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Review Article

The structural evolution of the deep continental lithosphere

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A B S T R A C T

Continental lithosphere houses the oldest and thickest regions of the Earth's surface. Locked within this deep and ancient rock record lies invaluable information about the dynamics that has shaped and continue to shape the planet. Much of that history has been dominated by the forces of plate tectonics which has repeatedly assembled super continents together and torn them apart - the Wilson Cycle. While the younger regions of continental lithosphere have been subject to deformation driven by plate tectonics, it is less clear whether the ancient, stable cores formed and evolved from similar processes. New insight into continental formation and evolution has come from remarkable views of deeper lithospheric structure using enhanced seismic imaging techniques and the increase in large volumes of broadband data. Some of the most compelling observations are that the continental lithosphere has a broad range in thicknesses (<100 to >300 km), has complex internal structure, and that the thickest portion appears to be riddled with seismic discontinuities at depths between ~80 and ~130 km. These internal structural features have been interpreted as remnants of lithospheric formation during Earth's early history. If they are remnants, then we can attempt to investigate the structure present in the deep lithosphere to piece together information about early Earth dynamics much as is done closer to the surface. This would help delineate between the differing models describing the dynamics of craton formation, particularly whether they formed in the era of modern plate tectonics, a transitional mobile-lid tectonic regime, or are the last fragments of an early, stagnant-lid planet. Our review paper (re)introduces readers to the conceptual definitions of the lithosphere and the complex nature of the upper boundary layer, then moves on to discuss techniques and recent seismological observations of the continental lithosphere. We then review geodynamic models and hypotheses for the formation of the continental lithosphere through time and implications for the formation and preservation of deep structure. These are contrasted with the dynamical picture of modern day continental growth during lateral accretion of juvenile crust with reference to examples from the Australian Tasmanides and the Alaskan accretionary margin.

Contents

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1. Introduction

Despite serving as home to human civilization, the continental lithosphere is arguably less well understood than the inaccessible oceanic lithosphere. This is because, while the first-order features of the oceanic floor (heat flux, bathymetry and thickness) can be directly attributed to the sea-floor age through plate-tectonics, that same theory fails to provide an equivalent context for continents.

Due to research initiatives such as EarthScope and LITHOPROBE (Clowes et al., 1999), along with greater computational power, there is now an unprecedented view of the sub-crustal interior of continental and cratonic lithosphere. Most of this increased coverage comes from a variety of seismic observations, which by their nature, provide not an exact replication of continental lithosphere, but a glimpse of properties that influence seismic parameters. As such, interpreting this expanded data requires not only a critical analysis as to what would cause variations in seismic properties, but also what geodynamic processes could produce the conditions required to produce an anomaly.

What does this new glimpse into the interior of continents tell us about past conditions and the history of deformation? How do we connect the new observations to other data sets? And more broadly, can we turn this internal gaze into a larger view of the process of building continents?

Global perspectives of the Earth's lithosphere highlight commonalities and differences that can help guide our interpretations of the controls on lithospheric structure and evolution. Fig. 1 shows two different global decompositions of the lithosphere that illustrate the first order domains using (A) the top-down analysis of Crust1.0 from all relevant geological and geophysical constraints (Laske et al., 2013), together with (B) an alternate regionalization based only on clustering

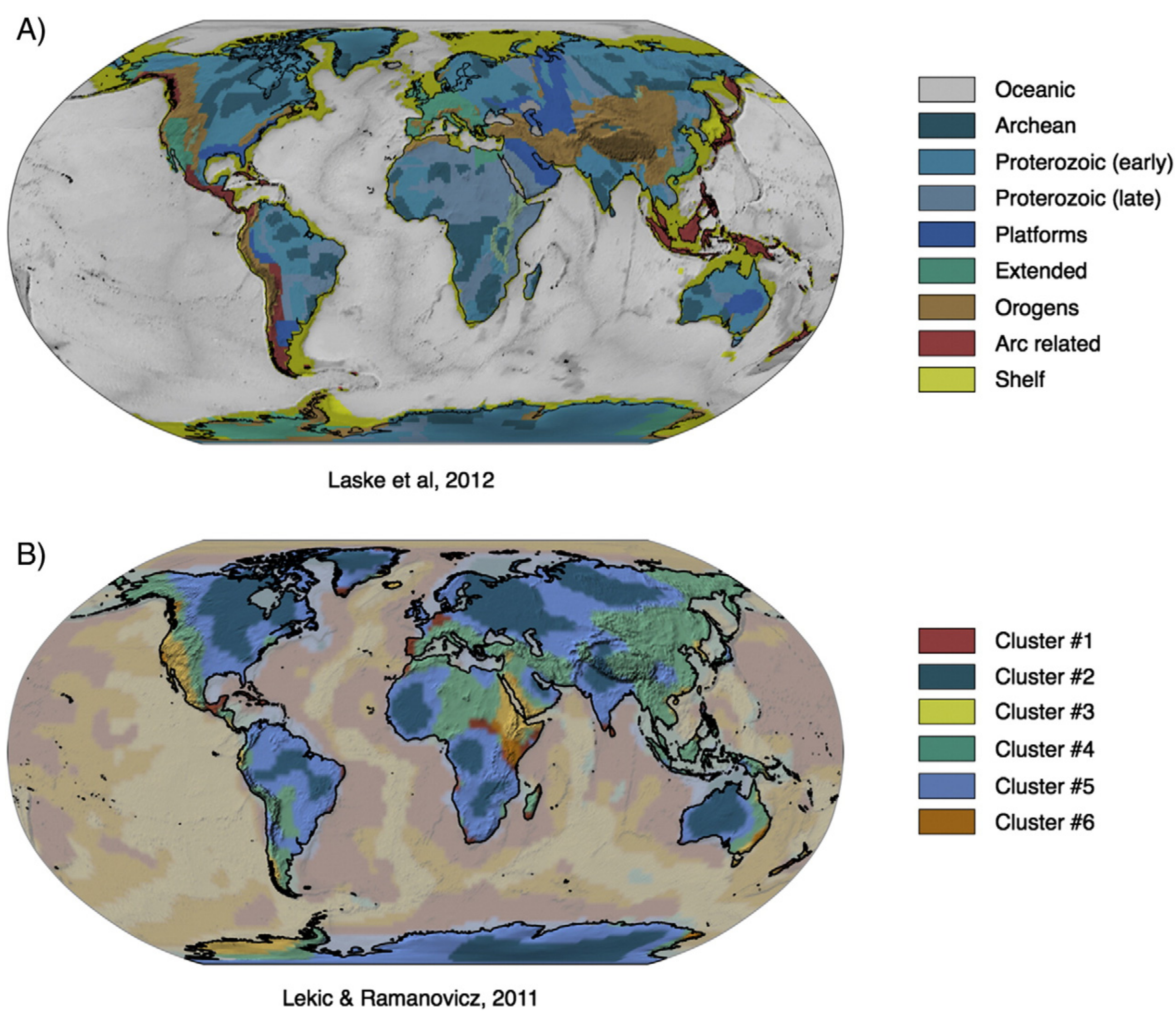

Fig. 1. Two examples of global tectonic regionalization overlain on a shaded relief map from the ETOPO1 (Eakins and Amante, 2009). (A) the Crust 1.0 model (Laske et al., 2013) based on 'shallow' geophysical observations and geological provinces where we have amalgamated the 36 distinct crustal types globally into 9 related types for the continents as described in the legend. (B) is the regionalization of (Lekic and Romanowicz, 2011) clustering analysis of global seismic tomography into six "similar" styles of lithosphere ("Clusters #1–6") with no a priori constraints from near-surface geology. In this model there is no distinction between continental and oceanic regions so we have applied a translucent mask to emphasize the distribution of the clusters within the continents. The numbering is arbitrary and matches that chosen by the authors of the study, and we have chosen a color scheme which emphasizes the similarity of old, stable regions in each map.

similarities in the top 350 km of a global, upper-mantle, seismic tomography model (Lekic and Romanowicz, 2011). Despite having been derived from very different means, some striking similarities exist that serve to highlight the vertical coherence of the lithospheric structure at this broad scale. For example, the cratons and the stable regions are picked out in both (darker blues), and the actively deforming orogens are largely identified with a single cluster (#4, shown in green). It is also not entirely unexpected that there is some similarity between the oceanic lithosphere and the regions of extension and rifting that is observed in the cluster analysis. The deep-seated clustering approach is least helpful in regions where the oceanic lithosphere is subducting beneath or close to the margin of the continent where the deep lithosphere is unrelated to the shallower crust and lithosphere. These clustering analyses suggest that, at a regional scale, deep and shallow lithospheric evolution are linked and deep structure can be related to the long term evolution and history of continental lithosphere.

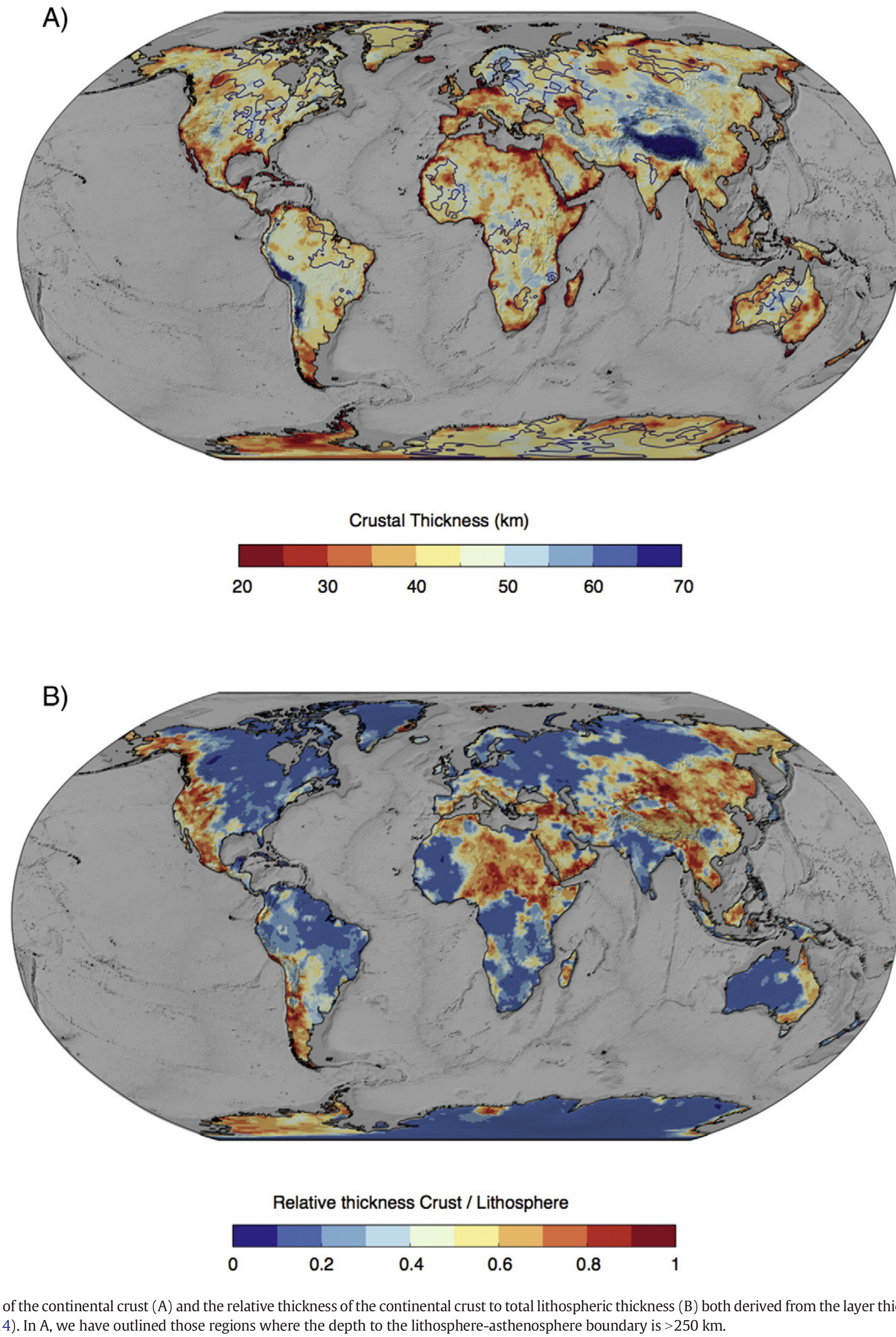

Fig. 2. The thickness of the continental crust (A) and the relative thickness of the continental crust to total lithospheric thickness (B) both derived from the layer thickness maps of litho1.0 (Pasyanos et al., 2014). In A, we have outlined those regions where the depth to the lithosphere-asthenosphere boundary is >250 km.

This is less obvious when looking at a similar global view of continental crust (Fig. 2A). The global map of continental crustal thickness, using the litho1.0 data of Pasyanos et al. (2014), is dominated by regions thickened by active orogenesis and broadly corresponds to relief as would be expected for long-wavelength topography supported by Airy isostasy. Crustal thickness appears to be less sensitive to past events (outlined in Fig. 2A). However, perhaps more interesting is the dichotomy apparent in the relative thickness of crust to that of the lithosphere overall (Fig. 2B) which clearly delineates the stable continental interior from the tectonically mobile regions (cf. clusters 3 and 4 of the Lekic and Romanowicz (2011) analysis). This could be indicating that, at least on the global scale, while the thickness of continental crust may not serve as an indicator of time or preserve deep time events, variations in sub-crustal continental lithospheric thickness (and corresponding lithospheric structure) may be potentially providing information about the deep history of deformation in the Earth. Certainly, this is connected, in part, to the preferential preservation of thick, cratonic lithosphere, but it could also be a hint that these regions could be informative in our attempts in interpreting the dynamic settings of the Earth's past.

