures, and

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evidence for paleo-subduction zones. Collectively, the geological–geochemical–seismic records endorse the operation of plate tectonics since the early Archean.

Keywords: Archean; Greenstone; Tonalite; Plate tectonics; Thermochemical convection models; Arc volcanism; Continental mantle lithosphere; Accretion; Plumes; LITHOPROBE

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1. Introduction and scope

Plate tectonic models have been constructed for Archean greenstone–granitoid terranes since the 1970s. These models were based primarily on field relationships, in which multiple lithotectonic terranes, each having distinct lithological associations, ages, structural history, and metamorphic grade(s) were juxtaposed along regional scale structures, and therefore were interpreted to have been accreted allochthonously (Talbot, 1973; Bridgwater et al., 1974; Burke et al., 1976; Tarney et al., 1976; Windley, 1976, 1993; Langford and Morin, 1976; Condie, 1981; Card and Cielsielski, 1986; Card, 1990; Hoffman, 1989; de Wit, 1998). From the geochemistry of sedimentary rocks, specifically rare earth element (REE: La to Lu) patterns, Taylor and McLennan (1985, 1995) interpreted the Archean upper crust to be composed of the bimodal association of arc basalts and dacites (or tonalities, the intrusive equivalent of dacite) erupted in convergent margins, with a transition to post-Archean upper continental crust having an andesite signature. It was explicitly realized that plate tectonics may have operated in a modified form compared to the Phanerozoic, and that mantle plume activity was more intense until ~1.9 Ga (Fyfe, 1978; Stein and Hofmann, 1994; Isley and Abbott, 1999). A contrary model involving largely vertical tectonics, ruling out operation of subduction zone geodynamic process in the Archean, was proposed by several studies (Grachev and Federovsky, 1981; Hamilton, 1998; McCall, 2003; Stern, 2005).

That debate, centered on field relationships and geochemistry, has been joined by contrasting numerical models. Based on a thermal model of the mantle, Abbott et al. (1994) calculated that shallow to flat subduction prevailed between 4.0 and 2.5 Ga, and transition from flat to well-established steep subduction occurred at about 2.5 Ga, coinciding with the major change to Proterozoic and Phanerozoic sedimentary REE patterns (Taylor and McLennan, 1985). Butler and Peltier (2002) modeled the thermal evolution of layered convection, and deduced that there has been minor secular change in near surface heat flow.

Given the higher mantle temperatures in the early Earth, extensive partial melting in the convective upper mantle is expected to have produced thicker oceanic crust (>20 km) than its modern counterpart at the mid-oceanic ridges (Sleep and Windley, 1982). From theoretically predicted thermal (higher temperature), viscosity (lower viscosity), and rheological (e.g., weaker strength) characteristics in the early Earth's upper mantle, van Thienen et al. (2004a) conclude that the operation of plate tectonics in the Archean is ruled out by buoyancy relationships. Rather, they construct numerical "thermochemical mantle convection" models to account for the duality of Archean greenstone–granitoid terranes. In their model, they used 600 × 600 and/or 1200 × 1200 km square-shaped domains as starting point for delamination, resurfacing, thickened crust, and mantle diaprisim. They assumed that the vertical boundaries of the domains are periodic, and the bottom of the boundary has a free slip condition, whereas the horizontal, tangential stress component is set to zero. The model involves two linked stages: initially, foundering of the entire section of 600 or 1200 km long sectors of thick basaltic crust, derived from convecting asthenosphere, triggered by the phase transition from basalt to denser eclogite at base of the oceanic crust (>30 km). In a second stage, upwelling of complementary mantle generates large volumes of basalt by decompressional partial melting, termed 'resurfacing'; forming the greenstone terranes. Melting at the base of the newly produced basaltic crust generates extrusive and/or intrusive felsic rocks, adding new material to the greenstones. In addition, partial melting of the foundered metabasalts under amphibolite grade metamorphic conditions produces voluminous tonalite–trondhjemite–granodiorite (TTG) batholiths intruding the greenstones (van Thienen et al., 2004a,b).

In this paper, we compile recent geochemical, field relationships, and seismic data for Archean cratons as 'boundary conditions' for evaluating the numerical models, which variously yield plate tectonic, non-plate tectonic, or plume outputs. Specifically, we address for Archean cratons: (1) The composite nature of Archean greenstone terranes; (2) Archean volcanic associations, including supracrustal boninites and island arc picrites, and Cenozoic-type arc association of tholeiitic to calc-alkaline basalts, adakites, Mg-andesites, and Nb-enriched basalts; (3) the thick, buoyant, refractory continental lithosphere mantle (CLM) "keel" that defines Archean cratons; (4) Archean ophiolites; (5) Precambrian norites; (6) tonalitic batholiths that have traversed a peridotitic mantle wedge; (7) mantle plumes; (8) oceanic plateaus; (9) regional strike-slip faults; (10) craton assembly by diachronous accretion of allochthonous terranes; (11) superfamilies of orogens (cf. Şengör and Natal'in, 1996, 2004); (12) the supercontinent cycle; and (13) seismic data.

2. Discussion

2.1. Archean greenstone terranes

Archean greenstone terranes (Fig. 1) are dominated by two volcanic associations. A komatiite–basalt association erupted from anomalously hot mantle plumes: these form intra-oceanic plateaus, or alternatively continental flood sequences where there is direct evidence of underlying continental crust, there are xenocrystic zircons, or there is geochemical evidence of contamination by crust (Sun et al., 1989; Arndt, 1994; Arndt et al., 2001; Condie, 1994, 2000; Dostal and Mueller, 2004).

The second association is bimodal tholeiitic to calc-alkaline basalts and dacites (Condie, 1994; Thurston, 1994; Sylvester et al., 1997). Primitive and evolved basalts that erupted in an intra-oceanic setting, and therefore are devoid of crustal contamination, are characterized by systematic negative anomalies of Nb, Ta, P, and Ti relative to neighbouring REE (see Section 2.2). This is a characteristic of all Phanerozoic and Recent intra-oceanic convergent margin magmatism (Perfit et al., 1980; Saunders et al., 1991; Hawkesworth et al., 1993; Pearce and Peate, 1995; Kelemen et al., 2004). Accordingly, the Archean bimodal basalt–dacite association has been interpreted as convergent margin magmatism (Taylor and McLennan, 1985, 1995; Condie, 1994; Polat et al., 1998; Manikyamba et al., 2004, 2005).

In the Abitibi greenstone belt the former association has spatially related high-Mg, Cr, Ni turbidites, interpreted as shed off komatiite–basalt ocean plateaus, whereas the latter has spatially related low-Mg, Cr, Ni greywackes with compositions consistent with mixing of ~60% basalt and 40% dacite shed off the bimodal arcs as trench turbidites (Feng and Kerrich, 1990). The volcanic–sedimentary associations are collectively intruded by TTG batholiths, over a duration of <5 to >40 Ma after termination of volcanism (Jackson and Fyon, 1991; Ayer et al., 2002).

Structural analysis of Archean greenstone terranes reveals that the first and second volcanic associations are tectonically imbricated (Naqvi and Rogers, 1987; Polat et al., 1998; Polat and Kerrich, 1999; Manikyamba et al., 2004). Accordingly, to a first approximation Archean greenstone supracrustal terranes on

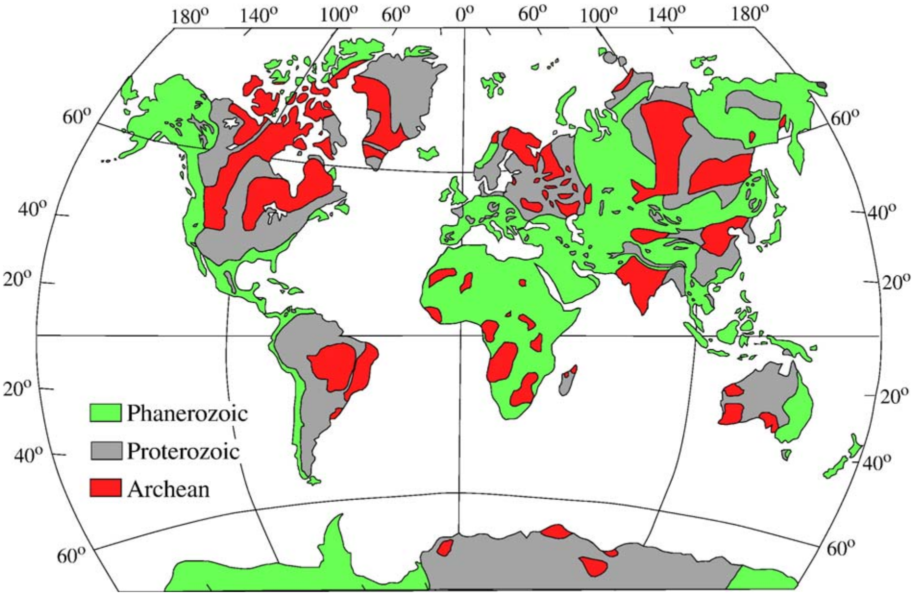

Fig. 1. World map showing the distribution of Archean, Proterozoic and Phanerozoic crusts. Modified from Kusky and Polat (1999).

many cratons are tectonically interleaved ocean plateau and arc crust (Kimura et al., 1993; Puchtel et al., 1998; Wyman et al., 1999; Shchipansky et al., 2004). A mixture of plateau and arc crust solves the question of "excess" Mg and Ni in Archean sedimentary compositions by virtue of incorporating komatiites (Wronkiewicz and Condie, 1989; Condie and Wronkiewicz, 1990; Rudnick, 1995; Polat and Kerrich, 2000).

