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Construction and destruction of some North American cratons

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10 Nov 2016

Abstract. Construction histories of Archean cratons remain poorly understood; their destruction is even less clear because of its rarity, but metasomatic weakening is an essential precursor. By assembling geophysical and geochemical data in 3-D lithosphere models, a clearer understanding of the geometry of major structures within the Rae, Slave and Wyoming cratons of central North America is now possible. Little evidence exists of subducted slab-like geometries similar to modern oceanic lithosphere in these construction histories. Underthrusting and wedging of proto-continental lithosphere is inferred from multiple dipping discontinuities, emphasizing the role of lateral accretion. Archean continental building blocks may resemble the modern lithosphere of oceanic plateau, but they better match the sort of refractory crust expected to have formed at Archean ocean spreading centres. Radiometric dating of mantle xenoliths provides estimates of rock types and ages at depth beneath sparse kimberlite occurrences, and these ages can be correlated to surface rocks. The 3.6–2.6 Ga Rae, Slave and Wyoming cratons stabilized during a granitic bloom at 2.61–2.55 Ga. This stabilization probably represents the final differentiation of early crust into a relatively homogeneous, uniformly thin (35–42 km), tonalite-trondhjemite-granodiorite crust with pyroxenite layers near the Moho atop depleted lithospheric mantle. Peak thermo-tectonic events at 1.86–1.7 Ga broadly metasomatized, mineralized and recrystallized mantle and lower crustal rocks, apparently making mantle peridotite more ‘fertile’ and more conductive by introducing or concentrating sulfides or graphite at 80–120 km depths. This metasomatism may have also weakened the lithosphere or made it more susceptible to tectonic or chemical erosion. Late Cretaceous flattening of Farallon lithosphere that included the Shatsky Rise conjugate appears to have weakened, eroded and displaced the base of the Wyoming craton below 140–160 km. This process replaced the old re-fertilized continental mantle with relatively young depleted oceanic mantle.

1. Introduction

Deconstructing an object is often an excellent way to understand how it was built, but the construction, modification, destruction and reconstruction of the most ancient parts of the continents remain generally undetermined (Lee et al., 2011). Rocks forming parts of these cratonic nuclei of continents date back to the very early Earth and thus become entangled in debates about early Earth tectonics, particularly when plate tectonics as it is known today began. Here we will use examples from four relatively

well-studied cratons in North America (Figure 1) to inform a more general review of how cratons formed and how their modification occurred throughout Earth history.

1.1 Some Terms and Conditions

Mutual understanding among authors and readers concerning specific, specialized word usage is important for clarity of concepts. Here we adopt the terminology in common usage throughout much of the diamond exploration industry in that cratons are long-lived stable blocks and Archean in age (Clifford's Rule: Clifford, 1966; Gurney et al., 2010; Shirey and Shigley, 2013). Long-stable lithosphere composed of amalgamated cratons, sutured together by Proterozoic orogens, typically with a considerable thickness of overlying sedimentary rocks is here termed a shield (Hoffman, 1988, 2014; compare with composite craton of Aulbach et al., 2013) (Figure 1). By long-term stability we mean at least 1 Ga since the last major tectonic disruption, because origin mechanisms older than 1 Ga are debated (Eaton and Perry, 2013). Cratons generally have only thin sedimentary cover, implying modern surfaces that have risen or always been near sea level. Archean cratons are also characterized by thick keels of melt-depleted mantle lithosphere (Boyd, 1989) which provides greater compositional buoyancy than that of most Proterozoic continental lithosphere. This buoyancy and the rheological stiffness due to the dry nature of depleted mantle are key factors in ensuring the survival of these ancient keels (Sleep, 2003; Jordan and Paulsen, 2013; Wang et al., 2014, 2015). Isolated from the convecting interior, the keels cool as the thermal structure evolves toward an equilibrium where the reduced conductive heating of the thick lithosphere is balanced by radioactive heating (Michaut et al., 2009; Jaupart et al., 2014). Cooling provides additional strength, results in the growth of a thermal boundary at the base of the depleted lithosphere (Cooper et al., 2004; Cooper and Conrad, 2009; Lee et al., 2011), and increases net lithospheric density. The gross layering from depleted mantle to a basal thermal lithosphere to the convecting interior is

supported by studies of xenolith composition (Rudnick et al., 1998; Kopylova et al., 1999) and texture (Kennedy et al., 2002), and by the continental-scale seismic imaging of upper mantle anisotropy (Yuan et al., 2011).

The plate tectonic paradigm explains much of what we understand about the geology of the present day Earth. Subduction is the most important tectonic process operating today, but most geoscientists agree that it could not have operated in its current fashion on the earliest Earth. The geologic time and manner in which early Earth geodynamics evolved to become modern plate tectonics remain major contentious issues (e.g., Hamilton, 1998, 2015; Kamber, 2015). Systematic variations in combined hafnium and oxygen isotope ratios in zircons of different ages provide one estimate of the relative proportions of reworked versus new crust (Dhuime et al., 2012). These variations indicate that 65% of the present-day volume of continental crust had formed by 2.8 Ga after which a sharp increase in the proportion recycled occurred. Before that time, the generation rate of new basaltic crust was nearly constant so that ~73% of all basaltic crust that was forming at ~3.2 Ga was juvenile. In contrast, only 20% of crust forming at ~2.0 Ga was juvenile, and today it approaches 0%, implying all new basaltic crust is currently composed of recycled crust (Dhuime et al., 2012; Spencer et al., 2014). Recycling basaltic crust with distinctive hafnium and oxygen isotope ratio signatures requires sinking of some early crust in plastic or solid state through the underlying less dense mantle. We use this isotope evidence to infer that an increasing percentage of basaltic crust that formed since 2.8 Ga was subducted in the solid state in order to be recycled, and thus assume that modern subduction-driven plate tectonics began at 2.8 ± 0.5 Ga (Roberts and Spencer, 2015), in agreement with the diamond inclusion record (Shirey and Richardson, 2011). Such chronology implies that oceanic crust did not subduct before that time although crust did rejoin the mantle in a fluid state.

1.2 Outstanding timescale issues

We assume that the Archean was a time in Earth's history during which crust and mantle lithosphere differentiated but was not recycled deep into the mantle prior to 2.8 Ga (Solomatov and Stevenson, 1993). An inferred diachronous transition in convection state from sluggish lid (Sleep, 2009) to local down-welling of pyroxene-rich melts (Lee et al., 2011) to solid-state plate subduction probably occurred at 2.7–2.5 Ga (Table 1). With the initiation of plate tectonics, the uppermost solid-state surface layer was incorporated into convective downwellings for the first time. Early plate tectonics then began to recycle crust, volatiles and carbon which reset the ages of some mantle rock (e.g., Pearson and Wittig, 2014).

Formation of oceanic lithospheric plates occurred within regions of divergent flow because of enhanced differentiation and cooling. Such plates made possible the full development of modern plate tectonics as more wide-spread subduction of these plates occurred during the Proterozoic. Subduction along the margins of cratons delivered volatiles into mantle beneath the cratons, volatiles that in turn rose to metasomatize cratonic lower lithosphere. This introduced volatiles such as carbon into the lithosphere, now observable as regions of enhanced conductivity. Observations of intact lithospheric plates of Archean or Proterozoic age within the mantle would demonstrate the onset of plate tectonics, but such observations remain inconclusive. Any slabs (Figure 3), or older drips (Figure 2), not incorporated in the craton lithosphere have long since sunk completely through the mantle, therefore any evidence would need to come from the structures within the craton itself (Bao and Eaton, 2015). Slab geometries defined seismically (e.g., Cook and Erdmer, 2005; Sigloch, 2011) can neither be traced directly to dated outcrops nor constrained by accurate source location of representative xenoliths.

1.3 Summary of assumptions

The general tectonic history of cratons adopted here as background for describing
 130 specific North American examples is as follows:

Early (>2.7 Ga) proto-continental blocks roughly 100-km thick form above down-
 132 welling instabilities in an early near-stagnant lid. Building blocks, large and small,
 collide and wedge into each other to form roughly 225-km thick proto-cratons between
 134 ~ 3.0 and 2.6 Ga (Cooper and Conrad, 2009); proximal extension and contraction
 accompanies this amalgamation.

The cratons develop a quasi-stable basal thermal boundary layer that is prone to
 136 convective instability, and its presence further isolates the overlaying mantle thermally
 138 (Cooper et al., 2004; Lee et al., 2011). This additional cooling strengthens the craton.

The transition to modern-style plate tectonics begins at about 2.8 Ga and Proterozoic
 140 (2.0–1.8 Ga) subduction zones weld cratons into shields (e.g., the Siberian craton –
 Rosen et al., 1994) and flux the lower lithosphere of cratons with carbon and water that
 142 metasomatize them.

Flat-slab subduction and small-scale convective instabilities erode this metasomatically
 144 weakened, lower lithosphere (Doin et al., 1997; Lenardic et al., 2003; Jaupart et al., 2007;
 Bao et al., 2014; Wang et al., 2015; Currie and van Wijk, 2016).

Isopycnicity keeps Archean cratons buoyant and near sea level (Jordan, 1978; Eaton and
 146 Perry, 2013) whereas Proterozoic shields are more susceptible to subsidence of tens of
 148 kilometers.

2. Methods

150 Understanding lithospheric-scale tectonic processes requires information derived
 from depths across at least 50–700 km of the mantle. This depth range can only be
 152 studied in situ today using geophysical techniques. Some methods provide spatially

continuous or homogeneous coverage. Geochemical observations are available locally and selectively where made possible by rare volcanic samples that include xenoliths or xenocrysts such as diamonds (including some from as deep as the upper-most lower mantle; e.g., Harte et al., 1999; Stachel et al., 2005) and garnets, or where igneous rocks can be determined to have originated from mantle melts originating from great depths (Mitchell, 2008; Tappe et al., 2013). Of most value to probing the lithospheric mantle are the xenoliths of peridotite and eclogite sampled by kimberlites and other small degree melts (Nixon, 1987). It is possible to obtain equilibration depth, bulk chemical composition and age from such samples (see review by Pearson et al., 2003). However, occurrences of mantle xenoliths are small samples of large volumes of lithospheric mantle that provide “point-source” information. The challenge thus is to robustly extrapolate the sparse geochemically derived knowledge across sometimes vast distances using geophysically derived knowledge.

