Formation and evolution of Archaean cratons: insights from southern Africa

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Abstract: Archaean cratons are the stable remnants of Earth's early continental lithosphere, and their structure, composition and survival over geological time make them unique features of the Earth's surface. The Kaapvaal Project of southern Africa was organized around a broadly diverse scientific collaboration to investigate fundamental questions of craton formation and mantle differentiation in the early Earth. The principal aim of the project was to characterize the physical and chemical nature of the crust and mantle of the cratons of southern Africa in geological detail, and to use the 3D seismic and geochemical images of crustal and mantle heterogeneity to reconstruct the assembly history of the cratons. Seismic results confirm that the structure of crust and tectospheric mantle of the cratons differs significantly from that of post-Archaean terranes. Three-dimensional body-wave tomographic images reveal that high-velocity mantle roots extend to depths of at least 200 km, and locally to depths of 250–300 km beneath cratonic terranes. No low-velocity channel has been identified beneath the cratonic root. The Kaapvaal Craton was modified approximately 2.05 Ga by the Bushveld magmatic event, and the mantle beneath the Bushveld Province is characterized by relatively low seismic velocities. The crust beneath undisturbed Archaean craton is relatively thin (c. 35–40 km), unlayered and characterized by a strong velocity contrast across a sharp Moho, whereas post-Archaean terranes and Archaean regions disrupted by large-scale Proterozoic magmatic or tectonic events are characterized by thicker crust, complex Moho structure and higher seismic velocities in the lower crust. A review of Re–Os depletion model age determinations confirms that the mantle root beneath the cratons is Archaean in age. The data show also that there is no apparent age progression with depth in the mantle keel, indicating that its thickness has not increased over geological time. Both laboratory experiments and geochemical results from eclogite xenoliths suggest that subduction processes played a central role in the formation of Archaean crust, the melt depletion of Archaean mantle and the assembly of early continental lithosphere. Co-ordinated geochronological studies of crustal and mantle xenoliths have revealed that both crust and mantle have experienced a multi-stage history. The lower crust in particular retains a comprehensive record of the tectono-thermal evolution of the lithosphere. Analysis of lower-crustal xenoliths has shown that much of the deep craton experienced a dynamic and protracted history of tectono-thermal activity that is temporally associated with events seen in the surface record. Cratonization thus occurred not as a discrete event, but in stages, with final stabilization postdating crustal formation.

Archaean cratons have long been prime targets for a broad array of scientific studies, partly because they form the oldest cores of the continents, but also because they have economic significance as a major source of the world’s mineral wealth. (The term craton is confined here to Archaean crust, although the word is commonly applied to Proterozoic shields as well. The Archaean–Proterozoic boundary is, as always, imprecisely defined, but in southern Africa is marked by a sufficient hiatus in time that there is no difficulty in distinguishing Archaean terranes from all others (e.g. Windley 1995).) Cratons are underlain by a thick mantle root that is both chemically and physically distinct from the rest of the mantle, suggesting formation by processes or under conditions unique to Archaean time (e.g. Jordan 1975, 1978; Richardson et al. 1984; Pearson et al. 1995; McDonough & Rudnick 1998; Rudnick et al. 1998; Shirley & Walker 1998; Carlson et al. 1999, 2000). Jordan (1975) adopted the term ‘tectosphere’ to describe the deep conductive...
(non-convecting) layer of the Archaean mantle that remains attached to the craton through geological time. The tectosphere is composed dominantly of highly depleted peridotites with low normative density and high seismic velocities (Boyd & McCallister 1976; Jordan 1979; Boyd 1987). These refractory roots extend into the mantle to depths of at least 200 km and perhaps more (e.g. Jordan 1975; Lerner-Lam & Jordan 1987; Van der Lee & Nolet 1997; Jaupart & Mareschal 1999; Rudnick & Nyblade 1999; Shapiro et al. 1999; Zhao et al. 1999; Ritsema & van Heijst 2000; James et al. 2001b). The tectosphere is characterized by low heat flow (Jones 1988) and a very low geothermal gradient relative to Proterozoic mantle (McDonough & Rudnick 1998; Jaupart & Mareschal 1999; Nyblade 1999; Rudnick & Nyblade 1999).

Although attention over the years has tended to focus on the distinctive character of cratonic mantle, the nature of the crust and the crust–mantle boundary further distinguishes Archaean mantle from post-Archaean terranes. First, Archaean crust is typically thinner than that of adjacent Proterozoic terranes (Durrheim & Mooney 1991, 1994; Clitheroe et al. 2000; Nguuri et al. 2001; Assumpção et al. 2001), although not all agree with this assertion (e.g. Rudnick & Fountain 1995). The lower crust beneath cratons may be less mafic than that beneath Proterozoic terranes (Griffin & O’Reilly 1987: Durrheim & Mooney 1994). Moreover, the Archaean crust–mantle discontinuity is characteristically sharper than that beneath post-Archaean regions (Nguuri et al. 2001; Assumpção et al. 2001). These observations, together with the unique character of the cratonic root, suggest that the processes of craton formation, or the physical conditions that controlled those processes, differed in important ways from continent-forming processes in post-Archaean times. In this paper we examine these issues in light of results from the Kaapvaal Project of southern Africa.

The Kaapvaal Project

The Kaapvaal Project was undertaken to study the formation, stabilization and evolution of cratons, and to image the deep structure of the tectosphere (Carlson et al. 1996) (see also http://www.ciw.edu/kaapvaal for participants and a description of the project). A cornerstone of the Kaapvaal Project was a large-scale broadband seismic experiment designed specifically for geological-scale imaging of the crust and upper mantle beneath the cratons and adjacent Proterozoic provinces of southern Africa (Fig. 1). The seismic studies were undertaken in concert with a host of complementary investigations in geology, geochemistry and petrology of the cratons of southern Africa and their relationship to adjacent Proterozoic belts (see project summary on http://www.ciw.edu/kaapvaal).

The Kaapvaal and Zimbabwe Cratons of southern Africa form one of the pre-eminent natural laboratories of the world for the study of early continental formation. Indeed, southern Africa is something of a ‘type locality’ for probing fundamental questions of cratonic formation and evolution. Preserved within the Kaapvaal Craton is a nearly continuous Archaean geological record, from 3.6 to 2.6 Ga (e.g. De Wi et al. 1992). From an experimental point of view, the terrain, coupled with the logistical and academic resources of the region, made it ideal for a large-scale seismic deployment. The seismic array covered a wide range of age provinces spanning more than 3 Ga (Fig. 1). A significant area of the region covered by the seismic array, both on and off craton, has been perforated by thousands of kimberlite pipes from which a wealth of crustal and mantle xenoliths were erupted. These xenoliths, arguably the most extensively studied of any in the world, have been derived from depths up to 200 km or more. The chemical and mineralogical compositions of these nodules provide powerful constraints on the interpretation of results from the seismic analyses.

This paper is, first, a compilation and summary of initial results drawn from the Kaapvaal Project studies published as a special section in Geophysical Research Letters (28(13), 2001) and, second, a preliminary attempt to integrate those results into a comprehensive set of constraints on craton formation and evolution. The multi-disciplinary studies discussed below represent the contributions of an army of students and scientists, cited in the text and in the acknowledgements section at the end of this paper. We first review recent geochemical and petrological results and then focus upon two central imaging aspects of the seismic investigations: (1) delay time (tomographic) analysis of mantle structure beneath southern Africa (James et al. 2001b); (2) receiver function analysis of the depth and topography of the Moho beneath the seismic array (Nguuri et al. 2001).

