Available online at www.sciencedirect.com
SCIENCE DIRECT®
Lithos 71 (2003) 575–589
LITHOS
www.elsevier.com/locate/lithos

Lithosphere development in the Slave craton: a linked crustal and mantle perspective

W.J. Davis[a,*], A.G. Jones[a], W. Bleeker[a], H. Grütter[b]

[a] *Geological Survey of Canada, 601 Booth St., Ottawa, Ontario, Canada K1A 0E8*
[b] *Mineral Services Canada, #1300-409 Granville Street, Vancouver, British Columbia, Canada V6C 1T2*

Abstract

The late tectonic evolution of the Slave craton involves extensive magmatism, deformation, and high temperature-low pressure (HT-LP) metamorphism. We argue that the nature of these tectonic events is difficult to reconcile with early, pre-2.7 Ga development and preservation of a thick tectosphere, and suggest that crust–mantle coupling and stabilization occurred only late in the orogenic development of the craton. The extent and repetitiveness of the tectonic reworking documented within the Mesoarchean basement complex of the western Slave, together with the development of large-volume, extensional mafic magmatism at 2.7 Ga within the basement complex argue against preservation of a widespread, thick, cool Mesoarchean tectosphere beneath the western Slave craton prior to Neoarchean tectonism. Broad-scale geological and geophysical features of the Slave craton, including orientation of an early F1 fold belt, distribution of ca. 2.63–2.62 Ga plutonic rocks, and the distribution of geochemical, petrological and geophysical domains within the mantle lithosphere collectively highlight the importance of an NE–SW structural grain to the craton. These trends are oblique to the earlier, ca. 2.7 Ga north–south trending boundary between Mesoarchean and Neoarchean crustal domains, and are interpreted to represent a younger structural feature imposed during northwest or southeast-vergent tectonism at ca. 2.64–2.61 Ga. Extensive plutonism, in part mantle-derived, crustal melting and associated HT-LP metamorphism argue for widespread mantle heat input to the crust, a feature most consistent with thin (< 100 km) lithosphere at that time. We propose that the mantle lithosphere developed by tectonic imbrication of one or more slabs subducted beneath the craton at the time of development of the D1 structural grain, producing the early 2.63–2.62 Ga arc-like plutonic rocks. Subsequent collision (external to the present craton boundaries) possibly accompanied by partial delamination of some of the underthrust lithosphere, produced widespread deformation (D2) and granite plutonism throughout the province at 2.6–2.58 Ga. An implication of this model is that diamond formation in the Slave should be Neoarchean in age.

Keywords: Archean; Slave province

1. Introduction

One of the defining features of Archean cratons is the presence of a thick (>150 km) lithospheric mantle keel, termed tectosphere by Jordan (1988), characterized by high P-wave velocities, low geothermal gradients and chemically depleted compositions. Debate

* Corresponding author. Tel.: +1-613-943-8780; fax: +1-613-995-7997.

E-mail address: bidavis@nrcan.gc.ca (W.J. Davis).

0024-4937/$ - see front matter © 2003 Elsevier B.V. All rights reserved.
doi:10.1016/S0024-4937(03)00131-2

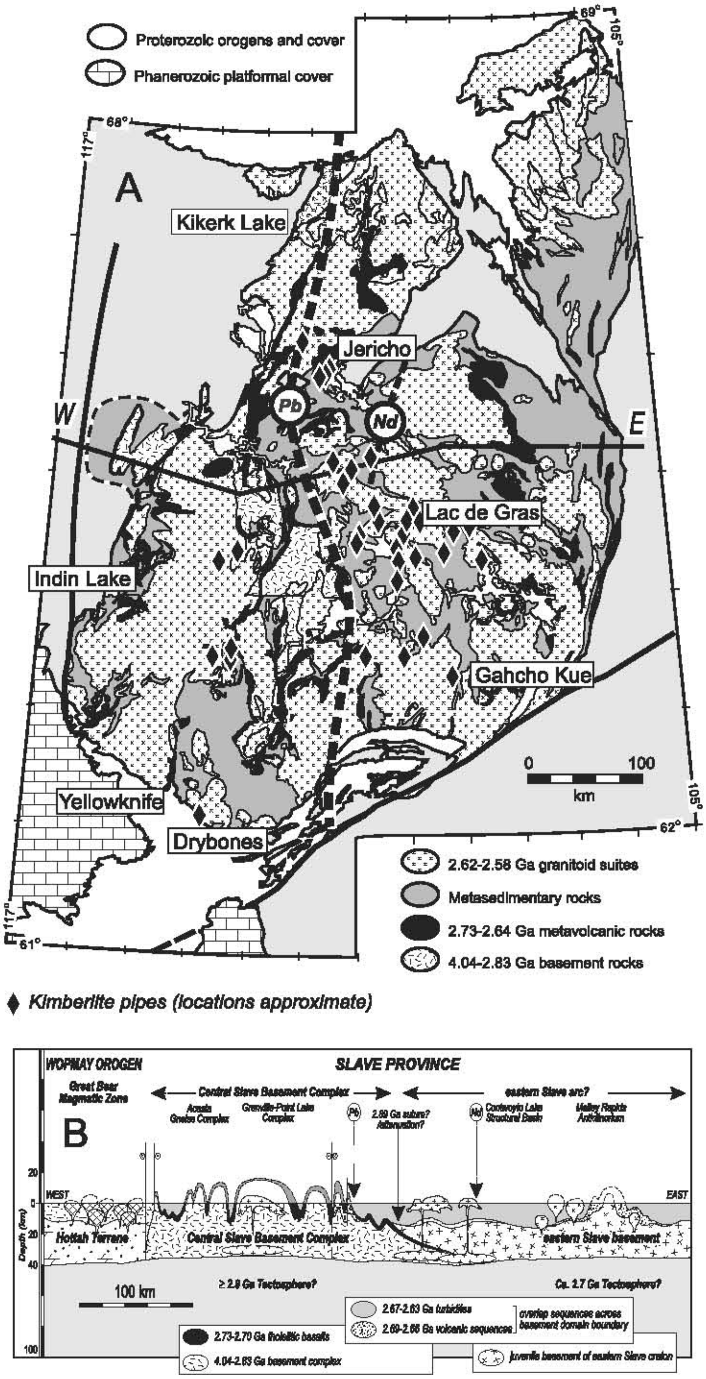

Fig. 1. (A) Geological map of the Slave craton showing distribution of Mesoarchean basement and isotopic boundaries defined by Pb in VMS deposits (Thorpe et al., 1992) and Nd in granites (Davis and Hegner, 1992). (B) E – W cross section of the central Slave craton, illustrating east-dipping boundary between Mesoarchean crustal block in west and Neoarchean crustal domain in east (Bleeker and Davis, 1999).

continues about the genesis of these keels, and models include repeated cycles of differentiation and collisional thickening (Jordan, 1975, 1988), collision of island arcs comprising depleted material (Ashwal and Burke, 1989), buoyant subduction and imbrication by lithospheric-scale stacks (Helmstaedt and Schulze, 1989), and basal accretion by cooling asthenospheric material (Thompson et al., 1996).

Equally important, however, and arguably less well understood, is the genetic relationship between these thick, depleted lithospheric Archean keels and their overlying crustal sections. Re–Os isotopic studies of xenolith samples from different Archean cratons indicate that significant portions of the tectosphere were initially depleted synchronously with, or within a short period following, formation of the overlying crustal section (Pearson, 1999). The broad similarity in timing of crust formation and mantle depletion is interpreted to indicate either (1) a temporal and genetic link and significant coupling between crust and subcontinental lithospheric mantle formation, or (2) that the Archean crust is preserved as a consequence of protection by deep lithospheric keels, which may be coupled to the crust somewhat later than the age of crust formation (e.g., Pearson, 1999; Moser et al., 2001). As it is often challenging to resolve lithospheric age differences at time scales of less than 200 my using Re–Os model age or isochron methods, establishing the direct temporal relationship between the crust and mantle at time scales appropriate to the cycle of orogenic processes is problematical. Therefore, it remains difficult to differentiate between these two competing possibilities.