2.1 Seismology

Three types of seismic data analyses are used in this review: teleseismic tomography, teleseismic receiver functions, and imaging using a controlled source. Seismic tomography provides a low-resolution, continental scale overview of mantle structures (e.g., Priestley and McKenzie, 2006; Schaeffer and Lebedev, 2014). Surface-, S- and P-wave velocity models provide increasingly better resolution because of the inherent wavelength of the seismic wave used: 500–10000 km, 50–1000 km, and 10–80 km, respectively. However, source and receiver array shape and sparseness influence the uniformity and specific resolution of anomalies in velocity more strongly than do seismic wavelength (Schmandt and Humphreys, 2010). Robust models to depths near the 410 km discontinuity can be achieved with regional arrays.

The other two methods depend on the conversion or reflection of seismic waves at discontinuities in rock properties and thus provide relatively high-resolution definition of the structures causing these discontinuities. Resolution can be less than several kilometers, but becomes poorer with increasing depth (Miller and Eaton, 2010; Levander and Miller, 2012). For teleseismic arrivals, isolation of converted phases generated at interfaces and timed against the direct arrival, enables estimation of discontinuity depth and contrast beneath the receiver station in what is traditionally called receiver functions (Burdick and Langston, 1977; Bostock, 1998). Here P-to-S (Ps) conversions at multiple azimuths were used to estimate mantle structures in the remote Slave and Rae cratons (Snyder et al., 2012).

The third method uses man-made seismic sources to map structures, primarily within the crust, but also in the uppermost mantle. Both reflected and refracted P-waves are used. The Lithoprobe SNORCLE survey crossed the southwest margin of the Slave craton and the adjacent Wopmay Orogen (Cook et al., 1999). The SAREX- Deep Probe surveys crossed the Wyoming craton (Henstock et al., 1998; Gorman et al., 2002) and the CD-ROM survey interpreted structures across the Cheyenne Belt at the southern margin of the Wyoming craton (Henstock et al., 1998; Crosswhite and Humphreys, 2003; Karlson et al., 2005). The Lithoprobe Western Superior survey provides structural geometries within the crust of the Superior craton (White et al., 2003; Percival et al., 2006).

2.2 Magnetotellurics

Conductivity estimates for the entire cratonic lithosphere depend on high-quality magnetotelluric measurements (e.g., Chave and Jones 2012; Jones et al., 2014; Spratt et al., 2014). Recent surveys have enabled 3-D inversions of magnetotelluric observations to produce 3-D conductivity models of most parts of the Slave, Rae and Wyoming cratons (Snyder et al., 2014; Snyder et al., 2015; Megbel et al., 2014). Not all of these

surveys use the same observational bandwidths. Recent Canadian surveys use
 204 broadband sensors and recorders with a range of 0.01–1000 s (Spratt et al., 2014). Older
 Canadian surveys and USArray MT surveys used so-called long-period sensors and
 206 recorders with a range of 10– 10000 s (Jones et al., 2003; Meqbel et al., 2014). The latter
 surveys can sense conductors to 400 km where overlying lithosphere is relatively
 208 resistive, but suffer poorer sensitivity and resolution of conductors within the crust.

210 2.3 Xenoliths

Xenolith suites that erupted in kimberlites and similar deeply derived magmatic
 212 rocks provide the only physical samples of the lower crust and mantle lithosphere in
 the areas we consider. Fortunately, the Slave, Rae, Wyoming and Superior cratons have
 214 all been subjected to intense diamond exploration over the past few decades, revealing
 several hundred kimberlites. Among these, less than a dozen locations provide mantle
 216 xenolith suites that allow estimates of the mantle geotherm, measurements of bulk
 composition, and age estimation using isotopic methods. Estimates of the peridotite and
 218 eruption age, degree of melt depletion, and metasomatic history exist for each craton.
 The Slave craton has three relatively extensive mantle xenolith suites (at the Jericho,
 220 Diavik/Ekati and Gahcho Kué diamond mines). These cover a depth range of 40 to 210
 km (Kopylova et al., 1999; Kopylova and Caro, 2004; Mather, 2013; Snyder et al., 2014).
 222 The Rae craton has two locations discovered so far – Pelly Bay and Repulse Bay - that
 yield xenolith suites suitable for petrological/geochemical study. At both locations
 224 samples are derived from limited depth ranges, but represent lower crust and the
 mantle at 90–150 km depths (Petts et al., 2014; Liu et al., 2015). The Wyoming craton has
 226 two main xenolith sources, near the Colorado-Wyoming border (Kelsey Lake, Sloan and
 Leucite Hills sites) and in central Montana (Homestead, Williams and Bearpaw sites)
 228 with 70 to 200 km depths sampled (Carlson et al., 2004; Downes et al., 2004; Mirnejad
 and Bell, 2006; Schulze et al., 2008).

It should be noted that the volcanic rocks that supplied this suite of mantle and deep-crustal samples erupted over a >500 Ma age range, from the early Cambrian (Pelly Bay/Repulse Bay: Liu et al., 2015; Gahcho Kué: Kopylova and Caro, 2004) through the Jurassic (Jericho: Kopylova et al., 1999; Heaman et al., 2006), to the Tertiary (Ekati/Diavik: Lockhardt et al., 2004; Sarkar et al., 2015). Hence, the picture of the state of the lithosphere supplied by these xenolith suites can be displaced in time more than 500 Ma from the present-day geophysical observables.

3. Building cratons

3.1 Ephemeral isopycnicity and survival

A relatively constant freeboard of cratons has been used to argue for cratonic construction in which the final phase of construction was cool (Flowers et al., 2004). The evolution of craton freeboard is affected by the buoyancy difference between ocean and craton lithosphere, the fractional area of the Earth's surface occupied by continent, and the volume of ocean water. On billion-year time scales, the influences of a cooling asthenosphere have a primary importance, both as a direct result of the cooling and indirectly because the oceanic crustal thickness diminishes as asthenosphere cools.

A 200-km thick craton would have cooled by an average of about 250°K. This amount of cooling assumes a 200°K cooling of the asthenosphere since the Archean, current radioactive heating to be 35–40% what it was in the Archean (e.g., Herzberg et al., 2010). For this calculation we reference the Priestley and McKenzie (2006) xenolith-based geotherms from Jericho in the Slave craton and Udachanaya in the Siberia craton, and assume the current temperature at 50 km depth to be 200–250°K above that given by a simple conductive geotherm. This cooling would have contracted the craton by 1.7 km (assuming all contraction to have occurred in the vertical direction, and using an expansion coefficient of $3.4 \times 10^{-4}/^{\circ}\text{K}$). The corresponding average density increase is 28

kg/m³, which is minor (6 kg/m³ more than that experienced by the asthenosphere) and
 256 justifying an assumption of no horizontal flow.

A similar calculation for a 200-km column of Earth beneath ocean lithosphere has
 258 it contracting by 2.0 km, assuming a thickening of average ocean lithosphere from 50 to
 90 km. Thus, thermal arguments alone predict ~0.3 km of sea-level drop and also an
 260 insignificant change in the craton buoyancy relative to the asthenosphere.

An evolution in the compositional buoyancy has also occurred. Assuming the
 262 Archean ocean crust to be 20 km thick compared to a modern thickness of 6 km, the
 ocean floor would lie ~1.7 km above its current depth. Subsidence of 0.1 km is due to
 264 decreased mantle depletion (assumed mantle density = 3300 kg/m³, basalt density = 2920
 kg/m³). These effects occur at relatively shallow depths and do not affect the
 266 isopycnicity of the craton with respect to adjacent asthenosphere. The net sea-level drop
 calculated from the above thermal and compositional effects is 2 km.

Note that the net load beneath the continent or ocean has not changed, only the
 268 density distribution. Thus no large-scale mantle flow occurs; a different density
 270 distribution can create pressure differences between the ocean and the continent,
 potentially driving flow within the upper 200 km. Because we calculate little change in
 272 the density difference between craton and ocean below ~40 km, the net effect is small
 and we ignore this effect. Also note that we have not considered the load of ocean water
 274 although the volume change requires surface uplift or subsidence. If significant
 amounts of water can flow over (or drain from) the continent, this would be a change in
 276 load; we ignore this relatively minor effect.

If freeboard remained about constant, then our calculations suggest that a sea-
 278 level drop of ~2 km would have been offset by a combination of: (1) an initially cool
 craton that warmed with time (Flowers et al., 2004), (2) continental growth, and (3)
 280 cratonic metasomatism. Significant densification by metasomatism may have had a
 destabilizing effect on the craton, and may not have been possible. Recent

geodynamical modeling suggests that the contrast in strength between cratonic lithosphere and the surrounding less depleted and hence weaker mantle is the key factor for the longevity of cratonic keels, irrespective of their buoyancy (Wang et al., 2015). Finally, we note that since cratons are now isopycnic, they would have been buoyant in the past (Eaton and Perry, 2013).

3.2 Making early building blocks---drips and greenstones

Does a consistent time frame exist for this craton building process across most cratons? Table 1 summarizes key dates known for the specific North American cratons reviewed here as well as for a few key global analogues. Most ages indicated are concordant, but a few suggest early initiation in the Pilbara and Kaapvaal cratons (Van Kranendonk, 2010; Shirey and Richardson, 2011). The following time scale is inferred from these ages.

A near-stagnant lid developed at 3.4–2.7 Ga (Figure 2a). Fractional crystallization produced an olivine-rich mantle, pyroxene-rich lower crust and plagioclase-rich upper crust (Elkins-Tanton, 2012; Hamilton, 2015). With progressive lid cooling, near-stagnant tectonics became organized but sluggish (Sizova et al., 2010; Crowley, in press). The lid cooled and thickened, and local down-welling initiated (Lee and Chin, 2014) (Figure 2b). The down-welling drove horizontal flow within the stagnant lid and hence necking or thinning of the lid distant from the down-welling site (Fig 2c). The thinning enhanced mantle differentiation and resulted in a depleted, harzburgite-rich lowermost mantle lithosphere and formation of tonalite-trondhjemite-granodiorite (TTG) crust (Bédard, 2006). Greenstone belts developed on many future cratons at 2.73–2.58 Ga (Table 1). Continued down-welling caused horizontal shortening above the site, and thickening, uplift and erosion of crust resulted in turbidite basins being created by about 2.66 Ga (Figure 2c). By 2.64–2.58 Ga crust-mantle differentiation had progressed sufficiently that dense pyroxene-rich lower crust also dripped into the mantle (Figure

2d). Coeval peak metamorphic conditions melted the crust, and craton stabilization
 310 occurred in that the lithosphere structure to 180–210 km depths formed at that time and
 has survived to the present. Pyroxene-rich drips not trapped in the lithosphere would
 312 not be observable as such today, having been recycled into the asthenosphere.