Tomographic results show that high-velocity mantle roots extend locally to depths of at least 250 km beneath undisturbed Archaean craton, with no comparable root structures beneath post-Archaean terranes. Neither body- nor surface-wave analyses have yet produced evidence of a low-velocity zone beneath the cratonic keel (Ritsema & van Heijst 2000; Freybourger et al. 2001;
Strong variations in crustal thickness based on receiver functions reveal significant differences in the nature of the crust and the crust–mantle boundary between Archaean and post-Archaean geological terranes. Both seismic and geochemical results show that the Kaapvaal keel was modified c. 2.05 Ga by a massive Bushveld event that affected both crust and mantle across a broad east–west swath of southern Africa.
Geological outline of southern Africa

Southern Africa is a complex collage of geological provinces. In the summary below we provide a thumbnail geological sketch of those provinces that bear directly on this study. In categorizing and describing the surface geology we have made a number of simplifications in the interest of characterizing features on a scale appropriate to our studies.

Cratons

The Archaean Kaapvaal and Zimbabwe Cratons form the nucleus of southern Africa. The Kaapvaal Craton, which is the better studied of the two, is composed of a mosaic of distinct geological terranes covering more than 10^6 km^2, with the oldest units generally in the eastern part of the craton and the youngest in the western part (de Wit et al. 1992). These terranes of disparate geological histories were assembled over a 1 Ga period from early Archaean (c. 3.6 Ga) to late Archaean time (c. 2.6 Ga) (de Wit et al. 1992; de Wit & Hart 1993; Carlson et al. 2000).

The nearly 0.5 x 10^6 km^2 continental mass that stabilized in the early Archaean time was termed the ‘Kaapvaal shield’ by de Wit et al., and forms the eastern part of the craton. The term shield was used by de Wit et al. to distinguish the geological character and mechanism of formation of the early Archaean regions from late Archaean stabilization. The oldest geological sub-domains, the Barberton and the Ancient Gneiss terranes, crop out in the northeastern part of the craton, close to the most easterly extent of the seismic array near the NW border with Swaziland (Fig. 1). The mafic and ultramafic volcanic rocks of the Barberton region have been interpreted by de Wit and coworkers (de Wit et al. 1992; de Wit & Hart 1993) as the products of mid-ocean ridge (MOR) magmatism, preserved by subsequent obduction onto an arc-like terrane. Recent field studies combined with laboratory experiments suggest, however, that the Barberton komatitites and associated basalts were the products of wet melting in an Archaean subduction zone, a process that can also lead to high degrees of depletion in the upper mantle (Parman et al. 1997, 2001; Grove et al. 2000).

The second period of cratonic development (onset c. 3.1 Ga) was viewed by de Wit and coworkers (de Wit et al. 1992; de Wit & Hart 1993) as a period of continental growth through a combination of tectonic accretion of smaller crustal terranes and subduction-related igneous and tectonic processes. Craton development culminated in the final mantle stabilization c. 2.6 Ga. The growth of the craton during this period was dominantly in the northern (including the Limpopo Belt) and in the western regions, which consist largely of agglomerated granite–greenstone fragments. The onset of this second period of craton evolution witnessed widespread extensional volcanism, followed by the development of enormous depositional basins. Notable among these was the Witwatersrand basin, supplied by sediments from a predominantly northern provenance, and the Pongola basin, at least the southern sector of which was part of a passive continental margin facing open ocean on the south (de Wit et al. 1992). A craton-wide extensional cycle of Ventersdorp volcanism and sedimentation between 2.7 and 2.6 Ga marked the end of craton formation in Archaean time.

Limpopo Belt

The Limpopo Belt is a high-grade metamorphic terrane formed by collision in late Archaean time between the Kaapvaal and Zimbabwe Cratons. The Limpopo Belt is divided into a Northern Marginal Zone, a Central Zone and a Southern Marginal Zone (Van Reenen et al. 1992). Seismic evidence summarized below confirms that the two marginal zones are overtrusts atop cratonic crust. The Central Zone, on the other hand, has a long and complex tectonic history that culminated in considerable crustal thickening during the collisional phase, followed by the uplift and exhumation of at least 20 or 30 km of crust (Van Reenen et al. 1987, 1992; Treloar et al. 1992; Windley 1995).

Proterozoic mobile belts

Mobile belts, accreted to southern Africa in mid- to late-Proterozoic time, surround the cratons and act to buffer them from later plate-tectonic events. The Kaapvaal is bounded on the south and east by the subduction-related Namaqua–Natal Belt of Proterozoic age and on the west by the Kheis overthrust belt (de Wit et al. 1992). The complex region flanking the Zimbabwe Craton west and NW of the Limpopo Belt is here termed the Okwa–Magondi terrane, although the regional geology remains enigmatic as a result of very complex geological relationships made all the more obscure by extensive Kalahari sand cover (Carney et al. 1994). The c. 2 Ga events of the Okwa and Magondi Belts were recorded in the western Limpopo Belt during the same time period and imposed a strong overprint on
the western sector of the Zimbabwe Craton where it extends into Botswana. The region we have termed Okwa–Magondi Belt is the site of the richest diamond mines in the world (Fig. 1), and from Re–Os studies of xenoliths and diamond inclusions the mantle beneath that region is known to be of Archaean age (Shirey et al. 2001). The presence of abundant diamonds with Archaean-age inclusions suggests that despite the Proterozoic overprint, the Archaean geotherm was not significantly disturbed. As we will show below, the Proterozoic-age events in the western sector of the Zimbabwe Craton are clearly recorded in the seismic crustal structure, and to a lesser extent in the mantle velocity structure.

**Bushveld Province**

The Bushveld Complex is the largest layered mafic intrusion in the world, with a total volume of magma intruded into the crust of the order of $0.6 \times 10^6$ km$^3$ (Von Gruenewaldt et al. 1985). The 2.05 Ga events associated with the Bushveld intrusion had a profound impact on the seismic and geological character of crust and mantle over a region far more extensive than that implied by outcrops of the Complex itself. Thus, we adopt here the term Bushveld Province in place of Bushveld Complex to connote the fact that Bushveld age correlations (2.05 Ga) are found along a broad east–west swath extending from the easternmost outcrops of the Complex westward well into southern Botswana (Hatton & Von Gruenewaldt 1990). For the purposes of discussing the seismic results presented here, the Bushveld Province encompasses the greater region of Bushveld age exposures.

**The Vredefort structure: a cross-section of Archaean crust**

The Vredefort dome is the deeply eroded remains of a large meteor impact crater formed c. 2.0 Ga in the central part of the Kaapvaal Craton. The crater is roughly 40 km in diameter and reveals a crustal section turned on edge, a virtually complete and well-preserved crustal cross-section exposed on the floor of the crater (e.g. Hart et al. 1990; Moser et al. 2001). The mid-crust (below about 14 km of Witwatersrand sedimentary rocks) is composed of about 10 km of granodiorite composition rocks (the outer granitgneiss domain), which shows a transition into an incomplete section of lower-crustal granulite facies gneiss of mixed felsic and mafic components. Peridotites believed to be mantle samples and with Re-depletion model ages of 3.3–3.5 Ga have been obtained from a drill hole near the centre of the impact structure (Tredoux et al. 1999).