Wyman et al. (1999, 2002) suggested a specific mechanism to account for these lithotectonic relationships: migrating arcs were "captured" by thick plateau crust that jams the arc; subduction is either stalled at the accretionary boundary, or "jumps" across the plateau. The Destor-Porcupine and Kirkland Lake-Cadillac accretionary boundaries, of the ~2.7 Ga Abitibi greenstone belt, are potential examples of subduction "jumping" across a plateau (Wyman et al., 1999). There is seismic evidence for a subducted slab "stalled" beneath the former structure (Fig. 2; Calvert and Ludden, 1999; see Section 2.12). Currently, the migrating New-Ireland arc has been captured and jammed by the Ontong-Java plateau (McInnes et al., 1999; Mann and Taira, 2004).

2.2. Greenstone volcanic associations

The geochemical characteristics of Phanerozoic and Recent mafic volcanic flows from different geodynamic settings are distinct in terms of rare earth element and high field strength element (HFSE: Th, Nb, Ta, Ti; high ratio of charge/ionic radius) interrelationships (Sun and McDonough, 1989; Saunders et al., 1991; Hawkesworth et al., 1993; Hofmann, 1997; Pearce, 2003). These differences stem primarily from source depletion (e.g. source of mid ocean ridge basalts: MORB) or enrichment (e.g. Ocean island basalts: OIB, Island arcs), depth of melting (garnet present or absent), and water activity (MORB, OIB anhydrous; arcs hydrous). Based on uniformitarian principles, many workers proceed on the basis that specified groups of elements will behave in a similar manner during geodynamic–petrogenetic processes in the Archean, especially given that REE (Ce, Eu excepted) and HFSE are resistant to modification by syn-magmatic hydrothermal alteration or greenschist facies metamorphism (Thurston, 1994; Condie, 1994, 2000; Sylvester et al., 1997; Arndt et al., 1998). We focus on volcanic sequences from largely intraoceanic settings, as determined from low Ce–Yb

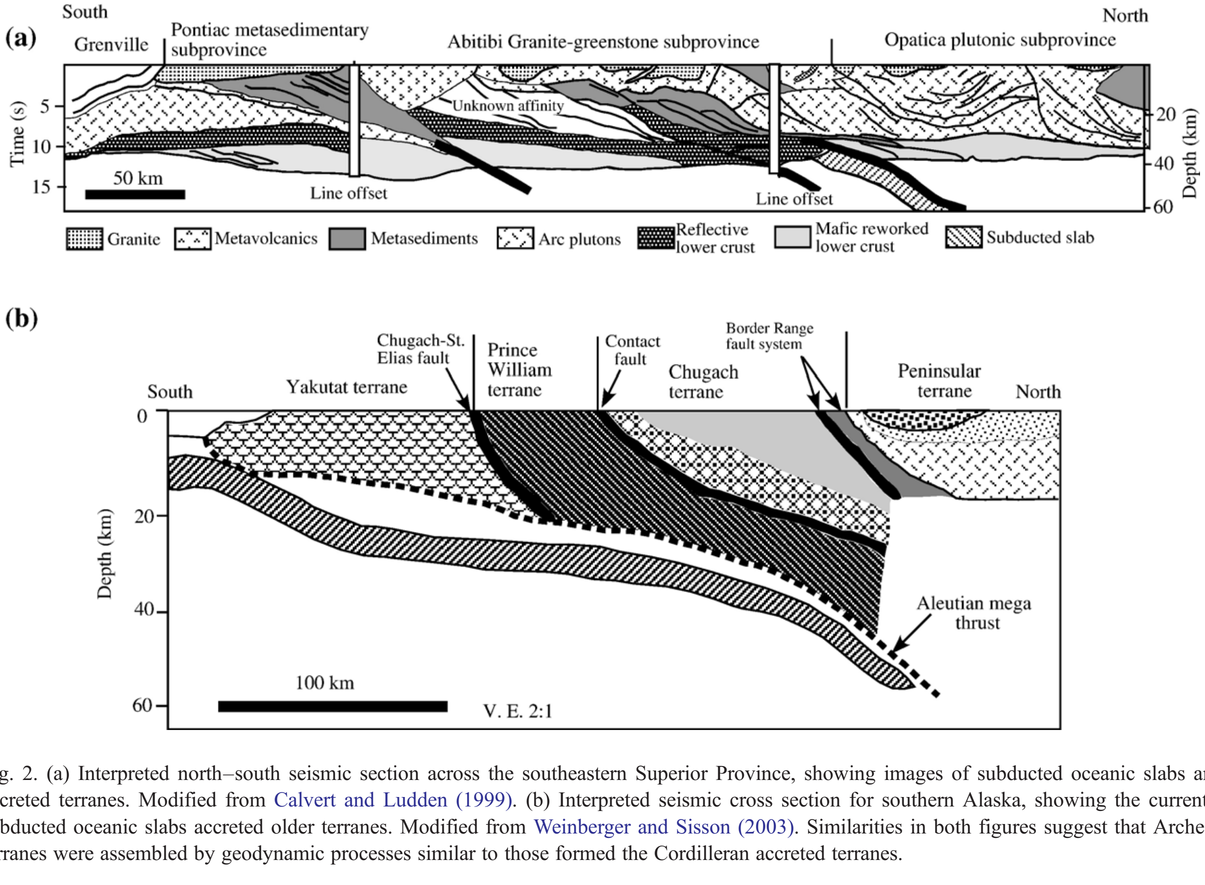

Fig. 2. (a) Interpreted north–south seismic section across the southeastern Superior Province, showing images of subducted oceanic slabs and accreted terranes. Modified from Calvert and Ludden (1999). (b) Interpreted seismic cross section for southern Alaska, showing the currently subducted oceanic slabs accreted older terranes. Modified from Weinberger and Sisson (2003). Similarities in both figures suggest that Archean terranes were assembled by geodynamic processes similar to those formed the Cordilleran accreted terranes.

trends (cf. Hawkesworth et al., 1993) combined with absence of xenocrystic zircons. This strategy circumvent complexities in magmas that have erupted through, and been contaminated by, continental crust (cf. Sun et al., 1989). Multi-element diagrams, that include selected REE and HFSE, are by convention normalized to the composition of primitive mantle such that elements at percent concentrations (e.g., Ti) can be compared to elements including Th and Yb at part per million levels, on a *y*-axis with only a few orders of magnitude (Figs. 3 and 4).

Tholeiitic basalts associated with komatiites in Archean terranes are geochemically comparable to Phanerozoic ocean plateau basalts erupted from plumes, such as Ontong Java basalts, and Recent plume-related basalts on Iceland (Hemend et al., 1993; Mahoney et al., 1995; Puchtel et al., 1998, 1999; Kerrich et al., 1999). Similarly, many tholeiitic basalts of the Archean bimodal association are indistinguishable from Recent intra-oceanic arc basalts that are generated by slab dehydration in a convergent margin, and mantle wedge hydration and melting (Taylor and McLennan, 1985, 1995; Hawkesworth et al., 1993; Polat and Kerrich, 2001; Hollings and Kerrich, 2004; Manikyamba et al., 2004, 2005). In contrast, some basalt sequences in the 2.7 Ga Abitibi belt are interpreted to have formed by hybridization between depleted plume asthenosphere and adakite melts generated by slab melting, based on trace element trends between the two end-members (Wyman, 2003; Wyman and Hollings, 2006).

Recently, types of volcanic rocks formerly thought to be restricted to Phanerozoic or contemporary

