Seismic tomography of the Arctic region: Inferences for the thermal structure and evolution of the lithosphere

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Abstract: New waveform tomography with very large datasets yields improved resolution in the upper mantle of the Arctic region. Computational petrology enables quantitative estimation of temperature from tomography in the lithosphere-asthenosphere depth range. Most boundaries of the mantle roots of cratons in the Arctic follow their geological boundaries at the surface. The roots of the Greenland and North American Cratons are separated by the Baffin Bay and a thin-lithosphere corridor through the middle of the Canadian Arctic Archipelago; the southern archipelago is a part of the North American Craton. The mantle root of the cratonic block beneath northern Greenland may extend westward as far as central Ellesmere Island. The Barents and Kara Sea show high velocities indicative of thick lithosphere, similar to cratons. Locations of intraplate basaltic volcanism of the High Arctic Large Igneous Province are all on thin, non-cratonic lithosphere. The lithosphere beneath the central part of the Siberian Traps is warmer and thinner than elsewhere beneath the Siberian Craton; this indicates the likely location of the lithospheric erosion associated with the Large-Igneous-Province volcanism. A corridor of relatively low seismic velocities cuts east-west across central Greenland and appears to delineate the track of the Iceland Hotspot.

Keywords: Arctic, tomography, lithosphere, mantle temperature, orogeny

Supplementary material: figures with comparisons of different tomographic models at 50 and 200 km depths are available at...

The Arctic region comprises lithospheric blocks with very different properties and very different tectonic history (Fig. 1). The continental units range in age from Archean cratons to Phanerozoic orogens (Pease *et al.* 2014). The deep-water High Arctic includes the Canada Basin, probably over 100 m.y. old, and the younger Eurasia Basin associated with modern spreading along the Gakkel Ridge (Herron *et al.* 1974). The intraplate basaltic volcanism of the Cretaceous High Arctic Large Igneous Province (HALIP) (Tarduno 1998; Saumur *et al.* 2016) has disrupted and modified pre-existing lithosphere across a broad area. The Cenozoic opening of the adjacent North Atlantic was accompanied by volcanism at and near the Iceland

Hotspot, with the hotspot activity likely to have affected the thermal evolution of both the oceanic lithosphere and the continental margins of the North Atlantic (e.g., White and Lovell 1996).

Understanding the structure and evolution of the different tectonic units—and of their inter-relationships within the Arctic region as a whole—requires understanding the structure and evolution of the entire lithospheric plates, from the crust to the lithospheric mantle. Two of the key parameters that determine the physical nature and mechanical behaviour of the lithosphere are its temperature and thickness. The two are closely related: the bottom of the thermal lithosphere (the lithosphere-asthenosphere boundary) is defined by the depth near which the steep, conductive geotherm that characterises the lithosphere meets the mantle adiabat that characterises the convecting asthenosphere below (Fig. 2).

Both the temperature and strength of the lithosphere also depend on the thickness and properties of the crust. Radiogenic heat production in thick continental crust makes continental lithosphere warmer than oceanic lithosphere of the same thickness, with the latter producing little heat within its thin basaltic crust (Fig. 2). Colder lithosphere is generally stiffer, although the mechanical behaviour of the lithosphere also depends on the strength layering within it (e.g., Burov & Watts 2006).

Because the lithosphere is cooled at the Earth's surface, it will tend to grow in thickness until its lower portion becomes unstable and delaminates, sinking into the underlying mantle. The thickness of the lithosphere thus reaches a maximum, and this maximum is determined by the composition of the lithospheric mantle. The lithosphere beneath Archean cratons is compositionally buoyant due to its depletion in basaltic components (e.g., Jordan 1975; 1988), so that it has a greater thickness than lithosphere elsewhere (e.g., Artemieva 2006). In seismic tomography models (which map variations in temperature-dependent seismic velocities), the cold, thick lithospheres of cratons around the world are manifest as prominent high-velocity anomalies, seen down to 200-250 km depths (e.g., French *et al.* 2013; Schaeffer and Lebedev 2015; Debayle *et al.* 2016).

Phanerozoic mantle lithosphere, in contrast, is less depleted and lacks the compositional buoyancy to grow as thick as the lithosphere of Archean cratons. The maximum thickness of Phanerozoic lithosphere has been estimated at around 150 km (Poudjom Djomani *et al.*, 2001), with a more typical thickness probably around 100 km. Tomographic models show seismic velocities within Phanerozoic lithosphere that are substantially lower than within the cratonic Archean lithosphere, reflecting much higher temperatures within the former.

Oceanic lithosphere appears to cool and thicken largely monotonically, from its origin at a mid-ocean ridge to its subduction (e.g., Maggi *et al.* 2006; Schaeffer and Lebedev, 2013; Burgos *et al.* 2014). The cooling can be interrupted by the plate passing over a mantle hotspot, which will re-heat the lithosphere; subsequently, the plate will resume its cooling and thickening. The flattening of the bathymetry of old (>80 Myr) oceans and the apparent independence of heat flow on their lithospheric age have suggested that the thickness of the oceanic lithosphere may reach a maximum when the seafloor is c. 80 Ma (the “plate model”, e.g., Stein and Stein 1992; 2015). Interestingly, seismic velocity structure and azimuthal anisotropy show evidence for the lithosphere continuing to grow in thickness

along the isotherms of the “half-space cooling model” with age (Davis and Lister, 1974), without reaching a maximum (Becker *et al.* 2014; Schaeffer *et al.* 2016a).