1.1. The complex, dynamic Earth

As Earth scientists, we often call upon the tenet - “the present is the key to the past” (Lyell, 1837) - to build analogies between geological observations across large spans of time, appealing to a commonality in the underlying processes. This uniformitarian approach has proven to be useful at explaining how the slow, but persistent nature of change shapes the surface (and interior) of Earth. However, there is an important caveat to the uniformitarian approach that applies to complex dynamical systems (the Earth is certainly included in this category): while the driving processes may not change, and the forcings may evolve gradually, the dynamic system does not necessarily evolve smoothly and may instead jump through one or more catastrophic transitions between distinctive regimes of behavior (e.g. Sanns, 2012).

We expect the early Earth to have higher overall heat production with a less evolved mantle and less differentiated crust, for example, but, even if we assume the same physical/chemical systems were operating under early-Earth conditions, should we expect to see modern plate tectonics as an outcome (e.g. Dewey, 2007)? If the Earth undergoes transitions in tectonic style, are the cratonic regions “survivors” from the previous regime, or are they a by-product of the transition?

This shapes the broad context within which we can question the formation and preservation of structure within the continental lithosphere. Is the old, stable lithosphere preserved because of some gradual stabilization process (e.g., Cooper et al., 2006) or because the conditions of its formation produced regions with a stronger lithospheric column (e.g., Poudjom Djomani et al., 2001)? If the latter, does the stable, cratonic lithosphere represent anomalous regions from the time of formation or typical ones which, by chance, avoided wholesale recycling or reworking (Lenardic et al., 2000)? Are ancient lithospheres truly “forever” (O'Reilly et al., 2001) or susceptible to changing mantle dynamics (Gao et al., 2004; Cooper et al., 2006; Abdelsalam et al., 2011; Guo et al., 2016)? Can we bridge ancient and modern processes to build a continuous conceptual model for continental evolution?

In the modern Earth, continental growth predominantly occurs through the lateral accretion of juvenile crustal material along existing convergent margins driven by the horizontal motion of the oceanic lithosphere (e.g., Cawood et al., 2009). The familiar “conveyer belt” style of subduction is thought to have become active at some point during the Archean (a time period spanning multiple billions of years), perhaps even earlier, though the exact timing is still heavily debated. Shirey and Richardson (2011) argue that it occurred at 3.0 Ga based on the appearance of eclogitic diamonds, which are indicative of modern style subduction. Prior to 3.0 Ga, the inclusions within the ancient diamonds point to a different compositional origin (one that samples the average mantle rather than metamorphosed crust) (Shirey and Richardson, 2011). Recently, Tang et al. (2016) demonstrated a change in the bulk upper crust composition from mafic to felsic also around 3.0 Ga further supporting the onset of plate tectonics around this time. Though Shirey and Richardson (2011) and Tang et al. (2016) argue that their evidence rules out an earlier start to plate tectonics, other studies push back this date even earlier in Earth's history based on a myriad of other compositional, structural and stratigraphic evidence (e.g., Betts et al., 2002; Chen et al., 2009; Polat and Kerrich, 2006; Kerrich and Polat, 2006; Furnes et al., 2007; Betts et al., 2011; Kusky et al., 2013; Smart et al., 2016).

From a geodynamic point of view, the early viability of modern-style subduction is also unclear. Due to higher mantle temperatures in the past, and the resulting thicker crust, the oceanic lithosphere may have been too buoyant to participate in subduction (Davies, 1992). This assumes a thicker oceanic crust (driven by increased melting due to the higher mantle temperatures) as well as a more vigorous convection and corresponding plate motion, both of which reduce the available negative buoyancy for subduction. However, van Hunen and van den Berg (2008) demonstrate that the rapid transformation to dense eclogite of the thicker oceanic crust, due to higher mantle temperatures, would have increased the negative buoyancy and allowed for subduction during early Earth conditions albeit in a slightly modified manner: subduction may have been episodic with slabs experiencing a significant amount of necking, thinning, delamination or breaking apart. These discrepancies in the style and reduction in the strength of subduction have been proposed to explain the lack of geochemical evidence of modern style subduction prior to 3.0 Ga (van Hunen and van den Berg, 2008). Some of the other hallmark processes of subduction might still have been in operation though occurring in sporadic bursts of activity (for more information about this debate, see van Hunen and Moyen, 2012). Davies (2006) has argued that an early and strong depletion of incompatible trace elements in the source region for mid-ocean-ridge melting would produce an oceanic crust of comparable thickness to the modern Earth and, correspondingly, a modern-style of subduction. However, the dynamics of subduction in the early Earth and corresponding dynamics continental collision are likely to have been different from their present form (e.g., Sizova et al., 2014).

The earliest continental material preserved in the geological record predates the transition to plate tectonics (the oldest rock on the Earth was dated as 4.28 Ga; O'Neil et al., 2008) and could serve as a witness to whichever mode of tectonic activity was present. Modern-style plate tectonics occurs primarily through lateral translation of continents across the surface of the Earth with a characteristic signature of progressive thickening from the margins inward. A pre-plate tectonic regime with weak or no subduction is likely to have been more dominated by vertical motions. Such conceptual models require that the entire surface of the Earth would have been a single, stagnant plate with vigorous convection beneath. Within this premise, heat from the interior of the early Earth would have been transferred through the lithosphere by conduction (a.k.a. stagnant lid; Moresi and Solomatov, 1995), by rapid and pervasive volcanic eruptions (a.k.a. heat pipe; Moore and Webb, 2013), or by episodic, intermittent, catastrophic bursts of subduction (O'Neill et al., 2007; O'Neill and Debaille, 2014). Each of these conceptual models makes specific predictions for the expected tectonic settings. Within stagnant lid and heat pipe regimes, deformation is driven primarily by vertical motion (small scale drips and delamination, and vertical advection) with the potential for regional, lateral motion as driven by gravitational collapse (Rey et al., 2014). Episodic subduction suggests a more complex setting that captures both the signatures of subduction, but also periods of stagnant lid convection.

As home to the oldest regions on the Earth (specifically the “cratons” which are defined as the regions of the crust and lithospheric mantle that have not undergone deformation in billions of years), the continental lithosphere should serve as a record for these past conditions. In this paper, we review the current status of the research on the structure and evolution of deep continental lithosphere with a focused gaze into what information can be gained from the similarities and differences between older and younger continental lithosphere. We begin with a discussion of the conceptual definitions used to describe and characterize continental lithosphere. From there, we review recent seismic observations of continental lithosphere and its complex internal structure. We also briefly describe the seismological techniques used to make these observations. We then move on to discuss a number of geodynamical models proposed for the formation of continental lithosphere and the corresponding structures induced during those processes. We follow that section with a comparison between the proposed ancient processes and modern day analogues of continental accretion in Alaska and Australia. Finally, we leave readers with potentially more questions that we initially posed as well suggestions for future work.

2. Continental lithosphere

The term lithosphere, while in use since 1914 (Barrell, 1914), remains an imprecise concept with different disciplines defining the "layer of strength" in a variety of ways. This is, in part, because it is unclear as to what is the primary control on the strength of Earth materials (e.g., temperature, composition, pressure) and also because the timescale of loading and response plays an important role in interpreting strength (e.g., Watts et al., 2013). One approach is to treat the lithosphere as the thermal boundary layer or the region encompassing the greatest temperature gradient near the surface of the Earth. In this construct, the base of the lithosphere is defined as the depth at which the temperature depth profile transitions from the steep gradient near the surface to align with the mantle adiabat. The strength is then due primarily to temperature as the mantle's viscosity is temperature dependent and below the lithosphere, the temperature is sufficiently high to consider the material "weak" and subject to viscous, convective deformation (for a full discussion see Eaton et al., 2009). To map out the base of the lithosphere in this manner requires an approach that is sensitive to temperature within the Earth's interior. One approach is to use surface heat flux measurements. The surface heat flux is proportional to the near surface temperature gradient (i.e., Fourier's law of heat conduction). If, then, the amount of heat production from the crust and mantle is known, a regional geotherm (steady state temperature with depth profile) can be derived using surface heat flux measurements (Fig. 3).

This approach depends on reliable surface heat flux measurements as well as knowledge of the composition of the crust and upper mantle (in particular, the amount of heat producing elements, which control the "curve" in the geotherm - for example, the geotherm is much steeper at shallow depths when there is a high concentration of heat producing elements in the crust) both of which are often lacking (e.g., Rudnick et al., 1998). Moreover, Rudnick et al. (1998) demonstrated the non-uniqueness in this approach; many geotherms "fit" the heat flux data available and it is unclear whether surface heat flux variations are more indicative of variations in crustal heat production or lithospheric thickness. As the Earth cools, the average thickness of the thermal boundary layer increases, and conversely, in the past when the Earth's mantle was warmer, the thermal boundary layer was thinner (Turcotte and Schubert, 2002). In addition, local thermal perturbations (for example, due to rifting or mantle upwellings) can thin and weaken the thermal boundary layer. Once the thermal perturbation has dissipated, the thermal boundary layer will thicken as the region cools. Thus, variations in the thickness of the thermal boundary layer can also indicate variations in age, either absolute age as in the case for the oceanic lithosphere (e.g., Müller et al., 2008) or in the amount of time since the last tectono-thermal event in the case of the continental lithosphere.

The lithosphere can also be described as a compositionally distinct region, often referred to as the chemical boundary layer. Specifically, the strength of the lithosphere can be explained not only by cooler temperatures, but also by increased buoyancy and viscosity driven by compositional differences (often explained as a consequence of melt depletion during crust formation (e.g., Pollack, 1986)). As such, the base of the lithosphere is demarcated by a compositional (and rheological) change and efforts to map out lithospheric thickness under this premise must be sensitive to composition. Mantle xenoliths have traditionally been used for this purpose (e.g., Rudnick et al., 1998; Lee et al., 2011). In addition to compositional information, mantle xenoliths also provide a direct sampling of the temperature and pressure conditions of the regions from which they erupted. Coupled with surface heat flux measurements (as discussed above), mantle xenoliths can be used to discriminate between the possible geotherms as derived from Fourier's law and energy conservation (Fig. 3B). The base of the lithosphere is then determined as the depth where the xenolith-determined geotherm intersects the mantle adiabat.

Defining the lithosphere in this fashion assumes that the layer is chemically distinct from the convecting mantle. Jordan (1975, 1978) pioneered this idea to explain the observation that cratons are neutrally buoyant despite their cold and old interiors. Following the thermal

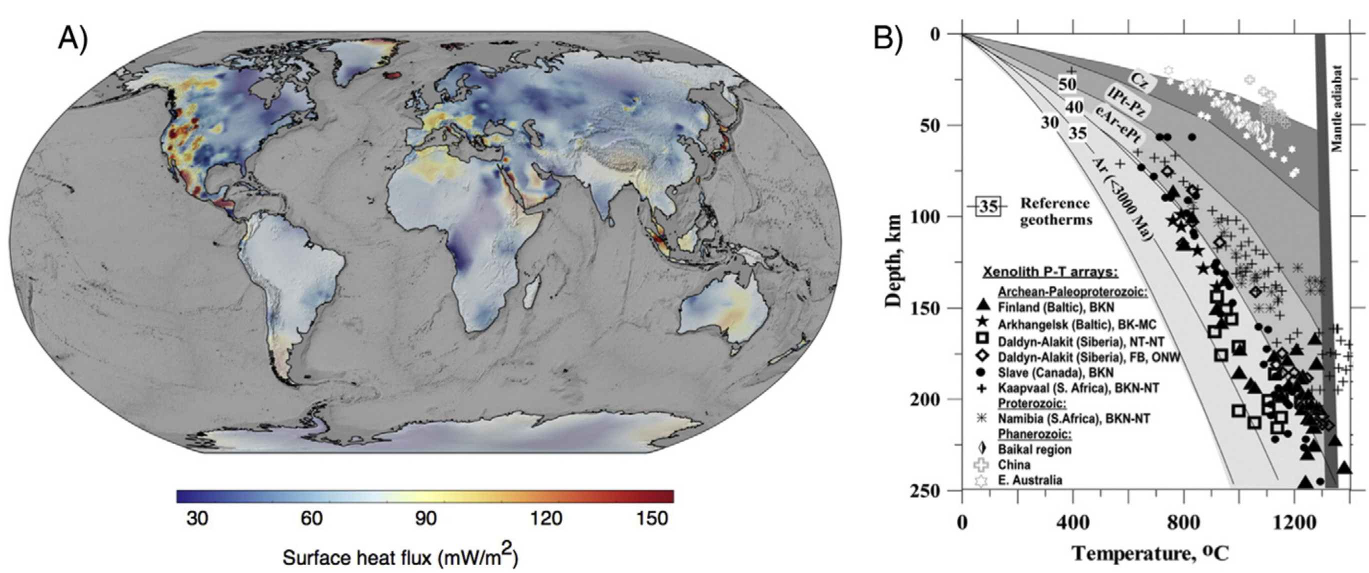

Fig. 3. (A) Global surface heat flow measurements (International Heat Flow Commission) plotted by binning the point-measurement datasets to an adaptively-refined triangulation with a minimum resolution of 50 km. Color saturation represents the resolution of the triangulation which washes out progressively such that the coarsest resolution is shown with only 5% saturation. (B) Lithospheric geotherms for continental regions (figure from Artemieva, 2009).

boundary layer approach, Jordan (1975, 1978) demonstrated that if continental and oceanic lithosphere were the same composition, there would be mass excess underneath continents that is not observed in the global free air gravity anomaly maps. The thermal analysis in cratonic regions indicated cool temperatures extending to great depths, thus necessitating thick continental lithosphere (Jordan's early estimates extended down to 400 km (Jordan, 1978)). To reconcile the thermally required thick continental lithosphere with the negligible free air gravity anomaly (and lack of mass excess), Jordan (1975, 1978) argued that continental lithosphere (in particular, cratonic lithosphere) has a different composition than oceanic lithosphere. This argument was formalized in the "isopycnic hypothesis" (*iso* = same, *pycnic* = density) which states that at every depth level in the cratonic lithosphere, the negative thermal buoyancy is exactly balanced by a positive chemical buoyancy rendering the region neutrally buoyant (Jordan, 1988). In other words, at each depth, the change in density due to the thermal contraction driven by the temperature difference between the colder cratonic lithosphere and the warmer surrounding mantle is equal to the change in compositional density between the two materials. As the temperature difference diminishes as the geotherm approaches the adiabat, the compositional difference between the continental lithosphere and surrounding mantle also decreases. We should note that isopycnicity is not the same as isostasy. Isostasy describes the vertical displacement driven by the average density of the entire lithospheric column. Isostasy does provide information about the composition of cratonic lithosphere (isostatic requirements also require positive chemical buoyancy; e.g., Lee et al., 2011; Cammarano et al., 2011), but as a weaker, integral constraint. Isopycnicity, on the other hand, prescribes a compositional stratification based on the thermal structure of cratonic lithosphere.