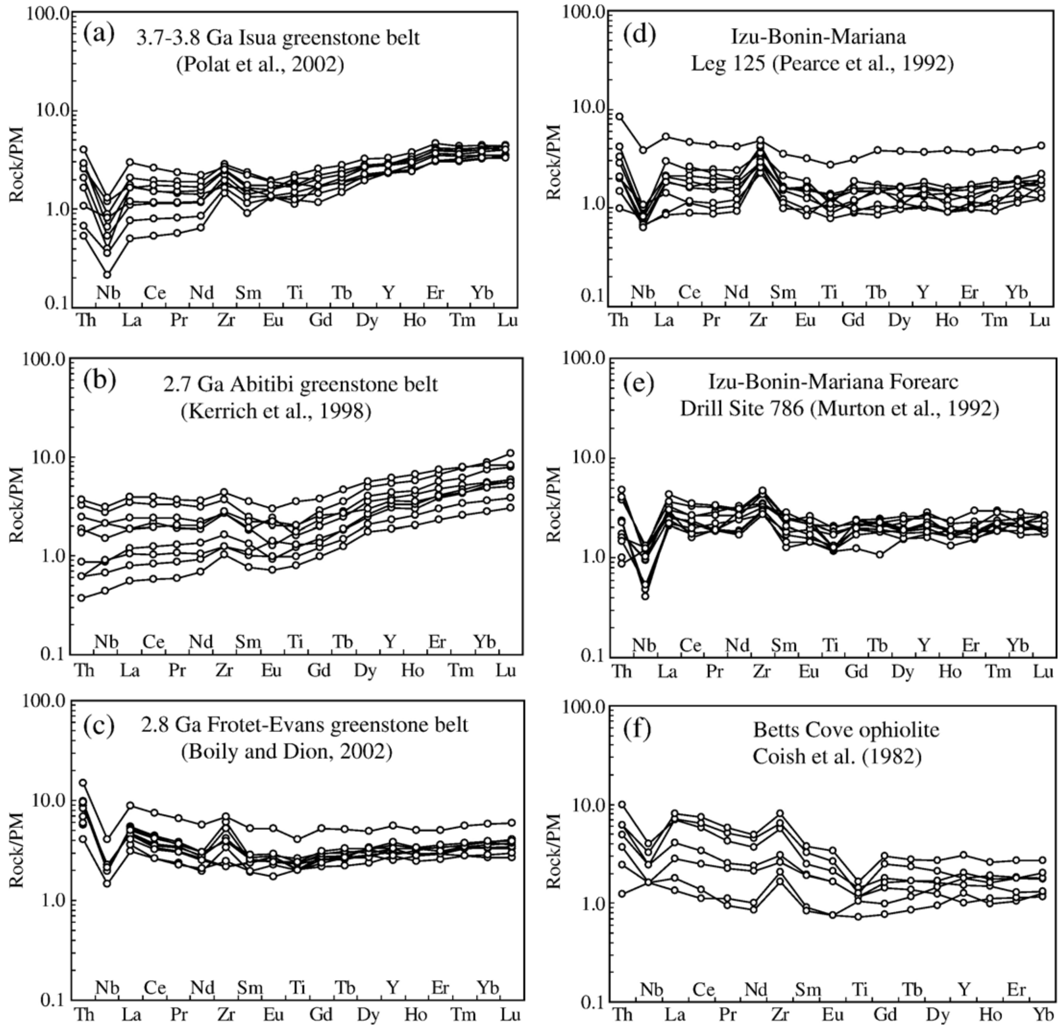

Fig. 3. Primitive mantle-normalized trace element diagrams for Archean boninitic volcanic rocks and Phanerozoic counterparts. Normalization values are from Hofmann (1988). Data for the Betts Cove ophiolite and Izu–Bonin–Mariana forearc are from Coish et al. (1982) and Murton et al. (1992), respectively.

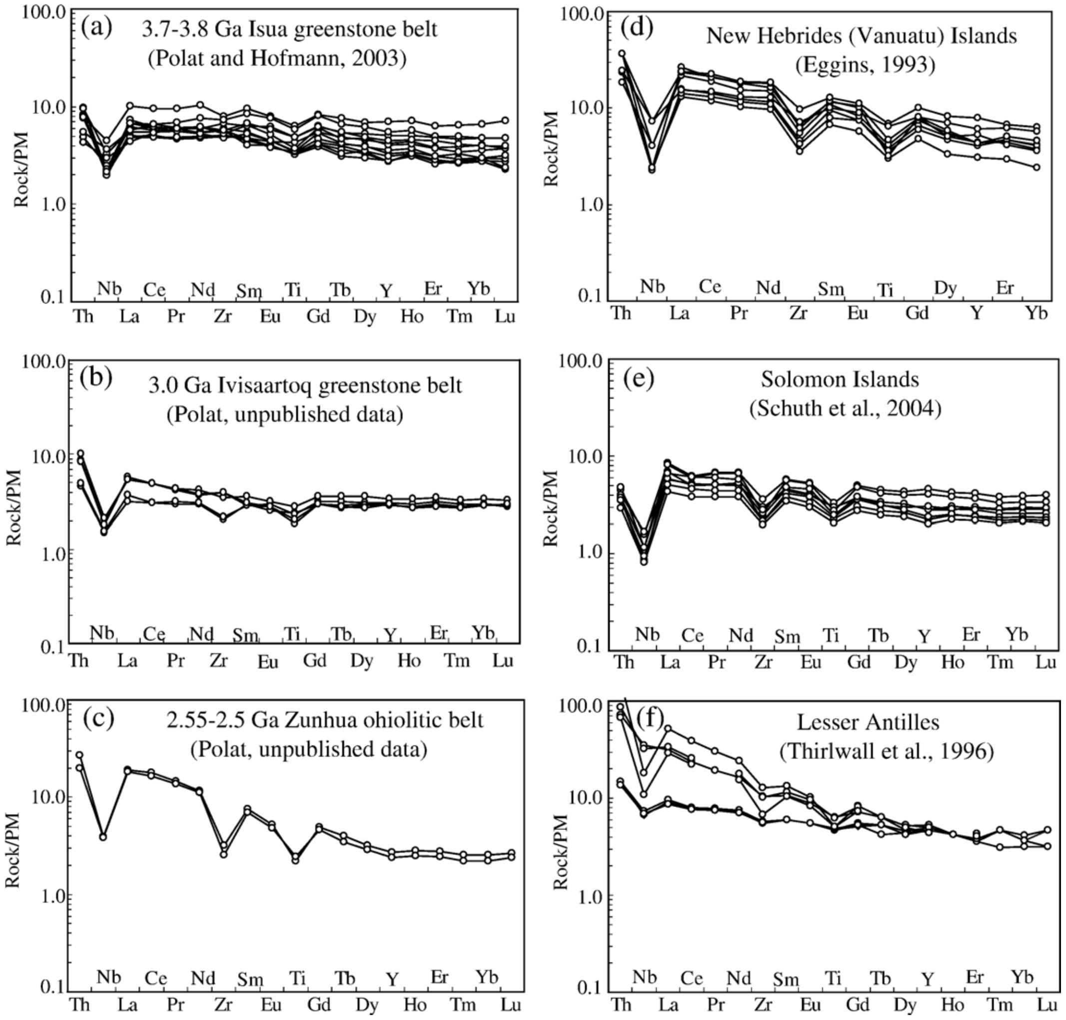

Fig. 4. Primitive mantle-normalized trace element diagrams for Archean picritic volcanic rocks and Phanerozoic counterparts. Normalization values are from Hofmann (1988).

convergent margin settings have been identified in Archean greenstone terranes. Boninites, or boninite series, refer to primary, or near primary, liquids compositionally defined as $SiO_2 > 53$ wt.% and Mg# > 69 (Crawford et al., 1989). Boninites feature concave-upward REE patterns, negative anomalies at Nb–Ta relative to light REE (LREE: La, Ce, Nd), in conjunction with low Ti content but high Al_2O_3/TiO_2 ratios (Fig. 3; Sun and Nesbitt, 1978; Hickey and Frey, 1982; Stern et al., 1991; Taylor et al., 1994). They form in a two stage processes: in stage 1, high degree anhydrous partial melting of mantle peridotite leaves a residue depleted in Ti and LREE–MREE; for stage 2, in a convergent margin, hydrous fluxing of the residue adds LREE generating melts with the characteristic concave-upward REE pattern (Fig. 3; Hickey and Frey, 1982; Crawford et al., 1989; Pearce and Peate, 1995; Pearce, 2003).

Boninites have not only been identified in western Pacific arcs such as the Izu-Bonin-Mariana arc, but have also been identified in the Tethyan, Appalachian, and Ural supra-subduction ophiolites (Beccaluva and Serri, 1988; Pearce et al., 1992; Bédard, 1999; Spadea and Scarrow, 2000; Ishikawa et al., 2002). Occurrences of flows with boninitic compositions in Archean terranes are compiled in Table 1. They are spatially related to a variety of volcanic types associated with convergent margins such as tholeiitic to calc-alkaline basalts.

Picrites occur in Phanerozoic plume and arc settings, the former without but the latter with the REE/HFSE fractionation characteristic of convergent margin settings (Eggins, 1993; Schuth et al., 2004; Thirlwall et al.,

Table 1
Archean hot subduction volcanic rocks and their Phanerozoic analogs

Rock type	Archean example	Phanerozoic analogs
Boninite	Abitibi (Kerrich et al., 1998); Isua (Polat et al., 2002); Frotet-Evans (Boily and Dion, 2002); Karelian (Shchipansky et al., 2004); Whundo (Smithies et al., 2004, 2005a); Dharwar Craton (Manikyamba et al., 2005)	Troodos ophiolite (Beccaluva and Serri, 1988); Papua New Guinea ophiolite (Hickey and Frey, 1982; Crawford et al., 1989); Izu-Bonin-Mariana forearc (Pearce et al., 1992); Betts Cove ophiolite (Bédard, 1999); Oman ophiolite (Ishikawa et al., 2002); Ural ophiolites (Spadea and Scarrow, 2000)
Low-Ti tholeiite (LOTI)	Abitibi (Kerrich et al., 1998; Wyman, 1999); Karelian (Shchipansky et al., 2004)	Mariana trench, Lau Basin, New Guinea (Beccaluva and Serri, 1988); Tasmania (Brown and Jenner, 1989)
Picrite	Isua (Polat and Hofmann, 2003); Wawa (Polat and Kerrich, 1999); Zunhua (Polat et al., submitted for publication); Ivisaartoq (Polat, unpublished data)	Solomon Islands (Schuth et al., 2004); Central Aleutians (Nye and Reid, 1986); Japan (Yamamoto, 1988); Vanuatu arc (Eggins, 1993); Eastern Kamchatka (Kamenetsky et al., 1995); Lesser Antilles (Thirlwall et al., 1996)
Adakite	Wabigoon (Tomlinson et al., 1999); Sumozero-Kenozero (Puchtel et al., 1999); Birch-Uchi (Hollings and Kerrich, 2000); Wawa (Polat and Kerrich, 2001); Abitibi (Wyman et al., 2002); Yellowknife (Cousens et al., 2002); Frotet-Evans (Boily and Dion, 2002); Ashuanipi complex (Percival et al., 2003); Wabigoon (Ujike and Goodwin, 2003); Wutaishan (Wang et al., 2004); Vedlozero-Segozero (Svetov et al., 2004)	Central America (Defant et al., 1992); Philippines (Sajona et al., 1993); Japan (Morris, 1995); Southern Volcanic Zone, Andes (Stern and Kilian, 1996); Baja California (Benoit et al., 2002); Northern Volcanic Zone, Andes (Bourdon et al., 2002)
High-magnesian andesite (HMA)	Birch-Uchi (Hollings and Kerrich, 2000); Wawa (Polat and Kerrich, 2001); Abitibi (Wyman et al., 2002)	Chile (Rogers and Saunders, 1989); Philippines (Sajona et al., 1996); Japan (Tatsumi and Maruyama, 1989); Western Aleutians (Yogodzinski et al., 1995); Antarctic Peninsula (McCarron and Smellie, 1998); Baja California (Calmus et al., 2003)
Nb-enriched basalt (NEB)	Birch-Uchi (Hollings and Kerrich, 2000); Wabigoon (Wyman et al., 2000; Ujike and Goodwin, 2003); Wawa (Polat and Kerrich, 2001); Uchi (Hollings, 2002); Karelian (Shchipansky et al., 2004)	Central America, Panama (Defant et al., 1992); Northern Kamchatka (Kepezhinskas et al., 1996); Philippines (Sajona et al., 1996); Baja California (Benoit et al., 2002)