Generally, the lithospheric temperature and thickness beneath both continents and oceans reflect the age and the evolution of the lithosphere. A craton, a Phanerozoic platform, and an active orogen will all have distinctly different lithospheric temperatures and thicknesses, and so will an old and a young part of an ocean basin. Lithospheric temperatures can, thus, inform us of key properties of tectonic terranes and of the locations of the boundaries between them.

Lateral variations of temperature within the mantle lithosphere can be inferred from seismic velocities. Apart from temperature, seismic velocities also depend on the composition of the rock and on the presence of water or partial melt within it. The effect of compositional variations in the mantle, however, is much smaller than that of temperature: it can account for less than 1% of shear-wave speed variations (e.g. Deschamps *et al.*, 2002; Schutt and Leshner, 2006), whereas the positive shear-wave speed anomalies beneath cratons, for example, reach 8-10% above the global averages (Lebedev *et al.* 2009). At the opposite end of the range of seismic-velocity variations, they can be over 10% below the average beneath mid-ocean ridges and intraplate volcanic areas (e.g., Schaeffer and Lebedev 2015). In these locations, the lithosphere is thin and hot, and seismic velocities are lowered further by partial melting, in particular in the top 100 km of the mantle. In the back-arc regions of subduction zones, partial melting is enhanced by the water released from the subducting crust into the mantle.

Shear-wave speed anomalies mapped by tomography can thus offer an immediate qualitative estimate for temperature heterogeneity of the lithosphere. In order to calculate temperature from seismic velocity at a given depth quantitatively, sensitivity of elastic and anelastic properties of realistic rock compositions needs to be evaluated (e.g., Karato, 1993; Sobolev *et al.*, 1996; Goes *et al.*, 2000; Cammarano *et al.*, 2003). Recently developed methods of computational petrological modelling (e.g., Connolly 2005; Afonso *et al.*, 2008; Fullea *et al.*, 2009) enable fast and accurate calculation of density, seismic velocities, attenuation, and other properties of the rock as a function of its composition, temperature and pressure based on extensive thermodynamic databases, and offer a self-consistent framework for relating seismic and other data to the physical properties of the entire lithosphere.

In this paper, we use new waveform tomography of the Arctic to analyse the variations of seismic velocities across the region. The seismic-velocity heterogeneity resolved by the new tomographic model presents abundant evidence on the nature of the various tectonic blocks in the region, the location of the boundaries between them, and the dynamic processes at depth. We also use computational petrology to calculate lateral variations in the temperature of the lithosphere and estimate lithospheric geotherms beneath selected locations in the High Arctic.

Seismic tomography

Thanks to the explosive recent growth of global and regional seismic networks, increasingly detailed models of the lithosphere and underlying upper mantle are now becoming possible, at scales from

regional to global. In order to take advantage of the resolving power of the very large available datasets, we must apply methods that extract highly complete structural information from each seismogram and that are automated, so that all relevant data can be processed.

Waveform Tomography

Waveform tomography is an effective approach for imaging the entire crust and upper mantle at a regional or global scale (e.g., Nolet 1990; Lebedev *et al.* 1997; Lebedev & Nolet 2003; Legendre *et al.* 2012; Schaeffer & Lebedev 2013, 2014). We invert seismic waveforms that include the fundamental-mode surface waves (which travel along the surface of the Earth and sense different depth ranges depending on their periods) and *S* and multiple *S* waves (which dive deeper into the mantle and have complementary depth sensitivities that depend on the source-station distance and the order of the multiple). Regional *S* and multiple *S* waves are often triplicated and, also, interfere with each other and with the fundamental mode on the seismogram. By inverting entire seismic waveforms, we can extract information simultaneously from all these waves without the need to separate their arrivals.

The Automated Multimode Inversion of surface and *S*-wave forms (AMI; Lebedev *et al.* 2005) performs automated, accurate processing of large quantities of broadband seismograms. Each successful waveform inversion produces a set of linear equations with uncorrelated uncertainties (Nolet 1990) that describe one-dimensional (1D) perturbations in elastic structure within a finite-width sensitivity volume between the earthquake and the station. The perturbations are with respect to a three-dimensional (3D) reference model, including, in particular, a realistic crust (Lebedev & van der Hilst 2008). All of the linear equations are then solved together for 3D perturbations in isotropic *S* and *P* velocity with respect to the 3D reference model, as well as for 180-degree-periodic azimuthal anisotropy of *S* velocity. The broad period range of the fundamental mode surface waves (10-400 s) and the complementary sensitivities of the regional *S* and multiple *S* waves provide resolution from the crust down to the deep upper mantle, including the entire lithosphere-asthenosphere depth range.

Our new tomographic model, AMISvArc, is a global, shear-wave speed model of the upper mantle and crust, constructed using the same waveform tomography methods and data handling approaches as the recently published SL2013sv, SV2013NA and SL2016SvA (Schaeffer & Lebedev, 2013, 2014, 2015; Schaeffer *et al.* 2016a), but incorporating additional data, particularly in and around the Arctic region (Fig. 3). In the High Arctic, there are no stations in its central, sea-covered part, but because we use surface and regional body waves traversing between the various sources and stations, we have dense coverage across the entire region. A specific technical description and validation of the tomographic model are presented elsewhere (Schaeffer *et al.* 2016b); a description of the waveform tomography methods can be found in Lebedev *et al.* (2005) and Lebedev & van der Hilst (2008). In the following, we focus on the inferences on the lithospheric structure of the Arctic region offered by the model.