Over the past decade the Slave craton, in northwestern Laurentia, has emerged as a major diamond producing province (Fipke et al., 1995; Rylatt and Popplewell, 1999). The extensive and well-documented geological record of the Slave craton (Fig. 1; Padgham, 1992; Isachsen and Bowring, 1994; Bleeker and Davis, 1999, and references therein) provides an important new crustal, as well as emerging mantle perspective (Grütter et al., 1999; Griffin et al., 1999; Bank et al., 2000; Kopylova and Russell, 2000; Carbno and Canil, 2002) on the development of diamond-bearing tectosphere. The late tectonic evolution of Archean cratons, such as the Slave, is complex and involves extensive rifting, magmatism, compressional deformation, and metamorphism that in many cases significantly post-dates the timing of initial crust formation by 10 to >100 my. The Slave's Neoarchean orogenesis is characterized by high temperature-low pressure metamorphic conditions (HT-LP) and the intrusion of voluminous granitoid plutons within a short time interval (Fyson and Helmstaedt, 1988; Thompson, 1989; van Breemen et al., 1992). In modern tectonic settings, the association of HT-LP metamorphism with compressional regimes is generally thought to require additions of mantle-derived heat to the crust, either directly through intrusion of mantle melts, or by delamination or lithospheric thinning processes (e.g., Midgley and Blundell, 1997). This implies at least partial removal of pre-existing mantle lithosphere, with the total replacement of the mantle section in extreme cases.

Such a tectonic style is difficult to reconcile with the notion of a relatively cool, thick mantle tectosphere coupled to the crust beneath the Slave craton throughout its Neoarchean evolution. Thus, the crustal perspective on tectosphere development and stabilization presents a fundamental paradox: Can extensive plutonism, including mantle-derived magmatism, and HT-LP metamorphism characteristic of the Slave craton and many other Neoarchean terrains develop above previously stabilized, thick tectosphere? This question is particularly relevant to understanding the development of the Slave craton, as initial Re–Os studies of xenoliths from kimberlites suggest that at least parts of the Slave mantle lithosphere may be Mesoarchean in age down to a considerable thickness and remained coupled with the overlying crust throughout the extensive tectonic reworking in the Neoarchean (Aulbach et al., 2001).

In this paper we discuss critical petrological, geophysical and geochemical observations and first-order geological observations that are relevant to this debate. We conclude that these observations can be best explained if thick tectosphere developed only relatively late during collisional orogenesis, most likely by tectonic imbrication (e.g., Helmstaedt and Schulze, 1989).

2. Geological background

The Slave is a small craton, ~ 700 × 500 km in exposed areal extent, bounded by Paleoproterozic belts to the south, east and west and covered by younger rocks to the north (Padgham, 1992; Isachsen and Bowring, 1994; Bleeker and Davis, 1999). The craton is characterized throughout its western part by a Mesoarchean basement (4.0–2.9 Ga), referred to as the Central Slave Basement Complex (Bleeker et al., 1999b), with isotopically juvenile (< 2.85 Ga?) but undefined basement in the east (Fig. 1A; Thorpe et al., 1992; Davis and Hegner, 1992; Davis et al., 1996). Isotopic data from granites and lower crustal xenoliths suggest that the Mesoarchaen basement dips to the east and underlies the central part of the craton at depth, although its eastern extent remains undefined (Davis et al., 1996, 2003; Davis and Hegner, 1992; Fig. 1B).

This east–west asymmetry has received considerable attention in tectonic models for the Slave's cratonic development. In part, it forms the basis for arc-continent collisional models of Kusky (1989) and Davis and Hegner (1992). The detailed structural and stratigraphic data to support these generalized models are lean, with the dominant structures being considerably younger and affecting equally the eastern and western parts of the craton (e.g., Fyson and Helmstaedt, 1988; Padgam, 1992; Padgham and Fyson, 1992; Isachsen and Bowring, 1994; Bleeker et al., 1999a; Bleeker, 2001). The origin of the asymmetry in crustal age domains remains uncertain. A collisional suture remains a possibility but such a structure must be early and predate 2.69 Ga (Bleeker et al., 1999a). Alternatively, the eastern Slave may represent highly attenuated and modified Mesoarchean lithosphere that developed during rifting at ca. 2.85–2.70 Ga (Bleeker, 2003). If one assumes that some thickness of mantle lithosphere was coupled to the isotopically distinct crustal domains, then mantle lithosphere under the western Slave could be significantly older, perhaps by up to 400 my, than that underlying the eastern Slave, regardless of the exact relationship between the domains (Grütter et al., 2000).

The composite basement preserves a complex polymetamorphic and magmatic history with at least 10 distinct magmatic and/or metamorphic "events" between 4.0 and 2.85 Ga (Isachsen and Bowring, 1994; Bowring and Williams, 1999; Bleeker and Davis, 1999; Ketchum and Bleeker, 2001). The extent and repetitiveness of this tectonic reworking on a ca. 100 Ma interval is uncharacteristic of the stability generally attributed to cratons underlain and protected by thick lithosphere. Development of a thin cover sequence consisting of fuchsitic quartzite and banded iron formation on the basement at 2850–2800 Ma marks the first indication of widespread, but transient stability within the basement (Bleeker et al., 1999b; Sircombe et al., 2001).

Thick, tholeiitic submarine volcanic sequences were extruded over the quartzites and Central Slave Basement Complex between 2.73 and 2.70 Ga, with no correlative volcanic sequences as yet documented in the eastern Slave (Padgham, 1992; van Breemen et al., 1992; Isachsen and Bowring, 1994; Bleeker et al., 2001). Mafic magmatic rocks cover an area of at least 100,000 km^2 with a typical thickness of 1–6 km, approaching proportions comparable to modern large igneous provinces (LIPs; Eldholm and Coffin, 2000). Such voluminous magmatism suggests it may be associated with large-scale mantle plume or mantle overturn events (Bleeker et al., 2001). Granitoids of similar age occur within the basement as a result of localized crustal melting.

Widespread calc-alkaline volcanism followed between 2.70 and 2.66 Ga in both the eastern and western Slave (van Breemen et al., 1992), and was terminated by deposition of thick turbidite sequences over the entire exposed craton at 2.66–2.63 Ga (Bleeker and Villeneuve, 1995; Pehrsson and Villeneuve, 1999) The post-2.69 Ga volcanic rocks represent the first sequence that can be correlated across the entire exposed craton, and provide the earliest evidence of linkage between the eastern and western Slave domains (Bleeker, 2001).

The dominant tectono-metamorphic structures recorded in exposed crustal rocks developed between 2.64 and 2.58 Ga, 20–80 my after deposition of the principal volcanic sequences, and at least several 100 my after development of the Mesoarchean Central Slave Basement Complex. Post-2.64 Ga structures are dominated by at least three regional folding events at shallow to mid-crustal levels (D1, D2, D3), accompanied by a systematic temporal variation in the composition of associated plutonic rocks (Relf, 1992; van Breemen et al., 1992; Davis and Bleeker, 1999; Pehrsson et al., 2000). The deformation events record large horizontal shortening and show little or no apparent spatial correlation with the location of known or inferred Mesoarchean basement. Pehrsson