**Primary results from geochemistry and petrology**

Comprehensive Re–Os isotope studies show that peridotite xenoliths brought to the surface in kimberlite pipes both on and off craton in southern Africa have ages similar to the overlying crust (Carlson et al. 2000; Irvine et al. 2001) (see Fig. 2). The mantle root of the craton is of Archaean age at least to the maximum depth (>200 km) from which nodules are derived. There is no correlation between the age of the xenolith and its depth of origin, an observation that rules out keel formation by progressive underplating of material to the continent over long time scales. Most peridotite nodules associated with undisturbed craton are now known to be ancient, including the sheared nodules once thought to have been part of the asthenosphere (Boyd 1987), but which are now considered to be metasomatized samples of highly depleted Archaean mantle (Carlson et al. 2000, with reference to previous work).

Recent studies have also shown that eclogite materials, occurring both as inclusions in diamond and as whole-rock samples, are as old as or older than the peridotites (Richardson et al. 2001; Shirey et al. 2001). This finding suggests that the eclogites and, by inference, subduction zones, were an integral part of the formation process of the craton. In addition, stable isotope analyses suggest a MOR basalt protolith for the eclogites, providing further evidence for a subduction-zone source for the eclogites. This evidence that subduction processes were involved in earliest craton formation is buttressed by laboratory petrological experiments (Parman et al. 1997, 2001; Grove et al. 2000). Those studies show that the high-MgO komatiite magmas of the Kaapvaal probably formed within an Archaean subduction zone via a single melt generation process similar to that by which high Mg-number boninites are formed in very young modern subduction zones. Petrological examination of Archaean ultramafic magmas (komatiites) from South Africa indicates that some komatiitic magmas contained substantial quantities of water (>4 wt%). This finding strengthens the possibility that the cratonic lithosphere formed initially in subduction-zone settings whose demise led to accretion of the arc crust and thickening of refractory mantle to create a stable, thick, continental
lithosphere (Carlson et al. 2000; Grove et al. 2000). The paucity of eclogite xenoliths (<1%) (Schulze 1989) indicates that essentially all of the descending oceanic plates during Archaean time were subducted into the deeper mantle, with little eclogitic material incorporated into the cratonic root.

The co-ordinated Kaapvaal Project geochronological studies of crustal and mantle xenoliths reveal that both crust and mantle have experienced a multi-stage history, and that a simple view of cratonicization as a discrete event is not a viable model for craton formation (Schmitz et al. 1998; Schmitz & Bowring 2000; Moser et al. 2001). The lower crust in particular retains a comprehensive record of the tectonothermal evolution of the lithosphere. The study of lower-crustal samples has shown that much of the deep craton experienced a dynamic and protracted history of tectonothermal activity that is temporally associated with events seen in the surface record, including late Archaean magmatism (Ventersdorp) and even Proterozoic deformation (Namaqua–Natal) (Schmitz et al. 1998). Thermal events are reflected in ages of 2 Ga or even less for some eclogitic diamond inclusions from tectospheric mantle originally of Archaean age (Carlson et al. 2000; Shirey et al. 2001).

Moser et al. (2001) suggested that stabilization of the mantle keel in the region of the Vredefort impact structure may have taken place roughly 100 Ma after formation of the overlying crust. They based this assertion on the notion that late Archaean thermotectonic events recorded in the lower crust argue against the presence of a thick mantle root that would have served to buffer the crust from such events. A separate evolutionary history for crust and underlying mantle is not inconsistent with the peridotite results that show the majority of peridotite xenoliths have Re–Os model ages younger than 3.0 Ga, compared with overlying crustal formation ages of 3.2–3.4 Ga. However, if this result is correct, and if it is craton-wide, it implies that cratonic crust and its underlying mantle root need not have a syngenetic relationship.

Lithospheric mantle xenoliths of Proterozoic age are derived from depths up to about...
150 km beneath the surrounding Proterozoic mobile belts, indicating crust–mantle coupling and long-term stability of the upper mantle beneath the Proterozoic belts (Carlson et al. 2000). The Proterozoic mantle root is not only thinner than that present beneath the cratons, but the peridotite components are more fertile in composition. Interestingly, however, the geothermal gradients beneath the southernmost Kaapvaal Craton and the adjacent Namaqua–Natal Belt are similar based on mantle xenolith geothermometry and geobarometry (P. Janney, pers. comm. 2000).

**Southern Africa seismic experiment**

The southern Africa seismic experiment is one of the largest broadband portable array seismic investigations ever undertaken. The experiment has produced images of crustal and upper-mantle structure in unprecedented detail beneath the cratons and adjacent Proterozoic provinces. Fifty-five portable broadband REFTEK/STS-2 seismic stations were deployed from April 1997 to July 1999 in a roughly 100 km grid along a NNE–SSW transect about 1800 km long by 600 km wide in southern Africa (Fig. 1). Approximately half the instruments were redeployed to new sites in April–May 1998 for a total of 82 station locations. The matched three-component sensors were installed in semi-permanent low-noise vaults on bedrock and signals were recorded with 24-bit dynamic range and a continuous sampling rate of 20 samples s⁻¹. The array data were supplemented by data from three global digital stations in the region of the array (Fig. 1). The experiment was augmented by a 6 month deployment of 32 broadband telemetered stations installed in a tight (c. 65 x 50 km) array in the region around and to the NW of Kimberley.

The seismic results discussed here are organized into two major topics: (1) tomographic images of the upper mantle beneath southern Africa; (2) receiver function characterization of crustal thickness and the crust–mantle interface beneath the stations of the array. Included also from work in progress is S-wave velocity structure from two-station surface-wave phase and group velocity inversion for a propagation path across undisturbed Kaapvaal Craton (Gore 2002; Nguuri 2002).

**Upper-mantle structure**

*Tomographic results.* Upper-mantle velocity structure (James et al. 2001b) was determined by tomographic techniques based on the analysis of delay times from teleseismic broadband waveform data. Relative arrival times of phases P, PKPdf, S and SKS were retrieved via a multi-channel cross-correlation procedure using all possible pairs of waveforms (VanDecar & Crosson 1990). This procedure produces highly accurate delay times, with typical standard errors for the

![Events for P phases](#)

![Events for S phases](#)

*Fig. 3.* Location map of events used for tomographic inversions. The map is centred on the approximate centre of the southern Africa array. Locations are from the National Earthquake Information Centre bulletin as archived by the Incorporated Research Institutions for Seismology (IRIS) Data Management Center (James et al. 2001b). Published in James et al., GRL28, 2001, fig. 2.
southern Africa data of c. 0.03 s for P waves and 0.06 s for S waves. Results for P waves are of higher resolution than those results for S with 8693 rays from 234 P-wave events compared with 4834 rays from 148 S-wave events. Epicentres of earthquakes used in the analysis are shown in Figure 3.

The inversion method for obtaining velocity structure has been fully described by (VanDecar 1991). By this method, P- and S-wave delay times are inverted independently for structure beneath the array. The model is parameterized identically for the P- and the S-wave inversion with splines under tension constrained by a series of regular knots (Fig. 4). Within the interior portion of the model, the knots are spaced 50 km apart in depth and ½ degree apart in latitude and longitude. Corrections for station elevation and crustal thickness from Nguuri (2002) are applied to the data before inversion. The data are inverted simultaneously for the slowness perturbation field, earthquake relocations and station corrections. The inclusion of earthquake relocations and station corrections ensures that the resulting velocity model will be constrained to contain the least amount of structure required to satisfy the observations within their estimated standard errors.