S-wave speed variations in the lithosphere and asthenosphere

Figure 4 shows map views of our model at depths in the shallow upper mantle. At 30-60 km, a dominant feature that stands out in the High Arctic is the cold lithosphere of the Canada Basin. *S*-wave speeds here are similar to those within old (>100 m.y.) oceanic lithosphere around the world (Schaeffer & Lebedev 2015). They are substantially higher than those beneath the younger Eurasia Basin (where the lithosphere is thinner and warmer) and also higher than beneath the continental margins of the Canada Basin (where the uppermost mantle is likely to be warmer due to the radiogenic heating within the continental crust).

At 80-110 km depth, cratons are the dominant high-velocity anomalies. In most cases, the deep boundaries of the cratonic roots in the mantle closely follow the boundaries mapped at the surface. A few significant exceptions to this are discussed below.

At 150 km and below (Fig. 5), there is no longer a visible contrast in seismic velocities between the central Canada Basin and the Eurasia Basin. Instead, a continuous low-velocity anomaly stretches from one basin to the other—an image of the warm asthenosphere beneath the High Arctic. This low-velocity anomaly is significantly smaller in amplitude than the one beneath the North Atlantic in the vicinity of the Iceland Hotspot, where the extremely low velocities indicate partial melting in the mantle.

In the SE corner of the Canada Basin, a high-velocity anomaly stands out at depths greater than 100 km, adjacent to the thick, cold lithosphere of the Mackenzie Craton, the NW extremity of the North American Craton (Schaeffer & Lebedev 2014). The anomaly is seen beneath the Beaufort Sea and also extends to the west and to the east of it (Fig. 5, 150-330 km).

Comparison of tomographic models

Figures 6 and 7 show a comparison of recent global upper-mantle and whole-mantle models at 100 and 150 km depths (see also the figures in the Supplementary material for 50 and 200 km depth comparisons). Each model is plotted with respect to its own global mean at the depth. (The reference model used in tomographic inversions will normally be somewhat different from the eventual global mean of the model, hence the slight difference in the appearance of AMISvArc in Figs. 4 and 5, where it is plotted with respect to its reference model, and in Figs. 6 and 7, where it is plotted with respect to its global mean.)

All the tomographic models were constrained by datasets that included surface waves, particularly sensitive to *S*-wave speed variations in the lithospheric depth range (Bartzsch *et al.* 2011; Lebedev *et al.* 2013). Some of the models also included waveforms of regional *S* and multiple *S* waves (including our AMISvArc, top left), surface-wave higher-mode measurements, or teleseismic body-wave arrival times. The quantity of data and station coverage vary substantially between the models (with AMISvArc using the most complete current station coverage in and around the Arctic, Fig. 3), and so do inversion schemes.

The comparison shows which features are consistently mapped by all the models. These include, for example, the mantle roots of the North American, East European and Siberian Cratons, where the thick, cold lithosphere is characterized by high seismic velocities, seen as dark blue and purple colours (Figs. 6, 7). All the models also show the warm High-Arctic asthenosphere and the hot asthenosphere beneath Iceland.

As we go to smaller spatial scales, the models differ. For example, all the models show a relatively thick lithosphere beneath the Barents Sea, but the distributions of the high-velocity anomalies beneath the sea, laterally and with depth, varies from one model to another (see also the Supplementary material figures). The V_s distribution here is also different in published regional surface-wave tomography (Levshin *et al.* 2007). The Barents Sea presents a challenge because of the very thick layer of sediments across much of it, not accounted for in sufficient detail in global reference crustal models. Fine-scale, regional crustal models will need to be incorporated into tomography with large, Arctic-scale datasets in order to resolve the smaller-scale lithospheric structure variations beneath and around the Barents Sea.

Some models (Savani, S40RTS) are very smooth, and others include smaller-scale structure. For the most part, the models are mutually consistent at the large scales. As we go to the finer scale structure—seen, in particular, in AMIScArc—many regional-scale features are seen clearer and present clearer evidence on the regional structure and evolution of the lithosphere. Examples of such regional-scale features include the corridor of relatively low velocities separating the cratonic roots of Greenland and North American Cratons, the low-velocity corridor cutting east-west across central Greenland, and the thinning of the lithosphere beneath the central part of the Siberian Traps.

Mapping lithospheric temperature

For a rock of a given composition at a given pressure (P) and temperature (T), we can compute the shear-wave speed within it (V_s), attenuation (important because it affects the V_s - T relationship) and other properties (e.g., Connolly 2005, Fullea *et al.* 2012). A V_s distribution such as in Figs. 4 and 5 can, therefore, be inverted for a distribution of temperature if we assume some reasonable, representative compositions for the mantle lithosphere and asthenosphere. If the composition and the rock properties that control attenuation are fixed, then there is a non-linear but unique, one-to-one relationship between T and V_s . The V_s - T inversion can thus be recast as a simple conversion by pre-computing V_s for the entire relevant ranges of temperatures, pressures and compositions.

The solutions of the tomographic inverse problem are non-unique: many different models can fit the data equally well. Specifically, our data constrain tightly the values of V_s in relatively broad depth ranges but constrain poorly small-scale radial V_s variations. Thus, instead of converting V_s values at a given single depth to T at this depth, we determine an average T over a depth range. The depth range needs to be sufficiently broad for the average temperature values within it to be robust. It also needs to capture relevant lithospheric property variations across the region. (Alternatively, one could reduce the non-uniqueness of seismic models in the course of their construction, either by introducing petrological constraints into the inversions or by formulating them as thermodynamic inversions from the beginning

(e.g., Khan *et al.* 2009, 2011; Fullea *et al.* 2012; Afonso *et al.* 2013). Development of thermodynamic inversion methods for large waveform datasets is the subject of current work.)