While mantle xenoliths from cratonic regions do possess significant compositional differences from the average mantle whether the compositional difference occur in a strictly isopycnic configuration is still debated (e.g., Lee et al., 2011). Many argue that the compositional distinction and positive chemical buoyancy of cratonic lithosphere were consequences of melt extraction (e.g., Pollack, 1986; Lee et al., 2011). Dehydration was also likely to occur during melt extraction increasing viscosity and contributing to the intrinsic strength of the lithosphere (e.g., Pollack, 1986; Hirth and Kohlstedt, 1996; Lee et al., 2011). This combination of the positive chemical buoyancy and increased viscosity are suggested to be responsible for lithospheric strength and stability, particularly so in cratonic regions (e.g., Richter, 1985; Pollack, 1986; Lenardic and Moresi, 1999; Sleep, 2003; Francois et al., 2013), thus indicating that a chemical boundary layer plays an important role in lithospheric structure.

The base of the chemical boundary layer and the thermal boundary layer need not coincide. Estimates for the thickness of the "lithosphere", may differ even in the same region depending on the techniques used (e.g., Jones et al., 2010; Eaton et al., 2009). Generally, geochemically based estimates (which may be sampling the base of the chemical boundary layer) give thinner values for lithospheric thickness than the seismic observations (which may coincide with the base of the thermal boundary layer) (Lee et al., 2011; Steinberger and Becker, 2016). These discrepancies indicate that the transition from the surface of the Earth to the convecting mantle is more complex than a single rigid, layer of strength (lithosphere) on top of a weak mantle (asthenosphere). Likely, the transition encompasses several boundary layers whose thicknesses depend on each other. For example, the chemical boundary layer can influence the thickness of the thermal boundary layer (Lenardic and Kaula, 1996; Cooper et al., 2004; Lee et al., 2005). The chemical boundary layer is assumed to be strong, rigid and resistance to deformation (e.g., Pollack, 1986; Lee et al., 2011). As such, heat is transferred through the chemical boundary layer solely by conduction. Small scale convection can occur within the base of the thermal boundary layer allowing for dripping, shearing and deformation of the material. In the scenario where a chemical boundary layer resides within the thermal boundary layer, the thicker the chemical boundary layer, the more it will dominate the conductive, rigid portion of the thermal boundary layer isolating the potentially deformable region to the base of the thermal boundary layer. This deforming region of the thermal boundary layer is often referred to as the convective sublayer. The thickness and temperature drop across the convective sublayer depends on the thickness of the chemical boundary layer (Cooper et al., 2004; Lee et al., 2005).

Numerical simulations and theoretical scalings demonstrated that the thicker the chemical boundary layer, the closer it approaches the thickness of the thermal boundary (Cooper et al., 2004; Lee et al., 2005) in a predictable manner. In other words, the thickness of the chemical boundary layer can be used to predict the thickness of the thermal boundary layer (and correspondingly, the thickness of the convective sublayer) and vice versa. For example, this theoretical relationship was observed, and independently confirmed, within the continental lithosphere of North America using seismic observations (Yuan and Romanowicz, 2010) indicating that the idea of nested boundary layers might be a good baseline representative for the structure of continental lithosphere. Recognizing this structure within the observations (and any variations both to and within the basic structure) can serve as the first step needed to illustrate processes that build and shape continental lithosphere. In particular, both a better view and conceptual framework for the complexity of the continental lithospheric structure can provide insight into early Earth and the progression into the present day tectonic regime.

We should note that there are several other historical approaches to defining and describing the lithosphere (e.g., "seismic lid" - Priestley and Tilmann, 2009, "tectosphere" - Jordan, 1978, "electric lithosphere" - Jones et al., 2001), but for brevity, we will limit our discussion in this paper to the thermal and chemical boundary layer descriptions of the lithosphere. For more information about the various definitions of the lithosphere, we suggest Eaton et al. (2009), Fischer et al. (2010) and Fischer (2015). Searching for the base of the lithosphere and structure within, whichever definition used, requires a proxy for temperature, composition and/or rheology. Often we turn to seismology to provide that information. In that vein, the lithosphere can also be defined seismologically (e.g., Fischer, 2015). The seismological lithosphere can also be described multiple different ways, using different techniques as described in the following section, but can generally be defined as the high-velocity outer layer of the Earth, which typically is underlain by a low-velocity layer and/or a distinct change in observed seismic anisotropy. Within this definition, the challenge becomes determining whether this high-velocity layer coincides with the thermal or chemical boundary layer or, perhaps, even some other physical properties.

In the next section, we describe several seismic imaging techniques as well as showcase observations and interpretations of the internal structure within continental lithosphere.

3. Seismic imaging

Broadband seismological observations of the continental lithosphere structure provide constraints on seismic velocity, but also estimates of crustal and lithospheric thicknesses, rheological contrasts, presence of anisotropic layers and internal structures that may be related to deformation and/or formation, such as the crustal-mantle boundary (Moho), mid-lithospheric discontinuities (MLDs, defined as seismic discontinuities occurring between the base of the crust and the asthenosphere), or lithosphere-asthenosphere boundary (LAB). Multiple seismological methods are used for imaging the lithosphere, and these techniques include a range of tomographic methods, velocity discontinuity imaging, and various procedures for mapping changes in seismic anisotropy. Different approaches in seismic imaging have become the subject of serious debate, as new methods are being developed and existing methodologies are improving in order to exploit the continuously growing, enormous volumes of readily available broad-band data for analysis. Varying approaches and tools are designed for different targets, and a few of those commonly used for lithosphere-scale imaging are briefly summarized here. For a more thorough review, see Fischer (2015).

3.1. Seismic tomography

Global body wave tomography studies robustly image long wavelength (~100 s of kilometers) lateral features in the upper (and lower) mantle (see Becker and Boschi (2002) for a summary) with travel-times, highlighting the contrast in velocities and attenuation not only beneath the oceans, but between the active tectonic regions of the continent and their stable cores (cratons). However, due to the near vertical incidence of teleseismic body waves used in P-wave inversions, the depth of the lithosphere is often over-estimated when only P-waves are used. Furthermore, crustal scale velocity perturbations can be difficult to resolve with teleseismic data and global inversions. In addition, the wavelength of structures imaged with global body wave tomography, whether or not they include surface waves in the inversion, are typically not at the scale needed (1–10 s of kilometers) to investigate the detailed internal structure of the lithosphere.

Global surface wave tomography, however, is often used to infer the absolute shear-wave velocity structure of the upper mantle and through analysis of various periods of horizontally propagating surface waves. Due to the inherent difference in the types of seismic waves used, there generally is not as much vertical smearing of the velocity structure as is common with body wave tomography. However, due to the integrative properties in surface wave tomography, near-horizontal boundary structures, such as the Moho and LAB can be difficult to image. Joint inversion of S-waves and surface waves (and also normal modes) for global shear wave velocity perturbations provide some of the most comprehensive tomographic models of the upper mantle (e.g., Schaeffer and Lebedev, 2013; French et al., 2013). These models clearly identify subduction zone structures, the difference in thickness of the lithosphere in cratonic environments versus active tectonic regions, and also clearly image the smaller-scale variations in structure within the continental lithosphere as shown in Fig. 4, which approach regional scale studies (Schaeffer and Lebedev, 2013).

Regional tomography studies can provide more detailed images of the structure of the continental lithosphere, and with the ever increasing number of seismic stations the images are constantly improving. Travel-time tomography using only body wave arrivals at densely spaced arrays, such as USArray, have provided "high" resolution images (~10 km) that show the incredible complexity of the continental lithosphere beneath the Moho (see Becker (2012) for a summary). Continent-scale surface wave models (of North America) also clearly illustrate the large (and regional) scale variations in structure (10–100 s of kilometers), but more focused studies that use either ballistic (e.g., Wagner et al., 2010; Pollitz and Mooney, 2016) or ambient noise (e.g., Lin et al., 2009; Porritt et al., 2011) surface waves techniques confirm this observation and further illustrate that the continental lithosphere is extremely heterogeneous and complex at the ~10 km scale within both the crust and mantle lithosphere. Similarly, surface wave images of the other continents show lithospheric complexity where there have been extensive seismic deployments such as in Europe (e.g., Zhu et al., 2012) or China (e.g., Bao et al., 2015), and Australia (e.g., Saygin et al., 2012; Fishwick and Rawlinson, 2012).

3.2. Discontinuity structures

Layering within the continental lithosphere has long been known (e.g. Mohorovičić, 1910; Gutenberg, 1948; Lehmann, 1961; Hales, 1969), but the complexity, layering, and structures within the outer layer of the Earth continues to be to revealed with seismic scattering and discontinuity imaging (e.g., Gaherty and Jordan, 1995; Bostock, 1998; Rychert et al., 2007; Savage et al., 2008; Rychert and Shearer, 2009; Abt et al., 2010; Miller and Eaton, 2010). Receiver function analysis (Langston, 1977; Vinnik, 1977; Farra and Vinnik, 2000) has become an important tool in mapping sub-horizontal velocity discontinuities in the crust and mantle. P receiver functions (PRFs), which use P-to-s conversions, are commonly used to map the crust-mantle boundary (Moho) and internal crustal structure and are less commonly used for deeper structure due to the potential contamination of crustal multiples at deeper lithospheric depths (Rychert and Shearer, 2009). The raw Ps signal is often weak and so the RFs are generally stacked to improve the signal-to-noise ratio. Then the stacked RFs represent the 1-D (generally) isotropic structure below a seismic station, where the phase delay between P and Ps indicates the depth of discontinuity it arises from, and the amplitude represents the strength of the impedance contrast at that boundary. S receiver functions (SRFs), which instead use S-to-p conversions, are also a powerful tool for imaging deeper subhorizontal structures such as the lithosphere-asthenosphere boundary (LAB) or mid-lithospheric discontinuities (MLDs) as the signals are not contaminated

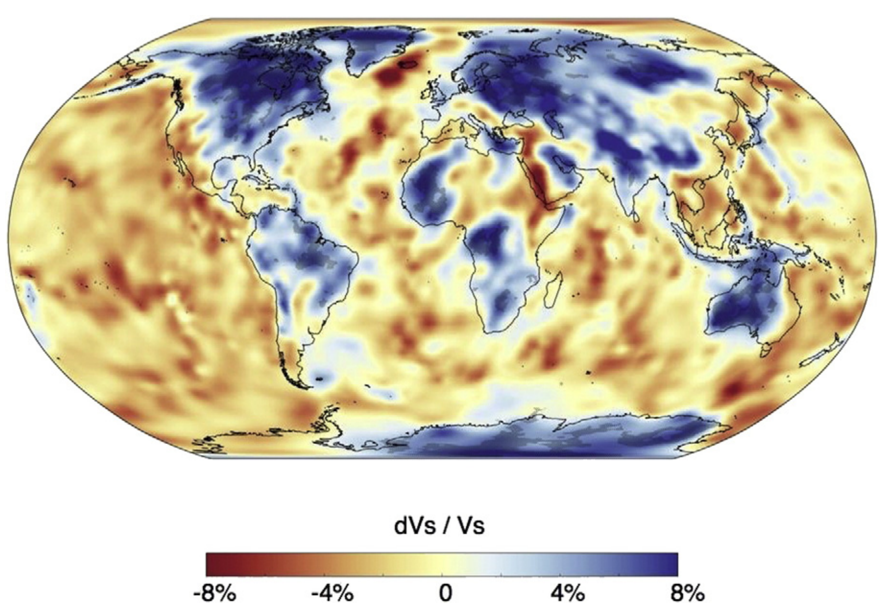

Fig. 4. Global S wave tomography model by Schaeffer and Lebedev (2013) plotted at 150 km depth. The additional shading indicates regions with very thick (>250 km) lithosphere in the litho1.0 model (Pasyanos et al., 2014).

by crustal multiples, but are limited by their lower frequency content and smaller distance range for appropriate teleseismic events (Yuan et al., 2006). With densely spaced seismometers, more sophisticated methodologies can be used with teleseismic scattered phases to produce either 2D (Bostock, 1998; Frederiksen and Bostock, 2000) or 3D images (Dueker and Sheehan, 1997) of seismic discontinuities (Rondenay et al., 2005). All these flavors of receiver functions are useful for mapping sharp velocity discontinuities and can provide tight constraints (~1 km) on the depth of these structures. Recent literature has provided global compilations of results that map the internal structure of the continental lithosphere (Fischer et al., 2010; Cooper and Miller, 2014; Hopper et al., 2014; Selway et al., 2015; Rader et al., 2015), which are shown in Fig. 6.

3.3. Anisotropic structures

Another seismic observable used to infer deformation within the continental lithosphere and its base is through the detection of seismic anisotropy, which is defined as the dependence of seismic velocities on the direction of wave propagation and polarization. Changes in anisotropy within the continental lithosphere have also been mapped with multiple methods which include, but are not limited to, receiver functions, shear wave splitting, and surface wave tomography. In the oldest centers of the continents, seismic anisotropy can often be linked to fossil mantle textures, dating from rifting, collision, and accretion events during the formation of the lithosphere (Park and Levin, 2002; Yuan and Romanowicz, 2010), but this interpretation is far from simple or widely-proven for different regions or even observed globally (Fouch and Rondenay, 2006; Long and Becker, 2010).