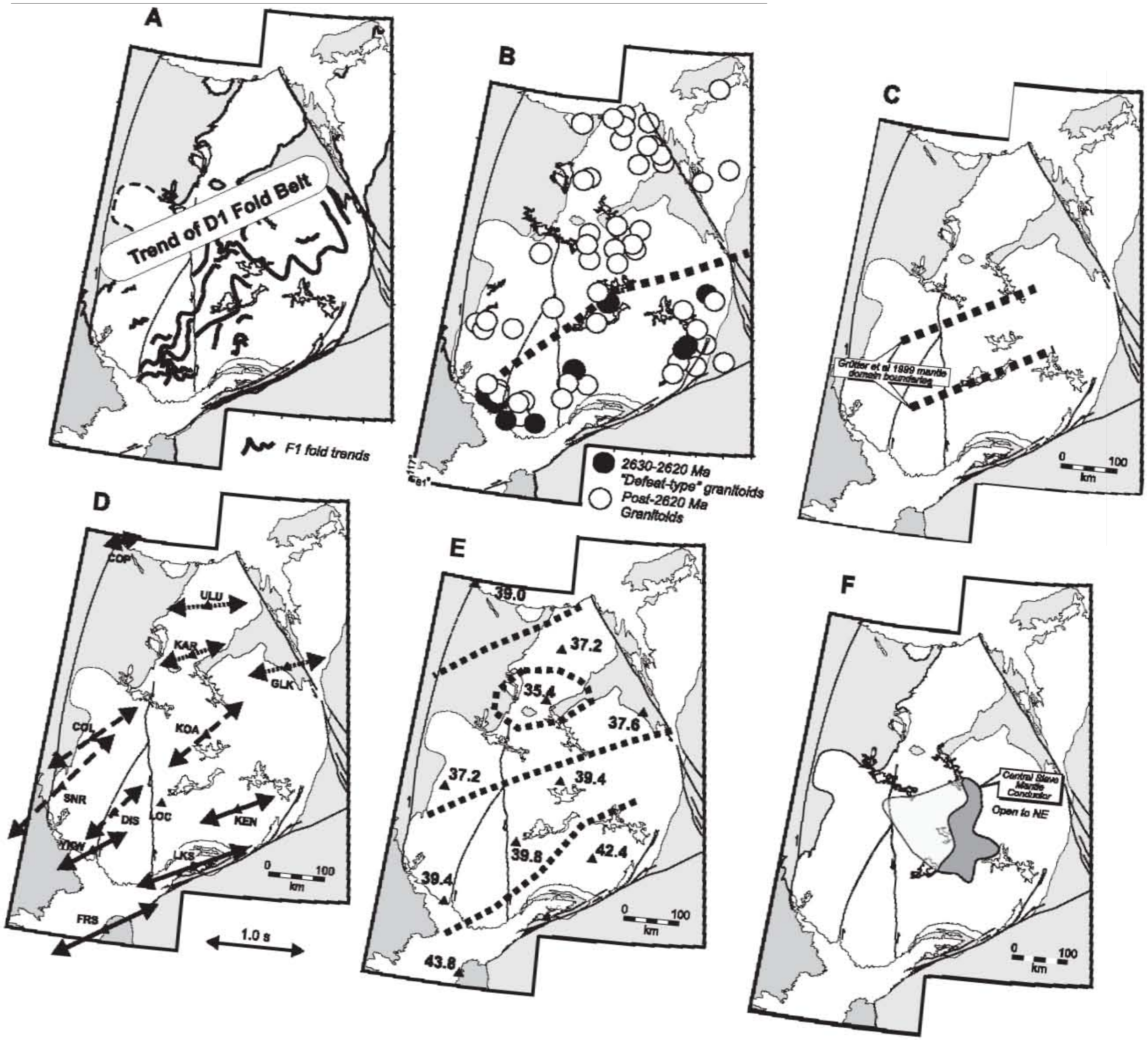

Fig. 2. Location and orientation of a number of geological, geochemical and geophysical characteristics of the Slave craton. (A) Inferred trend lines of the F1 fold belt (Bleeker et al., 1999b, 2001). (B) Distribution of dated plutons within the craton (open circles) with dated plutons between 2620 and 2635 Ma represented by filled circles documented only in the south and southeastern part of the craton (Davis and Bleeker, 1999). (C) Trends of geochemical mantle domains based on garnet chemistry (Grütter et al., 1999). (D) Summary of teleseismic anisotropy data (from Bank et al., 2000). (E) Crustal thickness estimates from seismic data (Bank et al., 2000). (F) Location and extent of mantle conductor in central Slave (Jones et al., 2001).

et al. (2000) suggest that widespread, medium-pressure granulite-facies rocks in the western Slave may be preferentially exposed owing to the presence of basement rocks in that area.

Folding cannot be related to events internal to the craton, such as previously inferred in arc/microcontinent collision models (e.g., Kusky, 1989), and is interpreted to reflect tectonic forces that originated outside the preserved area of the craton.

The orientation of D1 fold structures in the central and southern Slave province define an approximately NE–SW trending fold belt, after taking into account the effects of later D2 folding (Fig. 2A, Bleeker et al., 1999b). Padgham (1985, 1992) previously highlighted NE–SW trending zones within the craton. The orientation of the inferred fold belt is at relatively high angle to the inferred N–S trending boundary between contrasting basement domains (Bleeker et al., 1999b). The timing of D1 shortening is constrained in the Yellowknife area to pre-date intrusion of ca. 2.63 Ga diorite to granodiorite plutons of the Defeat plutonic Suite (Davis and Bleeker, 1999). In the north and central Slave, a minimum age for this event is only loosely bracketed to be older than ca. 2.615–2.608 Ga (e.g., Relf, 1992; van Breemen et al., 1992). The D1 event is established to be diachronous as sedimentary rocks in the Indin Lake area in the westernmost Slave craton were deposited after initiation of D1 folding in the Yellowknife area (Davis and Bleeker, 1999; Pehrsson and Villeneuve, 1999).

In the southeastern Slave, the post-D1 plutonism is characterized by diorite–granodiorite compositions (van Breemen et al., 1992; Davis and Bleeker, 1999). This plutonism is regionally diachronous, with >2.62 Ga plutonic rocks occurring in the south and southeastern parts of the craton, roughly paralleling the trend of the D1 fold belt, and younger, 2.62–2.60 Ga plutons to the north and northwest (Davis and Bleeker, 1999, Fig. 2B). Although the tectonic cause of this event remains uncertain the most primitive, gabbro to diorite compositions require a subduction-enriched mantle component, and thus a melting event in the mantle beneath the Slave craton at ca. 2.630–2.605 Ga (Davis et al., 1994; Yamashita et al., 1999). Geochemical signatures of these plutons are consistent with a 'subduction-modified' mantle source (Davis et al., 1994; Yamashita et al., 1999). Griffin et al. (1999) proposed a plume model to drive this event; however, the temporal and spatial relationships between regional deformation and plutonism are consistent with a subduction/collisional origin. The early 2.63–2.62 Ga plutons have compositional characteristics of arc-related plutons (Yamashita et al., 1999) and these are followed by intrusion of ca. 2.61 Ga diorites in the central and northern Slave with LREE-enriched high-Mg andesite compositions commonly found in arc or post-collisional settings, and interpreted to be related to lithospheric delamination (Davis et al., 1994; Sajona et al., 2000).

Major regional shortening continued through the interval 2610–2585 Ma and was accompanied by voluminous two-mica and K-feldspar granite plutonism throughout the craton (van Breemen et al., 1992; Davis and Bleeker, 1999). The D2 structures indicate east–west shortening, suggesting a change in the orientation of the principal shortening direction or an oblique geometry (Bleeker and Beaumont-Smith, 1995). Although spanning 20 my, the granite plutonism shows no resolvable regional diachroneity, regardless of the timing of the earlier ca. 2605–2630 Ma plutonism (van Breemen et al., 1992; Davis and Bleeker, 1999). Furthermore, the distribution of these younger granites shows no relationship to the distribution of basement domains, although the two-mica granites are certainly associated with areas of thickened sedimentary sequences. This intense craton-wide "granite bloom" argues for a widespread thermal disturbance, the exact cause of which remains speculative. Various models have been suggested for this event, including lithospheric delamination (Davis et al., 1994), post-collisional extension (Kusky, 1993), interaction with a mantle plume (Griffin et al., 1999) and crustal thickening of thinned, warm lithosphere (Thompson, 1989). These models predict a relatively thin (< diamond stability window) mantle lithosphere beneath the craton at 2.6 Ga.

3. Geophysical and geochemical mantle domains

As discussed above, prior deliberation of the Slave's tectonic history has been dominated by the obvious east–west disparity in exposed bedrock geology. However, we contend that this geometry is only a feature of the Slave's crust, and that its subcontinental lithospheric mantle exhibits a NW–SE mantle zonation comprising three regions with distinctive geochemical and geophysical characteristics.

3.1. Geochemical boundaries

The abundance, distribution and "stratigraphy" of lithologies within subcontinental mantle lithosphere can be constrained in space and time by detailed geochemical investigation of mantle-derived xenoliths and xenocrysts (e.g., O'Reilly and Griffin, 1996). Mantle lithologies are commonly defined with reference to garnet compositions because garnet shows extensive solid solution and is a stable mineral in a large variety of lithospheric bulk compositions at pressures exceeding 1.6–2.0 GPa (Boyd, 1970; Sobolev, 1977). Griffin et al. (1999) utilized minor and trace element compositions of Cr-pyrope garnet to identify and describe a unique ultradepleted layer (henceforth UDL) dominated by clinopyroxene-free, garnet harzburgite that underlies the shallow mantle lithosphere in the central Slave craton. This UDL occurs at mantle temperatures less than ~ 950 °C and is replaced by moderately depleted lherzolite-dominated lithologies at temperatures of ~ 950 to ~ 1200 °C. Xenolith thermobarometry constrains the base of the UDL at ~ 140 km depth and shows that the moderately depleted central Slave lithosphere extends to a depth of ~ 200 km (Pearson et al., 1999). The UDL contains Cr-pyrope garnets with distinctively low Cr_2O_3 subcalcic major element compositions (the G10-1 population of Grütter and Anckar, 2001) that are known to occur with regularity in kimberlites and till samples within a ~ 140 km wide and ~ 220 km long east–northeast trending zone in the central Slave craton (Fig. 2C). Similar low Cr_2O_3 subcalcic garnet compositions are extremely rare in kimberlite or till samples outside this zone (Grütter et al., 1999), indicating that the UDL occurs as a distinct east–northeast trending unit at shallow depth within the central Slave craton and that the stratigraphic relations and mutual proportions of garnet-bearing mantle lithologies below the crust of the northern and southern Slave craton differ from that in the central Slave craton (see also Kopylova and Caro, 2001). Carbno and Canil (2002) suggest that the ultradepleted layer may extend to the southeastern Slave (Drybones area) but the deeper lithosphere is of different composition than in the east. This may reflect modification during Paleoproterozoic craton margin events (Carbno and Canil, 2002).