A presence of active and passive up-flowing mantle probably delivered volatile-
 314 rich small-percentage melts to the base of the forming lithosphere (mechanical
 boundary layer of Tainton and McKenzie, 1994; Aulbach, 2012). These melts perhaps
 316 enhanced melting of the pyroxene-rich layer (Figure 2d), furthered differentiation, and
 promoted upward migration of more felsic components to the near surface to form TTG
 318 crust (Bédard and Harris, 2014). The result was a layered lithosphere about 120-km
 thick, similar to oceanic plateaus such as Ontong-Java today (Figure 3) (Ishikawa et al.,
 320 2011; Tommasi and Ishikawa, 2014). Some of these blocks accreted above zones of
 downwelling to form composite proto-cratons 200–250 km thick (Figure 3) (Snyder et
 322 al., 2014).

The products of this sequence are preserved on all cratons, with similar ages in
 324 all but a few cases (Table 1). Various cratons do preserve different fractions of each
 component: TTG-basement, greenstones, late granites, turbidite basins. Below we focus
 326 specifically on examples from the main North American cratons.

4. Constructing the cratons---tectonic wedging of peers

4.1 *Slave craton construction*

4.1.1 Crustal geology

330 The exposed Archean crust of the Slave craton is 500 km by 700 km, surrounded by
 Paleoproterozoic or younger orogens (Figure 4). An ancient gneiss complex (Central Slave
 332 Basement Complex of Bleeker et al. 1999) in the west hosts some of the oldest rocks on
 Earth, 4.04–2.83 Ga gneiss and granitoid rocks, that are overlain by the 2.8 Ga Central Slave

334 Cover Sequence of quartzite and banded iron formation (Table 1) (Bleeker, 2002). This
 complex comprises large antiforms overlain or flanked by 2.73–2.58 Ga basalt-dominated
 336 greenstones and 2.67 Ga turbiditic sedimentary belts of the Yellowknife Supergroup
 (Bleeker, 2003). The supracrustal and granitoid rocks are deformed by several discrete,
 338 intersecting fold sets to form an archetypical Archean “granite-greenstone terrain”. The
 oldest major folds trend NE-SW and, at 2.645–2.635 Ga, precede the abundant 2.635–2.605
 340 Ga tonalite-granodiorite plutons of the Defeat and Concession Suites (Davis et al., 2003;
 Table 1). The more evolved two-mica granites of the 2.59–2.58 Ga Prosperous Suite are
 342 coeval with peak metamorphism and craton-wide folding along N- to NW-trending axes
 (Davis and Bleeker, 1999). Gold deposits formed in the near-surface about 2.60–2.59 Ga
 344 (Siddorn et al., 2006). The associated high heat flux and craton-wide nature of this latter
 event appear to have been the last major phase of tectono-thermal activity before the crust
 346 and mantle experienced a long period of quiescence (Bleeker, 2002). Isotopic studies of
 large intrusive bodies show Nd and Pb isotope compositions that define a rough north-
 348 south trend in age; the 4.04–2.83 Ga ancient nucleus of the Central Slave Basement
 Complex (Bleeker et al. 1999) lies to the west, whereas isotopically juvenile basement rocks
 350 lie to the east (Davis et al., 2003).

Following the last major phase of tectono-thermal activity at 2.58 Ga, Slave crust was
 352 gradually exhumed until 2.00 Ga by 10–15 km of slow uplift and erosion as evidenced by
 Paleoproterozoic cover sequences that unconformably overlie neighboring Archean
 354 basement (Hoffman, 1980). Sediment-rich Paleoproterozoic sequences such as the
 Coronation Supergroup on the western margin of the craton probably formed on a passive
 356 margin. Numerous mafic dyke swarms within the craton may date rifting at 2.23–2.00 Ga
 (Bleeker, 2003). Regionally reconstructed oblique convergence of the Slave and Rae cratons
 358 during the Paleoproterozoic (2.0–1.8 Ga) assembly of NUNA (Hoffman, 2014) is consistent
 with the evolution of local structures observed along the southern margin of the Slave
 360 craton, e.g., the Great Slave Lake shear zone (Figure 4) (Hanmer et al., 1992). This common

boundary with the Rae craton is defined by the contact between Slave craton granitoids
 362 with granitoids and ortho- and para-gneisses of the Taltson and Thelon magmatic tectonic
 zone.

364 4.1.2 Mantle structure

366 The inferred Archean mantle of the Slave craton at 100 km depth extends laterally
 over 800 km by 1200 km (Figure 1). An ENE-WSW striking zonation within the mantle was
 368 first identified using population statistics of individual garnet crystals recovered from
 surface till samples taken downstream of glacier ice flow from kimberlites (Grütter et al.,
 370 1999). Geochemical analysis of these crystals revealed a zone of anomalously high
 concentrations of G10 garnets centered roughly on the Lac de Gras area where independent
 372 geochemical observations identified an ultra-depleted, olivine-rich layer at 70–145 km
 depth separated from a deeper, metasomatically re-enriched (reduced) layer by an
 374 unusually sharp transition (Griffin et al., 1999; Aulbach, 2012). Magnetotelluric
 observations indicated a prominent conductor at 80–100 km depths also centered on the
 376 Lac de Gras region (Jones et al., 2003). Structural and geochemical analyses of xenoliths
 from kimberlites indicate different depths for the base of the lithosphere at the time of
 378 kimberlite eruption (Kopylova and Caro, 2004). Estimates range from 214 ± 15 km in the
 north to 236 ± 6 km in the south (Mather et al., 2011), but it must be kept in mind that these
 380 estimates are made from kimberlites of disparate eruption ages and so small variations in
 depth could reflect the erosive effects of small-scale convective instabilities on the base of
 382 the lithosphere (Cooper and Conrad, 2009).

Three major mantle lithosphere discontinuity surfaces within the Slave craton have
 384 been mapped using multi-azimuthal receiver functions. The Moho at depths of 39 ± 4 km
 and mid-lithosphere discontinuities at depths of 140–155 km appear relatively flat
 386 regionally (Figure 5) (Snyder et al., 2014). A feature at 80–110 km depths found beneath the
 Lac de Gras kimberlite field, may connect to the west with underthrust lithospheric slabs

associated with the Hottah terrane within the Wopmay Orogen (Ootes et al., 2015). Because the mid-lithosphere discontinuity appears primarily on the transverse component of multi-azimuthal receiver functions, its implied seismic anisotropic behaviour will produce polarity flips at back azimuths coinciding with fast polarization directions (e.g., Wirth and Long, 2014). This characteristic makes changes in properties of the over- and underlying layers easier to observe than if only subtle changes in compositions caused the discontinuity. Within the southeast Slave craton such changes in discontinuity polarity are observed from northwest to southeast (Snyder et al., 2014). A polarity flip between the Diavik and Gacho Kué diamond mines correlates both with where xenocrysts from kimberlites show increasing eclogite/peridotite protolith ratios and where the boundary with the more juvenile eastern terrane expressed at crustal levels (Figure 6) (Snyder and Grütter, 2010; Helmstaedt and Pehrsson, 2012; Kopylova et al., 2016). In contrast, no polarity flip is observed along a profile between Diavik and the Snap Lake diamond mine, indicating that the uppermost mantle beneath both these locations is probably associated with the ancient Central Slave Basement Complex (Figure 4).

Comparison of diverse data sets for the Slave craton reveals that this mid-lithospheric discontinuity coincides with a vertical gradient in surface wave velocity, a relatively sharp change in seismic anisotropy, and with changes in rock types (harzburgite dominated to lherzolite dominated) as determined from xenoliths and xenocrysts (Figure 6) (Snyder et al., 2014). The 'Lac de Gras' discontinuity dips to the southeast between 85 and 110 km depths and lies 20–30 km deeper than the top of the broader zone of enhanced conductivity (Figure 7). Beneath the westernmost Slave craton, a subsidiary mantle discontinuity associated with the mapped Hottah terrane dips northeast, whereas the main western discontinuity surface dips eastward. Beneath Yellowknife, the Hottah surface is at 70 km depth and coincides with the top of Bostock's (1998) anisotropic ($\pm 5\%$) H discontinuity, which was interpreted as a 10-km-thick subducted oceanic crust. The main surface coincides with Bostock's (1998) X discontinuity at 135 km depth. Cook and Erdmer

(2005) used seismic reflection profiling to interpret the H discontinuity as separating a
 416 wedge-shaped westernmost Slave lithosphere from the over- and underthrust Hottah
 terrane. The same authors propose that the X discontinuity separates a wedge of Hottah
 418 terrane from subducted Proterozoic Ft. Simpson oceanic lithosphere (Figure 6).

Re-Os isotopic dating of central Slave craton mantle xenoliths and xenocrysts
 420 suggests the presence of some depleted ancient (ca. 3.1 Ga) harzburgite (Irvine et al., 2003;
 Aulbach et al., 2013), but a much greater abundance of 2.75 Ga peridotites (Heaman and
 422 Pearson, 2010; Mather, 2013). To the north, spinel peridotites are the oldest (ca. 2.9 Ga)
 xenoliths and lithosphere age decreases with depth; subducted eclogite (2.2-2.0 Ga) and
 424 ages of 1.8 and 1.3 Ga are interpreted as due to mantle metasomatism related to the
 Wopmay orogen and Mackenzie igneous event, respectively (Schmidberger et al., 2007;
 426 Heaman et al., 2006; Heaman and Pearson, 2010; Kopylova et al., 2016). More recent
 passage of subducted slabs beneath the base of the craton may also have induced
 428 metasomatism and diamond formation (Weiss et al., 2015). Changes in fabric orientation
 across seismic discontinuities and velocity gradients at 140–150 km depths suggested two
 430 ancient layers of peridotite with distinct uniform fabrics (Figure 6) (Snyder et al., 2014),
 whereas conductivity (Jones et al., 2003) and rare earth element trends (Aulbach et al., 2013)
 432 support overprinting of the layered structures by several phases of metasomatism.