The tomographic images presented in Figures 5 and 6 were determined using linear inversion and are therefore preliminary (James et al. 2001b). Results of the non-linear inversion and a detailed analysis of resolution tests will be provided in a future work. Although the non-linear inversion will undoubtedly affect results in the deeper parts of the model, the shallow (<300 km), relatively mild, velocity perturbations obtained from the linear inversion are unlikely to change significantly. We have designed simple resolution tests that approximately mimic the observed structures. The output from the tests and a description of our procedures can be found in the electronic supplement to James et al. (2001b), or as a downloadable file at http://www.ciw.edu/kaapvaal/pubs/tomography/grl_supplement.pdf. The resolution tests indicate that both laterally and vertically the cratonic roots are well recovered. As expected, recovered velocity perturbations are lower by varying degrees than those of the input model; this effect increases with depth in the keel. The maximum and minimum values of the recovered velocities are, however, approximately equal to those of the input model, suggesting that the maximum and minimum velocity perturbations obtained in the data inversions are not greatly different from those in the real Earth. Downward smearing of structure does occur, but the effect is relatively small and does not preclude reasonably accurate estimates of keel thickness. Because station spacing of the southern Africa array is about 100 km, shallow structures above about 50 km are sampled by relatively few crossing paths from teleseismic events and are thus included largely in the station terms. A further qualitative assessment of resolution is provided by the ray density as seen in Figure 7, which shows the ray 'hit' count in both P- and S-velocity horizontal and vertical sections shown in Figures 5 and 6. Of particular relevance is the high and rather uniform ray density along the central profile B–B'.

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**Fig. 4.** A 3D perspective view of grid knots for splines under tension that constrain the velocity perturbation model. Knot positions occur at line crossing. The yellow lines indicate regions to which the velocity maps and sections are confined.

**Fig. 5.** Cross-sections through the P-wave velocity perturbation models obtained by inversion of delay times corrected for elevation and crustal thickness. (a), (b), and (c) show plan views of velocity perturbations at depths of 150, 200 and 300 km, respectively. (d), (e) and (f) show vertical cross-sections along profiles A–A', B–B' and C–C', respectively, as shown in the horizontal sections. Surface topography is plotted at 20 times actual scale. Uppermost 50 km (shaded area) in vertical sections denotes regions where station delay time residuals are incorporated in model calculations. Colour scale shows the velocity perturbation in percent. Colours fade to black for ray hit counts less than 10 (see Fig. 7).
Fig. 7. Grey-scaled ray density maps of horizontal and vertical velocity sections shown in Figures 5 and 6. The scale grades from 0 (white) to 100 (black), with the scale reflecting the number of ray 'hits' in each node of the velocity perturbation model. Ray densities along the axis of the array are typically >50 hits per node.

Fig. 6. Cross-sections through the S-wave velocity perturbation models. (See caption for Figure 5 for a description of the cross-sections.)
The Proterozoic Namaqua-Natal Mobile Belt, extending at least into the southern part of the Okwa Belt of Botswana and probably into the Magondi-Limpopo zone in NE Botswana as well. Although these low mantle velocities are well resolved overall, the localized ‘patchiness’ of the low-velocity perturbations seen in Figures 5 and 6 is not. The tomographic results are consistent with the fact that whereas surface geology of the Okwa–Magondi region registers a strong Proterozoic overprint, the Re–Os signatures of the mantle xenoliths and diamond inclusions are clearly of Archaean age (Richardson et al. 2001; Shirey et al. 2001).

Within the resolution of the data, the mantle structure of the Archaean Limpopo Belt does not differ significantly from that of the adjacent cratons. The similarity with cratonic mantle structure contrasts sharply with the results of crustal structure determinations (Nguuri et al. 2001), which show the Central Zone of the Limpopo Belt to be characterized by thick crust and poorly developed Moho relative to the adjacent cratons. Interestingly, the SKS splitting results for the southern Africa array show that the Limpopo Belt exhibits a consistent east–west mantle fabric, presumably acquired at the time of craton collision (Silver et al. 2001).

The Proterozoic Namaqua–Natal Mobile Belt, thought to be the remnants of a major north–south convergent margin that extended as far north as the Zimbabwe Craton (De Wit et al. 1992), is characterized by velocity perturbations uniformly lower than those observed beneath the craton. The lower velocities are in keeping with the observation that the off-craton Proterozoic mantle tends to be somewhat more fertile (higher Fe) than that of the adjacent craton (Carlson et al. 2000; Pearson et al. 2002). Patches of higher-velocity material are seen in the 200–400 km depth range beneath the belt, however, and these higher velocities typically exhibit continuity with the high-velocity material beneath the adjacent Kaapvaal Craton.

Surface-wave results. Comprehensive studies of inter-station surface-wave dispersion across the southern Africa array are included in studies currently in progress (Gore 2002; Nguuri 2002). We cite here some preliminary results based on both phase and group velocity dispersion curves from the high-quality southern Africa data because they provide a relatively unambiguous test for low-velocity zones in the upper mantle. The question of a low-velocity zone at depths <200 km beneath the craton is an issue because it has purportedly been detected beneath southern Africa (Qiu et al. 1996; Priestley 1999). Initial results from the Kaapvaal experiment, including velocity estimates from mantle xenoliths to depths of 200 km, provide no support for a low-velocity zone beneath the craton. Surface-wave studies using much larger global and regional datasets (Zhao et al. 1999; Ritsema & van Heijst 2000) and the Kaapvaal array data using different methods (Freybourger et al. 2001) also failed to reveal a low-velocity zone beneath the craton (also see Discussion section below).

A typical example of inter-station Rayleigh-wave phase and group velocity dispersion from an Andean event for a pure cratonic path (between stations 16 and 32 (see Fig. 9)) across the southern Kaapvaal Craton near Kimberley is shown in Figure 8. Inversion of the surface-wave dispersion curves yields a model with high shear-wave velocities (>4.6 km s⁻¹) in the mantle to a depth of at least 200 km. No models incorporating a significant low-velocity zone fit the data. The results shown in Figure 8 are typical of many dozens of similar inversions of inter-station phase and group velocities of both Rayleigh and Love waves across the southern Africa array. The surface-wave data are entirely consistent with S-wave velocity estimates from mantle xenoliths (James et al. 2001a), which show velocities reaching a maximum of nearly 4.75 km s⁻¹ in the uppermost mantle, decreasing slightly to about 4.6–4.65 km s⁻¹ at 200 km depth. No samples, including several high-Fe lherzolites from
Fig. 8. (a) Rayleigh waves recorded on the vertical component at stations sa16 and sa32 for an on-azimuth magnitude 6.8 event on 15 October 1977 in central Chile. Total time 1350 s on the horizontal axis. Records have been decimated and instrument corrected for displacement. Seismograms have been low-pass filtered at 0.05 Hz. (b) Two-station Rayleigh wave phase and group velocities v. period for pure craton path across the southern Kaapvaal Craton (left panel). Left-hand panel shows measured phase velocities (RPF) and group velocities (RGF) for the 464 km interstation path between sa16 and sa32. Data, with phase velocities more heavily weighted than group velocities, have been inverted to obtain the best-fit S velocity–depth model shown as a continuous line in the right-hand panel. The dashed line was the starting model. The line curves in the right-hand panel are based on the best-fit S velocity model (adapted from Nguuri (2002)).
about 200 km depth, have calculated S-waves velocities $<4.6 \text{ km s}^{-1}$.