Mantle xenoliths from the diamondiferous Jericho kimberlite in the northern Slave craton show that garnet-bearing mantle lithosphere occurs within a depth range of ~ 80 to ~ 200 km and that eclogitic and pyroxenitic lithologies are comparatively common within a lherzolite-dominated lithospheric section (Kopylova et al., 1998). A relatively limited number of garnet xenocryst populations have been described from the northern Slave craton, but those that are available suggest the lithospheric section may contain an above-average proportion of low-Cr_2O_3 eclogite and that G10-bearing garnet harzburgite is very rare (e.g., Fig. 2(F) of Grütter et al., 1999). G10 garnets are also not described as a prominent xenocryst component in several recently discovered diamondiferous kimberlites within the Coronation district in the far northwestern Slave craton (data in Armstrong, 2002, but also based on an informal survey of press releases of various diamond exploration companies).

The southern Slave craton contains a number of ~ 530 Ma old diamondiferous kimberlites that have sampled garnet-facies mantle to extreme depths of ~ 250 km (Kopylova and Caro, 2001; McLean et al., 2001). Garnet xenocryst assemblages described from the Snap Lake (McLean et al., 2001), CL-25 (Pokhilenko et al., 1997), MZ dyke (Mountain Province Diamonds, 2001) and Gahcho Kue kimberlites (Grütter et al., 2000) document a lherzolite-dominated lithospheric section with subordinate eclogite and occasional G10 garnets with moderate-Cr_2O_3 which are different in composition to G10 garnets in the UDL. A compositionally distinct high-Cr_2O_3, moderate-CaO subcalcic garnet xenocryst population occurs with low frequency within these kimberlites (Grütter et al., 1999). Essentially identical garnet compositions are now also recognized as a low-abundance component derived from extreme lithospheric depths below the central Slave craton (the G10-3 population of Grütter and Anckar, 2001). These compositional and depth attributes indicate that the known lithospheric section of the southern Slave craton (east of longitude 111° W) is dissimilar to that of the central Slave craton at typical UDL depths, but that a mutually common high-Cr_2O_3 garnet harzburgite component exists at extreme depth. Hence, a combination of three different lithospheric sections is required to describe the geochemical features of the northern, central and southern Slave mantle. A schematic cross section of the geochemical architecture (Fig. 3) requires a threefold division at UDL depths, but shows a similar G10-3 component between the central and southern Slave mantles at extreme depths within the lithospheric keel.

3.2. Teleseismic SKS splitting observations

Determination of shear wave splitting (SKS) directions for stations on the Slave craton (Table 1) by Bank et al. (2000) were interpreted to show relatively uniform characteristics (Fig. 2D) similar in orientation to the North American plate vector motion. In particular, the northern two stations, COP and ULU, were considered to exhibit no evidence of deviation from other values on the Slave craton, and this was taken as lack of evidence for any MacKenzie plume modification of the underlying lithospheric mantle as suggested by Ernst and Baragar (1992).

Using statistics appropriate for directional data (Mardia, 1972), and taking the 90° ambiguity into account, the weighted average of the SKS directions for the northern two stations (14 data) is 074° with a standard error of 1.25°, and for all 13 Slave stations (84 data) is 055° ± 3.52°. The *t*-value to test whether the difference of these means is significant is 19.91, which indicates that the null hypothesis that these means are the same can be rejected. Similarly, the time delays show a statistically significant difference, with the two northernmost stations giving a weighted

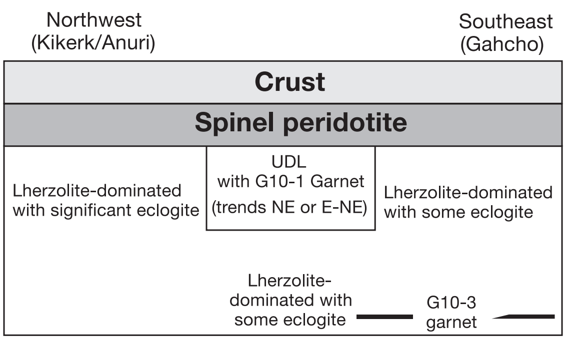

Fig. 3. Inferred geochemical architecture of the Slave craton lithosphere summarized in NW–SE schematic cross section through the central Slave province based on garnet and xenolith data referenced in text.

Table 1
SKS directions and time delays for Slave sites (taken from Bank et al., 2000) and statistical analyses

Site	No.	Phi	sd	Av	sd	dt	sd	Av	sd
COP	10	70	7	71	1.2	0.4	0.2	0.68	0.17
ULU	4	80	10			0.8	0.1		
KAR	3	66	9			0.7	0.4		
GLK	3	71	9			0.8	0.5		
KOA	8	43	9	42	0.8	1.0	0.2	1.09	0.19
COL	9	50	9			1.1	0.3		
SNR	3	40	2			1.5	0.6		
DIS	3	41	11			0.8	1.3		
KEN	3	65	10	62	0.9	0.8	0.5	1.01	0.19
YKW	24	56	10			0.8	0.3		
LKS	4	65	9			1.2	0.4		
FRS	5	59	12			1.2	0.3		
FPR	5	64	7			1.1	0.2		

average of 0.67 ± 0.2 s compared to the total Slave average of 0.90 ± 0.26 s, giving a *t*-value of 3.18 which rejects the null hypothesis at below the 0.5% level.

Closer inspection of the SKS azimuths (Table 1) shows a statistically significant three-part subdivision of the Slave SKS results into northern sites (COP, ULU, KAR, GLK), central sites (KOA, COL, SNR, DIS) and southern sites (KEN, YKW, LKS, FRS, FPR). The weighted azimuthal averages, and their estimated standard errors, are listed in Table 1. The time delays also show a similar subdivision, with the northern sites statistically different from the central and southern sites. The *t*-value for the northern and central groups is 7.41, which for 43 degrees of freedom is larger than the 0.1% *t*-distribution value of 3.55 and implies that the null hypothesis can be rejected with high confidence.

3.3. Crustal thickness

Crustal thickness was estimated by Bank et al. (2000) using receiver functions, and the estimated Moho depths are shown in Fig. 2E. There is a distinct NE–SW striking variation of crustal thickness through the Slave craton. The northwestern part of the exposed craton has crustal thickness of 37.3 ± 0.2 km (ignoring the anomalously low value for station KAR). The central Slave craton has crustal thicknesses of 39.5 ± 0.2 km, and the SE part of the craton has a crustal thickness in excess of 42 km. The thickest part of the craton occurs in the area of the early 2.63–2.62 Ga plutonic belt.

3.4. Electromagnetic anomaly

The mapped location of the central Slave mantle conductor (Jones et al., 2001, 2003) is shown in Fig. 2F, together with its inferred extension to the west to account for the high magnetotelluric phases observed there (Jones et al., 2003). The central Slave mantle conductor lies almost wholly within the NE-trending geochemical boundaries identified by Grütter et al. (1999) as shown in Fig. 2C. Although the cause of the observed enhanced electrical conductivity is unknown, the spatial association of the anomaly with Griffin et al.'s (1999) ultradepleted harzburgitic layer and with Grütter's mantle domain boundaries suggests an ancient origin, not one associated with the Eocene kimberlite emplacement event. Based on existing knowledge, Jones et al. (2003) interpret the central Slave mantle conductor as due to carbon in graphite form above the diamond stability field.

4. Discussion

The Slave craton has a well-documented crustal history from 4.0 to 2.6 Ga (Padgham, 1992; Isachsen and Bowring, 1994; Bleeker and Davis, 1999) but it is uncertain how persistent lithospheric mantle was during this interval. Did relatively thick mantle lithosphere stabilize at the same time as the crustal sections during the Mesoarchean, or was early formed mantle lithosphere modified and/or destroyed during the subsequent tectonic events? Based on the distribution of crustal age domains, the former hypothesis would predict older, Mesoarchean mantle depletion ages in the west beneath the Mesoarchean terrain and younger lithosphere in the east beneath the eastern domains (Fig. 1B; Grütter et al., 2000), with the structure within the mantle in part controlled by the distribution of Mesoarchean lithosphere. At present, the extent of Re–Os model age mapping of the lithosphere is insufficient to fully evaluate this possibility. Dominantly Mesoarchean depletion ages are determined beneath the central Slave area (Aulbach et al., 2001). Data from Jericho in the north-central Slave indicate dominantly Neoarchean or younger ages, with few samples having depletion ages >3.0 Ga (Irvine et al., 1999, 2001). This, in combination with the petrological differences described above argues for a lithospheric break or transition between these sites. The orientation of this boundary is not constrained, although it may correspond to NE–SW compositional boundaries shown in Figs. 2C and 3.