434 4.1.3 Summary

Available isotope dates indicate that the oldest Slave lithosphere (both crust and
 436 mantle) has a consistent age (>2.75 Ga) to depths of 100–140 km. In the south and east, a
 more juvenile Archean terrane was wedged on top of and below the more ancient block. In
 438 the north and west, metamorphism or metasomatism has overprinted the mantle below 100
 km with younger isotopic signatures, including partial replacement by Palaeoproterozoic
 440 lithosphere to 250 km depth (Figure 6). Everywhere deep within the central Slave craton,

reducing channelized metasomatism by melts has locally refertilized peridotite, enhancing
 442 conductivity and diamond prospectivity.

444 4.2 *Rae craton construction*

4.2.1 Crustal geology

446 Exposed Archean crust of the Rae craton is recognized about 2000 km along its
 primary southwest-northeast axis, with a circa 750 km breadth (Figure 8). The craton
 448 has a number of component blocks: Committee Bay, Repulse Bay, central Baffin, Queen
 Maud, SW Rae, Chesterfield and Cumberland (Hoffman, 1988; St-Onge et al., 2006;
 450 Berman et al., 2007, 2013; Eglington et al., 2013). Relatively juvenile Archean crust with
 2.74–2.69 Ga komatiite-quartzite sequences (Committee Bay, Woodburn Lake, Ege Bay
 452 greenstones) distinguishes the central part (Sanborn-Barrie et al., 2014). Major adjacent
 crustal terranes including Hearne, Meta Incognita, and Sugluk are here considered
 454 lesser terranes that have Archean crust but <150 km of lithospheric thickness (Snyder et
 al., 2012, 2015).

456 The central Rae craton is characterized by Neoarchean, amphibolite to granulite-
 grade, tonalitic to granitic orthogneisses as well as the distinctive northeast-striking
 458 2.730–2.687 Ga komatiite-bearing greenstone belts (Skulski et al., 2003; Sanborn-Barrie et
 al., 2014). These rocks are intruded by 2.72–2.68 Ga tonalite and voluminous 2.63–2.58
 460 Ga monzogranitic plutons (Hinchey et al., 2011 and references therein); Nd model ages
 of 2.87–2.52 Ga suggest predominantly juvenile isotopic source regions for these crustal
 462 rocks (Peterson et al., 2010). These plutons are recognized from northwestern
 Saskatchewan as far east as the Melville Peninsula (LaFlamme et al., 2014) and mark
 464 Rae craton stabilization (Berman et al., 2013).

The more ancient Repulse Bay block includes the Melville Peninsula and
 466 Southampton Island (Figure 8) and is characterized by 2.97 Ga volcanic and

metasedimentary rocks of the Prince Albert greenstone belt, intruded by 2.63–2.58 Ga
 468 monzogranitic plutons (LaFlamme et al., 2014). The plutons are locally metamorphosed
 to amphibolite or granulite grade and have 3.63–2.84 Ga Nd model ages. Igneous cores
 470 of zircon crystals from lower crust xenoliths have ages of 3.5–2.8 Ga (Petts et al., 2014;
 Eglington et al., 2013). The Repulse Bay block is separated from the Committee Bay
 472 block by a set of cryptic structures that include ancient northeast-southwest trending,
 steeply dipping structures (Sanborn-Barrie et al., 2014) that are apparently offset by
 474 neotectonic faults such as the Wager Bay shear zone (Snyder et al., 2015).

The Chesterfield block of the Rae craton (Berman et al., 2013) and central Hearne
 476 domain appear distinct from each other because of isotopically juvenile, 2.7 Ga mafic to
 intermediate volcanic rocks of the central Hearne supracrustal belt that are cut by 2.66
 478 Ga granitic plutons associated with a regional metamorphic event (Davis et al., 2006).
 The Hearne domain lacks any Archean magmatic or tectonic activity after 2.66 Ga. The
 480 Chesterfield block has 2.7 Ga supracrustal rocks and tonalite plutons. The block is
 intruded by 2.6 Ga granitic plutons (Davis et al., 2006) that tie it to the Rae craton after
 482 its ca. 2.64 Ga accretion (Berman et al., 2007). The Chesterfield fault zone separates the
 Chesterfield block from the Committee Bay and Repulse Bay blocks (Figure 8). The fault
 484 reactivated during the 2.56–2.50 Ga MacQuoid orogeny (Hanmer et al., 2006; Berman et
 al., 2013) and again when microcontinents accreted to the southeastern flank of the Rae
 486 craton at 1.9–1.87 Ga, the latter collectively designated as the Snowbird phase of the
 Hudsonian orogeny (Berman et al., 2007). The Chesterfield fault zone is a northwest-
 488 vergent thrust that juxtaposes ~10 kbar granulite facies rocks with 1.9 Ga hornblende
 cooling ages in the hanging wall against ~4 kbar, lower amphibolite-facies rocks with
 490 1.8 Ga cooling ages in the footwall (Berman et al., 2013; Pehrsson et al., 2013).
 Amphibolite-grade remnants of 2.1–1.9 Ga Paleoproterozoic sedimentary cover occur
 492 throughout the central Rae craton (Rainbird et al., 2010).

4.2.2 Mantle structure

Mantle xenolith samples are currently only available from two locations, one in each of the Repulse Bay and Committee Bay blocks (Figure 8). Based on PGE patterns and Os model ages, the peridotites from both localities can be categorized into three age groups. Archean ages of 2.8–2.6 Ga characterize Committee Bay mantle and 3.0–2.7 Ga Repulse Bay mantle (Liu et al., 2015). Paleoproterozoic ages of 2.1–1.7 Ga are observed from both locations, as are “Recent” (<1 Ga) model ages that are similar to the ca. 546 Ma kimberlite eruption age. As within the Slave craton, the Archean ages indicate that depleted Archean lithospheric mantle formed coevally with the overlying Archean crustal basement. The subtle difference in Os model ages between Committee Bay and Repulse Bay correlates with the age difference between overlying crustal basement and with assembly by collision of separate Archean blocks at 2.7–2.6 Ga.

Rae craton peridotite xenoliths have been variably altered by serpentinization, silicification and carbonation. Nonetheless, their high absolute PGE concentrations, mantle-like melt-depleted PGE patterns and low bulk rock Al_2O_3 contents indicate that these secondary processes, as well as mantle metasomatism recorded in pristine silicate minerals, did not significantly modify the relative compositions (Liu et al., 2015). The observed depletion in Pd and Pt, bulk Al_2O_3 contents of less than 1.5% as well as the depleted compositions of the silicate and oxide minerals all indicate that the peridotites are residues of considerable melt depletion. One group of peridotites clearly indicate extensive melt depletion in the Neoarchean. Re-PGE characteristics of the Paleoproterozoic age group can be interpreted as recording primary melt depletion, but at lower extents of melting than occurred within the Archean age group. Re-PGE characteristics are not consistent with Paleoproterozoic ages reset by melt refertilization of the Archean peridotites because higher Re enrichments would be expected (Liu et al., 2015). These Paleoproterozoic melt depletion ages may represent juvenile lithospheric

mantle that partially replaced lower portions of Archean lithospheric mantle (Liu et al.,
 520 2015) during the 1.77–1.70 Ga Nueltn event (Table 1) (Petersen et al. 2002).

The vertical distribution of the three age clusters in the central Rae lithospheric
 522 mantle is consistent with the observation of seismic lithospheric discontinuities (Snyder
 et al., 2012, 2015) that indicate two or more layers of fossil lithospheric mantle fabric
 524 beneath the central Rae craton. Seismic discontinuities from multi-azimuthal receiver
 functions map gently dipping discontinuities that project upward into major linear
 526 zones of deformation (Snyder et al., 2015). Most deformations are Paleoproterozoic (1.8
 Ga), but some rocks also record older deformation (Sanborn-Barrie, et al., 2014). The
 528 southeastward dip of the most prominent discontinuities indicate that lesser or more
 juvenile continental blocks such as the Hearne, Sugluk and Meta-Incognita were thrust
 530 northwestward onto the flank of the central Rae craton (Figures 9 and 10), reaching
 their current position following major horizontal shortening strain associated with the
 532 Trans-Hudson Orogeny (St-Onge et al., 2006). A shallow mantle lithosphere layer with
 predominately Archean depletion ages underlain by a layer of mixed Archean and
 534 Paleoproterozoic ages is indicated by joint interpretation of geophysical, geological and
 geochronological observations. Post-melting metasomatic enrichment is clearly
 536 indicated by complex zoning within garnets and sinuous garnet REE patterns (Liu et al.,
 2015) that may have generated the enhanced conductivity regions evident throughout
 538 the central Slave mantle at 100–200 km depths.

4.2.3 Summary

540 Within its central part, the Rae craton grew by successive accretion of terranes
 around an ancient nucleus that is today represented by the Repulse Bay block. Note that
 542 we do not here discuss the southern or northern parts of the rae craton due to limited
 data and observations. Accretion added the Committee Bay block by 2.7 Ga, the
 544 Chesterfield block by 2.64 Ga, the Hearne and other inter-Superior blocks by 1.9 Ga.

Proterozoic alterations included the possible replacement of the lower (100-200 km) lithosphere along gently east-dipping discontinuities and the metasomatism that enhanced conductivity at depths greater than 100 km.

4.3 Wyoming craton construction

4.3.1 Crustal geology

The exposed Archean basement rocks associated with the Wyoming craton occur within an area 700 by 500 km (Figure 11, Kilian et al., 2016). The oldest rocks come from its northern Bighorn block where 3.5–3.1 Ga tonalitic-granitic gneisses occur in and near the Beartooth Mountains (Mueller and Frost, 2006; Krogh et al., 2011). Isotopic evidence indicates that an ancient nucleus block with components >3.0 Ga lies between the Great Falls tectonic zone and Laramie Mountains (Chamberlain et al., 2003). Magmatism at 2.9–2.75 Ga produced tonalite-trondjemite plutons which evolved to high-potassium granites (Barker et al., 1979; Frost et al., 2006). The 3.30 Ga Sacawee Orthogneiss of this ancient northern TTG block is capped by the 2.85 Ga Barlow Gap cover sequence (quartzite, pelitic schist, iron formation) (Table 1). The Bighorn block is bounded by the Oregon Trail structure at its southern margin, and suffered little internal tectonism or magmatism after 2.75 Ga, with the exception of the 2.705 Ga Stillwater complex intrusion, (Chamberlain et al., 2003; Grace et al., 2006). Three discrete pulses of magmatism occurred in the southern Sierra Madre part of the craton at 2.72–2.67, 2.63–2.60, and 2.55–2.50 Ga; the pulses were interpreted as evidence of an accretionary margin (Frost et al., 1998). Three periods of greenstone basin development indicates that southward-migrating tectonism preceded each pluton bloom by 20–40 Ma; a large percentage of these greenstone belts contain meta-turbidites, but no komatiites (Houston et al., 1993).