1996). Both compositional types are documented in Archean greenstone belts (Francis et al., 1999; Polat and Hofmann, 2003). Island arc picrites occur in several early to late Archean greenstone belts (Fig. 4; Table 1; Polat and Kerrich, 1999; Polat and Hofmann, 2003; Polat et al., submitted for publication; and Polat unpublished data). High MgO, Ni, and Cr concentrations in these lavas are attributed to large degrees of partial melting of the mantle wedge during subduction of young, hot oceanic crust, or ridge subduction (see Schuth et al., 2004; and references therein). Slab-derived fluids likely play an important role for the initiation of partial melting and the generation of LREE-enriched, HFSE-depleted trace element patterns in these lavas.

Adakites are volumetrically minor flows initially reported from Cenozoic arcs, spatially and temporally related to tholeiitic to calc-alkaline basalts. Compositionally, they are distinctive in terms of high Al-, and Na-contents, fractionated REE with low HREE, but high Cr, Ni contents. They are generally interpreted to be melts of slab basalt with residual garnet, that acquire elevated Mg, Cr, Ni by hybridizing with peridotite traversing the mantle wedge (Drummond et al., 1996; Martin, 1999; Smithies, 2000; Condie, 2005a; Martin et al., 2005). They have been documented in the Aleutian, Baja California, Philippines, Southern Andes, and Central American arcs (Table 1; Sajona et al., 1996, Stern and Kilian, 1996; Benoit et al., 2002).

The geodynamic setting in which adakites are generated is considered to be flat subduction of young, hot oceanic lithosphere (Drummond et al., 1996). Currently, 10% of subduction is flat, notably in the Peru sector, central Chile, Ecuador, NW Columbia, Costa Rica, Mexico, southern Alaska, SW Japan, and western New Guinea (Table 1; Gutscher et al., 2000). According to van Hunen et al. (2002, 2004), three mechanisms may promote flat subduction; overthrusting of the subducting plate, subduction of a plume generated ocean plateau, or slab suction forces. Hargraves (1986) argued that the thermal regime in the Archean was such that there were smaller, slower moving, plates and accordingly greater ridge length. In this scenario, ridge subduction would be more common, a cause for high heat flow at convergent margins accounting for adakites (Polat and Kerrich, 2004). Ridge subduction at the Chile triple junction has yielded diverse magmatic types including adakites (Guivel et al., 2003). Given more intense plume activity in the Archean generating numerous oceanic plateaus (Campbell and Griffiths, 1992; Isley and Abbott, 1999, 2002), as well as higher heat flow, flat subduction was more prevalent.

In several occurrences of Cenozoic adakites, there is an association with high-Mg andesites (MA) and Nb-enriched basalts (NEB) (Table 1). MA are intermediate volcanic rocks having greater abundances of Mg, Cr, Ni than ‘normal’ andesites. NEB are mafic volcanic rocks possessing Nb ~7–20 ppm, as against ~2 ppm for ‘normal’ arc basalts. This association is generally interpreted as follows: adakites are slab melts mildly hybridized with wedge peridotite, Mg-andesites are more extensive hybrids with wedge peridotite, whereas NEB represent the residue of hybridization that melted at greater depth from induced convection in the wedge (Defant et al., 1992; Yogodzinski et al., 1995; Kelemen, 1995; Kepezhinskas et al., 1996; Benoit et al., 2002; Kelemen et al., 2004). Documented adakites, and the adakite–MA–NEB association from Archean terranes are compiled in Table 1. They are interpreted to have formed in similar geodynamic settings as Cenozoic counterparts (Hollings and Kerrich, 2000; Polat and Kerrich, 2001, 2004).

Fundamentally, arcs cannot generate melts over several Ma without migration, subduction of water and wedge fluxing or slab melting. De facto, arc volcanic sequences require plate translation (Hamilton, 1988).

2.3. Archean mantle lithosphere

The continental lithosphere has a crustal sector 30–80 km thick for Archean and younger eons. Continental lithospheric mantle (CLM) is 200–350 km deep in Archean cratons but ~150 km thick for the Proterozoic, and ~100 km for younger terranes (Fig. 5; Artemieva and Mooney, 2001; Plomerova et al., 2002). Archean CLM is also more buoyant and refractory than under younger continental regions; its thickness and thermal structure lead to the preservation of diamonds (Boyd, 1989; Griffin et al., 2003; for a recent review, see Sleep, 2003, 2005). There is a bimodal distribution to Archean CLM at 300–350 km and 220–200 km, where the former exists in terranes >6–8 × 10^6 km^2 (Artemieva and Mooney, 2002). Archean CLM is depth zoned from xenolith studies. Shallower levels are peridotite, metasomatized peridotite likely by Archean subduction, and eclogite, whereas deeper levels may have the composition of the residue of melting in anomalously hot mantle plumes, likely ejected from the core–mantle boundary. Younger plumes were too cool to generate refractory residues (Jordan, 1988; White, 1988; Boyd et al., 1997; Herzberg, 1999; Isley and Abbott, 2002; Artemieva and Mooney, 2001).

Given that Archean greenstone crust is tectonically imbricated bimodal arc and plateau komatiite–basalt sequences, it is not obvious why plume residue should form the CLM to arc-dominated crust (see Herzberg, 2004). Wyman et al. (2002; and references therein) suggested a process linked to formation of greenstone belts. Komatiitic liquids separated from anomalously hot mantle plumes at ~100→400 km (Herzberg, 1992, 2004; Xie et al., 1993), and with entrained upper mantle erupted as intra-oceanic plateaus or erupted through continental lithosphere as continental flood basalts (CFB). Migrating oceanic arcs were captured by thick arc lithosphere and jammed, forming composite plateau-arc crust (Polat and Kerrich, 2000). The residue of plume melting ascended to buoyantly couple to the composite crust (Fig. 5 of Kerrich et al., 2000; Wyman et al., 2002; Wyman and Kerrich, 2002).

Thermochemical mantle convection models do not predict Archean CLM having the composition of the residue of melting in an anomalously hot mantle plume. Rather, these models predict that all “resurfaced” greenstone terranes should have basalt compositions consistent with melting of peridotite in the upper convecting mantle (van Thienen et al., 2004a,b).

2.4. Archean ophiolites

Ophiolites are obducted fragments of oceanic crust, and dunites and harzburgites of upper mantle lithosphere (Anonymous, 1972; Coleman, 1977; Dilek, 2003). Given geochemical signatures of convergent margins, ophiolites are mostly from suprasubduction zone settings (Dewey, 2003; Dilek, 2003; Hawkins, 2003; Pearce, 2003). Recently, dismembered ophiolites have been described from Neoarchean greenstone terranes of the North China craton, conferring additional evidence for plate tectonic regimes in the Archean (Wang et al., 1996; Kusky and Li, 2003; Polat et al., 2005, submitted for publication).

Podiform shaped ore-bodies of chromite are present in some Phanerozoic ophiolites; they form from hydrous melting of peridotite in convergent margins (Edwards,