Similarly, studies of xenolith and xenocryst suites, along with geophysical imaging document important regional variations in the composition and structure of the Slave lithospheric mantle (Grütter et al., 1999; Griffin et al., 1999; Kopylova and Russell, 2000; Jones et al., 2001, 2003; Kopylova and Caro, 2001; Carbno and Canil, 2002). As described above, and originally proposed by Grütter et al. (1999), the Slave lithosphere can be divided into three approximately E–NE oriented zones, each defined by distinct garnet chemistry (Figs. 2C and 3). Importantly, the orientation of these zones is subparallel to the D1 structural grain of the craton (Fig. 2A, Bleeker et al., 1999a,b), and at high angle to north–south isotopic boundaries mapped in the crust (Fig. 1). Since the present distribution of the mantle domains appears to transect the east–west crustal age asymmetry, it is inferred to be a younger feature that probably developed after ca. 2.7 Ga. This would imply that at least the garnet-facies mantle beneath the craton was established late in its evolution, after the time of initial crust formation.

Absence of a pre-2.7 Ga, thick, buoyant lithosphere would be consistent with the repeated episodes of magmatism and metamorphism within the Central Slave Basement Complex throughout the 3.6–2.85 Ga interval (Isachsen and Bowring, 1994; Bleeker and Davis, 1999; Ketchum and Bleeker, 2001). As noted above, the Central Slave Basement complex does not exhibit the tectonic stability generally associated with continental areas underlain by thick tectosphere (Ketchum and Bleeker, 2001). At least two periods of extensional volcanism developed on the Mesoarchean crust; at 2.85 Ga, and perhaps more significantly at 2.73–2.70 Ga. Interpretation of the ca. 2.73–2.70 Ga tholeiitic volcanism in terms of LIP-scale basaltic volcanism (Bleeker et al., 2001) suggests that pre-existing lithosphere may have been substantially modified and/or thinned by the impinging of upwelling asthenosphere (plume?) during extensional magmatism. If the tholeiites were sourced beneath the Mesoarchean crustal block, then segregation at relatively shallow pressures within spinel facies is implied, consistent with a lithospheric thickness of less than 100 km at 2.7 Ga (e.g., White and McKenzie, 1995). Yamashita et al. (1999) suggested that the Mesoarchean basement terrains in the west-central Slave represent highly dismembered crustal segments with intervening dominantly juvenile ca. 2.70 Ga marginal basins. Their model equally suggests a high degree of lithospheric attenuation at 2.7 Ga, and in such a scenario, preservation of ancient mantle lithosphere is likely to be fragmentary, and relegated to the shallowest, spinel peridotite lithosphere.

The absence of thick lithosphere at ca. 2.7 Ga is consistent with the subsequent metamorphic and magmatic history of the craton. The metamorphic conditions attained at 2.6 Ga are characteristic of HT-LP metamorphic belts, with lower crustal temperatures of >700 °C at 0.9–1.1 GPa (Davis et al., 2003). Based on a conductive model with crustal heat production and metamorphic thermal conditions, Thompson et al. (1996) argued that a thermally stabilized lithosphere beneath the Slave could be no thicker than 100 km at 2.6 Ga, and suggested that the lithosphere grew by accretion of asthenosphere at its base between 2.6 and 1.8 Ga. Their model did not attempt to account for any chemical variation or lateral structure within the lithosphere, as is now indicated by geophysical and geochemical data sets.

Thermal models of shortening and thickening of continental lithosphere indicate that development of HT-LP metamorphism and widespread crustal melting are most sensitive to three parameters: (1) the total radiogenic heat production and its distribution in the crust; (2) the thermal structure of the crust prior to thickening; and (3) the reduced heat flow at the base of the crust (e.g., Midgley and Blundell, 1997). Lithospheric thickness and its control on reduced heat flow to the crust is arguably the most significant parameter in these thermal models and may be essential to generate high temperature conditions in modern orogens (e.g., Midgley and Blundell, 1997). Geologically, this may be the result of lithospheric thinning or delamination events, bringing hot asthenospheric material to shallow depths (e.g., Bird, 1979; Houseman et al., 1981; Nelson, 1992). These models argue against the presence of thick cool lithosphere beneath HT-LP metamorphic belts.

HT-LP metamorphism is by no means unique to a specific time period in Earth's history, but it is particularly common in the Archean (Sandiford, 1989). An important consideration is that Archean crust will have at least twice the heat production (e.g., Pollack, 1997) owing to the greater proportion of radiogenic heat-producing elements in the past, favouring higher metamorphic temperatures and crustal melting during shortening (e.g., McLaren et al., 1999). Certainly, the Slave crust is characterized by generally high heat production, particularly the late granites, although most units, such as the basement and volcanic rocks, are not anomalously rich in heat-producing elements (Thompson et al., 1996; Kopylova et al., 1999). Can greater heat production within the crust permit HT-LP metamorphic belts to develop above areas of thick, cool lithosphere? Although this possibility cannot be eliminated by thermal arguments alone (e.g., McLaren et al., 1999), it is not favoured for the Slave craton because it fails to account for the occurrence of the mantle-derived magmatism between 2.630 and 2.605 Ga (Davis et al., 1994; Yamashita et al., 1999). In many parts of the craton these plutons were intruded prior to, or early during D2 regional shortening and peak metamorphism, and thus argue for a role for transient heating within the crust (King et al., 1992). Certainly, greater crustal heat production would contribute to the observed steep metamorphic field gradients but the sequence of early mantle-derived magmatism followed by dominantly crustal melts argues for a significant mantle component to the heat budget.

4.1. Development of the Slave mantle lithosphere by subcretion

If the HT-LP metamorphism and magmatism at 2.6 Ga reflect a thinner lithosphere and transient heating, then thick, cool lithosphere must have developed sometime after ca. 2.6 Ga (e.g., Isachsen and Bowring, 1994; Thompson et al., 1996). Although admittedly speculative, our preference is for a model in which the mantle lithosphere developed by subcretion during NW, or possibly SE-vergent subduction beneath the Slave craton during D1 shortening and the early 2.63–2.61 Ga plutonism (Fig. 4). This would impart a NE–SW structural grain in the lithosphere during development of the D1 fold belt and early

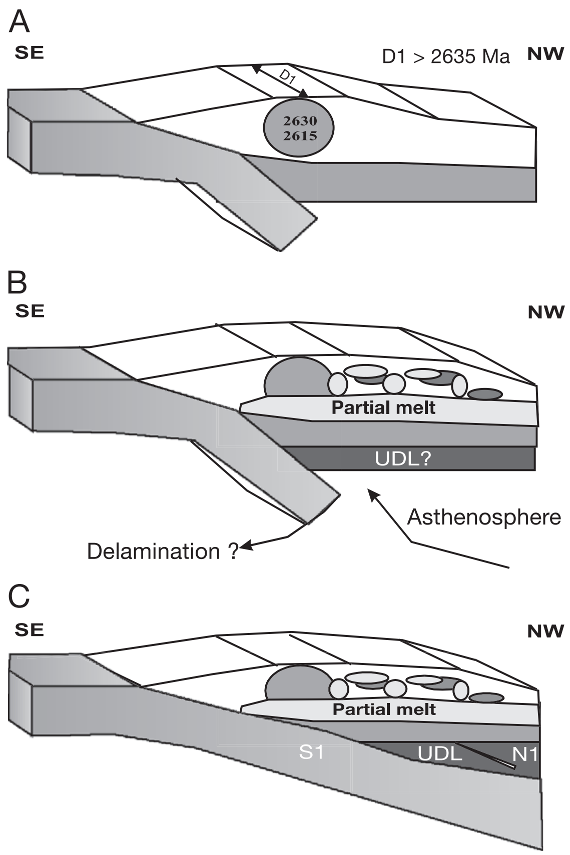

Fig. 4. Cartoon model for development of Slave lithosphere by tectonic imbrication of lithosphere during deformation and plutonism at ca. 2.6 Ga. (A) Subduction beneath the craton at ca. 2.64–2.61 Ga generates early SW–NE trending F1 fold belt and subparallel plutonic belt in SE Slave. Development of mantle domains may have been initially established at this time. (B) Subsequent collision (external to the present craton boundaries) possibly accompanied by partial delamination of some of the underthrust lithosphere produced widespread deformation (D2) and granite plutonism throughout the province at 2.6–2.58 Ga. C) termination of orogenesis and subcretion of deeper mantle lithosphere.