Figure 8 shows the average temperature in the 80-150 km depth range. It is determined as an average of the temperature values computed from V_s (Figs. 4, 5) at each depth within the range. Composition of the upper mantle at a point is assumed to fall into one of the three lithospheric compositional domains, corresponding to tectonothermal ages: Phanerozoic, Proterozoic and Archean. The domains are defined geographically by means of lithospheric regionalization based on our tomographic model (see Schaeffer & Lebedev 2015; Schaeffer *et al.* 2016b). A representative composition has been assigned to each domain, based on global average compositions from xenoliths and peridotite massifs (Griffin *et al.* 2009). Radially, mantle composition varies from that of a lithospheric type (any of the three mentioned before) to a typical sub-lithospheric, pristine upper mantle composition (PUM, McDonough & Sun 1995) (Table 1). For composition only, an average lithospheric thickness is assumed for each of the three compositional domains. In Phanerozoic and Proterozoic terranes the depth to the base of the layer with the lithospheric composition is set to 120 km. In cratons the composition is “Archaean” from the base of the crust down to 120 km, and “Proterozoic” from 120 km down to 200 km depth.

The V_s - T inversion is based on the thermodynamically self-consistent temperature-pressure tables (for each of the mantle compositions in Table 1) containing V_s values pre-calculated using the computational petrology software package Perple_X (Connolly 2005). V_s values are corrected to account for the effect of attenuation. We compute anelasticity as a pressure-temperature-dependent correction to the anharmonic output velocities from Perple_X (e.g., Afonso *et al.* 2005; Karato 1993; Minster & Anderson 1981). The values for grain size and activation volume on the anelasticity correction applied here are taken from Fullea *et al.* (2012).

White areas in Figure 8 show locations where partial melting is likely to occur at least at some of the depths in the 80-150 km range. Partial melting is inferred if the temperature obtained from the V_s inversion exceeds the mantle peridotite dry solidus (Katz *et al.* 2003) at the corresponding pressure.

A useful reference for interpretation of Figure 8 is given by steady-state geotherms, computed as described in the following section. A 260-km thick cratonic lithosphere, for example, has an average temperature within the 80-150 km depth range of 740°C. For a continent with a 100-km thick lithosphere (Fig. 2), the computed average temperature is 1277°C.

Lithospheric geotherms

We now use the integrated geophysical-petrological software LitMod (Afonso *et al.* 2008, Fullea *et al.* 2009) in order to compute the vertical profiles of temperature and density. The lithospheric geotherm is computed under the assumption of steady-state heat transfer in the lithospheric mantle, considering a P-T-dependent thermal conductivity in the mantle and prescribed thermal parameters in the crust (Table 2). In the convecting sub-lithospheric mantle, the geotherm is given by an adiabatic temperature gradient. A transitional buffer layer is located just beneath the bottom of the lithosphere, connecting

the lithosphere and the convecting sub-lithospheric mantle. This transitional buffer layer is characterized by a continuous linear super adiabatic gradient (i.e., heat transfer is controlled by both conduction and convection, see Fullea *et al.* (2009) for details). Stable mineral assemblages in the mantle are calculated using a Gibbs free energy minimization, as described by Connolly (2005). The composition is defined within the major oxide system NCFMAS (Na₂O-CaO-FeO-MgO-Al₂O₃-SiO₂). All the stable assemblages are computed using a modified and augmented version of the Holland and Powell's (1998) thermodynamic database (Afonso & Zlotnik, 2011). The density in the mantle is determined according to the elastic moduli and density of each end-member mineral as described by Connolly & Kerrick (2002) and Afonso *et al.* (2008).

The calculations are for the entire lithospheric column and, therefore, also produce values for the isostatic topography, gravity, surface heat flow and other observables (e.g., Fullea *et al.* 2012). The oceanic models in Figure 2 show a 5.5 km seafloor depth for the “old ocean” with a dense, 100-km thick lithosphere, and a 2 km seafloor depth for the “young ocean” with a 60-km thick lithosphere and a 4-km thick layer of sediment. The latter model was designed to be similar to that of the Nansen Basin, the southern half of the slowly opening Eurasia Basin adjacent to Eurasia's continental margin.

In Figure 9, we compare the profile of the central Canada Basin with those of a non-cratonic continent (100-km thick lithosphere) and a hypothetical profile for the Beaufort Sea. For the central Canada Basin, a 1D model with a steady-state geotherm and a 100-120 km lithospheric thickness can fit both the *V_s* values from our tomographic model and the observed bathymetry. For the SE part of the basin (the Beaufort Sea), in contrast, the observations are difficult to fit with a steady-state 1D model. Seismic velocities here are relatively low in the uppermost mantle (similar to those in a Phanerozoic continental region) but relatively high at 100-200 km depth and even deeper. We attempted to construct a 1D physical model that could reproduce these *V_s* variations, while keeping the bathymetry consistent with observations. A model with a 120-km thick lithosphere and a thick layer of sediments with high heat production goes some way towards explaining the relatively low temperatures in the uppermost mantle and relatively high temperatures at greater depths, but it cannot quite match the seismic velocity variations. This suggests that the high velocities at depths greater than 100 km may be due to the cooling of the upper mantle by the adjacent McKenzie Craton, the NW extremity of the North American Craton (Schaeffer & Lebedev 2014). The three-dimensionality of the thermal regime is why 1D steady-state models are insufficient to explain the *V_s* values beneath the Beaufort Sea.