For an isotropic, flat-layered Earth structure, tangential component receiver functions should be zero (or contain only incoherent noise in reality), and the radial component receiver function should have signals that are independent of back azimuth and then used to image the discontinuity structures beneath seismic stations as discussed above. However, in practice, coherent and azimuth-dependent signals are observed on both the radial and tangential component receiver functions, which can be used to infer anisotropic structures using a variety of techniques (Bostock, 1998; Savage, 1998; Frederiksen and Bostock, 2000; Schulte-Pelkum et al., 2014). Receiver function analysis of anisotropy can detect sharp vertical variations in anisotropy and dipping layers with better resolution than the other methods discussed here (Fouch and Rondenay, 2006), but are limited by non-linearity and densely spaced seismic stations.

Shear wave splitting analyses, in particular splitting of core phases (SK(K)S), are based upon the properties of birefringence of nearly vertically propagating shear waves beneath seismic stations. The lateral measurements (at multiple stations) of splitting of SKS phases allow for interpretation of the back-azimuthal variation of delay times and fast directions to infer the presence of anisotropic layers or dipping structures in the mantle, however the depth constraints of these layers is poorly resolved due to the path-integrated effect of anisotropy from the core to the crust (Silver and Chan, 1988; Fouch and Rondenay, 2006).

Two types of seismic anisotropy, azimuthal (horizontal) and radial (vertical) anisotropy, can be measured with surface wave methodologies. Radial anisotropy in the upper mantle has been detected with the discrepancy between Rayleigh and Love wave propagation, while the existence of azimuthal anisotropy is determined by changes in directional dependence of the propagation of Rayleigh waves. A major advantage of using surface waves to detect continental lithosphere anisotropy is that the depth of the anisotropic layers are much better constrained than when using body wave (e.g. SKS) measurements. For example, detecting azimuthal anisotropy in the upper mantle from surface waves can illustrate how the orientation of the fast axis of anisotropy changes with depth with more widely spaced stations than with receiver function methods. Many different methods can be used to infer both radial and azimuthal anisotropy within the continents (see Fouch and Rondenay (2006) and Fischer (2015) for reviews).

Further advances are rapidly being made to improve imaging capabilities by jointly inverting different types of data and results, integrating existing models as starting models, utilizing full waveform inversions, developing array based methods, and exploiting faster computational abilities (Julià et al., 2000; Sigloch, 2011; Yang et al., 2011; French et al., 2013; Shen et al., 2013; Porritt et al., 2014; Schmandt and Fan-Chi, 2014; Calò et al., 2016). These rapid developments have produced detailed structural images of the continental lithosphere that are providing new observations of the complex formation and deformation that has occurred throughout geologic history.

3.4. Seismic observations

Regardless of the seismic imaging technique, there is a first-order feature that is observed in continental lithospheric structure. The oldest cores of the continents have thicker, fast velocity structure; in contrast, the younger regions of the continents have much thinner lithosphere (Fig. 5). Global seismic observations (Fig. 4) illustrate that the thick cratonic mantle lithosphere has high velocity in comparison to the same depths beneath the Phanerozoic lithosphere. Some petrological studies suggest compositional variations between two age groups (e.g., Poudjom Djomani et al., 2001), however, recent seismic studies have suggested that there is no systematic difference in the velocity and overall thickness between the Proterozoic and Archean lithosphere (Darbyshire and Eaton, 2010).

More detailed, regional seismic observations illustrate the complexity of the deep continental lithosphere structure both in the cratons and in the tectonically active regions. Regional studies of active convergent margins, for example, image the subducted oceanic lithosphere but also image how the overriding continental lithosphere is deformed and formed by the subduction zone processes (e.g., Gutscher et al., 2000; Humphreys, 2009; Schmandt and Humphreys, 2010; Miller and Agostinetti, 2011; Porritt et al., 2011; Schmandt and Humphreys, 2011; Pearce et al., 2012; Levander et al., 2014; Abers et al., 2014; O'Driscoll and Miller, 2015). In the shallow mantle lithosphere a few high-resolution studies have imaged the dipping subducted oceanic lithosphere, mantle wedge and modified structure of the overriding plate (e.g., Bostock et al., 2001; Pearce et al., 2012; Kim et al., 2014; Bostock, 2013), accreted terranes such as the actively subducting Yakatut beneath Alaska (Ferris et al., 2003; Rondenay et al., 2010; Abers et al., 2014), the 55 Ma Siletzia terrane stagnating in the upper mantle beneath northwestern United States (Schmandt and Humphreys, 2011), and collision of continental lithosphere transferred to base of the overriding plate (Eurasia) in the Banda arc of southeast Asia (Porritt et al., 2016). Inboard of subduction zones, but in regions that are still considered tectonically active like the western Americas and the Mediterranean the continental lithosphere continues to be deformed and modified by tectonic processes (van der Hilst and Paul, 1994; Wortel and Spakman, 2000; Gilbert et al., 2006; Schmandt and Humphreys, 2010; Levander and Miller, 2012; Bao et al., 2014).

Away from convergent margins, within the continental interiors, structures imaged seem to be as complex and appear to have characteristics that are reminiscent of modern accretionary structures. Ancient accreted terranes and sutures have been mapped geologically (e.g., Oldow et al., 1989; Betts et al., 2011), and have been mapped by a few active source based studies (BABEL Working Group, 1990; Cook et al., 1999; Magnani et al., 2004). Large scale active source seismic experiments, such as Lithoprobe (Hammer et al., 2010), were able to image lithospheric structure across the entire North American continent from active subduction in the west to the cratonic interior. These efforts provide a link between the structures in the active tectonic regions and those in the stable continental interior. Passive source seismology has also been able to image the structures within the continental

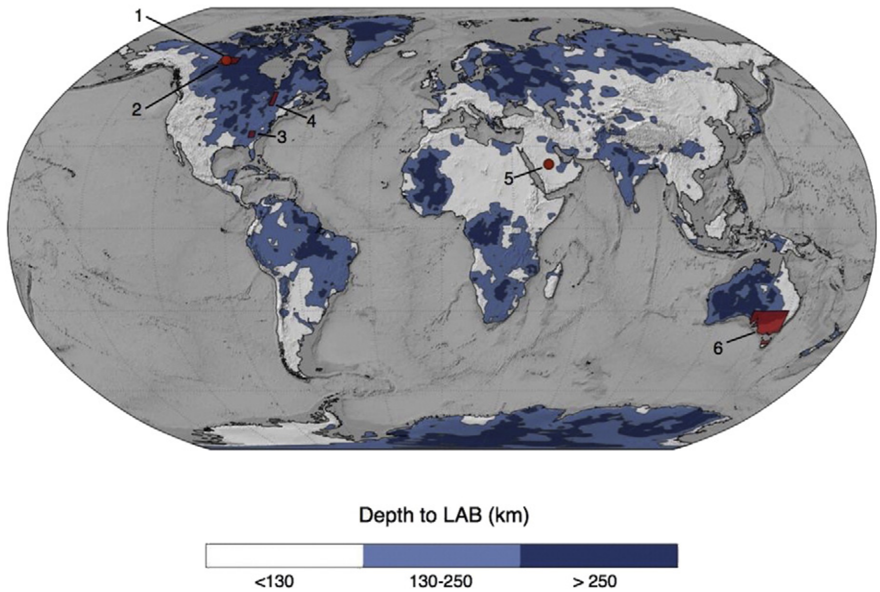

Fig. 5. Contours of continental lithosphere thickness taken from the litho1.0 model (Pasyanos et al., 2014). Dark blue are the regions thicker than 250 km and lighter blue are regions between 130 and 250 km in thickness. Thinner regions have not been colored. The background is shaded relief computed for ETOPO1 (Eakins and Amante, 2009). Red polygons indicate the locations of broadband seismic studies that observe structures that have been interpreted to be ancient accreted terranes/terrane boundaries preserved within the continental lithosphere (1 – Bostock, 1998; 2 - Cook et al., 1998; 3 - Parker et al., 2015; 4 - Rondenay et al., 2000; 5 - Levin and Park, 2000; 6 - Pilia et al., 2015).

lithosphere with impressive resolution (<1–1 km) in select locations (red polygons in Fig. 5).

One of the first studies linked the active and passive source seismological experiments in the Slave craton of Canada (Bostock, 1998). This study imaged dipping reflectors from P-to-s conversions that correlate with the shallow reflectors from the Lithoprobe-SNORCLE active source experiment (Cook et al., 1999) interpreted as tectonic boundaries or sutures, suggesting that subduction-like processes were fundamental in assembling the continental lithosphere. Similarly, on the southeastern edge of the Canadian shield, Rondenay et al. (2000) image a subduction zone suture at the Grenville Front using a combination of tomography, SKS splitting, and receiver function techniques. Recently, in the southern Appalachians, receiver functions were used to observe a décollement beneath the fold-thrust belt that transitions into a subhorizontal shear zone inward toward the orogen. This was interpreted as a set of accreted terranes overlying the basement rocks (Parker et al., 2015).

In southeastern Australia, Pilia et al. (2015) found evidence of discrete terranes within the continent that were accreted in the Paleozoic. The fabric of the accreted terrane geometries and deformation history is inferred from seismic anisotropy derived from ambient noise tomography (Fig. 9B). Their results suggest that exotic continental fragments from Paleozoic collision of terranes onto the Proterozoic and Archean core of the Australian continent are preserved and identifiable. Other less well sampled regions on Earth, also suggest the presence of dipping, anisotropic layers in other continental cratons. For example, one study of the Arabian shield also imaged dipping, anisotropic discontinuities in a similar depth range (~80 km) with one seismic station, which could be related to shear zones formed during Proterozoic continent-continent collision (Levin and Park, 2000).

More widespread surveys within the continents observe mid-lithospheric discontinuities (MLDs) with S receiver functions (Selway et al., 2015). The mid-lithospheric discontinuity refers to a seismic discontinuity recorded at a depth range of ~80–120 km, which may or may not be related to the Hales discontinuity or those imaged by the detailed scattered wave images mentioned above. Initially, the MLDs were only identified in sparse locations within the cratons of North America, Australia, and Africa, due to the wide, uneven distribution of broadband seismic stations ~10 years ago e.g. (Wittlinger and Farra, 2007; Olsson et al., 2007; Savage et al., 2008; Abt et al., 2010; Miller and Eaton, 2010; Ford et al., 2010; Cooper and Miller, 2014). As the number of evenly spaced stations increased and additional methods of imaging became available and adapted to the growing volume of data, the number and distribution of MLDs dramatically increased (Fig. 6). Furthermore, in some places are the MLDs are observed at multiple depths (Fig. 6, inset B, C) within the same region (Yuan and Romanowicz, 2010; Lekić et al., 2014; Hopper et al., 2014; Hopper and Fischer, 2015; Porritt et al., 2015; Hansen et al., 2015; Calò et al., 2016) and are represented by (most often) a negative velocity gradient, but occasionally a positive velocity gradient at similar depths. Fig. 6 (D, E) shows the distribution of the MLDs with lithospheric thickness and compares this to the global distribution of lithospheric thickness, showing that 80% these features are generally only seen in the thickest 20% (175 km or more) of the continental lithosphere.

Although the stable interiors of the continents are generally thicker (Fig. 5), the base of the lithosphere is not observed at a uniform depth. There are differences between each continent and even within an individual craton. For example, the North American craton, which is one of the most well-studied regions, illustrates this point. Beneath the core of the continent as seen in the high velocity structures in Fig. 4, around Hudson Bay there is significant variation in the depth (~200–325 km) to the LAB (Fig. 5) from seismic observations (Darbyshire and Eaton, 2010; Porritt et al., 2015). The variation in thickness and structure has been interpreted from broadband seismic studies to be a signature of Archean-Paleoproterozoic tectonics, and the formation of the Trans-Hudson orogen around 1.8 Ga when several Archean terranes were assembled into what is now the Canadian Shield (Bastow et al., 2010). Similarly, the African continent, particularly the South African craton, displays a significant variation (~100 to >300 km) in the lithospheric thickness (e.g. Wittlinger and Farra, 2007; Fishwick, 2010; Cooper and Miller, 2014) indicating that the sequential tectonic assembly of the continent is reflected in the structure of the individual cratons (Begg et al., 2009).

The continental lithosphere imaged with broadband seismic data suggests that the internal structure, including multiple layers (Moho and MLDs), variable depth to the base of the lithosphere (LAB), and dipping structures, is indicative of the long-term evolution and formation of both the youngest and oldest part of the continents.