Deformation is strongest along the northwestern and southern margins of the Wyoming craton (Figure 11). In the northwest, shear zones are dated at 2.56–2.45 Ga (Mogk et al., 1992). In the south, generally north-vergent thrusting occurred at 2.83, 2.70–2.67 and 2.62 Ga within the diffuse Oregon Trail structure (Figure 11). This structure separates the >2.7 Ga Sierra Madre block from the more ancient Bighorn block across a <2.65 Ga south-dipping crustal accretion zone (Chamberlain et al., 2003; Grace et al., 2006). Rock ages throughout this Sierra Madre block and eastward to the Black Hills (Figure 11) are predominantly Proterozoic (2.0–1.5 Ga) and therefore this block is not considered part of the Wyoming craton (Kilian et al., 2016). These Proterozoic ages date the docking of the Yavapai terrane along the 1.78–1.62 Ga Cheyenne Belt. The southern Wyoming craton was also modified by two periods of extension and mafic dyke swarm intrusion at 2.1–2.0 and 1.5–1.4 Ga (Snyder et al., 1990), that resulted in crust thinner by 5–10 km beneath the Sierra Madre block (Figure 12) (Gorman et al., 2002). In the north, the Great Falls tectonic zone was active at 1.85 Ga; the docking of the Medicine Hat block occurred at 1.76–1.72 Ga, and was coeval with final closure of eastern Wyoming and Superior cratons along the Dakotan orogen (Chamberlain et al., 2003; Kilian et al., 2016).

Seismic surveys indicate that unusually fast seismic wave speeds characterize the lower crust of the Wyoming craton and neighboring Medicine Hat block (Gorman et al., 2002). This layer is not found beneath upper crust younger than 2.75 Ga. Lower crustal xenoliths from the Medicine Hat block have 1.8–1.7 Ga ages (Davis et al., 1995), coeval with deformation along the Great Falls tectonic zone. Few xenoliths directly sample the lower crust within the Wyoming craton so these fast wave speeds may be speculated to derive from a pyroxenite-rich layer inherited from crust formation that was perhaps enhanced by mafic intrusions associated with the 2.705 Ga Stillwater complex that intruded shallower in the crust.

4.3.2 Mantle structure

Several xenolith suites help constrain the composition and history of mantle associated with the Wyoming craton (Figures 11 and 12). Homestead suite peridotites from 110–150 km depths represent the only clear sampling of the Wyoming craton mantle, and define a geotherm (60 mWm^{-2} ; Irving et al., 2003) that is significantly hotter than geotherms typical of cratonic mantle. The peridotite suite is 92% harzburgites with some G10 garnets and veins of clinopyroxene, websterite and phlogopite (Carlson et al., 2004). Peridotites from the Homestead kimberlite appear to have experienced melt depletion in the Archean (2.7–2.5 Ga TRD ages), with some samples recording isotopic evidence of metasomatic modification in Palaeo- to Mesoproterozoic times (Carlson et al., 2004). The nearby Williams peridotite xenolith suite erupted near the Great Falls tectonic zone (Figure 11). Although their Re-Os isotope systematics are complex, ‘low temperature’ xenoliths from 100–140 km depth have Re-depletion ages of 2.5–1.7 Ga indicative of a common Archean melt depletion history followed by Mesoproterozoic modification (Carlson et al., 1999). A 130–170 km deep ‘high temperature’ suite produces temperatures similar to those found at Homestead (Carter, 2004), but they yield Re-Os ages of <200 Ma (Carlson et al., 1999, 2004). Farther north and within the Medicine Hat block, the Bearpaw suite has Proterozoic (1.8–1.2 Ga Sm-Nd and Lu-Hf model ages) phlogopite-rich websterites, interpreted to represent metasomatism of the cratonic root (Downes et al., 2004). The Kelsey Lake and Sloan peridotite suites erupted ~600 km to the south, near the Cheyenne Belt (Figure 11). The Kelsey Lake mantle suite has Archean harzburgites with peridotitic diamonds typical of cratonic keels (Schulze et al., 2008). The Sloan mantle suite shows combined Nd-Hf-Os isotopic evidence of having been formed by melt depletion in the Mesoproterozoic (Carlson et al., 2004). The Sloan mantle suite also has eclogites and Fe-rich Proterozoic peridotites, typical of sutures and that appear to confirm (Schulze et al., 2008) an earlier model of a

southward dipping boundary between depleted Archean peridotite and overlying enriched Proterozoic peridotites with websterite veins (Eggler et al., 1988). The Leucite Hills suite from within the Sierra Madre block has 2.8 Ga depleted peridotites that were extensively metasomatized at >1 Ga by veins of phlogopite, richterite and clinopyroxene (Mirnejad and Bell, 2006).

The regional controlled-source seismic surveys SAREX, Deep-Profile and CD-ROM (Figure 11) revealed dipping seismic continuities within the crust and uppermost mantle (Henstock et al., 1998; Gorman et al., 2002; Karlstrom et al., 2005). Within the crust, seismic discontinuities are interpreted to dip away from the Wyoming craton as part of tectonic wedges whereas within the mantle, discontinuities dip beneath the craton. The overall geometry is of a tectonic wedge with its tip near the Moho. This conflicts with the xenolith interpretations because isotopic constraints from mantle xenoliths from the Kelsey Lake and Sloan kimberlites indicate a Proterozoic over Archean southward dipping boundary (Eggler et al., 1988) at the time of their eruption in the Devonian (Figure 12). As will be discussed further, this apparent relationship in the mantle may have been reversed in the Tertiary.

Recent teleseismic and magnetotelluric studies using USArray data provide deeper structural information within the upper mantle. Common conversion point stacking of Ps and Sp phases provide regionally consistent estimates of Moho depth and laterally consistent phases that indicate velocity decreases with depth at 75 and 125 km (Figure 13E and F) (Levander and Miller, 2012; Hopper et al., 2014). These deeper discontinuities were interpreted as the base of the lithosphere by the authors, but probably represent intra- or mid-lithosphere discontinuities related to mantle layering (Figure 12). Steeply dipping structures inferred from strong conductors associated with the Great Falls tectonic zone dip northwestward to about 80 km depth (Meqbel et al., 2014) (Figure 13D), a geometry consistent with an extensive, undeformed Wyoming craton keel such as characterized by depleted harzburgite xenoliths found at the

Williams and Homestead locales (Figure 12). Structure beneath the Cheyenne Belt is complex and possibly results from a southward-pointing crustal wedge above a northward-pointing mantle wedge (Karlstrom et al., 2005). Magnetotelluric conductors are consistent with a very steeply southeast dipping shear zone to about 200 km depth (Figure 13D).

4.4 *Western Superior craton*

4.4.1 Crustal geology that requires plate tectonics?

The western part of the Superior craton is cited as a type location for accretion of ribbon continents above subduction zones (Figure 14) (Percival et al., 2006).

Amalgamation of the western Superior craton is inferred from the similarity of modern geochemical arc signatures to its Archean calc-alkaline volcanic and plutonic rocks (Wyman and Kerrich, 2009). Geochemical signatures of many tonalite-trondhjemite-granodiorites indicate residual garnet and Nb-Ta-compatible titanate phases, but associated Archean basalts display characteristics of crustal assimilation and fractionation hybrids between crustal melts and alkali basalts (Bédard et al., 2013; Barnes and Van Kranendonk, 2014). The scale, volcanic facies, and geochemical systematics of Archean calc-alkaline suites also differ from those of modern arcs (Hamilton, 1998; Bédard, 2006; Bédard et al., 2013). Instead of the melting of subducted Archean oceanic crust, thick basaltic crust subcreted beneath a drifting continent's leading edge (Figure 3) could melt to create TTG-like melts geochemically similar to those formed during shallow subduction (Smithies et al., 2003). One difference is that a subducting slab is heated from above by forced convection in the mantle wedge so that only young slow-moving slabs can melt (van Keken et al., 2002). However, the predicted yield of TTG within such a model is inadequate to explain the observed volumes (Bédard, 2006). The alternative subcretion of a basaltic plateau-type crust formed in a near-stagnant lid regime could produce the observed larger volumes of

TTG as discrete events, assuming a combination of radioactive heating, thermal
 676 conduction from hot upwelling mantle, and magmatic underplating (Smithies et al.,
 2003; Foley et al., 2003; Bédard et al., 2013).

678 Dipping seismic reflectors on Lithoprobe sections previously interpreted as fossil
 subduction zones (Figure 15) (Calvert et al., 1995; White et al., 2003) may instead mark
 680 where crustal material was underthrust and subcreted, in a north to south sequence, to
 the leading edge of the drifting proto-craton (Bedard and Harris, 2014). This alternative
 682 interpretation assumes that the Northern Superior cratonic block (Figure 14) began
 drifting south at 2.72 Ga and that dipping crustal structures record tectonic imbrication
 684 ahead and beneath the drifting continental block rather than serial subduction zones.
 Late Neoproterozoic metasedimentary belts (e.g., Quetico and English River belts) record
 686 transient uplift and erosion during collision, followed by post-orogenic collapse
 (Calvert et al., 2004).

688 4.4.2 Mantle geology

The mantle xenolith record from the Superior craton is very sparse, especially
 690 considering the large size of the craton. The only extensive studies performed so far are
 both on kimberlites from the Attawapiskat region (Scully et al., 2004; Smit et al.,
 692 (2014a&b). Peridotites erupted by the Victor and Delta kimberlites have high Mg# (up
 to 93.1) indicating the presence of highly depleted dunite-harzburgite mantle stabilized
 694 in the Eo-Mesoarchean (3.6 Ga: Smit et al., 2014a). This age is older than any crustal age
 in the North Caribou Superterrane but broadly coincident with crustal ages (3.8– 3.5 Ga)
 696 at Assean Lake in Manitoba and the Inukjuak domain in Quebec (Böhm et al., 2000;
 David et al., 2009). Neoproterozoic mantle depletion ages are recorded by other similarly
 698 depleted peridotites, that may have formed synchronous with the crustal accretion and
 differentiation processes described above. Low-Mg eclogites at Victor have been shown

to have oceanic crustal precursors that were subsequently emplaced into the mantle (Smit et al., 2014a), likely during 2.7 Ga accretion of the western Superior craton.