**Crustal structure**

Initial results from the analysis of P- to S-wave converted phases (Ps) from the M-discontinuity were reported by Nguuri et al. (2001) and are summarized below. The results reported by Nguuri et al. were based upon 35 teleseisms processed at 75 stations (Fig. 9) to yield a comprehensive set of high-quality receiver functions (see Ammon (1991) for review and other references to receiver function techniques). Individual receiver functions were corrected for travel time moveout, binned by station and stacked by phasing depth at depth intervals of 0.5 km between 1 and 101 km. The resulting stacked receiver function images provide a display of the discontinuity structure beneath each station based on Ps conversions (e.g. Gurrola et al. 1994; Dueker & Sheehan 1997, 1998). The crustal model for the moveout correction is based on seismic refraction models for the Kaapvaal Craton, with an average P-wave velocity for the crust of 6.5 km s$^{-1}$, Poisson’s ratio of 0.25, and Moho depth of 38 km (Durrheim & Green 1992). Crustal multiples are suppressed in this procedure by virtue of the fact that the moveout correction

![Location map for the southern Africa seismic array, plotted on an outline map of southern Africa geology (from Nguuri et al. 2001). Published in Nguuri et al., GRL28, 2001, fig. 1 (corresponding author D. E. James).](http://sp.lyellcollection.org/Downloaded-from)
is appropriate only for the direct P- to S-wave conversion. Moreover, the amplitudes of the Ps signals in the stacked traces (phasing depth images) plotted in Figure 10 are not strictly interpretable, as they average events with differing ray parameters. As seen from Figure 10, only one consistent Ps signal occurs, and it is readily associated with the M-discontinuity. These results are summarized in the gridded map of crustal thickness shown in Figure 11 (from Nguuri et al. 2001). As the velocity–depth model was constructed by averaging seismic refraction results for the Kaapvaal Craton, the model may underestimate mean crustal velocity and crustal thickness for off-craton or modified cratonic regions (Nguuri et al. 2001).

Stations located within undisturbed Kaapvaal or Zimbabwe Craton typically have sharp, large-amplitude depth images for the Moho. Among the more distinctive results are those from stations located in the Zimbabwe Craton, where Moho depths with one exception cluster tightly

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**Fig. 10.** One-dimensional phasing depth images for southern Africa, organized by geological province. The number of events included in the stack and the station name are shown to the right of each trace. The dominant signal on most of the depth images shown is the Ps conversions from the Moho. Relatively thin (35–40 km) crust and sharp, well-defined Ps Moho conversions are associated with undisturbed craton. Ps arrivals associated with modified regions of the craton and post-Archaean terranes tend to be more diffuse and of smaller amplitude (from Nguuri et al. 2001, published in GRL28, 2001, fig. 2).
between 34 and 37 km. The one exception, sa71, is in the western craton near the zone of Proterozoic (c. 1–2 Ga) overprinting. Results for crustal thickness beneath the Kaapvaal Craton exhibit more variability, with thickness varying between about 33 and 45 km, and averaging about 38 km.

By comparison with stations within regions of undisturbed craton, those within post-Archaean and modified cratonic regions typically exhibit much more poorly defined Ps converted phases. Moreover, the crust beneath post-Archaean and modified cratonic regions is in almost all cases thicker than that beneath the cratons. Of particular interest is the broad region of the Bushveld and Okwa–Magondi terranes that extends from the Bushveld Complex in the northern Kaapvaal Craton westward into eastern and northeastern Botswana. The Archaean craton through this entire zone has been overprinted by tectonomagmatic events of Proterozoic age (e.g. Carney et al. 1994; Shirey et al. 2001) and the effects on crustal structure have been profound. The crust thickens systematically and the Moho signature degrades, for example, across the boundary between undisturbed craton and the Bushveld Province and between undisturbed Zimbabwe Craton and the Magondi Belt. The change in crustal structure from undisturbed craton into the Bushveld and Okwa–Magondi terranes is consistent with the results from upper-mantle tomographic studies described above, which show a significant low-velocity perturbation in the uppermost mantle extending beneath the entire region (James et al. 2001b).

The intercratonic Limpopo Belt is of particular interest. Depth images for the marginal zones are typical of cratonic structure, both in character of the Ps conversion and in crustal thickness. The northern marginal zone is underlain by crust about 37 km thick, typical of the adjacent Zimbabwe Craton, and the southern marginal zone has a crust around 40–42 km thick, consistent with that of the adjacent Kaapvaal Craton (see Fig. 10). The central Limpopo Belt, the site of pervasive deformation during the collision of the Kaapvaal and Zimbabwe Cratons in Archaean time, displays particularly complex structure. In some instances (e.g. stations sa66 and sa67) the identification of the Moho is ambiguous and could be resolved only with constraints from two-station surface-wave phase velocity inversions, which show unambiguously that thicker crust is required to satisfy the dispersion data (Gore 2002; Nguuri 2002). The broad, poorly defined Ps images have depth maxima occurring between 40 and 53 km and are indicative of a structurally complex Moho. These values for crustal thickness beneath the central zone of the Limpopo Belt obtained by Nguuri et al. (2001) differ significantly from those of some previous seismic and gravity studies, where a Moho depth of about 30 km was postulated (Stuart et al. 1986; De Beer & Stettler 1992; Gwavava et al. 1992). Although our results show that there may be a deep crustal discontinuity at about 30–35 km depth beneath some stations, inter-station surface-wave phase velocity inversion results show unambiguously that the crust must be at least 40–45 km thick beneath the Central Zone (Gore 2002; Nguuri 2002). As with the Bushveld Complex, the relatively thick crust beneath the Limpopo Belt does not translate to higher elevations relative to the adjacent cratons.

The Kheis Belt is similar to the Northern and Southern Marginal Zones of the Limpopo Belt in that it is an overthrust sheet atop Archaean crust of the Kaapvaal Craton. Not surprisingly, therefore, the crustal thickness of the Kheis Belt is similar to that of the adjacent Kaapvaal Craton. Indeed, the only evidence for crustal thickening beneath the Kheis Belt is in its northern extent near its contact with the Okwa Belt of southern Botswana. The large amplitudes of the Ps Moho conversions observed in the marginal zones of the Limpopo Belt, however, are not so apparent in the Kheis Belt. The lack of a purely 'cratonic' Moho signature is in some sense consistent with tomographic images of the upper mantle beneath the Kheis Belt, where large positive velocity perturbations observed in the cratonic mantle of the Kimberley region tend to decrease westward (James et al. 2001b).

Discussion

Tectospheric roots

The depth extent of cratonic roots has long been an issue of some controversy, dating back to Jordan’s seminal work in the mid-1970s (Jordan 1975). Although the tectosphere hypothesis of deep, cold, and chemically distinct keels beneath
continents was widely challenged when first proposed (e.g. Okal & Anderson 1975; Anderson 1979; Sclater et al. 1980), the model has gained widespread acceptance, buttressed by petrological and geochemical studies of mantle xenoliths, particularly those from southern Africa. Notable among these studies are Re–Os age determinations that show that mantle nodules erupted from even the greatest depths beneath the craton (about 200 km) are of Archaean age (Pearson et al. 1995; Carlson et al. 2000). Moreover, recent analyses of xenolith P–T data suggest that the intersection between the craton geotherm and the mantle adiabat occurs between depths of 160 and 300 km, with a best estimate in the range of 220–250 km beneath the cratons (Rudnick et al. 1998; Rudnick & Nyblade 1999). These estimates are entirely consistent with our results, which show that positive velocity perturbations beneath the cratons extend to depths of at least 250–300 km, and with results from other detailed regional seismic studies (VanDecar et al. 1995; Van der Lee & Nolet 1997; Ritsema et al. 1998; Ritsema & van Heijst 2000).