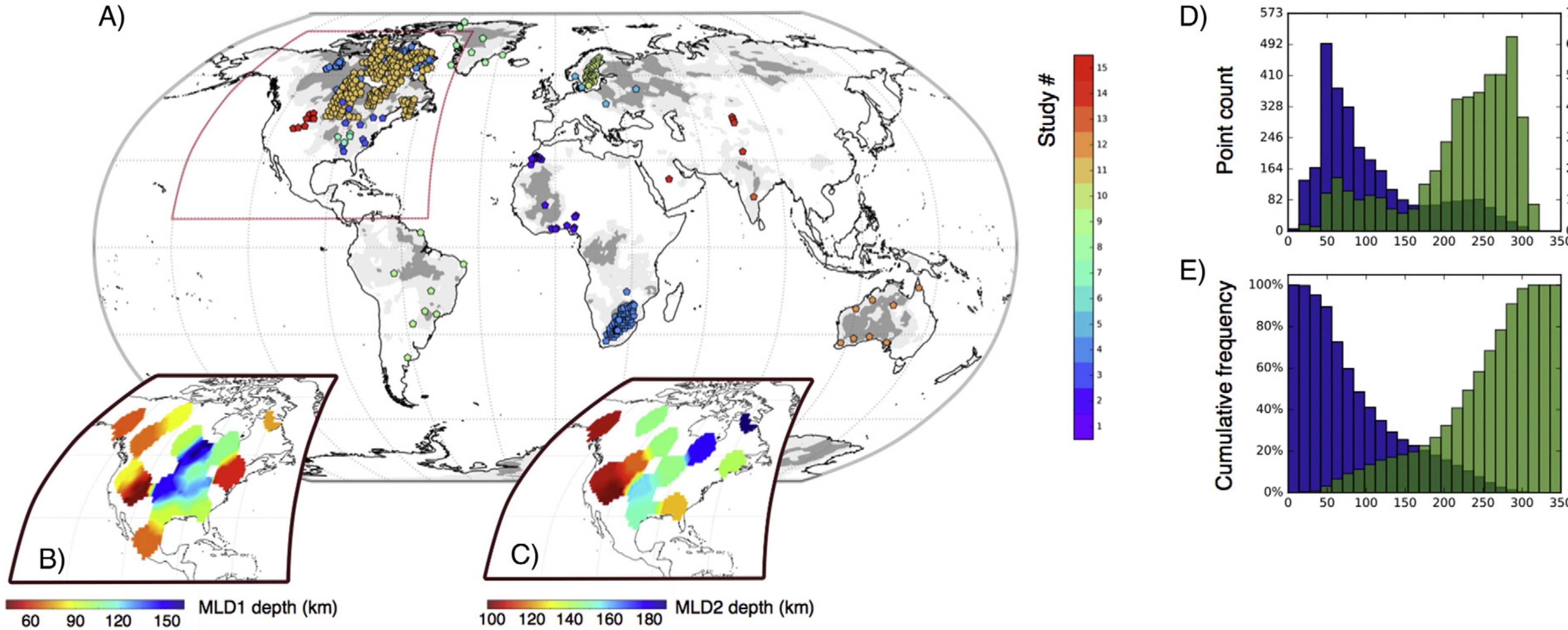

Fig. 6. (A) Global compilation of observations of S receiver function based mid-lithosphere discontinuities (MLD), color coded by provenance (see text for references) the shading indicates lithospheric thickness in two bands: light grey for lithosphere >150 km in thickness, and dark grey for >250 km thickness (litho1.0, Pasyanos et al., 2014). (B,C) are representations of the depths of two mid lithosphere discontinuities (MLD1 and MLD2) imaged by Calò et al. (2016). Gaps in the interior of this study indicate regions where the discontinuity surface is absent. (D) Frequency count for the thickness of the lithosphere taken at every point in the litho1.0 model (blue and left axis), and for those points where MLDs have been identified in (green and right axis). (E) Cumulative frequency diagram for the same information (global lithospheric thickness, blue, from thick to thin) and for the lithosphere at locations in (A) where an MLD observation has been made (green, cumulative from thinner to thicker). The litho1.0 model is an icosohedral triangulation of the sphere at approximately 1° separation. The MLD information was projected to the nodes through a weighted average and only those nodes with at least one observation within their local support were counted in the histogram. No additional weight was given to multiple observations near any point.

4. Dynamic interpretation of lithospheric structure

The techniques discussed above and observations from the global maps highlighted in Figs. 1–6, while deeply informative, offer only a static view of the continental lithosphere. What we “see” in these data sets is the sum of all of the geologic events and processes occurring over the history of a region (often spanning millions to billions of years). While the data sets are used to unravel the tectonic events through careful interpretation, further insight can be gained through geodynamic modeling. In particular, geodynamic modeling can help guide our interpretation and determine whether modern and ancient processes produce similar or different lithospheric structure.

As mentioned above, considerable debate revolves around the style of tectonics that operated during the Archean and Proterozoic, when much of the continental lithosphere was formed. The exact timing of the onset of plate tectonics and modern subduction is unresolved (e.g., Furnes et al., 2007; Chen et al., 2009; Smart et al., 2016). Therefore, it is unclear whether the processes that currently shape the continental lithosphere were active and operating in the same manner that they do today. Several hypotheses have been proposed for cratonic lithosphere formation, and they can be categorized into two groups: (1) those that invoke processes unique to the pre-plate tectonic Earth and (2) those that call upon modern day plate tectonic processes. Each offer an explanation for the observed stability of cratonic lithosphere as well as the basic structure of a chemical boundary layer residing within the thermal boundary layer. These hypotheses should also be able to provide context for the geochemical and seismological observations of cratonic lithosphere. Furthermore, each must also account for the longevity of cratonic lithosphere which is different than solely describing stability.

Regardless of the exact dynamic regime, the conditions experienced by cratonic lithosphere will change with time after formation. For example, convective stresses increase as the Earth cools. This is due to the dependency of the mantle's viscosity on temperature - as the Earth cools, the average mantle temperature decreases and the viscosity increases which nets in increased convective stresses (e.g., Cooper and Conrad, 2009; Sandu et al., 2011). In addition, the condition of isopycnicity, which explains the present day neutral buoyancy of cratonic lithosphere, may not have been met in the past when the mantle was hotter and the average thermal boundary was thinner (Eaton and Perry, 2013). Thus, creating stable and long-lived cratonic lithosphere requires producing a feature that can resist deformation shortly after formation and from that moment onward. The longevity issue prompts some to propose that the composition of cratonic lithosphere can only be produced during the early Earth (O'Reilly et al., 2001; Poudjom Djomani et al., 2001) or that cratonic lithosphere is preferentially buffered from deformation by surrounding, slightly weaker material (Lenardic et al., 2000; Cooper and Conrad, 2009).

We do note that there is evidence that some cratonic regions have not survived to present day (e.g., Gao et al., 2004; Abdelsalam et al., 2011; Guo et al., 2016). This does hint at the possibility that whatever processes form cratonic lithosphere may not always succeed in building long-lasting stable features. However, for the scope of this paper, we limit our discussion to creating and preserving, rather than destroying, cratonic lithosphere.

Here we outline the arguments for the formation hypotheses as well as outline the predicted lithospheric structure produced by each as well as their explanations for stability and longevity. We will consider these within the framework of the observations made in the previous section as well as critically evaluate their potential to form long lasting, stable lithosphere containing complex structures.

4.1. Hypotheses based on a pre-plate tectonic regime

Much of the debate surrounding the onset of plate tectonics depends on the uncertainty as to how different parameters respond to higher mantle temperatures in the past. Several of the ideas for cratonic formation call upon processes that would only be in operation during times of higher average mantle temperatures. This has the benefit of explaining the seemingly uniqueness of craton lithosphere, but, in some cases, removes the advantage of correlation to modern day processes. The majority of the pre-plate tectonic ideas centralize around primarily vertical motion (Fig. 7A & B), as large-scale lateral motion is more indicative of the plate tectonic regime. As such, the lithospheric structure produced in such settings would be indicative of processes driven by upwellings, drips and/or regional lateral motion (in response to isostatic adjustments or gravitational collapse).

An early idea for craton formation invoked large and deeply seated melting events caused by large mantle plumes (Fig. 7A) or upwellings (e.g., Pearson et al., 1995; Robin and Bailey, 2009; Aulbach, 2012). It is suggested that the melting occurred through a large single event triggered by a plume (e.g., Pearson et al., 1995). It is speculated that, in the early Earth, the hotter mantle temperatures allowed for incipient melting to occur at greater depths within a plume (e.g., Lee et al., 2011). This would provide a large enough melting event to produce residuum up to several hundreds of kilometers thick. In addition, the higher temperatures promote higher degrees of melting such that the residuum is strongly depleted and dehydrated. This provides the means to make thick, buoyant and highly viscous cratonic lithosphere in a single event. The argument for a single melting event centers around the similar crustal and lithospheric ages in some cratons indicating a coupled origin (e.g., Pearson et al., 1995; Carlson et al., 2000, 2005; Griffin et al., 2003). Other geochemical evidence, however, argues that the melting associated with cratonic lithosphere occurred at lower pressures (e.g., Lee, 2006) which calls into question the plume driven large, single melting event.

Others have suggested that the cratonic lithosphere was formed through diapirism and vertical tectonics where dense volcanic material is moved vertically deeper into the mantle in response to a density inversion (Fig. 7B). In this scenario, cratonic crust and lithosphere is made through successive sorting events. Robin and Bailey (2009) assumed that this process was driven by an unstable compositional stratification caused by the emplacement of mafic volcanics on top of felsic crust. The density inversion and incubation of the more radiogenic felsic crust could drive ductile deformation and the development of domes and keels (Sandiford et al., 2004). Robin and Bailey (2009) argue that the cratonic lithosphere could have formed by the successive vertical accumulation of the residuum generated from melting the mafic volcanics that moved downward during the keeling process or from melting driven from delamination events often referred to as “sagduction” (Bédard, 2006; Bédard et al., 2013). While the diapirism can explain the dome and keel structure of some Archean crust (e.g. Sandiford et al., 2004) and potentially provides an explanation for the coeval evolution of the crust and lithosphere, it has not yet been demonstrated that it is possible to form a root several hundred kilometers thick through the vertical migration of dense melts.

Similarly, another proposed tectonic setting for the early Earth, the heat pipe regime, allows for the rapid removal of heat through pervasive, localized volcanism as well as the development of thick lithosphere via vertical transport. It was first proposed to explain heat loss on Venus and Io (e.g., Turcotte, 1989; Armann et al., 2012; Moore, 2001), both terrestrial bodies demonstrating little horizontal motion, but significant volcanism. The heat pipe mode allows for efficient cooling of the hot interiors (e.g. as induced by tidal heating in the case of Io or heat from the decay of radioactive elements for Venus (e.g., Turcotte, 1989; Armann et al., 2012; Moore, 2001)) without invoking plate tectonics. This idea was extended to the early Earth to explain the removal of heat in conditions wherein plate tectonics might not be operating. The early Earth's mantle was hotter in the past due to the higher amount of radioactive elements and residual heat from the formation of the planet, yet early estimates of the temperature gradient across the lithosphere suggests surface heat flux values similar to present day (Burke et al., 1978). During a heat

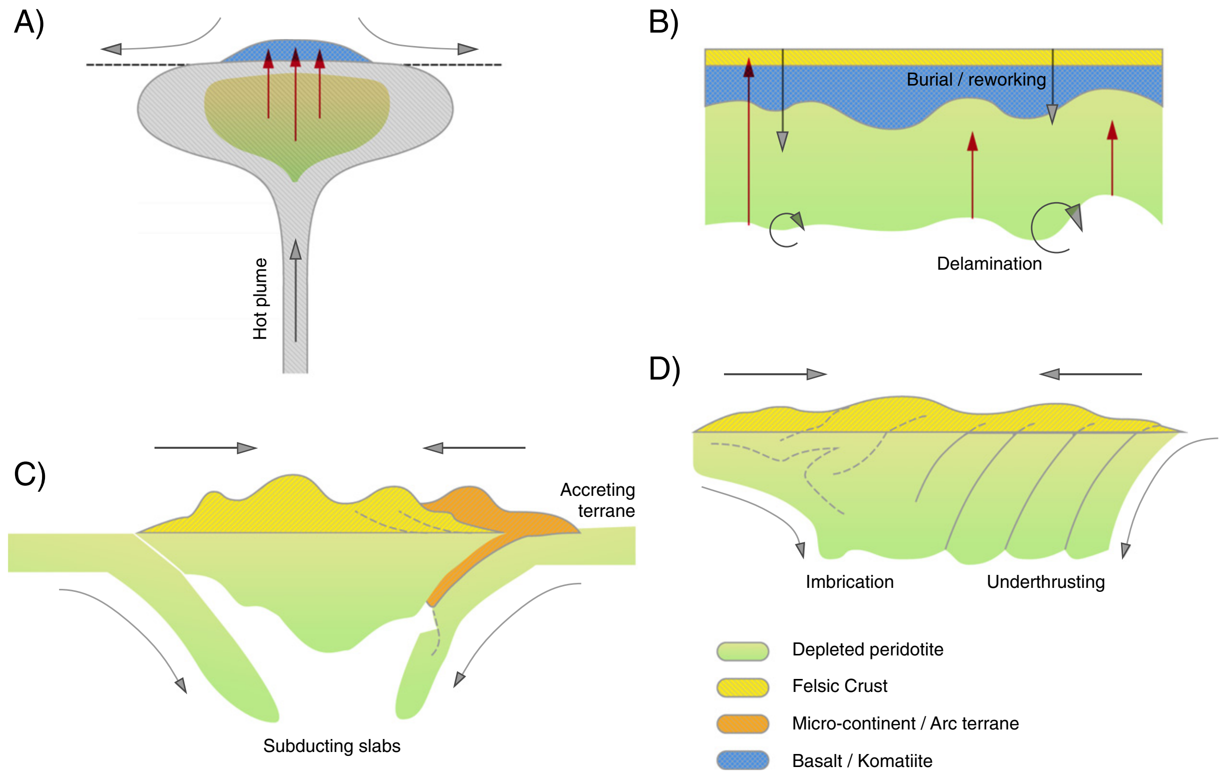

Fig. 7. Conceptual sketches of craton formation. Grey arrows show sense of motion. Red arrows indicate melt transport. (A) Melting event trigger by large mantle plume as described in Pearson et al. (1995) and Lee et al. (2011). (B) Burial/reworking and sorting of dense materials (e.g., Robin and Bailey, 2009; Moore and Webb, 2013; Sizova et al., 2015). Though not explicitly shown in this sketch, diapirism can further aid the resorting of dense materials. (C) Lithospheric thickening through the amalgamation of arc material (e.g., Jordan, 1978). (D) Thrust stacking and imbrication of buoyant oceanic lithosphere (e.g., Bostock, 1998).

pipe mode of convection, the entire surface acts as a single plate and rapid, localized volcanic eruptions connect, as "pipes", the hot, convecting mantle to the surface, effectively removing heat. The volcanic eruptions pile up on the surface burying cooler regions of the lithosphere and thus, building a cool, thick lithosphere even in the midst of a hotter mantle. Depending on eruption rates, the rapid and successive volcanism described in Moore and Webb (2013) might also explain the observation of coeval cratonic crust and lithosphere (e.g., Pearson et al., 1995; Carlson et al., 2000, 2005). This process also would promote melting at shallow depths satisfying the geochemical requirement for low pressure melt production (Lee, 2006). In a modified form of the heat-pipe hypothesis, which included the possibility of intracrustal melting in addition to rapid volcanism, Sizova et al. (2015) demonstrated that eruption of melts followed by their burial and dynamic overturning could generate the observed Archean crustal composition and depleted mantle lithosphere.