2.63–2.61 Ga plutonism, and would imply decoupling of Mesoarchean–Neoarchean crustal boundaries from the underlying deeper (garnet-bearing) subcontinental lithospheric mantle (Grütter et al., 1999). It is important to note that the Slave craton, as exposed, represents only a small fragment of a presumably much larger craton (Bleeker, 2003; Isachsen and Bowring, 1994), and the location of the preserved craton within a framework of possible Neoarchean plate boundaries is unconstrained. The subcreted mantle component may include oceanic lithosphere or arc-wedge material. The ultradepleted component characteristic of the central part of the Slave lithosphere may represent the latter material, as suggested by Griffin et al. (1999), structurally separated from adjacent zones. Subsequent collision (external to the present craton boundaries) possibly accompanied by partial delamination of some of the underthrust lithosphere, produced widespread deformation (D2) and granite plutonism throughout the province at 2.6–2.58 Ga, with continued metamorphism (extension?) in the lower crust to at least 2.56 Ga (Davis et al., 2003).

A prediction of this model is that the Slave mantle lithosphere was dominantly stabilized in the latest Archaean or younger times. It is however at odds with the documentation of extensive regions of Mesoarchean lithosphere beneath the central Slave to depths of 150–200 km (Aulbach et al., 2001). Certainly, accreted oceanic lithosphere is expected to be somewhat older than the time of its emplacement, perhaps by up to 150 my in modern systems. Significantly older components (i.e., >2750 Ma) could represent older parts of the oceanic lithosphere that were decoupled from their crust and imbricated beneath the craton during collisional events, or perhaps remnants of ancient Slave lithosphere caught up in the subcreted collage. Greater buoyancy of ultradepleted oceanic lithosphere in the Archean may permit longer cycles for recycling of oceanic lithosphere.

Further modification and addition to the mantle is thought to have occurred through imbrication accompanying Proterozoic accretion to the western craton margin (Cook et al., 1999; Bostock, 1997; Carbno and Canil, 2002), which may have disturbed a primary lithospheric architecture of Neoarchean age.

4.2. On the occurrence of Archean diamonds

If a thin lithosphere and elevated reduced heat flow is required to account for the metamorphic and magmatic history of the craton at ca. 2.6 Ga, what does this imply for the age of diamonds? Thompson et al. (1996) argued, on the basis of paleogeotherms and crustal heat production, that diamonds could not be stabilized within the Slave lithosphere until >500 my after the last tectonothermal event to have affected the craton. Although not yet proven to be present in the

Slave, diamonds of Archean age have been identified in other cratons, implying that thermal conditions appropriate for diamond stability were established relatively early, perhaps within 100–200 my of the last major tectono-metamorphic event recorded in the crust (e.g., Richardson et al., 2001). A logical implication of the model presented above is that diamond growth in eclogite and/or peridotite occurred contemporaneously with the subcretion event, or at younger times (e.g., Kesson and Ringwood, 1989a,b). Subcretion of relatively cool mantle will serve to cool the lithospheric section permitting the preservation or growth of diamond. A prediction of the model is that diamonds beneath the Slave craton formed at or after 2.6 Ga, within slightly older mantle lithosphere.

One question that can be posed is whether older diamonds in subcreted lithosphere can survive the thermal pulse from the overlying hot crust. This may be specifically relevant to the case of subcretion or other addition of a significantly older, cold buoyant lithosphere to the Slave after ca. 2.6. Using estimates of the thermal structure of the crust during the ca. 2.6 Ga granite event we have modeled the crust, with an elevated geotherm, being instantaneously underlain by a lithospheric mantle with a conventional cratonic geotherm. The approach used was a standard conductive 1-D solution (Wang, 1999). Fig. 5 shows the initial geotherm, with the base of the 50-km-thick crust at 850 °C juxtaposed against cold mantle at 450 °C, i.e., a 350 °C step, and the relaxation of that geotherm over successive intervals. Also shown on the figure is the experimentally determined graphite-diamond stability field (Kennedy and Kennedy, 1976). Over a relatively short interval, ~ 10 my, the thermal pulse relaxes to the continental geotherm. Note that its effects do not diffuse into the subcreted lithosphere much beyond ~ 75 km depth, and at the graphite-diamond boundary (~ 140 km), there is less than a few degrees increase in temperature.

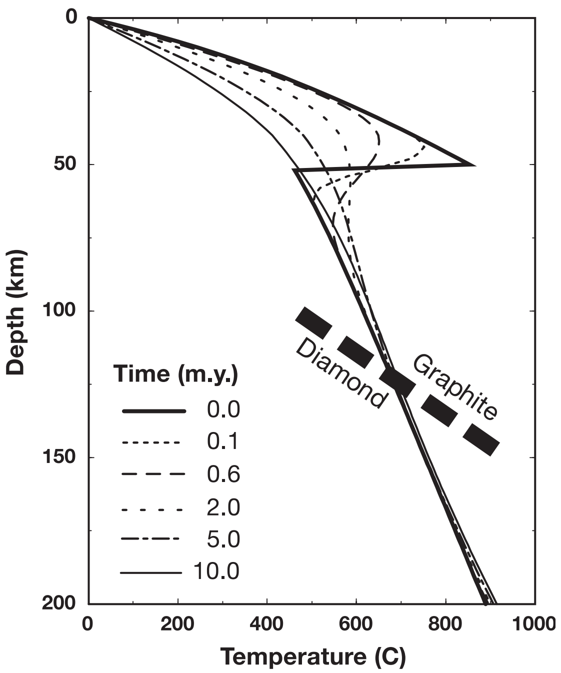

Fig. 5. Conductive thermal relaxation modelling of cold lithosphere, with a cratonic geotherm, subcreted beneath hot crust. The geotherms at time intervals of 0.1, 0.6, 2, 5 and 10 my after subcretion are shown. Also shown is the boundary between stability fields of graphite and diamond (Kennedy and Kennedy, 1976).

5. Conclusions

Broad-scale geological and geophysical features of the Slave craton, including orientation of an early F1 fold belt, distribution of ca. 2.62–2.63 Ga plutonic rocks, and the orientation of geochemical and geophysical domains within the mantle lithosphere collectively highlight the importance of a NE–SW structural grain to the craton. This structural grain is oblique to the north–south crustal age domain boundaries directly mapped by exposures of Mesoarchean crust and indirectly by the isotopic composition of VMS deposits and late granites. We interpret this to indicate that the subcontinental lithospheric mantle architecture post-dates events that lead to the crustal age asymmetry (suture?) as well as the extensive plume or rift-related LIP-type volcanism at 2.7 Ga. The lithosphere developed beneath the craton late in the orogenic cycle, most likely as a result of tectonic imbrication of buoyant lithosphere. An implication of the model is that diamond formation occurred at the earliest in the latest Archean, within only slightly older lithosphere. Improving the resolution of mantle domains and reconciling their age and structural geometry with crustal structures is essential to develop more refined models of tectosphere formation.

Acknowledgements

We wish to acknowledge the logistical and financial assistance of all the many organizations and companies that made our work possible in the Slave craton. These include LITHOPROBE, the Geological Survey of Canada (GSC contribution number 2001219, under the NATMAP, LITHOPROBE, EXTECH-III and Targeted Geoscience Initiative programs), the Department of Indian and Northern Development (DIAND), the U.S. National Science Foundation's Continental Dynamics Program, DeBeers Canada Exploration, Kennecott Exploration and BHP Billiton Diamonds, Diavik, DeBeers, Winspear, Royal Oak Mines and Miramar Mining. Secondly, we wish to acknowledge our many colleagues, within government, academia and industry, for their insightful discussions that prompted us to carefully construe our arguments. Kelin Wang is thanked for providing us with his thermal modelling code, and John Percival for comments on an earlier version of this manuscript. Sam Bowring, an anonymous journal reviewer, and volume editor Ric Carlson, provided useful comments.

References

Armstrong, J.P., 2002. Diamond exploration data in the North Slave Craton, Nunavut. DIAND NU Open Report 2002-01 (CD-ROM).

Ashwal, L.D., Burke, K., 1989. African lithospheric structure, volcanism, and topography. Earth and Planetary Science Letters 96, 8–14.

Aulbach, S., Griffin, W.L., Pearson, N.J., O'Reilly, S.Y., Doyle, B.J., Kivi, K., 2001. Re–Os isotope evidence for Meso-archean mantle beneath 2.7 Ga Contwoyto Terrane, Slave craton, Canada: implications for the tectonic history of the Slave Craton. Abstract. In: Carlson, R., Jones, A.G. (convenors), The Slave Kaapvaal Workshop, Merrickville, Ontario, September 5–9, 2001.

Bank, C.G., Bostock, M.G., Ellis, R.M., Cassidy, J.F., 2000. A reconnaissance teleseismic study of the upper mantle and transition zone beneath the Archean Slave Craton in NW Canada. Tectonophysics 319, 151–166.

Bird, P., 1979. Continental delamination and the Colorado Plateau. Journal of Geophysical Research 84 (B13), 7561–7571.

Bleeker, W., 2001. The ca. 2680 Ma Raquette Lake Formation and correlative units across the Slave Province, Northwest Territories: evidence for a craton-scale overlap sequence. Current Research - Geological Survey of Canada 2001-C7 (26 pp.).