In general, low geotherms and refractory mantle compositions contribute to the high velocities associated with the tectospheric keels of cratons (Jordan 1975, 1981). The rocks that make up the bulk of the tectospheric mantle are highly depleted peridotites. The sheared nodules formerly believed to be enriched are now known to be depleted peridotites that were heavily metasomatized shortly before eruption (Pearson et al. 1995; McDonough & Rudnick 1998; Carlson et al. 2000). Although it is possible that metasomatized rocks are widespread in the deep cratonic mantle, pervasive metasomatism would significantly reduce seismic velocities, which is not observed in the seismic data. Jordan showed, for example, that fertile cratonic samples contain significant weight percentages of both clinopyroxene and garnet, resulting in seismic velocities up to 1% lower and densities 1–2% higher than the depleted nodular peridotites (Jordan 1979). Similarly, eclogite mantle, if present (Shirey et al. 2001), will have both lower velocity and higher density than depleted peridotic mantle at the same temperature, although an inventory of mantle xenoliths of the Kaapvaal Craton suggests that the cratonic mantle is <1% eclogite by volume (Schulze 1989).

Results to date from the southern Africa seismic experiment, including the surface-wave results cited above, fail to reveal a low-velocity asthenospheric layer beneath the Archaean keel. This result is particularly relevant in light of studies published by Qiu et al. (1996) and modified by Priestley (1999), which fit surface-wave form data from regional earthquakes recorded at global digital stations in southern Africa. Their interpretation of the data indicated a high-velocity lid in the uppermost mantle to a depth of 160 km, underlain by a significant low-velocity zone with S-wave velocity about 4.32–4.45 km s⁻¹ (Qiu et al. 1996; Priestley 1999). As indicated in the discussion of tomographic results, however, these results are inconsistent with other surface-wave results from the Kaapvaal experiment as well as studies based on much more comprehensive global and regional datasets (Zhao et al. 1999; Ritsema & van Heijst 2000; Freybourger et al. 2001). Perhaps more significantly, velocity estimates based on modal abundances in mantle xenoliths and P–T conditions determined from geothermometry and geobarometry reveal not a single sample out of the 50 or so studied that produces a predicted S-wave velocity less than 4.6 km s⁻¹ at ambient mantle conditions to depths of 200 km (James et al. 2001a).

The low P-velocity anomaly (c. 0.5%) in the mantle associated with the extended Bushveld and Okwa–Magondi terranes suggests considerable modification of cratonic mantle by Proterozoic tectonomagmatic events. That the mantle has been modified chemically is evidenced in the Re–Os data, where mantle nodules from the Bushveld Province have been reset to Proterozoic (about 2.05 Ga) ages (Carlson et al. 2000; Pearson et al. 2001). The isotopic resetting of a large volume of Archaean Kaapvaal mantle apparently required material addition (Carlson et al. 2000). The observed seismic velocity reduction is small, perhaps c. 0.5% in P and c. 0.8% in S. Although a thermal anomaly of c. 100°C could produce the velocity effect observed (Christensen 1982), there is no evidence for higher geotherms in the region of the Bushveld Province either from the observed heat-flow measurements or from thermobarometric determinations in mantle nodules (Danchin 1979; Jones 1988). For standard P–T conditions, Jordan showed that observed chemical variations in rocks of the cratonic mantle, including ‘refertilized’ samples, can account for up to 1% total velocity variation, with typical heterogeneity of the order of 0.5% (Jordan 1979). The seismic velocity perturbations that are observed between undisturbed Archaean cratonic mantle and cratonic mantle that has been modified by Proterozoic events are comfortably within this velocity range. The lower velocities observed in the adjacent Proterozoic belts are probably the result of more fertile compositions and higher geothermal gradients combined. The regions of modified cratonic mantle such as that associated with the Bushveld Province are cautionary in terms of a central assumption of
the geochronological studies discussed in a previous section (see Results from geochemistry and petrology). There it was postulated that a thick mantle root, had it been present during the whole of late Archaean time, would have shielded the crust from tectonomagmatic activity (Moser et al. 2001). Although plausible, this is not proven and may not be generally applicable, as we have seen from the Bushveld example. The abundance of diamonds brought to the surface in kimberlite pipes of post-Bushveld age within the Bushveld Province indicates that massive volumes of mafic magma can rise through cratonic mantle and intrude the crust without raising the geotherm of the cratonic root sufficiently to move it out of the diamond stability field. The inference from this observation, therefore, is that widespread magmatic events recorded in the lower crust do not necessarily preclude the presence of a tectospheric root. On the other hand, in the case of the Bushveld Province, the through-flow of magma seems apparently to have refertilized the mantle involved. If late Archaean events similar to the Bushveld event had occurred within the craton, their geochemical signature should remain visible in the seismic image, as it does beneath the Bushveld Province. This is not observed elsewhere in the craton.

**Crustal structure**

Among the most significant findings presented by Nguuri et al. (2001) is evidence for pervasive Proterozoic (c. 2 Ga) modification and thickening of Archaean crust across a broad east–west zone bounded on the east by the Bushveld and on the west by the Okwa–Magondi terranes. The area of thickened crust corresponds closely to the zone of reduced upper-mantle velocities found from body-wave tomography (James et al. 2001b). Moho Ps conversions for stations in this region of disturbed craton tend to be low in amplitude and in some cases ambiguous, suggesting that the Moho is a weak and/or transitional (e.g. >5 km) boundary. One possible interpretation of the poor Ps signals is that they reflect Proterozoic age magmatic underplating or reworking of Archaean crust (e.g. Griffin & O’Reilly 1987). Although magmatic addition to the crust is plausible beneath the Bushveld Complex, the cause of increased crustal thickness elsewhere in the region of Proterozoic overprinting is not so apparent. Both crustal thickness and the Moho signature observed in the region of modified Archaean crust are similar to those observed at stations in the Proterozoic Namaqua–Natal Belt, although a comparative analysis of structure between disturbed Archaean and Proterozoic terranes has yet to be carried out.

In a review of Precambrian crustal structure worldwide, Durrheim & Mooney (1994) found that Archaean crust is typically 27–40 km thick, whereas Proterozoic crust is about 40–55 km thick, with higher-velocity material (P waves >7 km s⁻¹) at its base. For southern Africa, the thinnest crust (35–40 km) is found beneath those regions of the Kaapvaal and Zimbabwe Cratons that have been undisturbed since Archaean time. A prominent exception is the crustal thickness beneath the Limpopo Belt. The Central Zone of the Limpopo Belt is not only characterized by thick crust (up to 50 km or more) and complex Moho, but there also is strong geological evidence that a total of 20–30 km of crustal uplift and exhumation has taken place since Archaean time (Treloar et al. 1992). If correct, this implies that the crustal section beneath the Limpopo Belt in Archaean time was comparable with the thinnest crust observed today in the Himalayan and Andean convergent margins. It also may indicate that those Archaean terranes where the crust is thin and the Moho simple were not shaped to any significant degree by continent–continent collisional processes.