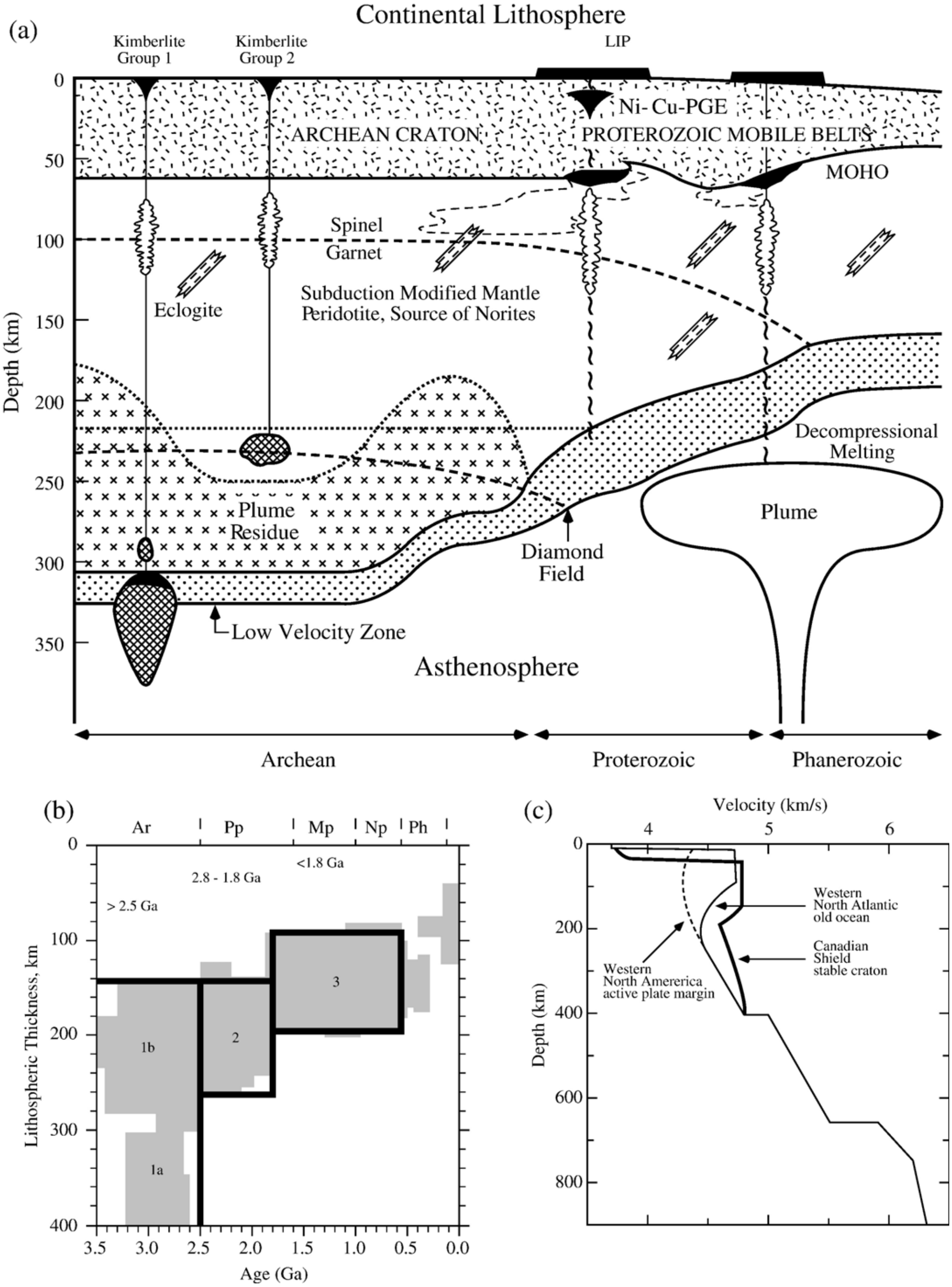

Fig. 5. Cross section through continental lithosphere illustrating the thick, refractory CLM "keel" distinctive of Archean cratons. The CLM includes sub-accreted plateau and ocean floor lithosphere, locally metasomatized by subduction, at shallower levels, the source of Neoarchean and Proterozoic cratonic norites. Deeper levels of CLM are the residues of melting in mantle plumes buoyantly coupled to overlying CLM and crust. Spinel, garnet, and diamond facies indicated. Modified from Nixon and Davies (1987), Artemieva and Mooney (2001), and Wyman and Kerrich (2002). (b) Age and thickness relationship of CLM from velocity structure, after Artemieva and Mooney (2001). (c) Depth–shear wave velocity relationships of different geodynamic settings. Modified from Kearey and Vine (1996).

Li et al., 2002; Kerrich et al., 2005). Podiform chromitites have recently been documented in harzburgites and dunites of 3.0 Ga Ukrainian and 2.5 Ga Chinese terranes (Li et al., 2002; Gornostayev et al., 2004), consistent with the emplacement of Archean subarc mantle wedge onto continental crust by arc–continent collision.

Sylvester et al. (1997) viewed many mafic to ultramafic Archean volcanic associations as dismembered Archean oceanic ophiolites. Şengör and Natal'in (2004) provided a detailed review of ophiolites (a near-complete Penrose sequence) and ophirags (dismembered pieces of ophiolites, oceanic plateaus, island arcs, ocean islands) in the Paleozoic Altaid orogenic system. They also compared the Altaid ophiolites and ophirags with Archean greenstone belts, and concluded that both orogenic systems are composed largely of ophirags. Many of these rocks originated as oceanic island arcs and forearcs that were dismembered and incorporated into the accretionary complex by orogen-parallel strike-slip faults, as in the case of Cordilleran orogeny. Collectively, these geological processes result in linear map patterns similar to those of many Archean greenstone belts. Occurrence of boninites and island arc picrites in several major Archean greenstones belts (Table 1), are consistent with the formation of some Archean greenstone belts as forearc oceanic crust (Kusky, 2004).

2.5. Precambrian norites

Voluminous norites were emplaced into Archean cratons during the Neoarchean and Proterozoic (Hall and Hughes, 1993; and references therein). They occur as dykes, sills, and mafic intrusive complexes such as the 2.05 Ga Great Dyke of Zimbabwe that represents a failed rift (Aspler and Chiarenzelli, 1998; Bekker et al., 2003; Aspler et al., 2004; see Section 2.11). Compositionally, norites are characterized by the subduction zone signature of Nb, Ta, and Ti depletions relative to fractionated and enriched REE. Hall and Hughes (1990) critically evaluate possible models for norites. From spatial and temporal distribution, and composition, they concluded that norites are partial melts of Archean CLM peridotite metasomatized by fluids driven off subduction zones during the Archean (Fig. 5; see Bell et al., 2005; and references therein). Melting of metasomatized CLM was subsequently triggered either by later decompressional melting from lithosphere extension, by impingement of a mantle plume, or some combination of these processes. This scheme accounts for their restriction to Archean cratons and Precambrian timing.

2.6. TTG batholiths

Batholiths of TTG are characteristic of Archean greenstone–granitoid terranes. Drummond et al. (1996) model TTG by melting of slab basalt crust in Archean convergent margins featuring shallow subduction, whereas in Phanerozoic arcs having steeper subduction magmatism of the basalt–andesite–dacite–rhyolite (BADR) association is triggered by slab dehydration–subarc mantle wedge melting. In a second model, Smithies (2000) argue that low-Mg, Ni TTG may be melts of very shallowly subducting oceanic lithosphere, such that the mantle wedge is thin to absent, and slab melts do not interact significantly with high-Mg, Ni wedge peridotite. Alternatively, low-Mg, Ni TTG may represent melts of the base of thick basaltic crust (Whalen et al., 2002; Condie, 2005a), which does not require subduction, a position adopted by van Thienen et al. (2004b) in their thermochemical mantle convection and resurfacing model for Archean greenstone-TTG duality.

Both slab melts and melting of crustal basalt sequences may occur under different geodynamic conditions. Martin and Moyen (2002) assembled a large database for TTG of 4.0–2.5 Ga. They showed an overall secular trend of increasing Mg and Ni in progressively younger suites; this result is consistent with the general case of steepening subduction with time such that slab melts transited through, and hybridized with, progressively thicker subarc wedge. Tarney and Jones (1994) note that Archean batholithic TTG are characterized by abundances of highly incompatible elements (Cs, Th, LREE) too great to be melts of typical LREE depleted MORB-type basalt. Rather, they suggest that TTG are melts of more fertile ocean plateau crust (cf. Kerrich et al., 2000). If correct, this idea meshes with the scheme of Wyman et al. (2002) for greenstone belts and CLM: thick plateau crust (>50 km fertile basalts, and mostly REE depleted komatiites) jams migrating arcs inducing orogenesis; plateau and arc crust imbricate; sub-accreted plateau crust melts to form TTG; and the residue of plume melting couples buoyantly as refractory CLM.

Generation of Archean TTG requires melting of hydrated oceanic crust at the depths of garnet and/or hornblende stability (Defant and Drummond, 1990; Rapp et al., 1999). As pointed out by Campbell and Taylor (1983), without hydration of oceanic crust no TTG or continent would have formed. Plate tectonic-driven oceanic spreading and subduction processes are the most efficient known mechanisms for hydration and transformation of the hydrated oceanic crust to the melting zone, respectively, to form TTG.

Recently, Polat and Frei (2005) proposed a ridge subduction model to explain the origin of the Paleoarchean TTG in the Isua region, southwestern Greenland (Fig. 6). According to this model, partial melting of laterally accreted hydrothermally altered oceanic crust under amphibolite to eclogite metamorphic conditions by upwelling of asthenospheric windows, resulting from ridge subduction produced TTG melts. The diapirically upwelling TTG melts formed the early Archean continental crust and the complementary eclogitic residues contributed to the sub-continental lithospheric mantle (Fig. 6).

Phanerozoic magmatic arcs confer additional constraints on the geodynamic setting of Archean TTG. As a general case, there is a switch from TTG to granites at the Archean–Proterozoic transition (Taylor and McLennan, 1985, 1995). However, Phanerozoic-type granites are known from Archean terranes (Vearncombe and Kerrich, 1999), and TTG batholiths occur in Phanerozoic arcs. Gromet and Silver (1987) report west to east zoning of oxygen and strontium isotope compositions, and REE fractionation in the Cretaceous Peninsular Ranges Batholith, California, generated above the subducting Pacific oceanic plate. The central and eastern zones are tonalitic, with no Eu anomalies, high-Sr, fractionated REE, and low-Yb, as for Archean TTG; these tonalities are interpreted to be melts of oceanic basaltic underplate as the arc migrated east.