The Arctic lithospheric structure: A brief overview

The seismic-velocity and thermal structure of the Arctic upper mantle offers abundant evidence on the architecture and evolution of the lithosphere across the region. In the following, we highlight a few selected, robust observations, with a bias for more unexpected—or less well-understood—features and for more provocative inferences.

296 *Canada Basin*

297 The seismic-velocity and thermal structure of the Canada Basin lithosphere is similar to that of old
298 oceans. The origin of the basin is debated, with end-member models invoking normal sea-floor
299 spreading, stretching of continental crust, and exhumation and serpentinization of the upper mantle; all
300 three mechanisms may have contributed to the formation of the basin at different locations and
301 different times (Pease *et al.* 2014; Chian *et al.* 2016). Our 1D lithospheric-scale physical modelling shows
302 that the inferred temperature within the lithosphere, bathymetry of the basin, and the thickness of its
303 sedimentary cover (Jakobsson *et al.* 2012; Petrov *et al.* 2016) are consistent with it having a normal-
304 oceanic, basaltic crystalline crust with a thickness of around 7 km. On the basis of the 1D modelling, we
305 cannot rule out, however, that such crust could have been formed by a mechanism other than normal
306 seafloor spreading. Regarding the age of the lithosphere, the temperature profile inferred from the high
307 seismic velocities within it indicates that it has been cooling for over 100 Ma.

308

309 *Eurasia Basin*

310 The younger and narrower Eurasia Basin shows lower seismic velocities and higher temperatures,
311 compared to the Canada Basin. In contrast to the Mid-Atlantic Ridge in the North Atlantic, where partial
312 melting is inferred in a broad corridor along its entire length, the ultra-slow spreading Gakkel Ridge in
313 the Eurasia Basin (e.g., Coakley & Cochran 1998) shows partial melting beneath only a few of its
314 segments, one of them just north of the Fram Strait that separates the North Atlantic and Eurasia Basins.

315 A continuous low-velocity anomaly extends through the Fram Strait, from the North Atlantic to beneath
316 the adjacent, westernmost portion of the Gakkel Ridge. The extent of this anomaly may indicate how far
317 the hot North Atlantic asthenosphere flows northward. Schmidt-Aursch & Jokat (2016) used 3D gravity
318 modelling to infer that the separation of Greenland from Svalbard (forming the Fram Strait at chron
319 C5/C6) led to the inflow of North Atlantic mantle into the western Eurasia Basin, creating a distinct
320 pattern of magmatism along the Gakkel Ridge.

321 Our tomographic model presents an image of the inflow of hot North Atlantic asthenosphere into the
322 Eurasia Basin. It suggests a major effect of the inflow on the character of the sea-floor spreading: the
323 northern limit of the low-velocity anomaly that we observe (Fig. 4, 80-110 km) is at around 84N, and this
324 is where a boundary is observed between the Western Volcanic Zone of the Gakkel Ridge to the south,
325 with basalts covering the seafloor and with well-developed magmatic characteristics similar to those of
326 slow-spreading ridges elsewhere, and the Sparsely Magmatic Zone to the north, with little evidence for
327 magmatism and a predominantly peridotitic crust (Thiede *et al.* 2002; Schlindwein *et al.* 2007).

328

329 *Craton-ocean boundaries*

330 Both the southern half of the Eurasia Basin (the Nansen Basin) and the southern portion of the Canada
331 Basin (the Beaufort Sea) are bordered to the south by thick, cold continental lithospheres. Both

locations show high-velocity anomalies at 150-200 km depths, indicating upper mantle that is colder than beneath the rest of the basins further north. This is likely to be due to cooling from the adjacent cold continental lithosphere. The resulting thermal anomaly in the deep upper mantle must have caused an extra amount of subsidence of the basins in the course of their evolution and contributed to the development of their thick sedimentary layers.

Cratons

The boundaries of the East European and Siberian Cratons, as seen in the mantle (Figs. 4, 5: 110 and 150 km depths), generally follow their geological boundaries at the surface. To the north, the Barents and Kara Sea also show high velocities indicative of thick lithosphere, similar to that of the cratons.

Seismic-velocity heterogeneity beneath the Baltic Shield reveals pronounced lateral variations of the lithospheric thickness within it. These have been reported previously in a number of studies; different models, however, continue to show substantially different lithospheric-thickness variations (e.g., Bruneton *et al.* 2004; Artemieva *et al.* 2006; Weidle & Maupin 2008; Legendre *et al.* 2012; Pedersen *et al.* 2013; this study).

In the NW corner of the North American Craton, our tomography delineates the westward lateral extent of the cratonic lithosphere. The recently identified Mackenzie Craton, unexposed at the surface, forms the northwestern extremity of the North American Craton and extends roughly to the front of the Rockies (Schaeffer & Lebedev 2014).

The deep cratonic root of Greenland is separated from the root of the North American Craton by tectonic units without a thick cratonic lithosphere. West of central Greenland, the two are separated by the Baffin Bay, with its warmer, thinner oceanic lithosphere. The thin-lithosphere corridor defined by the Baffin Bay stretches further west into the Canadian Arctic Archipelago.