Bleeker, W., 2003. The late Archean record: a puzzle in ca. 35 pieces. Lithos 71, 99–134 (this issue).

Bleeker, W., Beaumont-Smith, C., 1995. Thematic structural studies in the Slave Province, Northwest Territories; the Sleepy Dragon Complex. Current Research - Geological Survey of Canada 1995-C, 87–96.

Bleeker, W., Davis, W.J., 1999. The 1991–1996 NATMAP Slave Province Project: introduction. Canadian Journal of Earth Sciences 36, 1033–1042.

Bleeker, W., Villeneuve, M., 1995. Structural studies along the Slave portion of the SNORCLE Transect. In: Cook, F., Erdmer, P. (compilers), Slave-NORthern Cordillera Lithospheric Evolution (SNORCLE). Report of 1995 Transect Meeting, April 8–9, University of Calgary, 8–14 pp.

Bleeker, W., Ketchum, J.W.F., Davis, W.J., 1999a. The Central Slave Basement Complex: Part II. Age and tectonic significance of high-strain zones along the basement-cover contact. Canadian Journal of Earth Sciences 36, 1111–1130.

Bleeker, W., Ketchum, J.W.F., Jackson, V.A., Villeneuve, M.E., 1999b. The Central Slave Basement Complex: Part I. Its structural topology and autochthonous cover. Canadian Journal of Earth Sciences 36, 1083–1109.

Bleeker, W., Davis, W.J., Ketchum, J.W., Sircombe, K., Stern, R.A., 2001. Tectonic evolution of the Slave craton. In: Cassidy, K.F., et al. (Eds.), 4th International Archaean Symposium 2001, Extended Abstracts. AGSO-Geoscience Australia, Record 2001/37, pp. 288–290.

Bostock, M.E., 1997. Anisotropic upper-mantle stratigraphy and architecture of the Slave Craton. Nature 390, 392–395.

Bowring, S.A., Williams, I.S., 1999. Priscoan (4.00–4.03 Ga) orthogneisses from northwestern Canada. Contributions to Mineralogy and Petrology 134, 3–16.

Boyd, F.R., 1970. Garnet peridotites and the system CaSiO3–MgSiO3–Al2O3. Special Paper - Mineralogical Society of America 3, 63–75.

Carbno, G.B., Canil, D., 2002. Mantle structure beneath the southwest Slave craton, Canada: constraints from garnet geochemistry in the Drybones Bay kimberlite. J. Petrol. 43, 129–142.

Cook, F.A., van, d.V.A.J., Hall, K.W., Roberts, B.J., 1999. Frozen subduction in Canada's Northwest Territories; lithoprobe deep lithospheric reflection profiling of the western Canadian Shield. Tectonics 18, 1–24.

Davis, W.J., Bleeker, W., 1999. Timing of plutonism, deformation, and metamorphism in the Yellowknife domain, Slave province, Canada. Canadian Journal of Earth Sciences 36, 1169–1187.

Davis, W.J., Hegner, E., 1992. Neodymium isotopic evidence for the accretionary development of the Late Archean Slave Province. Contributions to Mineralogy and Petrology 111, 493–503.

Davis, W.J., Fryer, B.J., King, J.E., 1994. Geochemistry and evolution of late Archean plutonism and its significance to the tectonic development of the Slave craton. Precambrian Research 67, 207–241.

Davis, W.J., Gariepy, C., van Breemen, O., 1996. Pb isotopic composition of late Archaean granites and the extent of recycling early Archaean crust in the Slave Province, northwest Canada. Chemical Geology 130, 255–269.

Davis, W.J., Canil, D., Mackenzie, J.M., Carbno, G.B., 2003. Petrology and U–Pb geochronology of lower crustal xenoliths and the development of a craton, Slave Province, Canada. Lithos 71, 541–573 (this issue).

Eldholm, O., Coffin, M.F., 2000. Large igneous provinces and plate tectonics. In: Richards, M.A., Gordon, R.G., van der Hilst, R.D. (Eds.), The History and Dynamics of Global Plate MotionsGeophysical Monograph, vol. 121. American Geophysical Union, Washington, DC, pp. 309–326.

Ernst, R.E., Baragar, W.R.A., 1992. Evidence from magnetic fabric for the flow pattern of magma in the Mackenzie giant radiating dyke swarm. Nature 356, 511–513.

Fipke, C.E., Dummett, H.T., Moore, R.O., Carlson, J.A., Ashley, R.M., Gurney, J.J., Kirkley, M.B., 1995. History of the discovery of diamondiferous kimberlites in the Northwest Territories, Canada. Sixth International Kimberlite Conference, Extended Abstracts. Proceedings of the 6th International Kimberlite, pp. 158–160.

Fyson, W.K., Helmstaedt, H., 1988. Structural patterns and tectonic evolution of supracrustal domains in the Archean Slave Province, Canada. Canadian Journal of Earth Sciences 25, 301–315.

Griffin, W.L., Doyle, B.J., Ryan, C.G., Pearson, N.J., O'Reilly, S.Y., Davies, R., Kivi, K., van Achterbergh, E., Natapov, L.M., 1999. Layered mantle lithosphere in the Lac de Gras area, Slave craton: composition, structure and origin. Journal of Petrology 40, 705–727.

Grütter, H.S., Anckar, E., 2001. Cr–Ca characteristics of mantle pyropes from the central Slave craton—a comparison with the type areas on the central Kaapvaal craton, with implications for carbon in peridotite. Presented at the Slave-Kaapvaal Workshop, Merrickville, Ontario, Canada, September 2001.

Grütter, H.S., Apter, D.B., Kong, J., 1999. Crust–mantle coupling; evidence from mantle-derived xenocrystic garnets. In: Gurney, J.J., et al. (Eds.), Proceedings of the 7th international Kimberlite Conference. J.B. Dawson Volume 1 Red Roof Design, Cape Town, pp. 307–313.

Grütter, H.S., Davis, W.J., Jones, A., 2000. Chemical and physical images of the Slave craton lithosphere. Pan-Lithoprobe 2 Moho Workshop Abstract, Banff, Alberta.

Helmstaedt, H.H., Schulze, D.J., 1989. Southern African kimberlites and their mantle sample: implications for Archean tectonics and lithosphere evolution. In: Ross, J. (Ed.), Kimberlites and Related Rocks: Volume 1. Their Composition, Occurrence, Origin, and Emplacement. Geological Society of Australia Special Publication, vol. 14, pp. 358–368.

Houseman, G.A., McKenzie, D.P., Molnar, P., 1981. Convective instability of a thickened boundary layer and its relevance for the thermal evolution of continental convergent belts. Journal of Geophysical Research 86, 6115–6132.

Irvine, G.J., Kopylova, R.W., Carlson, R.W., Pearson, D.G., Shirey, S.B., 1999. Age of the lithospheric mantle beneath and around the Slave craton: a rhenium–osmium-isotopic study of Peridotite Xenoliths from the Jericho and Somerset Island kimberlites. Ninth Annual V.M. Goldschmidt Conference LPI Contribution No. 971. Lunar and Planetary Institute, Houston, pp. 134–135.

Irvine, G.J., Pearson, D.G., Carlson, R.W., 2001. Lithospheric mantle evolution of the Kaapvaal craton: a Re–Os isotope study of peridotite xenoiths from Lesotho kimberlites. Geophysical Research Letters 28, 2505–2508.

Isachsen, C., Bowring, S.A., 1994. Evolution of the Slave craton. Geology 22, 917–920.

Jones, A.G., Ferguson, I.J., Evans, R., Chave, A.D., McNeice, G.W., 2001. The electric lithosphere of the Slave craton. Geology 29, 423–426.

Jones, A.G., Lezaeta, P., Ferguson, I.J., Chave, A.D., Evans, R.L., Garcia, X., Spratt, J., 2003. The electrical structure of the Slave craton. Lithos 71, 505–527 (this issue).

Jordan, T.H., 1975. The continental tectosphere. Reviews of Geophysics and Space Physics 13, 1–12.

Jordan, T.H., 1988. Structure and formation of the continental tectosphere, in Oceanic and continental lithosphere; similarities and differences. In: Menzies, M.A., Cox, K.G. (Eds.), Journal of Petrology, Special Lithosphere Issue, pp. 11–37.

Kennedy, C.S., Kennedy, G.C., 1976. The equilibrium boundary between graphite and diamond. Journal of Geophysical Research 81, 2467–2470.

Kesson, S.E., Ringwood, A.W., 1989a. Slab–mantle interactions: 1. Sheared and refertilised garnet peridotite xenoliths; samples of Wadati–Benioff zones? Chemical Geology 78, 83–96.