Of particular note is that although the Victor and Delta kimberlites are located well to the North of the midcontinent rift, they record a prominent 1.1 Ga event in their Re-Os isotope systematics that indicates significant modification by that major Mesoproterozoic event (Smit et al., 2014a). In contrast, the geotherm defined by peridotites and xenocrysts from these locations reveals a mantle that had thermally equilibrated to a typical mantle geotherm by the Jurassic, with a total lithospheric thickness of circa 200 km. In contrast, mantle xenocrysts erupted by the 1.1 Ga Kyle Lake kimberlite sample a much shallower lithosphere (~150 km), probably as the result of thermal disturbance by the Mid-continent rift (Smit et al., 2014b). The implied generation of new diamonds at mid-lithosphere levels after the thermal pulse associated with the Midcontinent Rift had subsided is further substantiated by studies of diamonds by Smit et al. (2014b) who found much lower nitrogen aggregation states and a distinct carbon isotopic signature for diamonds in Jurassic versus Mesoproterozoic Attawapiskat area kimberlites.

Lithoprobe seismic surveys were interpreted to show shallowly dipping thrust-accretion surfaces in the lower crust (Figure 15) (White et al., 2003) and more steeply dipping features in the mantle to 250 km depths without clear lateral or structural continuity between the two (Percival et al., 2006). No seismic attributes resemble those of modern subduction zones such as northern Cascadia (Nicholson et al., 2005).

5. Constructing the Canadian Shield---Proterozoic plate tectonics

After Archean lithosphere became stabilized it probably had a thermal boundary layer greater than 120 km thick that was compositionally layered with depleted

peridotite at the bottom grading upward into a pyroxenite-rich layer and plagioclase-rich TTG layer. These buoyant, strong cratons circulated on the surface of the Earth and interacted with similar cratons and smaller, weaker continental blocks. Lateral motion was driven by continuing sites of down-welling that gradually evolved into what we today recognize as subduction zones when thinner, more rigid lithosphere formed in divergence zones or rifts and began to cool, densify and sink. Where this oceanic lithosphere subducted along the margins of an established craton, cratonic detritus and volatiles began to recycle into the interior and eventually re-incorporated into the bases of cratons that are now 180–250 km thick (Dhuime et al., 2012). The melts and volatiles derived from the subduction zones locally metasomatized the lithosphere and generally enhanced its conductivity by introducing carbon-rich fluids that also enhanced the probability of eclogitic diamond formation (Shirey et al., 2013).

The Slave craton was affected in this way. The 1.88–1.84 Wopmay orogen along its western margin (Figure 4) includes a dipping reflective surface within the uppermost mantle observed by the Lithoprobe transect in the south (Cook et al., 1999) and by teleseismic studies along its shared border with the Slave craton (Bostock, 1998; Snyder et al., 2014). The Great Bear arc associated with this orogeny is recognized as a volcanic arc, but its implication for subduction polarity remains uncertain (Hoffman, 1980; Jackson et al., 2013; Ootes et al., 2015, 2016). Widespread metasomatism and recrystallization of the mantle occurred at this time in the central and northern Slave (Heaman and Pearson, 2010; Aulbach et al., 2013), but less so in the south (Kopylova and Caro, 2004; Kopylova et al., 2016). The quality of peridotitic diamonds is observed to increase to the south, suggesting that Proterozoic metasomatism may have reset rock ages and also degraded ancient diamonds within the Slave craton whilst introducing newer eclogitic ones (Helmstaedt, 2009; Gurney et al., 2010).

The mutual borderland of the Rae and Superior cratons is occupied by the 1.9–1.87 Ga Trans-Hudson orogen (Figures 1 and 8)(Corrigan et al., 2009). At mantle depths of 100–200 km the cratons may interact (Snyder et al., 2012). At the surface a collage of terranes with evolving names (Hearne, northern Hearne, Sugluk, MetaIncognita), a micro-craton (Sask), and two batholiths (Wathaman and Cumberland) separate these cratons by several hundred kilometers (Ross et al., 2000; Snyder et al., 2012; Pehrsson et al., 2013). Neither long, linear batholiths/plutons are observed, nor foreland basins (but see St-Onge et al., 2006 for discussion). Lithoprobe seismic reflection profiling did not reveal dipping reflectors or other uncontested evidence of subducted oceanic plates along the Trans-Hudson orogeny (Lucas et al., 1994), leading instead to interpretation of the Hearne terrane as a collisional plateau entrapped in a cratonic vise (Ross et al., 2000). Lower crustal and mantle ages throughout the Rae craton lithosphere reset at 1.9–1.87 Ga (Petts et al., 2014; Liu et al., 2015) attest to the widespread thermo-mechanical effect of this collisional orogeny. Enhanced conductivity at 100–120 km depths is also attributed to metasomatism at this time (Snyder et al., 2015), metasomatism that is clearly recorded in mantle xenoliths (Liu et al., 2015).

Proterozoic-age deformation also affected much if not all of the Wyoming craton. The most thoroughly documented deformation is associated with the 1.78–1.62 Ga Cheyenne Belt along its southern margin (Figure 11) (Karlstrom et al., 2005). Xenolith suites and seismic surveys indicate that both Archean and Proterozoic rock wedges occur beneath the surface trace of this belt, but considerable disagreement continues as to whether the structural boundary generally dips north or south and whether it includes a coherent slab (Figure 12) (Eggler et al., 1988; Karlstrom et al., 2005; Schultze et al., 2008). Recent 3-D conductivity models suggest a near-vertical boundary (Figure 13D) (Megbel et al., 2014). The 1.81–1.71 Ga Great Falls tectonic zone separates the northern margin of the Wyoming craton from the Medicine Hat block of similar age,

776 and is generally thought to dip northward (Condit et al., 2015; Megbel et al., 2014). The
 eastern margin of the Wyoming craton is less distinct, but the Black Hills (Figure 11)
 778 have ages of 1.9–1.7 Ga (Chamberlain et al., 2003). The most significant metasomatic
 activity in the craton’s early history, prior to Laramide disruption, is documented by
 780 well-defined 1.8–1.7 Ga U-Pb ages from monazites within highly metasomatised mantle
 assemblages (Carlson and Irving, 1994; Rudnick et al., 1999), which can be clearly linked
 782 to the above crustal events. The overall resistive mantle lithosphere of the Wyoming
 craton suggests that this metasomatism was localized and not widespread within the
 784 mantle (Figure 13D) (Megbel et al., 2014).

6. Destroying cratons: the Wyoming craton and the Farallon slab

786 Seismic tomography models indicate mantle with abnormally fast seismic wave
 speeds beneath the Wyoming craton to depths of 300 km (Schmandt and Humphreys,
 788 2010; Humphreys et al., 2015). In map view, the elongate shape to the anomalous
 velocities at depth does not match the outline of the Wyoming craton at the surface
 790 (Figure 11). The volume of relative fast seismic wave speeds, typical of cratons and
 shields, is most obvious on depth slices of shear wave models at 80–160-km (Figures 1
 792 and 11). In surface wave models, it appears as a southwest-projecting ‘peninsula’ to the
 North American shield bounded to the northwest by prominent slow wave-speeds
 794 associated with the Snake River Plain (Shen et al., 2013). Body-wave tomography
 (Schmandt and Humphreys, 2010) shows a similar association of fast P-wave speeds,
 796 but with more detail and lateral variability. In all seismic models, the western part of
 the Wyoming craton, now occupied by the Basin and Range and Snake River Plain, is
 798 underlain by anomalously slow wave speeds compared to other cratons.

Xenoliths and isostatic arguments suggest that Wyoming craton mantle is
 800 depleted of basalt (Carlson and Irving, 1994; Carlson et al., 2004; Schultze et al., 2008),

but that its deep part is not typical Archean craton. Xenolith suites suggest that the base
 802 of the original Archean mantle was eroded from 180–200 up to 140–150 km depth at
 some time between the Devonian and Eocene (Humphreys et al., 2015), and that the
 804 deeper lithospheric mantle today is younger than 200 Ma (Figure 12) (Carlson et al.,
 1999, 2004). The same xenolith suites from the Wyoming craton region discussed
 806 previously help to understand its mantle replacement. The Kelsey Lake and Sloan
 xenoliths from the Cheyenne Belt (Figure 11) were erupted during the Devonian and
 808 show no indication of Phanerozoic ages (Schulze et al., 2008). Three xenoliths from the
 Williams kimberlite erupted from ~150 km depth at 48 Ma, have Os model ages reset to
 810 380–260 Ma, and a single pyroxenite has been dated at 67–57 Ma (Carlson and Irving,
 1994). Peridotites from the nearby 54–50 Ma Bearpaw minettes retain some Proterozoic
 812 signatures but were heavily metasomatized by water-rich fluids during the Cretaceous
 and Tertiary; their chemistry suggests that potassic parent magmas reacted with
 814 phlogopite-rich veins (Downes et al., 2004). This metasomatism is assumed to have
 weakened the mantle peridotite and made it more susceptible to tectonic erosion from
 816 below (as discussed by Li et al., 2008, in the context of the Colorado Plateau). The
 essential effect of metasomatic weakening on otherwise highly robust cratonic mantle
 818 was shown by recent geodynamical modeling (Wang et al., 2015). In contrast, modeled
 plume impact cannot erode strong anhydrous cratonic mantle without pre-cursor
 820 hydrous metasomatism that weakens rocks of the stable root. Abundant evidence of
 hydrous metasomatism of the Wyoming craton root exists, as outlined above.