Canadian Arctic Archipelago

The southern part of the Canadian Arctic Archipelago—the islands to the north of the North America mainland and between North America and Greenland—is clearly a part of the North American Craton, with high seismic velocities indicative of cold, thick lithosphere.

The Sverdrup Basin in the northern part of the archipelago (Fig. 1) is a major Carboniferous to Paleogene depocentre (Stephenson *et al.* 1987; Embry & Beauchamp 2008, Midwinter *et al.* 2016), with sedimentary thicknesses in the 5-15 km range (Petrov *et al.* 2016). According to our models, the lithosphere is relatively thick in the southern part of the basin and thins to the north; the axis of the basin (Embry & Beauchamp 2008) coincides roughly with the 1000°C isotherm (the transition from green to yellow) in Figure 8. The area to the south of the Sverdrup Basin (comprising the Devon, Somerset, Cornwallis, Bathurst and Melville Islands) shows thinner lithosphere and mostly thin (<1 km) or absent

sedimentary cover (Petrov *et al.* 2016). The subsidence and sedimentation within the Sverdrup Basin are thus likely to reflect the cooling and thickening of the Phanerozoic mantle lithosphere, whereas just south of the Sverdrup Basin, the lithosphere is thin and there has been little subsidence and sedimentation—a causal relationship seen in other parts of the world as well (Meier *et al.* 2016).

The thin-lithosphere corridor that cuts through the archipelago south of the Sverdrup Basin connects to the Baffin Bay near Devon Island. Here, Burke & Dewey (1973) postulated a plume-generated, Devon Island triple junction, noting that a junction in this location has also been recognised earlier by Wegener (1929) and Wilson (1963). The three arms of this proposed junction were the Baffin Bay (between Greenland and North America) to the southeast, the Nares Strait (between Greenland and the Ellesmere Island) to the north, and the Lancaster Sound (between Devon and Baffin Islands) to the east. Burke & Dewey (1973) suggested that the Lancaster Sound arm became inactive before reaching the spreading stage—unlike the Baffin Bay arm, where seafloor spreading has formed the oceanic lithosphere now separating Greenland from North America. The Nares Strait arm, underlain by continental crust (e.g., Altenbernd *et al.* 2016), was proposed to accommodate left-lateral transform motion.

Our results show pronounced thinning of the lithosphere beneath the corridor between the Sverdrup Basin and the North American Craton. No lithospheric thinning is observed beneath the Nares Strait between Greenland and the Ellesmere Island.

Ellesmere Island

At 150 km depth (Fig. 5), the entire southeastern-central part of the Ellesmere Island shows high velocities that appear to form a single anomaly with cratonic northwestern Greenland. Northern Ellesmere Island, in contrast, is underlain by lower velocities, indicative of thinner lithosphere. At 110 km (Fig. 4), seismic velocities beneath the central-southeastern part of the island are also high, but not as high as beneath Greenland. The inferred average lithospheric temperature (Fig. 8) suggests that the central Ellesmere lithosphere is not quite as cold as that of northern Greenland. However, it is as cold as cratonic lithosphere in many locations elsewhere.

While the southeastern-most part of the island is well known to be a part of the Canadian-Greenland Shield, the central part of the island is covered by Cambrian-Devonian shelf deposits (e.g., Saalman *et al.* 2005). Our images suggest that the cratonic lithosphere—forming a single lithospheric block with that of northern Greenland—may extend beneath much of the Ellesmere Island. The inferred lateral extent of cratonic lithosphere beneath the island can put limits on where the relative motions between Greenland and (parts of) Ellesmere Island may occur (the “Nares Strait debate”, e.g., Dawes & Kerr 1982; Saalman *et al.* 2005). It provides evidence against major displacement along Nares Strait, which is consistent with many Precambrian and Paleozoic structures continuing without offset from Greenland across the Nares Strait to Ellesmere Island (e.g., Dawes *et al.* 1982; Dawes & Kerr 1982).

403 *Thin lithosphere and intraplate volcanism*

404 All locations of intraplate basaltic eruptions associated with the HALIP are underlain by relatively thin,
405 non-cratonic lithospheres. This is evidenced by seismic velocities beneath them being much lower than
406 beneath cratons (Fig. 4, 80 and 110 km depths). For example, northern Ellesmere Island has thin, warm
407 lithosphere with low seismic velocities. It is in this northern part of the island that the Cretaceous
408 alkaline basaltic volcanism associated with HALIP has been identified (Fig. 1). The presence of thick
409 lithosphere beneath the central and southeastern part of the island, as shown by our tomography, is
410 consistent with the volcanism being limited to the northern part of the island. Even if anomalously hot
411 asthenosphere was placed beneath the entire island—or beneath its central part—it would flow
412 towards locations with thinner lithosphere along the lithosphere-asthenosphere boundary and undergo
413 significant partial melting where it reaches relatively shallow depths (Thompson & Gibson 1991; Morgan
414 & Morgan 2005; Lebedev *et al.* 2006), i.e., beneath the northern part of the Ellesmere Island.

415

416 *Siberian Traps*

417 The Permo-Triassic Siberian Traps represent one of the largest continental Large Igneous Province (LIP)
418 on Earth, erupted rapidly at around 248 Ma and believed to be located on thick cratonic lithosphere
419 (e.g., Renne & Basu 1991; Dobretsov *et al.* 2013). Our model shows that the lithosphere beneath the
420 central part of the LIP is, in fact, warmer and thinner than beneath the surrounding parts of the Siberian
421 Craton.