The unexpected relationship between thin cratonic crust and high elevations and large negative Bouguer gravity anomalies has been the subject of a number of recent studies (Webb et al. 1999; Webb 2002). Although thin crust and high elevation have been observed elsewhere in the world (Braile et al. 1989; Durrheim & Mooney 1991, 1994; Assumpção et al. 2002), the question remains of how the higher elevations are gravitationally compensated at depth. Compensation produced by the low-density upper mantle beneath the craton is only part of the story, particularly if the lower crust beneath cratons is less mafic on average (and hence less dense and lower velocity) than is post-Archaean lower crust (Griffin & O’Reilly 1987; Durrheim & Mooney 1994; Rudnick & Fountain 1995). Poisson’s ratios (a measure of the elastic parameters of a solid) for the crust determined both from receiver function analysis and from the relationship between P-wave and S-wave travel times (Wadati diagrams) tend to be low for undisturbed regions of the Kaapvaal, around 0.25, whereas values for off-craton areas and disturbed craton are in the range 0.27–0.29 (Nguuri 2002). The average Poisson’s ratio of 0.25 for the cratonic crust suggests that it may have an intermediate or even felsic bulk composition. Similarly, the large-amplitude Ps conversions reported by Nguuri et al. (2001) are consistent with a less mafic lower crust in the Kaapvaal Craton. Whereas high-amplitude
Ps conversions are not alone sufficient to establish an intermediate composition for the lower crust beneath the cratons, that feature coupled with low Poisson’s ratio favours a less mafic lower crust relative to that of the adjacent Proterozoic belts. The Ps signatures for the Proterozoic terranes or overprinted Archaean terranes, on the other hand, are characteristic of a complex and/or gradational Moho discontinuity, a significantly smaller velocity contrast across the Moho, or both. It has been suggested that such regions may reflect a history of underplating through successive or episodic intrusion of basaltic melts into the lowermost crust (e.g. Griffin & O’Reilly 1987), so that the M-discontinuity itself becomes a complex interlayering of mafic and ultramafic rocks. Such a model is consistent with the results from southern Africa. A mafic or mafic-to-ultramafic lower crust would also mean a smaller density contrast across the Moho, perhaps as little as 300 kg m\(^{-3}\) (Webb et al. 1999), helping to account for the apparently anomalous negative correlation between elevation and crustal thickness in southern Africa.

The evidence reported by Nguuri et al. (2001) for a thick crust and a relatively diffuse Moho in the Bushveld Province supports the argument for a broadly continuous mafic body at depth beneath the Bushveld Complex (Cawthorn et al. 1998; Webb et al. 2000; Cawthorn & Webb 2001). The substantial load of material added to the crust during the Bushveld event was interpreted by Webb and coworkers to have produced a downward crustal flexure of up to 6 km with a resultant crustal thickness of 45–50 km beneath stations located in the central Bushveld Province (Webb et al. 2000).

Conclusions and implications for formation of cratons

The results described above place a number of important constraints on the structure and composition of cratons and provide insights into the processes by which they formed. Several conclusions may be enumerated with some confidence from the results cited above.

(1) Re–Os studies of mantle nodules demonstrate that the craton keel is Archaean in age and that there is no correlation with depth in the keel. The keel, therefore, did not thicken over time. Many eclogite inclusions in diamond are the same age as depleted peridotite or even slightly older. The age data on eclogite inclusions, coupled with evidence from stable isotope measurements that the eclogites formed from seawater-altered basalts, indicate that subduction processes were probably important in Archaean time. Subduction processes have also been invoked from both laboratory and field studies for the formation of komatiites by wet melting, suggesting that subduction was active at least as far back as early Archaean time, c. 3.6 Ga. The paucity of eclogite in the xenolith samples, however, suggests that the descending plate in Archaean time was, as it is today, almost always subducted into the deeper mantle.

(2) Tomographic images of the upper mantle exhibit a clear correspondence between seismic structures and geological boundaries. The boundaries of the higher-velocity root structures that define the cratons are typically well correlated with surface geological contacts. Adjacent Proterozoic mobile belts are in almost all cases characterized by slightly lower mantle velocities. Deep keel structures within the cratons are irregular, with evidence for maximum keel depths of at least 250–300 km in the southern and apparently most undisturbed part of the Kaapvaal Craton and in regions of the Zimbabwe Craton. The contrast in velocity perturbation within cratonic regions is <1% and thus could be accounted for entirely by variations in mantle composition.

(3) The northern part of the Kaapvaal Craton was profoundly affected by the massive magmatic events associated with the emplacement of the 2.054 Ga Bushveld mafic igneous complex. The mantle beneath the larger Bushveld Province exhibits lower seismic velocities, particularly in the P-wave results. The seismic data, which suggest widespread compositional modification (metasomatic enrichment) of the mantle beneath the Bushveld Complex, are consistent with evidence of younger (reset) Re–Os model ages of mantle nodules brought up in the Bushveld region.

(4) The thickness of the crust in undisturbed Archaean terranes is typically 35–40 km, some 5–15 km less than that found for adjacent Proterozoic terranes and in disturbed craton. Perhaps the most remarkable result of the receiver function studies to date in southern Africa is the pervasive evidence for crustal thickening and Moho disruption of Archaean terranes associated with tectonomagmatic disturbances of Proterozoic age. The most significant example of crustal modification is in the broad region of the Bushveld and Okwa–Magondi terranes. Although the evidence is not conclusive, the relatively large amplitude of Ps conversions at the Moho beneath undisturbed craton may suggest lower seismic velocities in the lowermost crust. This observation, coupled with low Poisson's ratios, may in turn indicate less mafic compositions for the lower crust in Archaean terranes relative to those of Proterozoic age. Systematic
variations in lower-crustal composition may be one reason why crustal thickness across southern Africa does not correlate with elevation (Webb et al. 1999, 2000).

(5) The Limpopo Mobile Belt appears to be underlain by a mantle root of cratonic seismic character, consistent with its Archaean age. The Northern and Southern Marginal Zones, which are thrust belts overlying cratonic crust, exhibit typical cratonic Moho signatures. These marginal zones of the Limpopo Belt contrast markedly in terms of both crustal thickness and Moho structure with the Central Zone of the Limpopo Belt, which is a major collisional terrane deeply exhumed through uplift and erosion (Treloar et al. 1992). Here, despite at least 10–20 km of erosional unroofing, the crust remains anomalously thick, suggestive of a Himalayan-scale crustal thickening at the time of the Kaapvaal–Zimbabwe Craton collision in the Archaean eon. The Ps signal from the Moho is complex, as would be expected for a highly foreshortened crust, and could be due to mantle–crust interlayering, dipping layers, or a host of other structural complications in the lower crust.

(6) Geochronological studies demonstrate that a simple view of cratonic root formation (depleted peridotite) as a discrete event is not a viable model for craton formation. Crust and mantle have both experienced a multi-stage tectono-magmatic history. Measurements on lower-crustal xenoliths and exposed rocks of the Vredefort crustal section reveal an intricate record of lithospheric evolution clearly associated with events seen in the surface geology. Of considerable relevance are the late Archaean magmatic events (c. 3.1–2.6 Ga), which overprinted earlier ages in much of the lower crust over vast regions of the western part of the Kaapvaal Craton. Widespread magmatism and lower-crustal heating events seem to preclude the presence of an underlying cold tectospheric root that would act to shield the crust from thermal events (Schmitz et al. 1998; Schmitz & Bowring 2000; Moser et al. 2001). A possible implication is that the crust and mantle may share an allochthonous relationship and that stabilization of a tectospheric keel (cratonization) and crust–root fusion occurred rapidly in late Archaean time (Moser et al. 2001). Although many other lines of evidence support the independent formation of crust and tectospheric mantle of the present-day craton, there are caveats, including the possibility that significant volumes of magma can rise through cratonic mantle and intrude the crust without altering the original geotherm sufficiently to move it out of the diamond stability field, as observed in the Bushveld Province.