822 The base of the Archean Wyoming craton was probably removed and replaced
 by newly accreted Jurassic oceanic mantle about 75–65 Ma in a three-stage process
 824 associated with the Sevier–Laramide orogeny (Humphreys et al., 2015). Flat-slab
 subduction and its consequent release of fluid into the overlying mantle first eroded 40–
 826 50 km from the base of the Archean Wyoming craton lithosphere. Basalt-depleted ocean

plateau mantle lithosphere of the Shatsky Rise conjugate part of the Farallon slab was
 828 then emplaced during the early Laramide orogeny. Stratigraphically recorded vertical
 movement in this region suggests that the plateau's dense oceanic crust diffused into
 830 the underlying asthenosphere at 70–75 Ma. Initiation of Colorado Mineral Belt
 magmatism about this time may mark a slab rupture in the Farallon slab along the
 832 margin of the Shatsky Rise conjugate through which the dense crust descended and
 buoyant mineralizing fluids rose (Chapin, 2012; Humphreys et al., 2015).

834 7. North American Summary

Construction histories of Archean cratons remain poorly understood; their
 836 destruction is even less clear because of, by definition, its rarity, but metasomatic
 weakening is an essential precursor (Foley, 2008; Wang et al., 2015). By assembling
 838 geophysical and geochemical data in 3-D lithospheric models, a clearer understanding
 of the geometry of major structures within the Rae, Slave, Wyoming and western
 840 Superior cratons of central North America is now possible. Little evidence exists of
 subducted slab-like geometries similar to modern oceanic lithosphere in these
 842 construction histories. Underthrusting and wedging of proto-continental lithosphere is
 inferred from multiple dipping discontinuities, emphasizing the role of lateral accretion.
 844 Archean continental building blocks may resemble the modern lithosphere of the
 Ontong-Java-Hikurangi oceanic plateau in geometry, size and shape. Compositionally,
 846 such blocks best match either the highly depleted peridotites found in Phanerozoic arc-
 mantle settings (Canil, 2004; Pearson and Wittig, 2008) or the sort of refractory
 848 peridotite that would be predicted to form at Archean ocean spreading centres with hot
 mantle potential temperatures that initiate melting deeper and create thicker oceanic
 850 lithosphere (Herzberg and Rudnick, 2012; Pearson and Wittig, 2014). Whatever the
 mode of formation, refractory mantle formed the proto-cores of cratons that thickened
 852 by lateral accretion. Radiometric dating of mantle xenoliths provides estimates of rock

types and ages at depth beneath sparse kimberlite occurrences, and these ages can be correlated to surface rocks. The 3.6–2.6 Ga Rae, Slave and Wyoming cratons comprise smaller continental terranes that stabilized during a granitic bloom at 2.61–2.55 Ga. This stabilization probably represents the final differentiation of early crust into a relatively homogeneous, uniformly thin (35–42 km), tonalite-trondhjemite-granodiorite crust with pyroxenite layers near the Moho atop depleted lithospheric mantle 100 to 200 km thick. A long period of passive tectonics, diabase dyke intrusion and failed rifting followed. Peak thermo-tectonic events at 1.86–1.7 Ga broadly metasomatized, mineralized and recrystallized mantle and lower crustal rocks, apparently making mantle peridotite more ‘fertile’ in the lower sections of the cratonic roots and enhancing conductivity by introducing or concentrating sulfides or graphite throughout the lithosphere at 80–120 km depths. This metasomatism also weakened the lithosphere or made it more susceptible to tectonic or chemical erosion. The arrival of the subducted Shatsky Rise conjugate part of the Farallon plate at the Wyoming craton at 65–75 Ma appears to have completely eroded and displaced the thus weakened base of the craton below 140–160 km. This process replaced the old re-fertilized continental mantle with relatively young depleted oceanic mantle.

8. A global perspective

Lithospheric structure discussed here for the four North American cratons provides interesting contrasts among each other, but also with structure observed elsewhere in the world (Table 1). A few will be briefly highlighted here.

The Yilgarn craton of western Australia is recognized to be constructed by amalgamation of at least two superterrane that can be defined by their distinct ϵ_{Nd} signatures (Blewett et al., 2010 and references therein). The individual histories of the component terranes have similar relative timing intervals to those of North American cratons already discussed, but also indicate two distinct construction phases (Table 1).

This may be due either to their membership in a different supercontinent grouping of
 880 cratons or else to an earlier phase of continent formation at 2.9 Ga (van Kranendonk et
 al., 2013).

882 The Slave and Rae cratons appear to have undergone widespread Proterozoic
 metasomatism whereas the central Wyoming craton did not, based largely on its less
 884 extensive enhanced mantle conductivity. The heavily silicified and metasomatized
 mantle of the Kaapvaal craton and the largely pristine dunite/harzburgite of the North
 886 Atlantic (western and eastern Greenland) craton provide a similar contrast where they
 have been much more thoroughly studied as of extensive diamond exploration. Simon
 888 et al. (2007) describe the processes that led to the magmaphile element depletion of the
 Kaapvaal cratonic lithospheric mantle and its subsequent re-enrichment in Si and
 890 incompatible trace elements. Most samples are characterized by Archean Os model ages
 and a decoupling between Nd and Hf isotope ratios, features most consistent with
 892 initial melting at 3.2 Ga preceding metasomatism by hydrous fluids and further melting
 to produce a harzburgitic residue. Fractional melting models suggest either 30%
 894 melting in the spinel field or else 20% melting in the garnet field followed by 20%
 spinel-facies melting. Garnet Nd–Hf isotope characteristics similarly indicate
 896 metasomatic trace element enrichment during melting followed by metasomatism of
 the Kaapvaal upper mantle, probably during amalgamation of smaller pre-existing
 898 terranes in the Late Archaean (2.9 Ga).

Thin lithosphere beneath the eastern North China Craton (NCC) is now widely
 900 recognized, similar to the Archean Wyoming craton, but the mechanism and timing of
 the thinning are more contentious (Menzies et al., 2007). Peridotitic xenoliths from the
 902 Cretaceous (100 Ma) Fuxin basalts at the northern edge of the craton (Zheng et al., 2007)
 can be subdivided into three types, representative of other xenolith suites across the
 904 craton. The dominant (Type 1) lherzolites with olivine Mg# 90 represent fertile mantle
 (5–12% partial-melt extraction) that makes up much of the Late Mesozoic–Cenozoic

lithosphere beneath the craton. Type 2 consists of harzburgites (olivine Mg# >92) interpreted as shallow relics of the Archean cratonic mantle. Type 3, minor lherzolite xenoliths with olivine Mg# 86 reflect the interaction of the lithosphere with magmas. The regional synthesis suggests that Mesozoic–Cenozoic lithospheric thinning and mantle replacement was heterogeneously distributed across the North China Craton in space and time. Lateral spreading of the lithosphere, accompanied by asthenospheric upwelling and melt–peridotite interaction, is the most probable mechanism for lithospheric thinning beneath the eastern part of the craton. Subsequent cooling of the upwelled asthenosphere caused some re-thickening of the lithosphere; this overall more fertile and hence denser lithosphere resulted in widespread basin formation.

Rayleigh wave tomography of the central and western NCC resulted in a S-wave velocity model that shows two major low-velocity zones (Jiang et al., 2013). These two zones extend to depths greater than 200 km and coincide with sites of Cenozoic magmatism and the narrow trans-tensional zone of the Cenozoic Shanxi–Shaanxi Rift, respectively. The Ordos Block in the western NCC is associated with high-velocity anomalies, interpreted as intact cratonic mantle root, but with a discernible low-velocity layer at depths of 100–150 km due to metasomatism. Most workers conclude that the current highly heterogeneous lithospheric structure results from multiphase reworking of pre-existing mechanically weak zones since the amalgamation of the craton, but no lithosphere replacement via flat-slab subduction is hypothesized.

9. Discussion

What role do cratons play in subsequent plate tectonic geodynamics? It could be argued that cratons are key to preserving entire continents because cratons and shields form a central part of every continent today and during past supercontinent assemblies (Pearson and Wittig, 2008; Griffin et al., 2009). The North American continent has a uniquely large shield because east of the Rockies the entire continent is underpinned by

932 Archean-Proterozoic mantle keel. This large shield emphasizes the geodynamic roles
 cratons can play in perturbing net rotation or causing slab flattening via the craton's
 934 effect on mantle convection (O'Driscoll et al., 2012).

Cratonic roots in North America and indeed worldwide appear to have
 936 maximum depths between 200 and 250 km, as indicated by both seismic and mantle
 geotherm constraints (Priestley and McKenzie, 2006; Jordan and Paulsen, 2013). The
 938 thermally and mechanically robust nature of cratonic mantle roots preserves the record
 of melting from an early and protracted part of Earth's history – the Archean. This
 940 reservoir, today comprising about 2% of the volume of the upper mantle (Pearson and
 Nowell, 2002), generally exhibits extreme depletion in incompatible components,
 942 including water. As such it played a small role in subsequent melt generation on Earth,
 other than to act as a "lid" that controlled the depth of melting. Only the most deep-
 944 seated, volatile-rich, small-volume melts managed to pierce this canopy, bringing with
 them a unique archive of the root. While the prolonged isolation of cratonic mantle
 946 from the homogenizing effects of mantle convection led to them accumulating some of
 the most extreme isotopic signatures measured in Earth's reservoirs (e.g., Carlson and
 948 Irving, 1994; Pearson et al., 1995), the extremely low abundances of Hf, Nd and Sr in the
 peridotites mean that they contribute little to the isotopic signatures in the rest of the
 950 mantle. Only where the lithosphere becomes metasomatised can it be classified as "re-
 enriched" in terms of its magma-generation capacity. Such metasomatised lithosphere
 952 in turn becomes the most vulnerable to removal via thermo-mechanical erosion.

10. Conclusions

954 Each North American craton was affected by each of four main processes to
 some degree. These include the building of basic continental cratonic blocks during the
 956 Archean, assembly of cratons into the North American shield during the Proterozoic,

coeval or subsequent weakening by metasomatism, and partial erosion or delamination
 958 of the lithospheric base.

Initial ancient (4.0-2.8 Ga) continental lithosphere blocks formed via fractional
 960 differentiation of an early Earth semi-stagnant lid into plagioclase-, pyroxene- and
 olivine-rich layers. These continental nuclei, several hundred kilometers wide and 90–
 962 120-km thick, grew by lateral tectonic accretion of similar, but more juvenile blocks.
 Seismic observations document wedge-shaped discontinuity surfaces that
 964 accommodated horizontal shortening, but no clear seismic evidence currently exists of
 surviving pyroxene-rich drips or deeply subducted lithospheric slabs from this period.
 966 Isotopic evidence does indicate coeval recycling of near-surface rocks into sub-
 lithospheric mantle, possibly via pyroxene-rich drips. These composite blocks cooled
 968 sufficiently by about 2.6 Ga to possess the strength and buoyancy to survive subsequent
 collisions and become stabilized cratons.