422 The thinning of cratonic lithosphere by a hot mantle plume has been modelled recently specifically for
423 the Siberian Traps case, with the model invoking extensive plume melting and erosion of the thick
424 cratonic lithosphere at the time of the emplacement of the traps (Sobolev *et al.* 2011). Our images map
425 the location where lithospheric thinning associated with the LIP probably took place.

426

427 *Iceland Plume track across Greenland?*

428 A corridor of relatively low seismic velocities cuts across central Greenland from west to east (Figs. 4, 5:
429 110, 150 km; Figs 6, 7: top left), indicating that the lithosphere within this corridor is thinner and
430 warmer than normal cratonic lithosphere. We cannot exclude the possibility that the lithosphere here
431 has been non-cratonic since the assembly of the Greenland landmass. Tectonic boundaries within ice-
432 covered Greenland's interior are, unavoidably, somewhat uncertain. It is generally thought, however,
433 that central Greenland—through which the low-velocity corridor cuts—is occupied by the Archean Rae
434 Craton (e.g., St-Onge *et al.* 2009). The thin-lithosphere corridor may thus be a result of modification of
435 the lithosphere within the craton that post-dated the assembly and stabilization of Greenland. Such
436 modification could occur due to the interaction of the lithosphere with hot sub-lithospheric mantle
437 (Foley 2008).

The track of the Iceland Hotspot is expected to cross Greenland, although its trajectory across it is uncertain (Lawver & Müller 1994; Lawver *et al.* 2002; Mihalfy *et al.* 2008; Ganerød *et al.* 2010; Torsvik *et al.* 2015). Near the eastern coast of Greenland, the thin-lithosphere corridor we observe terminates where Torsvik *et al.* (2015) put the hotspot at 50-60 Ma. The Cape Hold With Hope—a location of HALIP basaltic volcanism—is just to the north. Just to the south along Greenland’s coast, abundant basaltic volcanism occurred at the Blossville Kyst; this is where the Iceland Hotspot moved (with Greenland as a reference) by 40 Ma, according to the reconstruction of Torsvik *et al.* (2015).

Near Greenland’s western coast, the low-velocity corridor we observe is roughly where Lawver & Müller (1994) placed the hotspot at 70 Ma—also close to the location of significant intraplate volcanism. (The trend of Lawver and Müller’s track after 70 Ma is NW-SE, however, compared to the roughly W-E trend of the thin-lithosphere corridor mapped here.)

Previously, Yakovlev *et al.* (2012) inferred thinned lithosphere beneath parts of central Greenland from their regional, travel-time, *P*-wave tomography and attributed this to the passage of the area over the Iceland Hotspot. In particular, they inferred thin lithosphere roughly beneath the eastern half of the area occupied by the thin-lithosphere corridor seen in our model (see also Petrunin *et al.* 2013). Steinberger *et al.* (2015) used global *P*-wave tomography (Amaru 2007; Bijwaard & Spakman 2000) and regional waveform tomography (Rickers *et al.* 2013) to argue that the lithosphere beneath eastern Greenland was thinned by means of thermal and mechanical erosion by the Iceland Plume material underneath.

The thin-lithosphere corridor seen in our models may show the complete extent of Greenland’s lithosphere modified by the Iceland Hotspot, as the Greenland landmass moved across it. It thus provides important new evidence on the location of the hotspot track, from the east to the west of Greenland.

Conclusions

Upper-mantle tomography reveals lateral variations in the temperature and thickness of the lithosphere. Because the variations in seismic velocities reflect primarily variations in temperature, tomographic maps offer a proxy for the lateral variations of temperature in the upper mantle.

New waveform tomography (Schaeffer *et al.* 2016b; this study) provides improved resolution in the upper mantle of the Arctic, thanks to the highly complete waveform dataset used to construct the model—taking advantage of the recent expansion of broadband seismic networks around the region—and the elaborate waveform inversion and error analysis procedures.

Computational petrology enables self-consistent joint modelling and inversion of various geophysical and geological data. Assuming the composition of the lithosphere and asthenosphere and the parameters controlling anelasticity, seismic tomography models can be converted to thermal models of the upper mantle. A temperature estimate at a point, however, may be biased due to the non-

uniqueness of the 3D tomographic models. Temperature averages over a depth range represent more robust and reliable temperature estimates. Our new map of the average temperature in the 80-150 km depth range reveals the lateral thermal heterogeneity in the lithosphere-asthenosphere depth range across the Arctic region.

The seismic and thermal models map the deep lithospheric boundaries of tectonic blocks with different properties and different age—including cratons, Phanerozoic continents, locations of intraplate volcanism, and old and young oceans—and offer new insights into the dynamics and evolution of the Arctic lithosphere.

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Table 1: Bulk mantle compositions used in this work from xenoliths suites and peridotite massifs.

	1) Av. Phanerozoic (wt%) ^a	2) Average Proterozoic (wt%) ^a	3) Average Archean (wt%) ^a	4) PUM M&S95 (wt%) ^b
SiO ₂	44.5	44.6	45.7	45
TiO ₂	0.14	0.07	0.04	0.201
Al ₂ O ₃	3.5	1.9	0.99	4.45
Cr ₂ O ₃	0.4	0.4	0.28	0.384
FeO	8.0	7.9	6.4	8.05
MnO	0.13	0.12	0.11	0.135
MgO	39.8	42.6	45.5	37.8
CaO	3.1	1.7	0.59	3.55
Na ₂ O	0.24	0.12	0.07	0.36
NiO	0.26	0.26	0.3	-
Mg#	89.9	90.6	92.7	89.3

^a Global average taken from Griffin et al.(2009).