An additional formation hypothesis that could have also operated within (or as a consequence from) a heat pipe or other early Earth tectonic regimes calls upon the gravitational collapse of thick lithosphere (Rey et al., 2014). This idea proposes the surface of the Earth, though immobilized by stagnant lid convection, is divided into regions of thick, continental lithosphere and oceanic lithosphere. The authors propose that variations in the thickness of the early lithosphere could have driven horizontal gravitational forces sufficient to cause yielding of the adjacent thinner oceanic lithosphere simultaneously initiating subduction and collapse within the continental lithosphere (Rey et al., 2014). The gravitational collapse initiates decompression melting at the base of the continental lithosphere. This melting event introduces depleted residue at the base of the thinned continental lithosphere. In this model, the lithosphere continues to thicken and strengthen via cooling. The authors also propose that as subduction develops, the cratonic lithosphere along the newly formed margins will experience metasomatism further characterizing the geochemical signature of cratonic lithosphere (Rey et al., 2014). This hypothesis depends on the assumption of pre-existing regions of thick and buoyant cratonic lithosphere. The authors argue that these regions would have formed by differentiation within buoyant oceanic plateaus (possibly akin to the dome and keel behavior) (Rey et al., 2014). There is also significant tradeoff between rheological parameters of the initial cratonic lithosphere and oceanic lithosphere necessary to allow for the gravitational collapse and subsequent initiation of subduction. The initial cratonic lithosphere needed to be strong enough to maintain a thickness differential, but weak enough to eventually, catastrophically collapse. Regardless, this mechanism does provide a unique and one-off scenario to form thick and stable cratonic lithosphere without invoking plate tectonics.

All of these hypotheses are capable of forming a strong, thick chemical boundary layer residing within the thermal boundary layer. Stability and longevity within these scenarios are provided by the unique composition and rheology induced by the high degree melting. The lithosphere produced by a single, plume melting event would exhibit compositional stratification due to the gradual decline of melt-depletion in the plume structure with depth (e.g., Lee et al., 2011) whereas the others might produce compositional stratification due to the progressive layering of dense residuals (e.g., Rey et al., 2014). All would produce planar structures related to the melting event(s). Recently, there has been observational evidence for a mid-lithospheric discontinuity (MLD) in Ontong Java Plateau interpreted as a remnant of the melting event that formed the oceanic plateau (Tharimena et al., 2016). This suggests a mechanism for producing MLDs observed within cratonic lithosphere produced by large melting events associated with plumes or successive volcanic eruptions. However, these hypotheses fail to describe the more complex nature, mentioned above, of some of the MLDs observed within the cratonic regions that show continuous dipping surfaces (e.g., Bostock, 1998; Miller and Eaton, 2010) and even multiple MLDs in a single location (e.g., Hopper and Fischer, 2015; Calò et al., 2016) suggesting a differing origin than melting events alone, at least in those areas.

4.2. Hypotheses based on early development of plate tectonics

There are several lines of evidence that suggest that plate tectonics could have been in operation during formation of the cratonic lithosphere. These range from geochemical signatures of subduction within cratonic mantle xenoliths (e.g., Lee, 2006) to seismic imaging of complex structure deep within the cratonic lithosphere (Bostock, 1998; Miller and Eaton, 2010; Yuan and Romanowicz, 2010; Cooper and Miller, 2014; Wirth and Long, 2014; Hopper and Fischer, 2015; Calò et al., 2016). Indeed, seismological observations seem to be converging on the idea that the pervasive mid-lithospheric discontinuity observed in the thick, Precambrian lithosphere is a likely relic from formation (Bostock, 1998; Bastow et al., 2010; Miller and Eaton, 2010; Yuan and Romanowicz, 2010; Cooper and Miller, 2014; Wirth and Long, 2014; Hopper and Fischer, 2015; Calò et al., 2016). In the framework of plate tectonics, cratonic lithosphere was formed by thickening pre-existing lithosphere in a subduction zone setting. Early proponents of this idea suggest that cratons were formed through the thrust stacking of buoyant oceanic lithosphere incapable of participating in subduction (Bostock, 1998) or through the amalgamation of arc material (Jordan, 1978). Both of these processes are dominated by lateral motion and large-scale deformation (Fig. 7C & D). Indeed, a puzzling component of this hypothesis is the capacity for high-deforming processes to form stable, long-lived features.

Cooper et al. (2006) demonstrated the feasibility of forming stable lithosphere either through thrust stacking of buoyant oceanic lithosphere or accretion/accumulation of arc lithosphere. The authors used a combination of numerical simulations and theoretical scaling relationships to map out the parameters that allowed for thickening of lithospheric material over a downwelling in a convection cell. Two scenarios arose: the lithospheric material thickening via localized deformation (akin to thrust stacking) or through viscous deformation (more indicative of arc amalgamation) (Cooper et al., 2006). Both required the initial material to be buoyant so as to resist entrainment into the mantle downwelling. Indeed, this limitation is what drove Jordan (1978) and Bostock (1998) to suggest either arc lithosphere or buoyant oceanic lithosphere as the thickened proto-cratonic lithosphere. The material must be sufficiently buoyant to explain the observed compositional buoyancy, but also remain neutrally (or positively) buoyant (Jordan, 1978) both at the time of formation and subsequent evolution in a cooling Earth.

The differing styles of thickening (thrust stacking vs amalgamation) observed within the simulations depended on the rheological properties of lithosphere (Cooper et al., 2006). Deformation occurred through a localized manner along shear bands in simulations with stronger, more viscous lithospheric material whereas those with weaker, less viscous material deformed in a distributed fashion with no distinguishable shearing (Cooper et al., 2006). Thus, the manner of lithosphere thickening invoked to explain craton formation is dependent on the rheology of the proto-cratonic lithosphere. Gray et al. (2010) also demonstrated the controls that rheology plays on simulations of continental collision, though this study focused more on the crustal influence on deformation rather than the entire lithosphere. Regardless of the differing rheological framework, both styles of thickening produced thick, stable features within the simulations wherein deformation within the lithosphere ceased, or significantly decreased (small-scale dripping occurred along the base of the viscously thickened lithosphere, but large-scale deformation ended), even though convection continued (Cooper et al., 2006).

The act of the thickening itself drove the transition to stability within these simulations. As the lithospheric thickness increased, the effective yield stress increased (necessary to cease localized shearing) as well as the integrated buoyancy (suppressing entrainment, delamination or dripping) (Cooper et al., 2006). Both of these conditions depend not only on the material properties of the lithosphere, but also on the dynamic setting of the convecting mantle. In other words, to turn off yielding within the lithosphere the effective yield strength must be higher than the convective stresses. To resist entrainment, the integrated buoyancy must exceed the negative thermal buoyancy of the mantle downwelling. Thus to achieve longevity, the lithosphere must exceed a critical thickness to ensure stability not only in the past dynamic setting when formed (e.g., Lenardic and Moresi, 1999), but also sufficiently thick to offset the increase in convective stresses and negative thermal buoyancy of the cooling, thermal boundary layer (Cooper et al., 2006). The progressive cooling of the cratonic lithosphere itself will contribute to an additional strengthening due to the temperature-dependent rheology. This is a necessity for the more distributed thickening regime which does allow for small drips to form at the base of the lithosphere. Cooling of the temperature-dependent lithosphere will help suppress potential thinning from these drips (Molnar et al., 1998). The localized thickening (or thrust stacking) behavior does not experience these drips due to the high viscosity required for the behavior to develop, but stability through this process does assume healing of shear zones once the stresses to promote yielding are no longer present (Cooper et al., 2006).

Much like the Rey et al. (2014) continental collapse model, all of these models requires assuming a priori an initial lithospheric material that is buoyant from the onset. This is where these lines of models run into difficulty, in particular, the thrust stacking of buoyant oceanic lithosphere. The buoyancy in the buoyant oceanic lithosphere is supplied by thick basaltic crust (a consequence of greater melt production driven by higher mantle temperatures in the past). This buoyancy is most likely not a permanent condition as basalt transitions to the denser eclogite under pressure. In present day subduction zones, the density change associated with this metamorphic reaction helps drive the downward motion of the subducting oceanic lithosphere (Peacock, 1993). If cratonic lithosphere is formed by the progressive thickening of buoyant oceanic lithosphere, buoyancy would be lost if a significant portion of the basaltic crust transforms to eclogite once at depth. Arc lithosphere is also problematic as mafic cumulates are the complementary residue to the more felsic or intermediate buoyant crust (e.g., Rudnick, 1995). Thus, it is unclear that the buoyancy required to thicken either material through lateral motion can be achieved without preferentially removing the dense material either during thickening (Kay and Kay, 1993; Krystopowicz and Currie, 2013; Wood, 2014) or post formation through the overturn of the entire thickened lithosphere (Percival and Pysklywec, 2007).

Regardless whether thrust stacking of buoyant oceanic lithosphere, arc accumulation or continental collision, the thickening of the lithosphere as driven by lateral motion introduces complex structure at depth. Cooper and Miller (2014) expanded upon the earlier models in a manner that highlighted structures that would arise from such large-scale deformation. The authors show that if deformation occurred in a more distributed manner, horizontal planar features would be produced, but if deformation localized into shear zones, dipping structures developed in the lithospheric interior (Cooper and Miller, (2014) - see also Fig. 8C). Similar studies (e.g., Gray et al., 2010) also show the dependency of lithospheric structure on deformation style. This provides an explanation for the varied nature of lithospheric structure in differing cratons regions; thickening driven by lateral motion can produce a range of features depending on the rheological conditions. This suggests that some form of lateral accretion was likely at work during the formation of cratonic lithosphere given the observed complexity within the deep structure, though there are still many unanswered questions about the origin of the material that has been thickened and whether the thickening process was driven by true plate tectonic behavior or by the transition toward modern day processes.

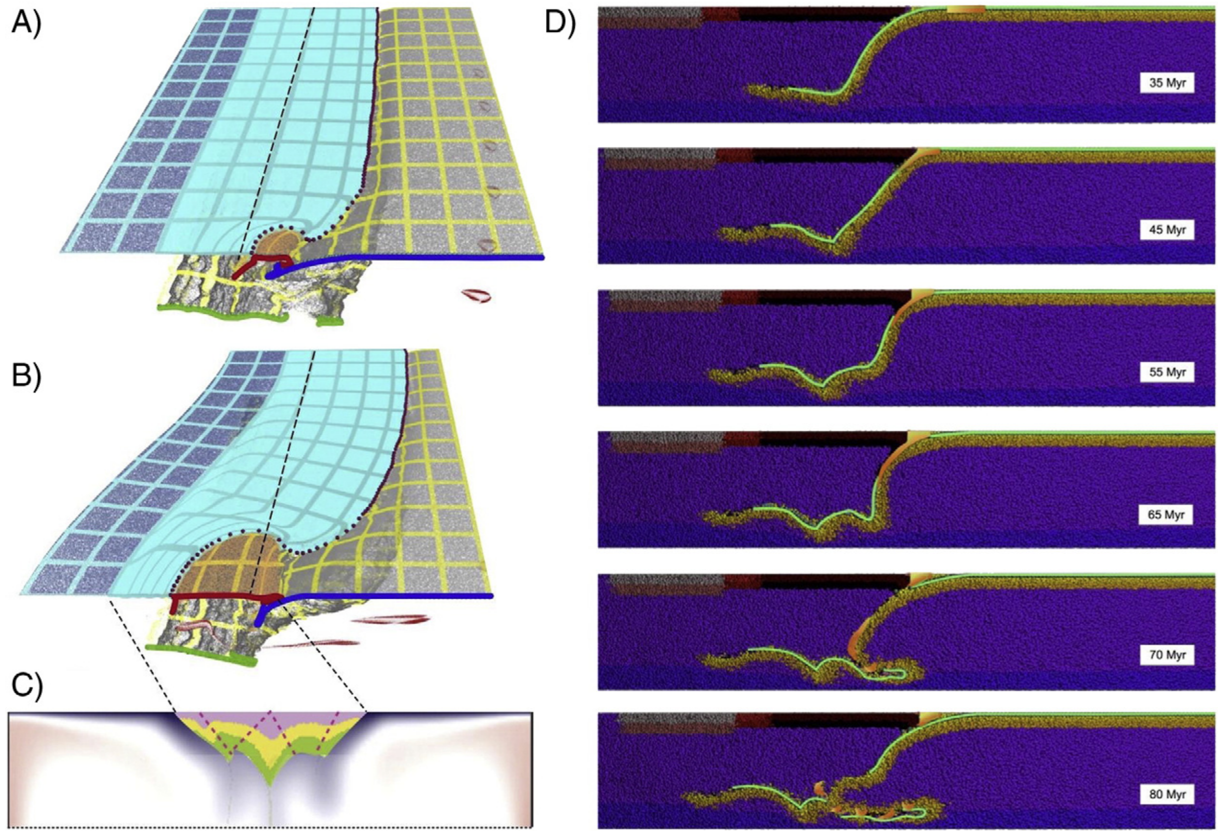

Fig. 8. Three dimensional models of accretion of (A) a 250 km and (B) a 500 km radius oceanic plateau (shown with orange markers) into a deformable margin (snapshots of movies from Moresi and Willis (2015) showing only their marker surfaces). The oceanic plate is grey with a yellow strain-marker grid. The continental crust is dark blue with a light grid. Transitional crust is cyan with a grey grid. The indenting plateau is orange. The plate boundary is a purple dotted line. The dark, dashed line indicates the position of the trench at the onset of collision. The colored lines indicate the cross-sectional view of the pre-collision lower-plate lithosphere (green), the plateau (red), and the post-collision lower-plate lithosphere (blue). (C) shows the correspondence between the high-resolution 2D, thermal-mechanical models of Cooper and Miller (2014) and the compositional-mechanical 3D models. (D) is a cross section through the collision of a 125 km radius plateau showing all material elements (Betts et al., 2015) and illustrating a geometry similar to that of the Yakutat subduction and accretion.