Once subduction started about 2.8 Ga, Archean cratonic blocks eventually
 interacted and collided along mostly Proterozoic orogenic belts to form a North
 972 American continental shield that has largely survived to the present. Today this shield
 has layered lithosphere that generally gets younger with depth and is 180–220 km thick.
 974 Seismic discontinuities beneath many Proterozoic orogenic belts document subduction
 of intact (oceanic?) lithospheric slabs as is observed in modern subduction zones.

Sub-cratonic lithosphere is pervasively metasomatized and locally melted or
 recrystallized numerous times wherever it has been studied to date via xenoliths or as
 978 inferred from enhanced conductivity. Upwardly migrating small percentage melt
 intrusions apparently introduce pyroxene-garnet (eclogite) assemblages, often with
 980 associated diamonds. Metasomatic fluids are often reducing and significantly weaken
 the lithosphere. These fluids are apparently rich in carbonates, silica or brine and
 982 therefore widely enhance conductivity as shallow as 90–120 km depths.

Old, stacked, weakened sub-cratonic lithosphere is variably eroded or
 984 underplated by asthenospheric convection. Sparse indicators of lithosphere thickness
 during the Phanerozoic suggest thicknesses of 150–220 km. Subsidence or uplift
 986 indicated by surface basins is modest, only a few kilometers. One prominent exception
 is the Wyoming craton beneath which the Farallon flat slab was subducted. This process
 988 removed the Archean lithospheric base below about 140 km and replaced it with
 Mesozoic oceanic lithosphere.

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Acknowledgments

992 B. Kjarsgaard and J. Bedard provided comments on an early draft. This collaboration
 grew out of a Meierjürgen Fellowship awarded to DBS at the University of Oregon. GH
 994 acknowledges support from the CIDER program (EAR-1135452) and from National
 Science Foundation grant EAR-0952194. DGP acknowledges funding from the Canada
 996 Excellence Research Chairs Program. This is contribution #20150111 to Natural
 Resource Canada's Open Geoscience Program.

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- 1498

Figure captions

1500 Figure 1. Map of the North America shield as outlined at 90-km depth by positive
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 1502 Lebedev, 2014). The four principal Archean cratons of North America are outlined at 90
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1512 Figure 3. Schematic cross section (a) showing amalgamation of Archean cratons into
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 1514 thick, and similar in composition to oceanic plateaus (b) today (modified from Tommasi
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 1516 low-angle shear zones (MLD). Note presumed absence of thin oceanic lithosphere.

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 1522 sections of Figure 6.

Figure 5. Two north-looking perspectives of a lithospheric-scale 3-D model of Slave
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 1530 seismic anisotropy, but also coincides with a marked increase in surface wave velocity
 (Yuan et al., 2011). Picks for two other discontinuities are represented by yellow and
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 1536 Central Slave nucleus grew tectonically outward by wedging apart first the accreting
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 1538 1.85 Ga. WF and MDF are the Wopmay and MacDonald strike-slip faults. Numbers are
 rock ages in Ga (Heaman and Pearson, 2010). The vertical columns labelled J, D, and
 1540 QG are xenolith suite rock types from Jericho, Diavik/Ekati, and Gahcho Que,
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 1542 town of Kugaruk. Dashed lines mark seismic discontinuities. LAB is the lithosphere-
 asthenosphere boundary inferred from geotherms (Mather et al., 2011). H, X, and L are
 1544 discontinuities as interpreted by Bostock (1998) and discussed in the text.

Figure 7. Lithospheric-scale 3-D conductivity model of Slave craton (modified from
 1546 Snyder et al., 2014). View is from the south. Parts of the Hottah and LDG surfaces are
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 1548 enhanced by the extensive reducing metasomatism that affected the lithosphere at

depths greater than about 100-km. In a few locations (e.g., D) equally conductive
 1550 regions continue upward into the crust, mostly near known kimberlite fields or
 mineralized zones. Yellow dots, J, D, and GQ are as in Figure 5. Tiny red dots are other
 1552 discontinuity picks.

Figure 8. Simplified basement geology map of the Rae craton (modified from Berman et
 1554 al., 2013). Dashed lines locate sections of Figures 6 and 10. C is the Committee Bay
 block, QM the Queen block, CB the Cumberland batholith, Cb the Chesterfield block, M
 1556 the Melville Peninsula, MI the MetaIncognita terrane, Rb the Repulse Bay block, and
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Figure 9. Lithospheric-scale 3-D model of Rae craton structures, viewed from the
 1558 northwest. Numerous surfaces are determined from seismic discontinuities in Ps
 receiver functions, most project upward toward major crustal structures such as the
 1560 Central Structural Domain (CSD) or Chantrey fault (Ch) (Snyder et al., 2015). QM
 1562 indicates a mantle surface that projects up to the Queen Maud block; Mel projects up to
 the central Melville Peninsula.

Figure 10. Summary cross sections of the Rae craton. Dashed lines indicate seismic
 1564 discontinuity surfaces here interpreted as thrust faults that thickened the Rae
 lithosphere. Solid black lines are faults interpreted from surface structures. Numbers
 1566 are ages in Ga; italics indicate tectonic events. Late metasomatism at depth is inferred
 from enhanced conductivity. Ch is the Chantrey fault, CSD is the Central Structural
 1568 Domain, QM is the Queen Maud block, and CB is the Cumberland batholith. Q is the
 1570 Qilaluqaq kimberlite with its xenolith suite rock and lithosphere base (LAB)
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Figure 11. (a) Location map of the Wyoming craton with topography as background.
 1572 Circles labeled BP, H, W, LH, K, S are xenolith suite locations Bearpaw, Homestead,

1574 Williams, Leucite Hills, Kelsey Lake and Sloan, respectively. Heavy dashed lines mark
 major deformation zones. Lighter dashed line shows location of cross sections in Figure
 1576 12. Dotted line marks a regional seismic traverse SAREX/Deep Probe (Gorman et al.,
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 1578 al., 2016), green line outlines interpreted limits of the Shatsky conjugate part of the
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 1580 surface (b) and S-waves (c) (modified from Humphreys et al., 2015).

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 1582 Wyoming craton nucleus (Bighorn block) grew tectonically outward by wedging apart
 first the accreting Sierra Madre block at 2.62 Ga, and then the Green Mountain belt of
 1584 the Yavapai terrane at 1.75 Ga. Numbers are rock ages in Ga (Chamberlain et al., 2003).
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 Schultze et al., 2008). CMB is the Colorado Mineral Belt; THO is the Trans-Hudson
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Figure 13. Lithospheric-scale 3-D seismic tomography (A-C), conductivity (D), and Sp
 1590 receiver function models of the Wyoming craton region (modified from Levander and
 Miller, 2012; Megbel et al., 2014; Humphreys et al., 2015), presented here as cross
 1592 sections. Locations are shown on 160-km depth slice of surface wave model. CB is
 Cheyenne belt, GFTZ is Great Falls tectonic zone, MHB is the Archean Medicine Hat
 1594 block. Red triangle locates the Yellowstone caldera.

Figure 14. Simplified basement geology map of the western Superior craton (modified
 1596 from Percival et al., 2006). Dashed lines locate sections of Figure 15.

Figure 15. (a) Crustal model of western Superior craton structures as defined by seismic
 1598 reflections (modified from White et al., 2003). (b) Summary cross section of the western

Superior craton. Ages are in Ga (modified from Percival et al., 2006). The most ancient
1600 block, the Northern Superior, grew southward by the wedging apart of and subsequent
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1602 the English River and Quentico metasedimentary belts.

Table 1. Comparative radiometric ages of some cratons

AGES (Ga)	Slave craton ¹	Rae craton ²	Wyoming craton ³	Superior craton ⁴	Yilgarn craton ⁵	Kaapvaal craton ⁶
Basement gneiss	3.5-3.0 (4.03 Acasta) 3.4-2.7 ReOs	3.6-2.6 3.0-2.5 ReOs	3.5-3.1 Bighorn block 3.3-3.1 TTGs	3.9-2.67 TTG	3.3-2.67 TTG	3.55-3.2 TTG
Cover sequence	2.8 C. Slave C.S.	2.97 Prince Albert Gp.	2.853 Barlow Gap			2.68-2.43 Black Reef (Transvaal)
Juvenile basement	2.75 East Slave	2.7-2.66 Hearne	2.65-2.60 Sierra Madre			
Greenstones	2.73-2.58 Yellowknife Gp.	2.74-2.69 Committee Bay	(2.701 Stillwater) 2.55-2.45 mafic dykes	2.75-2.695	2.96-2.93 2.825-2.7 Murchison 2.72-2.655 E Goldfields	2.710 Ventersdorp
Turbidite basins	2.66 Burwash		2.73-2.67 in south	2.69-2.685 Porcupine	2.665-2.65 "late basins"	2.642-2.432 Prieska
Granites or dykes	2.635-2.615 Defeat 2.61-2.605 Concession 2.59-2.58 Prosperous (2 mica)	2.72-2.68 2.63-2.58 juvenile	2.72-2.67 2.63-2.60 2.55-2.50 (2 mica)	2.719-2.67 TTG 2.684- 2.677 "late"	2.67-2.55 High-Ca	
Peak metamorphism	2.60-2.59 Au	2.5-2.3 in mantle	2.55-2.45	2.673 Au 2.645	2.65-2.63 Au	3.23
Major deformations	2.645-2.635 D2 1.88-1.84 Wopmay	2.56-2.5 MacQuoid 2.1-1.9 Pilings Gp. 1.9-1.87 Hudsonian 1.87-1.77 Nuelton	2.65-2.60 Oregon Trail 2.56 1.78-1.62 Cheyenne Belt	<2.6	2.67 shortening 2,72-2.66 doming	2.47 Limpopo

¹Bleeker, 2002; Helmstaedt and Pehrson, 2012

²Berman et al., 2013

³Chamberlain et al., 2003

⁴Percival et al., 2006; 2012

⁵Blewett et al. 2010; Van Kranendonk et al., 2013

⁶Shirey et al., 2013

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