In summary, Phanerozoic granites and tonalitic batholiths occur in well-constrained plate tectonic settings of steep and flat subduction, respectively. Both compositional types are present in Archean greenstone–granitoid terranes, although tonalities are prevalent. Compositional and timing relationship reveal a pattern where voluminous Archean batholithic TTG are syn-, to late-tectonic melts of >50 km fertile plateau basalt, whereas Cenozoic and Archean syn-arc adakites are low volume melts of thin (~7 km) LREE depleted MORB-type basaltic oceanic crust.

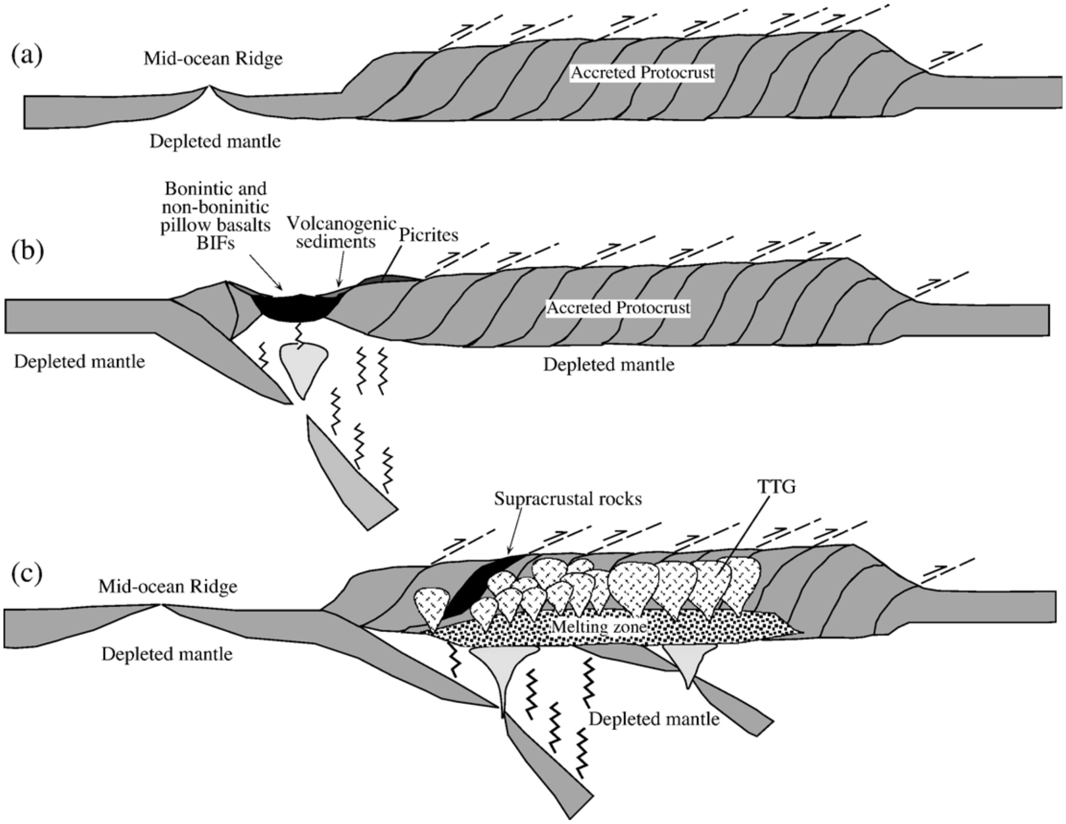

Fig. 6. Ridge subduction model for the origin of the Isua boninitic and picritic pillow basalts, BIFs and early Archean continental crust (TTG suites). Higher mantle geotherms in the early Earth implies much greater oceanic ridge length, which suggests that the early Earth was covered by many small slowly moving plates. Accordingly, subduction of oceanic ridges would have been much more common in the Archean than the present-day. A number of mantle windows developed as a consequence of a series of ridge subductions may have caused the melting of laterally accreted hydrated thickened oceanic crust under eclogitic metamorphic conditions to produce TTG. From Polat and Frei (2005).

Several recent studies documented the presence of adakite-like volcanic and intrusive rocks in intra-continental settings (Xu et al., 2002; Gao et al., 2004; Wang et al., 2005). On the basis of evolved Nd and Sr isotopes, the generation of these rocks is attributed to the melting of mafic lower continental crust by either underplating or foundering (delamination). The origin of the late Jurassic adakites and high-magnesian andesites in the North China craton has been attributed to the melting of the foundered Archean mafic lower crust in the underlying convective mantle. High-Mg numbers, and Cr and Ni contents in these lavas are interpreted to reflect interaction between melts derived from the foundered lower crust and the overlying mantle (Gao et al., 2004). Although the thermochemical convection model proposed for Archean TTG originated in intra-oceanic settings (van Thienen et al. 2004b), it may account for the generation of Phanerozoic intra-continental adakitic rocks through delamination and re-melting of the lower continental crust in the convecting upper mantle.

2.7. Mantle plumes

Komatiites are erupted from anomalously hot mantle plumes and are mostly restricted to Archean greenstone terranes (Fig. 7; Arndt, 1994; Dostal and Mueller, 2004; Herzberg, 2004; Smithies et al., 2005b). Stratigraphically associated basalts are Mg- to Fe-rich tholeiites characterized by near flat REE and no HFSE/REE fractionation, in contrast to fractionated LREE in concert with depletions of Nb, Ta, P, and Ti of Archean primitive tholeiitic arc basalts (Kerrich et al., 2000). Aluminum-depleted komatiites are characterized by negative normalized anomalies of Zr and Hf relative to MREE, indicative of melting with residual β-majorite garnet at ~400 to 670 km, and consequently be a product of shallow resurfacing (Xie et al., 1993; Kerrich and Xie, 2002). The Iceland plume has been seismically “imaged” extending to the core–mantle boundary (Bijwaard and Spakman, 1999), and similarly Archean mantle plumes were likely ejected from this boundary layer.

Isley and Abbott (1999) demonstrated common time series for mantle plumes and banded iron formation (BIF) from 3.8 to 1.9 Ga. This work has been extended by Condie (2004a) to show common time series for mantle plumes, BIF, black shales, and weathering intensity. Mantle plumes release CO_2 and other gasses generating a greenhouse that promotes silicate weathering, displace oceans across continents, and generate Fe-, and S-rich hydrothermal fluids in the submarine lava fields that precipitate as BIF (Aubaud et al., 2005; Polat and Frei, 2005; Polat et al., 2005).

The numerical models of Butler and Peltier (2002) explicitly recognize the role of mantle plumes in planetary geodynamic evolution. The thermochemical mantle convection models of van Thienen et al. (2004b) do not reproduce the geological relationships of Archean greenstone-tonalite terranes, including plume-related magmas and their products, and the deposition of some greenstone belts on the older continental basement (see Arndt et al., 2001; Blake et al., 2004; Hartlaub et al., 2004).

2.8. Oceanic plateaus

Oceanic plateaus (e.g., Ontong-Java, Caribbean, Kerguelen) are often considered as the closest modern

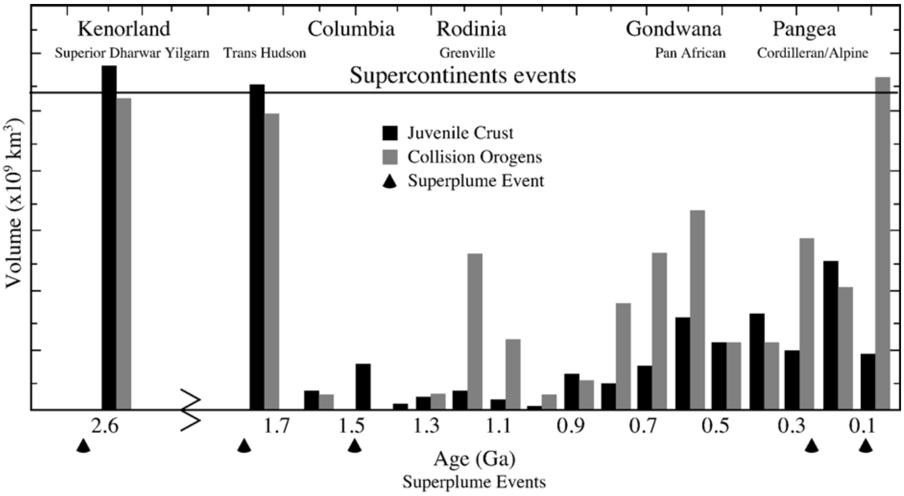

Fig. 7. A histogram showing the distribution of juvenile oceanic crust, orogenic events, super plume plumes, and supercontinents since 2.7 Ga.