^bPUM stands for Primitive Upper Mantle, M&S95 refers to McDonough and Sun (1995).

854 **Table 2:** Assumed geophysical properties of different crustal layers.

Layer	Density (kg/m ³)	Heat production (W/m ³)	Thermal conductivity (W/m K)
Oceanic sediments	2200	$1.2 \cdot 10^{-6}$	3
Oceanic crust	2920	$1 \cdot 10^{-7}$	2.1
Continental Crust	2850	$0.7 \cdot 10^{-6}$	2.5
Beaufort Sea sediments	2400-2700	$2 \cdot 10^{-6}$	3
Beaufort Sea crust	2950	$1 \cdot 10^{-7}$	2.1

855

856

857

858 Figure captions

859 Fig. 1. Topographic map of the Arctic region. Ridges on the High-Arctic seafloor are shown as green
860 hatched areas, locations of intraplate volcanism—as red hatched areas (Gaina *et al.* 2014; Schiffer *et al.*
861 2016). AB, Amundsen Basin; AR, Alpha Ridge; BI, Baffin Island; BR, Brooks Range; BS, Baltic Shield; CB,
862 Chukchi Borderlands; CDF, Cordilleran Deformation Front; DI, Devon Island; EI, Ellesmere Island; FJL,
863 Franz Josef Land; HB, Hudson Bay; LR, Lomonosov Ridge; LS, Laptev Sea; MAR, Mid-Atlantic Ridge; MB,
864 Makarov Basin; MR, Mendeleev Ridge; NB, Nansen Basin; Nvl, Novosibirsk Islands; NZ, Novaya Zemlya;
865 SB, Sverdrup Basin; SZ, Severnaya Zemlya; UM, Ural Mountains; VF, Verkhoyansk Fold Belt.

866

867 Fig. 2. Type profiles of temperature and density beneath an old ocean with a 100-km thick lithosphere, a
868 young ocean with a 60-km thick lithosphere and a layer of sediments atop the crystalline crust, and a
869 Phanerozoic continent with a 100-km thick lithosphere.

870

871

872 Fig. 3. Broadband seismic stations (inverted triangles) and earthquakes (small squares) in and around
873 the Arctic region. These sources and stations were used in the waveform tomography of the Arctic along
874 with our global set of stations and earthquakes (Schaeffer & Lebedev 2014; Schaeffer *et al.* 2016b).

875

876 Fig. 4. Map views of our tomographic model at four depths in the shallow upper mantle. The reference
877 values of the vertically-polarized S-wave speed in the mantle are: 4.46 km/s at 36 km depth, 4.42 km/s
878 at 56 km depth, and 4.38 km/s at 80 and 110 km depths. Where the depths shown fall within the crust,
879 the reference values are those of CRUST2.0 (Bassin *et al.*, 2000;
880 <http://igppweb.ucsd.edu/~gabi/crust2.html>), smoothed at its block boundaries. The reference values
881 and deviations from them are at the reference period 50 s.

882

883 Fig. 5. Map views of our tomographic model at four depths in the deeper upper mantle. The reference
884 values of the vertically-polarized S-wave speed in the mantle are: 4.39 km/s at 150 km, 4.45 km/s at 200
885 km, 4.62 km/s at 250 km, and 4.75 km/s at 330 km depth. The reference values and deviations from
886 them are at the reference period 50 s.

887

888 Fig. 6. A comparison of vertically-polarized S-wave speed distributions at 100 km depth in the Arctic
889 region according to different recent global tomographic models, AMISvArc (Schaeffer *et al.* 2016b);
890 3D2015_07Sv (Debayle *et al.* 2016); Savani (Auer *et al.* 2014); SEMum2 (French *et al.* 2013); S40RTS
891 (Ritsema *et al.* 2011); CUB (Shapiro & Ritzwoller 2002). Each model is plotted with respect to its own
892 global mean at the depth.

893

894 Fig. 7. A comparison of vertically-polarized S-wave speed distributions at 150 km depth in the Arctic
895 region according to different recent global tomographic models (as in Fig. 6). Each model is plotted with
896 respect to its own global mean at the depth.

897

898 Fig. 8. Average temperature in the 80-150 km depth range, derived from our tomographic model.

899

900 Fig. 9. Estimated profiles of temperature and density beneath the central Canada Basin and the Beaufort
901 Sea, compared with those of a Phanerozoic continent with a 100-km thick lithosphere.

902 Fig. S1. A comparison of vertically-polarized S-wave speed distributions at 50 km depth in the Arctic
903 region according to different recent global tomographic models (as in Fig. 6). Each model is plotted with
904 respect to its own global mean at the depth.

905

906 Fig. S2. A comparison of vertically-polarized S-wave speed distributions at 200 km depth in the Arctic
907 region according to different recent global tomographic models (as in Fig. 6). Each model is plotted with
908 respect to its own global mean at the depth.

909

AMISvArc

-12.55/11.05

3D2015_07Sv

-5.50/9.63

Savani

-5.27/8.00

SEMum2

-6.93/10.07

150 km

S40RTS

-3.66/6.71

±8%

CUB

-6.58/10.44

S-wave speed anomaly, %

80–150 km

whole lithosphere

zoom-in on the crust