5. Comparison with modern analogues

While geodynamic models clearly suggest a mobile-lid, laterally-driven model of thickening of the Archean lithosphere was responsible for the widely observed, dipping structures, the extent to which this process is comparable to modern accretion is still an open question. It is useful, then, to examine what we know about the structural aftermath of the lateral assembly of continental lithosphere in the modern Earth.

5.1. Accretion dynamics

The nature of plate tectonics is to draw any material on an oceanic plate inexorably into collision with a subduction zone. Whether such material is recycled into the mantle or accreted onto the overriding plate is determined by dynamic considerations related to the relative buoyancy and strength of the ingested material, the subducting slab and whether the overriding plate is sufficiently deformable to accommodate the incoming material.

For example, two-dimensional models of oceanic plateaus colliding with subduction zones by Arrial and Billen (2013) show that eclogitization of the plateau crust inhibits accretion by increasing slab negative buoyancy with the potential for underplating of the plateau after slab break-off. Tetreault and Buiter (2012) argue that the strength of the incoming material is important in allowing a buoyant anomaly to be sliced at the base from the down-going plate and emplaced on the over-riding plate. van Hunen and Allen (2011) show how congestion of the subduction zone by a strongly-buoyant continental ribbon can lead to slab break-off and stalled convergence.

In a two dimensional cross-section, the structural signatures of these forms of accretion are integrated in a study by Vogt and Gerya (2014) who propose three distinct modes of accretion. (1) Frontal accretion in which the terrane butts up against the overriding plate, and remains vertically coherent with the underlying oceanic lithosphere after slab breakoff. (2) Basal accretion in which the plateau is significantly underthrust below the overriding plate, is sheared from the oceanic, downgoing plate, and stranded after subduction restarts. This produces a dipping structure akin to that in Fig. 8C. (3) An underplating mode in which the buoyant plateau rebounds after partial subduction and is emplaced vertically into the lithosphere behind the arc. These modes are described in terms of plateau collision but can be generalized to the subduction of any buoyant anomaly carried by the sea-floor.

Some accretionary orogenic belts remain active for hundreds of millions of years during which time they accumulate a collage of terranes including arcs, back-arc, and micro-continents or continental ribbons (Wilhem et al., 2012). The age of the subducting plate will vary in time and along the strike of the margin, each of the multiple incoming terranes will have different physical characteristics that influence the style of accretion, and the nature of the overriding plate is itself evolving. Examples of Phanerozoic and Mesozoic accretionary orogens include the Altaïds in central Asia (Wilhem et al., 2012), Tasmanides along the eastern margin of Gondwana (Glen, 2005) and the North American Cordillera (Johnston, 2001).

The along-strike variability nature of the margin in North America (see map from Hildebrand (2013) for an example) reminds us that modeling in cross section is a significant simplification of the true Earth. Three-dimensional dynamic models that include laterally varying structures in both the overriding and incoming plate are essential in trying to understand these complex structures observed on Earth.

Moresi et al. (2014) and Betts et al. (2015) developed 3D models of the accretion of plateaus, oceanic large igneous provinces, and continental ribbons. Their models trace the entire lifecycle of an accreting terrane from collision, to underthrusting/suturing, indentation into the overriding plate, and finally to the point where subduction re-starts with the terrane becoming part of the upper plate. A representative of such a model is illustrated in Fig. 8 (A, B & D) and shows snapshots of the accretion of a small (125 km radius, shown in cross section), medium (250 km) and large (500 km radius) buoyant anomaly representing an oceanic plateau (Moresi and Willis, 2015; Betts et al., 2015). In all of the simulations, the plateau dimension is a small fraction of the overall length of the convergent margin allowing the subduction system to evolve smoothly through the accretion event and recover.

The smallest plateau (125 km radius - shown with orange markers) is partly eroded from beneath and the remainder is accreted at the edge of the continental margin (Fig. 8A & D). The suture is vertical near the surface, then transitions to a dipping feature at depth. A small slab window forms as the plateau is stripped from the oceanic lithosphere. The dipping section of the suture is stripped away once subduction restarts (in Fig. 8D this occurs between 65 and 70 Myr). For both the larger plateaus, the structure of the suture is similar, but the buoyant material indents significantly inboard into the margin which undergoes shortening close to the indentor, lateral escape, and modest extension in the lithosphere adjacent to the collision forming an orocline in the overriding plate. The restarting of subduction behind the indentor is through a buckling of the oceanic lithosphere driven by the adjacent, intact slab and this buckling overturns the rear of the accreted terrane and entrains a small section of the oceanic crust. The original suture is inboard of the new slab and remains undisturbed in this case.

The congestion of the subduction zone drives horizontal compression of the margin lithosphere ahead of the indentor/plateau similar to the geometry modeled by Cooper and Miller (2014) as shown in Fig. 8C, and their analysis of the conditions for localized versus distributed deformation can be applied in the plane perpendicular to the collision. However, as the indentor/plateau also produces lateral extrusion of the lithosphere, there is significant strike-slip deformation which, when localization occurs, forms the classical escape-tectonics pattern of faulting (e.g., Tapponnier et al., 1982 and shown in Moresi et al., 2014). This has the potential to greatly complicate the analysis of a long-lived accretionary margin with multiple cycles of interpenetrating collision.

5.2. Examples from Australian Tasmanides and the Alaskan accretionary margin

To place these models in context, we briefly discuss two examples: 1) southeast Australia: the collision of the Selwyn block with the extensive, Paleozoic Gondwana accretionary margin (Fig. 9) which has remained unmodified by a subsequent ocean closure (Foster and Gray, 2000); and 2) southern Alaska: the ongoing accretion of the Yakutat

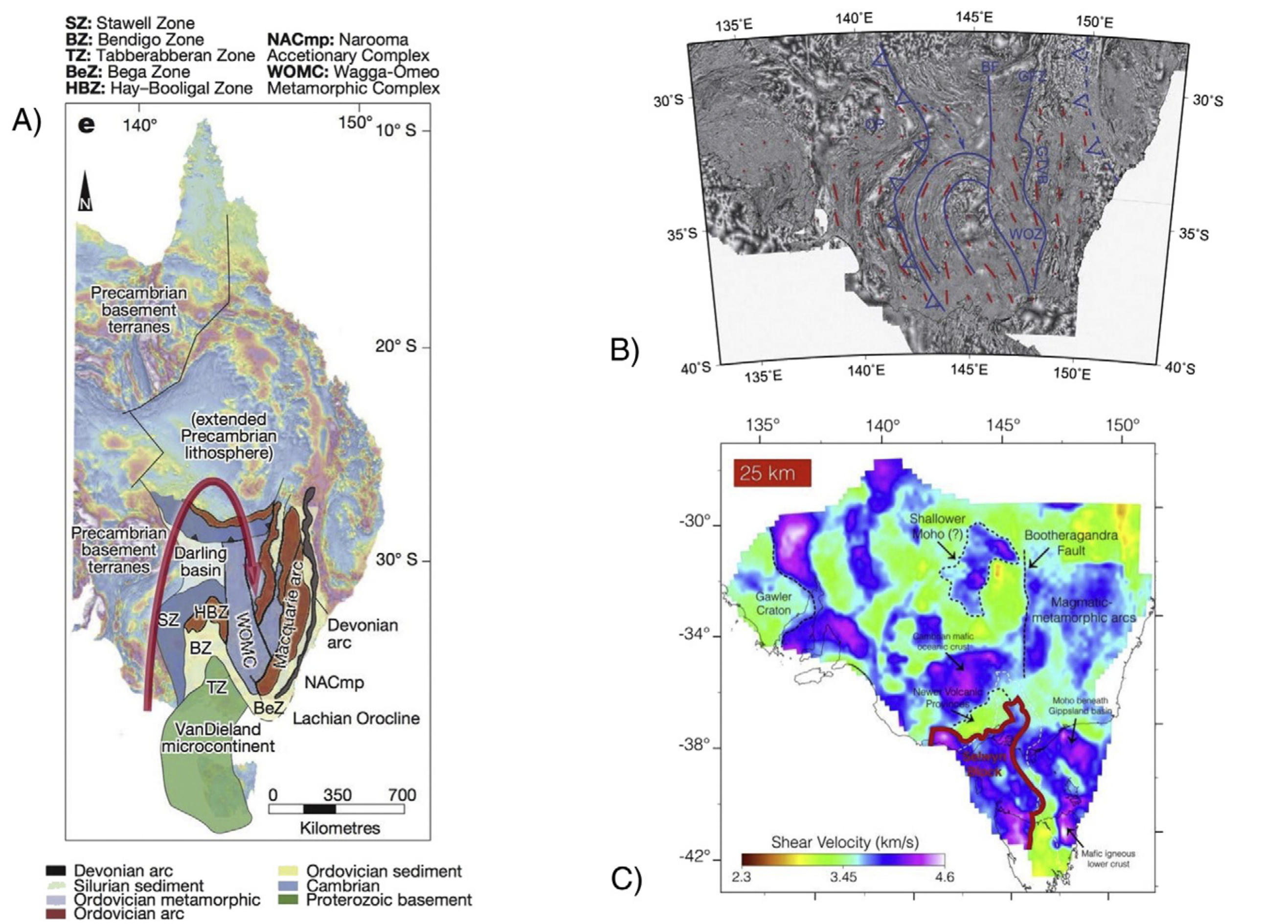

Fig. 9. (A) is a simplified regional tectonic map of the southern and central Tasmanides superimposed on a magnetic intensity map of eastern Australia (figure from Moresi et al., 2014). The curved red arrow shows the sense of overall motion predicted by numerical models and inferred from the geological and geophysical interpretation of Cayley (2011). (B) the orientation of the seismic fast direction (dark red vectors) superimposed on magnetic data with an interpretation of the inferred fabric shown in blue (figure from Rawlinson et al., 2014). (C) Seismic shear-wave velocity anomalies at 25 km from the regional tomographic model of Pilia et al. (2015) with their outline of the seismic signature of the Selwyn block at depth (red contour).

block in the eastern end of the Aleutian/Alaskan subduction zone (Fig. 10).

The structural evolution of the accretion of the Selwyn block, and the associated Lachlan fold belt and orocline system in Australia has been extensively documented by Cayley (2011) and coworkers, and supports the dynamic framework for accretion outlined above. Recent observations of seismic anisotropy and interpretation of magnetic fabric, together with regional tomographic imaging further support the causal

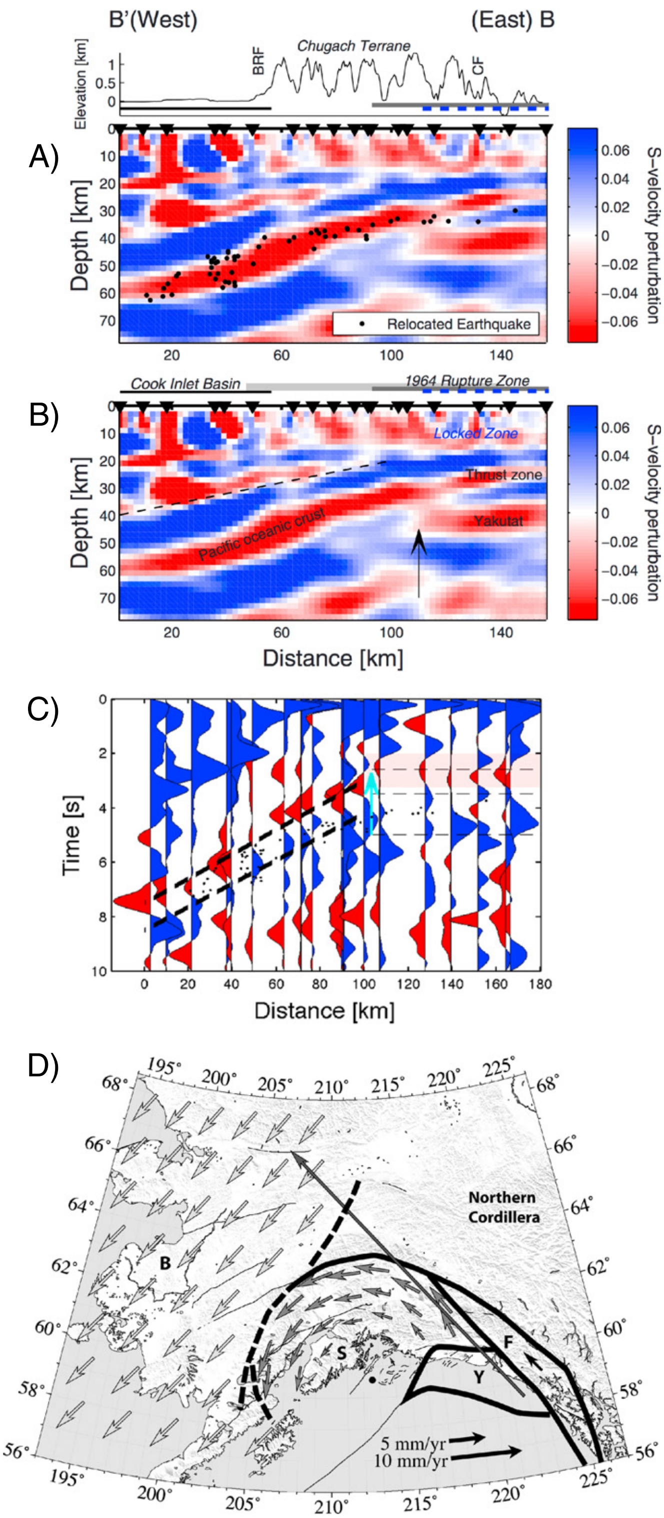

Fig. 10. (A-C) Seismic discontinuity imaging from (Abers et al., 2014) in southern Alaska showing a profile (B-B′) that cuts obliquely across the subducting portion of the Yakutat terrane. The interpreted seismic images show the low-velocity Pacific oceanic crust and the westernmost extent of the Yakutat terrane being subducted beneath the Chugach terrane on the North American plate. The arrows indicate the position of the Yakutat in both the GRT based images (see Abers et al. (2014) for details) and the stacked P receiver functions. (D) Freymueller et al. (2008) block rotation model for the Alaskan syntaxis showing the Yakutat block indentor and the escape of material westward and southward. (E) shows a simplified tectonic map of Alaska (Colpron et al., 2007) that highlights the long history of accretion, the present day orocline, and the dissection of the margin by the incoming Yakutat block. The yellow patch indicates the approximate extent of the subducted Yakutat crust from (Eberhart-Phillips et al., 2006) and the red line is the location of the profile (B-B′) in the left hand column (A-C). (Eberhart-Phillips et al., 2006).