analogs for the sites of the eruption of komatiite–tholeiitic basalt associations in Archean greenstone belts (Storey et al., 1991; Xie et al., 1993; Desrochers et al., 1993; Puchtel et al., 1998). The origin of both oceanic plateaus and Archean komatiite–tholeiitic basalt sequences is generally attributed to mantle plumes (Saunders et al., 1996; Kerr et al., 2000). Accordingly, the geological evolution of oceanic plateaus can be used to evaluate various geodynamic models proposed for the origin of Archean greenstone belts. Seismic studies indicate that the thickness of oceanic plateaus reaches up to 40 km (for a review, see Mann and Taira, 2004), suggesting that they are comparable to Archean oceanic crust predicted by thermochemical mantle convection models (van Thienen et al., 2004a,b). From seismic studies of modern oceanic plateaus, no evidence has yet been reported for the delamination of the base of these plateaus, or sinking of the entire plateau into the mantle, during their evolution (see Mann and Taira, 2004, Kerr et al., 2000). Instead, their initial high temperatures and positively buoyant nature result in the survival of oceanic plateaus for longer periods of time relative to normal oceanic crust. According to Saunders et al. (1996), oceanic plateaus become negatively buoyant only after $\gg 100$ Ma. In contrast to normal oceanic crust, oceanic plateaus, particularly the young ones, cannot easily subduct into the mantle (Burke et al., 1978; Cloos, 1993). Accordingly, majority of oceanic plateaus is eventually carried to subduction zones where they collide with island arcs or continental margins. If fluids are available for transformation of basalt to eclogite at the base of oceanic plateaus above the subducting oceanic crust, plateaus may be able to subduct (Saunders et al., 1996). When an oceanic plateau encounters a subduction zone, it may choke the subduction zone, causing either subduction “flip” or “backstep” (Saunders et al., 1996; Mann and Taira, 2004). Consequently, the upper parts of oceanic plateaus are sliced off, tectonically imbricated with, or accreted onto the overriding magmatic arcs (Table 2). This interaction between subduction zones and oceanic plateaus leads to the development of an island arc at the margins of the plateau, resulting in calc-alkaline magmatism (e.g. tonalites intrusion; Kerr et al., 1997). Collectively, structurally imbricated tholeiitic to calc-alkaline (island arc) and komatiite–tholeiite (oceanic plateau) volcanic associations, and intrusive tonalites in Archean greenstone belts can be explained by oceanic plateau and subduction zone interaction (Kimura et al., 1993; Polat et al., 1998; Puchtel et al., 1998).

2.9. Strike-slip faults

In Phanerozoic orogenic belts, strike-slip faults with lengths of hundreds of kilometers accommodated large displacements (Sylvester, 1988; Şengör, 1990; Kearey and Vine, 1996). Based on this observation, Sleep (1992) argued for Archean plate tectonics (see also Mueller et al., 1996; Polat et al., 1998).

2.10. Accretionary tectonics

Detailed structural mapping of diverse lithologies of the ~3.8 to 3.7 Isua greenstone belt and 2.7 Ga Abitibi-Wawa greenstone terrane reveal that these composite terranes were assembled as accretionary complexes (Feng and Kerrich, 1992; Polat et al., 1998; Komiya et al., 1999; Wyman et al., 1999; Nutman et al., 2002; Hanmer and Greene, 2002). An intrinsic feature of Phanerozoic accretionary complexes, such as the North American Cordillera, is allochthonous exotic blocks.

On the basis of detailed field observations, contrasting ages, and metamorphic histories, it is shown that the southern West Greenland Archean craton is a collage of Paleo- to Neoarchean terranes, assembled in several accretionary Archean tectonothermal events, involving accretion of exotic terranes (e.g., arcs, continental fragments) by horizontal tectonics (Bridgwater et al., 1974; Friend et al., 1996; McGregor et al., 1991; Nutman et al., 2005; Friend and Nutman, 2005; and references therein). Structural, geochronological, and trace element geochemical studies in the early Archean Isua greenstone belt are consistent with the ~3800 Ma island arc picrite–chert–BIF sequence, and the ~3700 Ma boninite–chert–BIF sequence, having been juxtaposed by accretionary Phanerozoic-like plate tectonic processes operating in the early Archean (Nutman et al., 2002; Polat and Hofmann, 2003). Recently, Friend and Nutman (2005) have shown that the Paleoarchean Archean Isukasia (ca. 3800–3600 Ma) and Mesoarchean Kapisilik (ca. 3000 Ma) terranes were tectonically juxtaposed at about 2960 Ma. This was followed by the accretion of the Neoarchean (ca. 2800 Ma) Brødre terrane to the composite Isukasia-Kapisilik terrane at about 2700 Ma.

Horizontal displacement, characterized by thrusting, imbrication, and overturning of folded-strata, has been documented in a number of greenstone belts in several Archean cratons. Examples of thrust tectonics have been documented from greenstone belts in the Superior Province, Zimbabwe Craton, Pilbara Craton, Yilgarn Craton, Kaapvaal Craton, and Baltic Shield (Table 2; Poulsen et al., 1980; Davis et al., 1988; Jirsa et al., 1992;

Table 2
Interpreted geodynamic processes in Archean greenstone–granitoid terranes and their Phanerozoic analogs

Geodynamic process	Proposed Archean examples	Phanerozoic analogs
Plume–subduction interaction	Superior Province (Dostal and Mueller, 1997; Wyman, 1999; Hollings et al., 1999; Ayer et al., 2002)	Tonga forearc (Danyushevsky et al., 1995); Mexican arc (Márquez et al., 1999)
Plume–lithosphere interaction	Superior Province (Tomlinson et al., 1999); Pilbara Craton (Arndt et al., 2001; Blake et al., 2004); Baltic Shield (Zozulya et al., 2005).	Gondwana-Land (Storey, 1995; and references therein) Siberian traps, Iceland-North Atlantic, Basin and Range (Coffin and Eldholm, 1994; Dalziel et al., 2000)
Subduction accretion (Arc–arc, arc–continent, continent–continent; arc–oceanic plateau accretion)	Superior Province (Card, 1990; Skulski and Percival, 1996; Mueller et al., 1996; Polat et al., 1998; Calvert and Ludden, 1999; Daigneault et al., 2004; Percival et al., 2004); Baltic Shield (Puchtel et al., 1998, 1999; Shchipansky et al., 2004); Dharwar Craton (Saha et al., 2004); Greenland (Nutman et al., 2002; Friend and Nutman, 2005); Kaapvaal Craton (de Ronde and de Wit, 1994); Zimbabwe Craton (Kusky and Kidd, 1992; Jelsma and Dirks, 2002); Yilgarn Craton (Şengör and Natal'in, 1996)	Himalayas, Alps, Altaids, Alaska, Appalachians, North American Cordillera (Samson and Patchett, 1991; Burchfiel et al., 1992; Kearey and Vine, 1996; Şengör and Natal'in, 2004); Japan (Taira et al., 1992); Southwestern Pacific (Hamilton, 1988; Mann and Taira, 2004); Caledonides (Andersen et al., 1990); Andes (Vergara et al., 1995).
Thrusting and imbrication	Abitibi (Lacroix and Sawyer, 1995);Wawa (Gill, 1992; Jirsa et al., 1992; Corfu and Stott, 1988); Rainy Lake (Poulsen et al., 1980); Wabigoon (Davis et al., 1988); Hearne Domain (MacLachlan et al., 2005), Bulawayo, Zimbabwe (Jelsma and Dirks, 2002); North Karelian (Shchipansky et al., 2004); Warrawoona (Kloppenburg et al., 2001); Barberton (de Ronde and de Wit, 1994); Yilgarn Craton (Davis and Maidens, 2003)	Canadian Cordillera (Crowley and Brown, 1994; Umhoefer and Schiarriza, 1996; Tardy et al., 2001); American Cordillera (Burchfiel et al., 1992; Wakabayashi, 1992); Taurides (Polat et al., 1996); Oman (Robertson and Searle, 1990); Altaids (Şengör and Natal'in, 1996, 2004); Appalachians (van Staal, 1994); Ontong Java (Mann and Taira, 2004); Caribbean (Kerr et al., 1997); Wrangelia (Richards et al., 1991)
Strike-slip faulting	Superior Province (Percival et al., 1994; Mueller et al., 1996); Dharwar Craton (Chadwick et al., 2000); Pilbara Craton (Zegers et al., 1998); Yilgarn Craton (Swager, 1997; Chen et al., 2003)	Alps and Himalayas (Şengör et al., 1985; Şengör, 1990) Altaids (Şengör and Natal'in, 1996); Canadian Cordillera (Umhoefer and Schiarriza, 1996); Andes (Dewey and Lamb, 1992); Yukon (Stevens and Erdmer, 1996)
Continental rifting	Superior Province (Tomlinson et al., 1999); Yilgarn Craton (Swager, 1997); Pilbara Craton (Blake et al., 2004); Slave Province (Bleeker, 2002); Rae Province (Hartlaub et al., 2004); Dharwar Craton (Srivastava et al., 2004) Baltic Shield (Zozulya et al., 2005)	East Africa, South and North Atlantic, Basin and Range, Rio Grande (Kearey and Vine, 1996; and references therein)
Orogenic collapse	Superior Province (Calvert et al., 2004); Greenland (Hanmer and Greene, 2002); Barberton (Kisters et al., 2003); Yilgarn Craton (Davis and Maidens, 2003)	Alps, Himalayas, Appalachians (Dewey, 1988); North American Cordillera (Vanderhaeghe and Teyssier, 2001); Basin and Range (Dilek and Moores, 1999); Urals (Knapp et al., 1998); Caledonides (Andersen et al., 1994)

de Ronde and de Wit, 1994; Lacroix and Sawyer, 1995; Jelsma and Dirks, 2002; Shchipansky et al., 2004; Daigneault et al., 2004; Table 2). Formation of these structures was followed by regional- to continental-scale strike-slip faulting and orogenic collapse (Fig. 8). Many Archean greenstone belts display non-coaxial, poly-

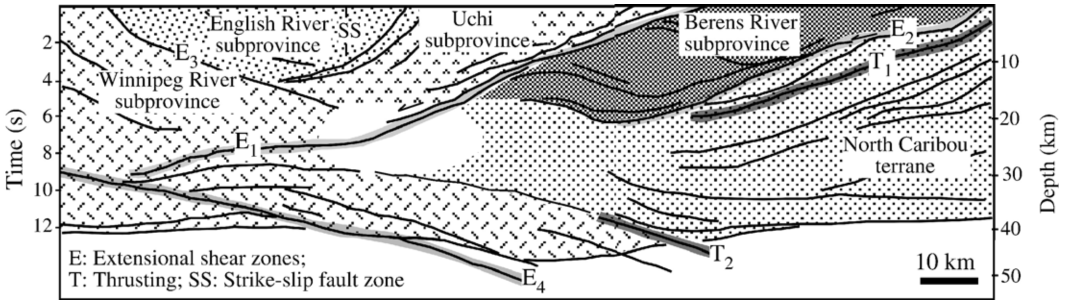

Fig. 8. Interpreted seismic section the late greenstone–granitoid Uchi and North Caribou gneissic terranes. Crustal extension is likely have resulted from collapse of a late Archean orogenic belt. Modified from Calvert et al. (2004).

phase deformation, characterized by significant transposition, asymmetric folds, and steep foliation, suggesting that horizontal stresses played an important role in the generation of these structures (Polat et al., 1998). Such structures have been attributed to Phanerozoic-style accretionary and collision processes (see Polat et al., 1998; de Wit, 1998; Blewett, 2002; Mueller et al., 1996; Daigneault et al., 2004; Polat and Kerrich, 2005).