Kesson, S.E., Ringwood, A.W., 1989b. Slab–mantle interactions: 2. The formation of diamonds. Chemical Geology 78, 97–118.

Ketchum, J.W.F., Bleeker, W., 2001. Genesis of the 4.03–2.85 Ga Salve protocraton, northwestern Canada. In: Cassidy, K.F., et al. (Eds.), 4th International Archaean Symposium 2001, Extended Abstracts. AGSO-Geoscience Australia, Record 2001/37, pp. 514–515.

King, J.E., Davis, W.J., Relf, C., 1992. Late Archean tectono-magmatic evolution of the central Slave Province, Northwest Territories. Canadian Journal of Earth Sciences 29, 2156–2170.

Kopylova, M.G., Caro, G., 2001. Lithospheric mapping of the Slave craton: contrasting North and South. Extended Abstract, Slave-Kaapvaal Workshop, 5–9 September, Merrickville, Ontario.

Kopylova, M.G., Russell, J.K., 2000. Chemical stratification of cratonic lithosphere; constraints from the northern Slave Craton, Canada. Earth and Planetary Science Letters 181, 71–87.

Kopylova, M.G., Russel, J.K., Cookenboo, H., 1998. Upper-mantle stratigraphy of the Slave craton, Canada: insights into a new kimberlite province. Geology 26, 315–318.

Kopylova, M.G., Russell, J.K., Cookenboo, H., 1999. Petrology of peridotite and pyroxenite xenoliths from Jericho Kimberlite; implications for the thermal state of the mantle beneath the Slave Craton, northern Canada. Journal of Petrology 40, 79–104.

Kusky, T.M., 1989. Accretion of the Archean Slave Province. Geology 17, 63–67.

Kusky, T.M., 1993. Collapse of Archean orogens and the generation of late- to postkinematic granitoids. Geology 21, 925–928.

Mardia, K.V., 1972. Statistics of Directional Data. Academic Press, London. ISBN 0-12-471150-2.

McLaren, S., Sandiford, M., Hand, M., 1999. High radiogenic heat-producing granites and metamorphism; an example. Geology 27, 679–682.

McLean, R.C., Pokhilenko, N.P., Hall, A.E., Luth, R., 2001. Pyropes and chromites from kimberlites of the Snap Lake area, southeast Slave craton: garnetization reaction of depleted peridotites at extremely deep levels of the lithospheric mantle. Extended Abstract, Slave-Kaapvaal Workshop, 5–9 September, Merrickville, Ontario.

Midgley, J.P., Blundell, D.J., 1997. Deep seismic structure and thermo-mechanical modelling of continental collision zones. Tectonophysics 273, 155–167.

Moser, D.E., Harte, R.J., Flowers, R., 2001. Birth of the Kaapvaal Tectosphere 3.08 Billion Years Ago. Science 291, 465–468.

Mountain Province Diamonds, 2001. De Beers reports on MZ Lake kimberlites: three sills diamondiferous. Public news release dated September 10, 2001.

Nelson, K.D., 1992. Are crustal thickness variations in old mountain belts like the Appalachians a consequence of lithospheric delamination? Geology 20, 498–502.

O'Reilly, S.Y., Griffin, W.L., 1996. 4-D lithosphere mapping; methodology and examples. Tectonophysics 262, 3–18.

Padgham, W.A., 1985. Observations and speculations on supracrustal successions in the Slave Structural Province. In: Ayres, L.D., Thurston, P.C., Card, K.D., Weber, W. (Eds.), Evolution of Archean Sequences. Geological Association of Canada, Special Paper, vol. 28, pp. 133–151.

Padgham, W.A., 1992. The Slave structural Province, North America: a discussion of tectonic models. In: Glover, J.E., Ho, S.E. (Eds.), Proceedings Volume for the Third International Archaean Symposium. Geology Dept. and University Extension, The University of Western Australia, Publication, vol. 22, pp. 381– 394.

Padgham, W.A., Fyson, W.K., 1992. The Slave province: a distinct Archean craton. Canadian Journal of Earth Sciences 29, 2072–2086.

Pearson, D.G., 1999. The age of continental roots. Lithos 48, 171–194.

Pearson, N.J., Griffin, W.L., Doyle, B.J., O'Reilly, S.Y., Van Achterbergh, E., Kivi, K., 1999. Xenoliths from kimberlite pipes of the Lac de Gras area, Slave craton, Canada. In: Gurney, J.J., et al. (Eds.), Proceedings of the 7th International Kimberlite Conference. P.H. Nixon Volume 2 Red Roof Design, Cape Town, pp. 644–658.

Pehrsson, S.J., Villeneuve, M.E., 1999. Deposition and imbrication of a 2670–2629 Ma supracrustal sequence in the Indin Lake area, southwestern Slave Province, Canada. Canadian Journal of Earth Sciences, 1149–1168.

Pehrsson, S.J., Chacko, T., Pilkington, M., Villeneuve, M.E., Bethune, K., 2000. Anton Terrane revisited; late Archean exhumation of a moderate-pressure granulite terrane in the western Slave Province. Geology 28, 1075–1078.

Pokhilenko, N.P., McDonald, J.A., Melnik, U., McCorquodale, J., Reimers, L.F., Sobolev, N.V., 1997. Indicator minerals from the CL-25 kimberlite pipe, Slave craton, Northwest Territories, Canada. Russian Geology and Geophysics 38, 550–558.

Pollack, H.N., 1997. Thermal characteristics of the Archaean Greenstone belts. In: De Wit, M.J., Ashwal, L.D. (Eds.), Greenstone Belts. Oxford Monographs on Geology and Geophysics, vol. 35, pp. 223–232.

Relf, C., 1992. Two distinct shortening events during late Archean orogeny in the west-central Slave Province, Northwest Territories, Canada. Canadian Journal of Earth Sciences 29, 2104–2117.

Richardson, S.H., Shirey, S.B., Harris, J.W., Carlson, R.W., 2001. Archean subduction recorded by Re–Os isotopes in eclogitic sulfide inclusions in Kimberley diamonds. Earth and Planetary Science Letters 191, 257–266.

Rylatt, M.G., Popplewell, G.M., 1999. Ekati diamond mine—background and development. Mining Engineering 51, 37–43.

Sajona, F.G., Maury, R.C., Pubellier, M., Leterrier, J., Bellon, H., Cotton, J., 2000. Magmatic source enrichment by slab-derived melts in a young post-collision setting, central Mindanao (Philippines). Lithos 54, 173–206.

Sandiford, M., 1989. Secular trends in the thermal evolution of metamorphic terrains. Earth and Planetary Science Letters 95, 85–96.

Sircombe, K.N., Bleeker, W., Stern, R.A., 2001. Detrital zircon geochronology and grain-size analysis of a approximately 2800 Ma Mesoarchean proto-cratonic cover succession, Slave Province, Canada. Earth and Planetary Science Letters 189, 207–220.

Sobolev, N.V., 1977. Deep Seated Inclusions in Kimberlites and the Problem of the Composition of the Upper Mantle (English Translation by D.A. Brown). American Geophysical Union, Washington, DC, pp. 1–279.

Thompson, P.H., 1989. An empirical model for metamorphic evolution of the Archaean Slave Province and adjacent Thelon tectonic zone, north-western Canadian Shield. In: Daly, J.S., Cliff, R.A., Yardley, B.W.D. (Eds.), Evolution of Metamorphic Belts. Geological Society Special Publications Geological Society of London, London, UK, pp. 245–263.

Thompson, P.H., Judge, A.S., Lewis, T.J., 1996. Thermal evolution of the lithosphere in the central Slave Province: implications for diamond genesis. In: LeCheminant, A.N., et al. (Eds.), Searching for Diamonds in Canada. Geological Survey of Canada, Open File 3228, pp. 151–160.

Thorpe, R.I., Cumming, G.L., Mortensen, J.K., 1992. A significant Pb isotope boundary in the Slave Province and its probable relation to ancient basement in the western Slave Province. Project Summaries, Canada Northwest Territories Mineral Development Subsidiary Agreement. Geological Survey of Canada Open File Report 2484, pp. 279–284.

Wang, K., 1999. Personal communication.

White, R.S., McKenzie, D., 1995. Mantle plumes and flood basalts. Journal of Geophysical Research 100B, 17543–17585.

van Breemen, O., Davis, W.J., King, J.E., 1992. Temporal distribution of granitoid plutonic rocks in the Archean Slave Province, northwest Canadian Shield. Canadian Journal of Earth Sciences 22, 2186–2199.

Yamashita, K., Creaser, R.A., Stemier, J.U., Zimaro, T.W., 1999. Geochemical and Nd–Pb isotopic systematics of late Archean granitoids, southwestern Slave Province, Canada; constraints for granitoid origin and crustal isotopic structure. Canadian Journal of Earth Sciences 36, 1131–1147.