(7) Although the results reported here represent only an initial examination of a vast seismic dataset, there is little evidence to date in bodywave tomography, surface-wave inversion, or receiver function analysis of deep discontinuity structure for a low-velocity zone beneath the cratonic root. High-resolution seismic studies of the upper mantle are continuing, however, and more sophisticated analysis may yet reveal a velocity reversal at the base of the tectosphere.

Remarks on craton formation

The constraints summarized above have numerous and important ramifications for the formation and stabilization of cratons. It is now clear that at least three things must occur to form tectospheric crust and mantle in Archaean time: (1) continental crust of intermediate composition must be generated in substantial volume, probably through arc-like processes; (2) significant volumes of upper mantle must be thoroughly depleted of both their basaltic fraction and volatile content; (3) fragments of continental crust and blocks of buoyant depleted mantle must be accreted together in late Archaean time, with the refractory mantle agglomerated into a coherent tectosphere at least 250 km thick.

Two key features of the cratonic mantle are critical to understanding its formation: its very low geotherm relative to that in other parts of the upper mantle, and the absence of any radiogenic age progression in mantle xenoliths with depth. The Archaean age of many diamonds suggests that the cratonic geotherm has always been low, as has been postulated previously (Burke & Kidd 1978; Jordan 1988). The fact that the keel in any given region is the same age within error over the depth range from which nodules were derived confirms that the chemical boundary layer (CBL) that forms the tectospheric keel did not thicken by progressive cooling. By inference, then, any process advanced to explain the formation of tectospheric mantle must include a mechanism by which the resultant keel possessed a low geotherm from the time of its formation.

Jordan (1988) investigated much this same question in considerable detail. The conclusions he drew more than a decade ago from available seismic, geochemical, petrological and geological data remain viable today. Jordan concluded that the most plausible mechanism of keel formation involved subduction-zone depletion of mantle rocks and subsequent advective thickening through large-scale compressional (collisional) processes where the refractory residues are swept
together to form a thickened CBL beneath a consolidated Archaean crust. The model he proposed is a multi-stage one, in which basalt (or komatiite) (Boyd 1987) melt depletion and upper-mantle cooling precede cratonization. From the results of the Kaapvaal Project, it appears most plausible that the bulk of basaltic and komatiitic melt extraction and a considerable degree of mantle cooling must have occurred in supra-subduction zones, although mantle differentiation along mid-ocean ridges may also have been a factor. A corollary to the model of very wet subduction-zone melting for formation of komatiites and the depletion of the mantle is that mantle temperatures need have been only slightly higher in Archaean time than they are today. The relatively low temperature required to produce komatiite in wet subduction-zone melting (Parman et al. 2001) means both that komatiites are not necessarily evidence for a hot mantle in Archaean time (as suggested, for example, by Nisbet & Fowler (1983)) nor that they had an origin in Archaean hotspots (e.g. Abbott et al. 1994). (Our conclusion here is that mantle temperatures were not necessarily much higher in Archaean time than at the present day, perhaps c. 150°C. There are suggestions, however, based on thermal modeling that even in a hot Archaean mantle thermal blanketing could maintain a cool tectospheric root within the central zone of a cratonic volume substantially more massive than that remaining today (e.g. Ballard & Pollack 1988). The principal question remains in that case of how large cool cratonic masses stabilized in Archaean time in the presence of very hot mantle.) In light of our current understanding of cratonic structure and composition, Jordan's advective thickening model provides a satisfactory explanation for the geothermal gradient and the observed seismic mantle velocities.

Abbott (1991) faulted Jordan's advective thickening model, citing lack of geological evidence for an abundance of large-scale collisional terranes in the Archaean geological record. Thus, whereas the high-grade metamorphic gneiss terranes are consistent with crustal thickening and subsequent exhumation during major tectonic events, the more abundant low-grade greenstone belts provide no comparable evidence of either the crustal thickening or large-scale shortening characteristic of continental collisions. Although this remains a serious objection to the hypothesis of advective thickening via plate-scale compressional events, it may be that successions of smaller accretionary events, such as those found in the westward-stepping subduction zones of the northern Andes or in the allochthonous collage along the margin of western North America, played a larger role in the process of root thickening than is generally recognized.

Two principal alternatives to advective thickening for formation of the CBL have garnered varying degrees of support in the past (Abbott 1991). The first, the so-called 'plum pudding' model (Davies 1979), involves the gradual accretion of buoyant refractory material from the deep mantle to a thickening cratonic root. The model depends on conductive cooling, however, and can reasonably be excluded, mostly on the grounds that the keel exhibits no evidence of temporal growth vertically. The second model, which retains a strong following, is based on the notion of progressive imbrication or 'trench jumping' of subduction zones (Helmstaedt & Schulze 1989; Abbott 1991; de Wit et al. 1992) and the successive stacking of layers of oceanic lithosphere beneath the cratonic crust. Several variants of this mechanism have been put forward, two of which we discuss here. As envisioned by Helmstaedt & Schulze, continental lithosphere proto-plates progressively override continental margin subduction zones, shearing off the descending oceanic plate and leaving a trail of stacked and shallowly dipping remnants beneath the continent to form a root structure composed of imbricated slabs. Although little is certain about how subduction processes operated in Archaean time, the Helmstaedt and Schulze slab imbrication model finds no support in modern subduction analogues. Continents simply do not override subduction zones in the absence of external factors, such as collision of a continental mass with the trench. The trenchward motion of a continental plate is readily compensated by trench rollback, a common, dynamically predictable and unremarkable phenomenon (e.g. Forsyth & Uyeda 1975; Meijer 1995). Abbott (1991) proposed another model involving imbricated slabs, but suggested instead that the Archaean mantle was sufficiently hot to melt the basaltic crust of the descending plate before it reached the eclogite stability field. Under those conditions, the buoyant refractory (harzburgite) mantle of the oceanic lithosphere would survive to underplate the continent rather than sink under the gravitational effect of the basalt-to-eclogite conversion. Thus, the model postulates that successive layers of depleted, relatively cool, oceanic lithospheric slabs are stacked atop one another and at low angles to form a thickened root beneath the cratonic crust. A host of difficulties can be cited for this model, among them the virtual improbability that the cold oceanic crust would melt before its water had infiltrated and fluxed the hot overlying mantle (as in modern subduction zones). Full melting of the basaltic
crust creates an even more fundamental problem: it eliminates gravitational sinking induced by the basalt-to-eclogite transition as a driving force in subduction (Forsyth & Uyeda 1975). Without gravitational sinking the principal force that drives subduction goes to zero and subduction ceases. The issue is further muddied by the fact that the MOR basaltic crust may have been thicker in Archaean time (Bickle 1986), making wholesale crustal melting an even less viable hypothesis. Finally, it should be noted that for southern Africa it is estimated that <1% of the volume of the cratonic mantle is made up of eclogite (Schulze 1989).

We conclude by reiterating and re-emphasizing the fact that the above discussion leaves many fundamental questions unanswered. Among the outstanding issues touched on above but that remain yet to be fully resolved, we include the following.

1. Why and how was the Archaean different, and what does it mean in terms of lithosphere formation, plate processes and differentiation in the early Earth?

2. Is cratonization (i.e. mantle stabilization) separate from crust formation? How and over what time period did craton stabilization occur?

3. How do images of mantle heterogeneity and crustal composition and thickness relate to the assembly history of the cratons?

4. What is the influence of post-Archaean events on the composition, seismic structure, and geothermal gradient of the craton?

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References


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