E)

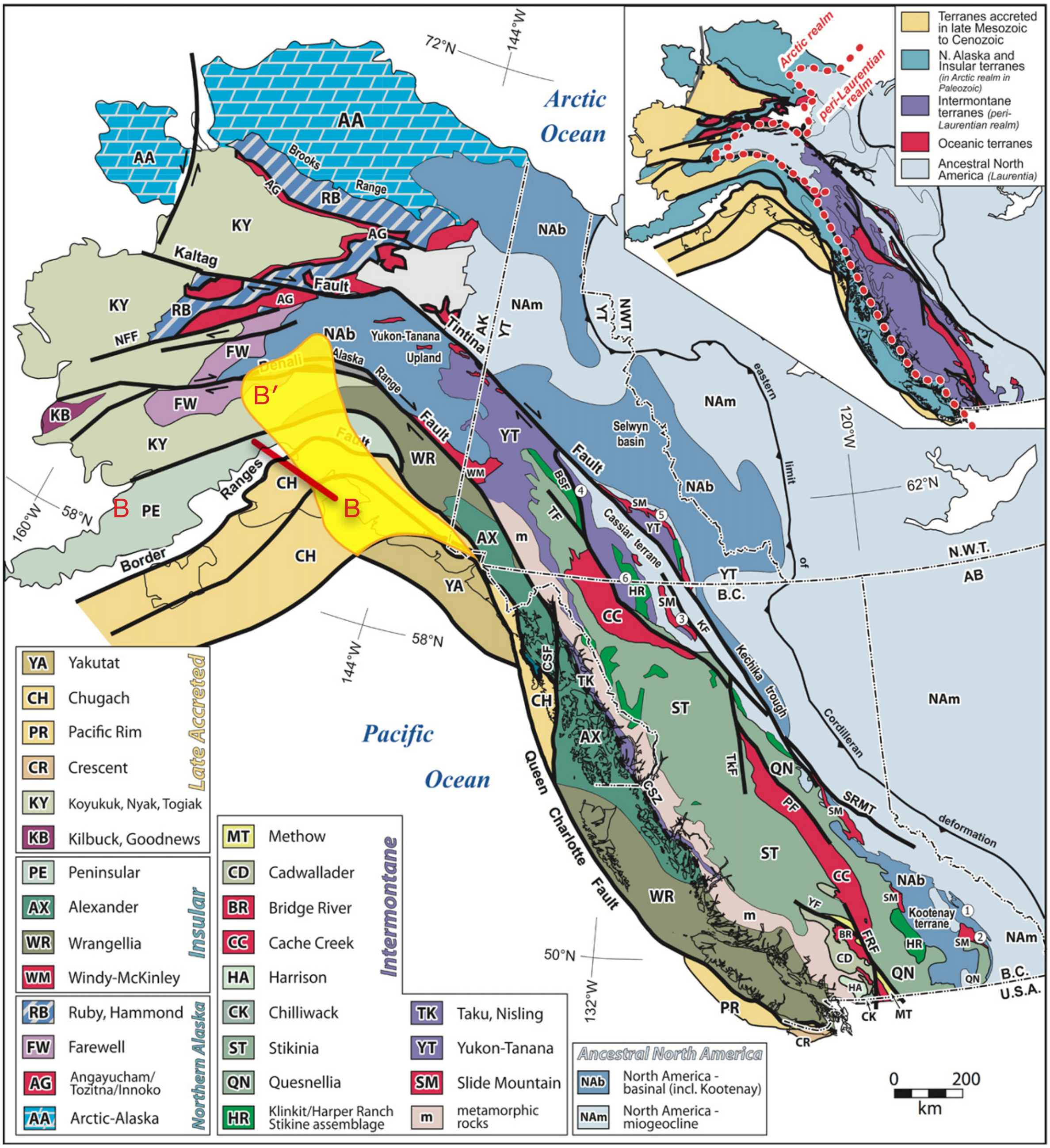

Fig. 10 (*continued*).

connection between the indentation of the Selwyn block, the associated fold belt, and the surrounding orocline (Rawlinson et al., 2014; Pilia et al., 2015).

Fig. 9 shows a simplified geological map of the accreted Selwyn block (the northern tip of the VanDieland microcontinent as outlined in red in Fig. 9C) showing the Lachlan orocline which is thought to be composed of the dissected remnants of the pre-accretion margin and arc. The curved, red arrow shows the approximate sense of motion of the tectonic escape during collision. Note the later (Devonian) arc that indicates where the subduction zone re-established itself post-accretion. The seismic evidence for the oroclinal fabric (Fig. 9B) and the indented block (Fig. 9C) is taken from Rawlinson et al. (2014) and Pilia et al. (2015), respectively.

A modern example of collision of a buoyant plateau at a continental convergent margin is at the eastern end of the Aleutian subduction zone. This collision is the most recent in a series of accreted terranes that have formed the northern Cordillera (Fig. 10E). This collision results in a section of flat-slab subduction due to the underthrusting of the Yakutat Plateau (Eberhart-Phillips et al., 2006). The Yakutat terrane continues to actively collide with the margin, (Fig. 10D: Freymueller et al., 2008), bisect the Chugach terrane and results in accentuating the curvature of the Alaskan orocline. Previous sutures have been reactivated in strike-slip deformation (Fig. 10D&E) and now form part of the orocline and associated lateral escape (Glen, 2004).

P receiver functions results (Fig. 10A-C) concentrated across the southern half of central Alaska effectively image the Yakutat Plateau offshore (Worthington et al., 2012; Christeson et al., 2013), through the forearc (including the Yakutat-Pacific boundary) (Abers et al., 2014), and northward below the Alaska Range and major strike-slip Denali

Fault (Ferris et al., 2003; Rondenay et al., 2010; Brennan et al., 2011) illustrating sharp structures at quite fine resolution (a few kilometers).

As the models show in Fig. 8, the slab can flatten and/or break off in response to congestion by anomalously buoyant crust. This will result in trench advance in this section. The majority of the subduction zone is not congested and will continue to subduct although with an enhanced tendency to retreat (a geometrical requirement to accommodate the escape of material from the collision). The Yakutat collision (Fig. 10) appears to follow these general observations and has a similar morphology to the Lachlan orocline (Fig. 9). Of particular note is that, in both cases, there has been considerable lateral movement of the previously-accreted terranes through re-activation of the original suture zones. If the Yakutat collision is typical of an accreting terrane, the present shallow-dipping, underthrust section associated with flat-slab subduction, would be highly modified and, presumably, detached from the colliding section of the terrane by strike-slip reactivation of the shallow suture (Denali Fault).

6. Conclusions and suggestions for future work

There are some commonalities between modern accretion processes, the structures they produce and the structures preserved within older, cratonic lithosphere (Fig. 8). In particular, modern accretion introduces features (Fig. 10) reminiscent of the dipping structures observed in cratonic regions. It is interesting to note that a significant portion of deformation in modern accretion settings is accommodated by strike-slip motion which may not be captured in deeper lithospheric structure. Tracking anisotropy patterns, however, seems to be a way to locate the signature of accretion once the orogeny is complete. This approach may also be applicable for mapping out fabric associated with ancient orogenic events (e.g., Wirth and Long, 2014; Pilia et al., 2015). In addition, while vertical tectonics are still in operation on modern Earth (e.g., lithospheric drips in the western United States, see Zandt et al., 2004 and West et al., 2009), it is unclear as to whether these events are part of the accretionary process or a consequence of other controls on continental deformation. If the latter, these more modern events could also serve as modern analogues to test against the hypotheses of craton formation that invoke vertical tectonics.

Regardless, continued mapping of a more detailed view of the structure within continental lithosphere can help continue to determine the similarities between modern and ancient accretion as well as provide valuable information about the composition and rheology of cratonic lithosphere. Within models of lithospheric deformation, the structure produced is dependent on its material properties as well as the dynamic setting (Cooper et al., 2006; Gray et al., 2010; Cooper and Miller, 2014). Dipping structures are more indicative of localized deformation occurring in shear zones in strong lithosphere. Planar features result from distributed deformation in weaker lithosphere. Planar features could also be produced by melting events as described in the dynamics section. Placing the structural observations within the geodynamic context can help winnow the candidates for proto-cratonic lithosphere (i.e., buoyant oceanic lithosphere, arc lithosphere or potentially another unexplored option).

The observational similarities between modern and ancient lithosphere suggests that lateral accretion of lithospheric material played a role in craton formation especially when coupled with geodynamic models that demonstrate the processes that produce similar structures. However, whether this lateral motion was a consequence of fully developed plate tectonic behavior or of a potentially, unstable transition period between regimes still remains to be resolved. The transition from whichever early tectonic regime in operation in the early Earth to plate tectonics was most likely not instantaneous and probably tumultuous. Understanding this transition could help provide a solution for the observed differences between modern and ancient accretion. We do note that most of the geodynamic simulations for craton formation and/or continental collision during the early Earth are limited to the two-dimensional geometric framework. Further work would be advanced by exploring three-dimensional studies.

While the more complex structure of continental lithosphere does share similarities with modern accretion, there are still many unanswered questions, setting the stage for exciting avenues of future work. For example, modern accretion does not seem to produce as thick a lithosphere; why not? Is making thick lithosphere limited to the Archean and Proterozoic? If the proto-cratonic lithosphere is either buoyant oceanic or arc lithosphere, what are the steps of removing dense material while still maintaining the stability and thickness associated with the present day state of cratonic lithosphere? What would accretion in the early Earth look like if driven by weaker subduction zones as proposed by van Hunen and van den Berg (2008) or if it occurred as a consequence of the breakdown of the heat pipe regime (Moore and Webb, 2013)?

Another intriguing question is the ability for the lithosphere to retain or lose evidence of deformation. Cratonic lithosphere appears to have a long memory of past deformation. Is this dependent on unique material properties or the lack of exposure to additional deformation? Interestingly, younger lithosphere, at times, also retains evidence of past events. Recently, McKenzie et al. (2015) observed that when global lithospheric thickness maps were reconstructed to the configuration of Pangea that the thicker regions aligned in a continuous manner. This suggests that today's regions of thicker lithosphere (minus cratonic regions) are holding on to the evidence of the last continental collision. However, perplexingly, most of the observed mid-lithospheric discontinuities (MLDs) are in cratonic regions. Is this because younger lithosphere does not retain them as easily or because the deformation is accommodated differently? Furthermore, there is significant evidence that suture zones and other features are often reactivated in more modern settings (Holdsworth et al., 2001). Yet, if some of the observed structures in cratonic lithosphere are suture zones, what makes them resistant to reactivation? Does this indicate healing or lack of sufficient and/or appropriately oriented stresses? And does this connect to other pressing questions about craton stability such as why have some cratonic regions have deformed and been "destroyed" (Gao et al., 2004; Abdelsalam et al., 2011; Guo et al., 2016) while other seems to deflect deformation to their exteriors (e.g., Koptev et al., 2015)?

We should note that there are other interpretations for the source of the mid-lithospheric discontinuities including phase transitions (Selway et al., 2015) or partial melting (Thybo and Perchuc, 1997) indicative of late stage (or near present day) modification to the continental lithosphere or grain boundary sliding (Selway et al., 2015; Karato et al., 2015). Regional studies, as proposed below, would help further delineate between these differing explanations for the mid-lithospheric discontinuities. Even if the presence and nature of the mid-lithospheric discontinuities prove to be less than relevant to the tectonic evolution story, there are plenty of other techniques, several of which mentioned, that can be used to investigate the deep structure of the continental lithosphere.

We can start addressing many of these unanswered and pressing questions with additional focused, regional studies of both cratonic and accretion settings that couple seismic imaging, gravity, and magnetotelluric studies. Similar approaches such as (Jones et al., 2002; Afonso et al., 2016) can help further identify and detail characteristics of modern and ancient deformation. Focus should also be made to chronicle the differences between cratonic lithosphere from different regions (both geochemically and seismologically). Too often it is tempting to group regions based on similarities into one conceptual model (such as in the cluster approach of Fig. 1) and leave it at that. Perhaps, however, the differences in cratonic lithosphere may further highlight the processes at play during their formation and may suggest that not all cratonic lithosphere was formed in the same manner. Finally, we argue for continued collaboration between observationalists and geodynamicists. Some of the early mysteries of the Earth are trapped deep within the continents. We are moving closer to piecing the clues together, but it will certainly take a team effort.

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