Examples of accretionary geodynamic processes recognized in Archean greenstone–granitoid terranes are compiled in Table 2, endorsing independent evidence for large-scale horizontal tectonics.

2.11. Superfamilies of orogens

Şengör and Natal'in (1996) classified orogenic belts into two superfamilies. Accretionary, also known as Turkic or Cordilleran (e.g., Alaska) type orogens, and continent–continent collisional also termed Alpine–Himalayan or Tethyan type orogens. Cordilleran orogens represent continental growth via accretion of oceanic and continental fragments (Dickinson, 2004). The accretionary orogens are dominantly juvenile crust of subduction–accretion complexes, with extensive vertical accretion above a subducting slab. Accretionary collision of lithotectonic units is dominantly in an oblique manner, with the partitioned compressional and strike-slip components responsible for much of the deformation. Seaward growth of Cordilleran orogens involves multiple lithotectonic packages, bounded by diachronous sutures, over long durations of 100–400 Ma (Dickinson, 2004; Şengör and Natal'in, 2004). Rather than a single curvilinear magmatic arc, as in continent–continent orogens, broad (1000 km scale) arcs develop. Cordilleran type orogens show a commensurately broad pattern of subduction, where the axis of arc magmatism migrates oceanward through the accretionary prism. Examples are: the Cordillera of N. America (370 Ma–present); Altaids of Central Asia (610–250 Ma); Pan African orogen (900–630 Ma); and Trans Hudson orogen of North America (1.9–1.75 Ga; Şengör and Natal'in, 1996, 2004; Dickinson, 2004).

Şengör and Natal'in (1996) specifically compared the characteristics of the broad 2.7–2.6 Ga Yilgarn orogen, having multiple sutures, to the Cordilleran-type. Similarly, the Superior Province, with its E–W trending volcanic-, plutonic- and sedimentary-dominated subprovinces, formed by diachronous accretion of the diverse terranes across multiple sutures over 2.74–2.65 Ga (Langford and Morin, 1976; Card and Cielsielski, 1986; Hoffman, 1989; Thurston et al., 1991; Percival et al., 1994; Calvert and Ludden, 1999), with seismic evidence for subducted slabs beneath some of the sutures (Fig. 2).

Similarities between Neoarchean orogens (composite greenstone–granitoid terranes) and Phanerozoic Cordilleran orogens include: (1) accretionary tectonics; (2) mélanges (3); subduction–accretion complexes; (4) multiple sutures; (5) dismembered forearcs; and (5) magmatic arcs that migrate outboard (Kusky and Vearncombe, 1997; Polat et al., 1998; Kusky and Polat, 1999). All such orogens, in the Phanerozoic, Proterozoic, and Archean are best accounted for by plate tectonics. As a corollary, thermochemical mantle convection models do not predict the diachronous accretion of Neoarchean greenstone–granitoid composite terranes across broad multiple sutures as recorded by precise geochronology (Card and Cielsielski, 1986; Hoffman, 1989; Percival et al., 1994; Stott, 1997; Calvert and Ludden, 1999).

2.12. The supercontinent cycle

There is a consensus that the continents were assembled into the supercontinent Pangea ~230 Ma (Murphy and Nance, 2004). Earlier proposed supercontinents are Rodinia (~1100 to 750 Ma), Columbia (1.6 to 1.4 Ga), and Kenorland or Ur (2.7 to 2.4 Ga) (Fig. 7; Rogers, 1996; Condie et al., 2001; Condie, 2002, 2004b; ; Dalziel et al., 2000; Rogers and Santosh, 2003, 2004; Zhao et al., 2004). There is progressively greater uncertainty in reconstructing older supercontinents, as geological processes are intrinsically stochastic.

Evidence for assembly of Kenorland are the ~2.7 to 2.6 Ga composite greenstone Cordilleran style orogens on all Archean cratons of Laurentia, Baltica, India, Southern Africa, and Australia. Evidence for the dispersal of Kenorland includes: (1) giant mafic dyke swarms on those same cratons over 2.4–2.2 Ga, (Hall and Hughes, 1990); (2) ~2.4 Ga extensive iron formation on passive margins of Africa and Australia (Vaalbara), and South America (Trendall and Blockley, 2004); and (3) ~2.4 to 2.2 Ga siliciclastic–carbonate sequences on the passive margins of Laurentia and Baltica (Bekker et al., 2003).

2.13. Seismic evidence

High precision veibroseis transects have been conducted in the Archean Superior Province as part of Canada's LITHOPROBE program. Calvert and Ludden (1999) reviewed data for the eastern Superior Province. There are three salient features. Shallow dipping crustal reflectors that coincide with surface faults are interpreted as craton scale shortening by thrust tectonics (Fig. 2). Geophysical evidence is supported by geological relationships. For example, the entire Abitibi terrane is thrust over the Pontiac terrane, as indicated by inliers of Pontiac in the Abitibi (Feng and Kerrich, 1992; Feng et al., 1993; Benn, 2006). Second, shallow dipping reflectors extend into the lithospheric mantle. Third, a high velocity zone, likely a detached slab, is "stalled" beneath one of the major sutures (Fig. 2). Alternatively, the regional scale fold structures in the Abitibi belt have recently been interpreted to have formed as detachment folds in a single tectonic event (Benn and Peschler, 2005).

In the western Superior Province, shallow dipping reflectors coincident with the Winnipeg River-North Caribou micro-continents are interpreted as their accretion above a N-dipping subduction zone. The Uchi greenstone belt was preserved by lithospheric extension involving crustal collapse and upwarping of the MOHO (Fig. 8; Calvert et al., 2004; Musacchio et al., 2004). Preservation of shallow dipping reflectors in this Province rules out the Archean vertical tectonic model of Hamilton (1998).

In northern Canada, veibroseis transects reveal a lower crustal seismic structure consistent with shallow imbrication of the Slave Province at ~2.65 Ga, and a paleo-subduction zone between the 1.86–1.84 Ga Great Bear magmatic arc and the Phanerozoic Cordilleran orogen to the west (Cook et al., 1999).

3. Conclusions

Subsolidus convection in the present-day mantle is the main driving force of modern plate tectonics and associated geological processes operating at plate boundaries, including magmatism, deformation, seismicity, and metamorphism (Tackley, 2000; Condie, 2005b). Mantle convection and plate tectonics are one system, because oceanic plates are cold upper thermal boundary layers of the convection cells (Tackley, 2000). Accordingly, vigorously convecting mantle in the hotter Archean Earth is expected to have resulted in the operation of plate tectonics. Both geological relationships and numerical models have been used to argue for or against operation of plate tectonics in the Archean. A recently developed numerical mantle convection model, using the present-day thermal conditions as starting point, to characterize the thermal history of the Earth's mantle predicts slowly moving plates in the Archean and operation of subduction zone geodynamic processes since 2.7 Ga (Korenaga, 2006). The numerical model of Abbott et al. (1994) for the thermal evolution of the mantle predicts flat to shallow subduction between 4.0 and 2.5 Ga, and well-established steep subduction at 2.5 Ga. Recent field and geochemical data for 3.8 to 2.5 Ga Archean terranes show many commonalities with Phanerozoic convergent margins, albeit with more intense plume activity, suggesting that the numerical model of Abbott et al. (1994) is consistent with numerous geological, geochemical, and seismic relationships for Archean terranes as set out above. Butler and Peltier's (2002) numerical model is also in accord with several sets of observations for these terranes. Thermochemical mantle convection models involving resurfacing of van Thienen et al. (2004a,b), which predict no plate tectonics in the Archean, do not account for the geological record.

Archean TTG were emplaced pre- to post-volcanic stages in greenstone belts (Condie, 2005a; Martin et al., 2005; and references therein). Although the thermochemical models of van Thienen et al. (2004b) can account for the emplacement of syn- to post-volcanic TTG into greenstone belts, it cannot readily explain the formation of pre-volcanic TTG gneisses unconformably underlying greenstone sequences, because, according to the models, mafic volcanism should predate the TTG intrusion. There are several compelling lines of evidence in the geological record indicating that some greenstone belts were deposited as autochthonous sequences (CFB) on older continental basement (see Arndt et al., 2001; Blake et al., 2004; Hartlaub et al., 2004; and references therein). On the other hand, the assumptions made for the van Thienen models are consistent with the production of thicker oceanic crust, hence resulting in shallow subduction in the Archean.

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