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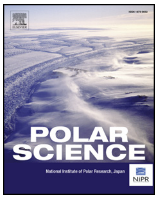

Effects of the Iceland plume on Greenland's lithosphere: New insights from ambient noise tomography

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ABSTRACT

Ambient noise tomography is used to image Greenland's lithosphere, which passed over the Iceland plume between ∼70 and ∼40 Ma. Cross-correlations from 21 stations from GLISN seismic network were used to invert for 2-D Rayleigh wave phase velocity maps for 14 periods between 8 and 40 s. We find that Rayleigh wave phase velocities substantially vary across Greenland, with slow velocities coinciding with NW-SE trending Iceland plume track. In east Greenland the detected velocity reduction at longer periods (33–40s) reflects substantially thinned lithosphere, thermally ablated by the plume. From the east, the reduced velocities shift NW at shorter periods (12–20s), indicating shallowing of the plume-related slow anomaly. In north-central Greenland, the reduced velocities appear in the proximity of the plume ∼60 Ma, reflecting lithospheric weakening in the presence of residual heat that still persists within the lithosphere. Our results provide important new constraints on variations in the seismic velocity structure of Greenland's crust and uppermost mantle, revealing prolonged effects of the mantle plume on the overpassing craton.

1. Introduction

Rayleigh wave tomography using ambient noise as a source is widely used for imaging of the crust and the uppermost mantle (e.g. Yang et al., 2008; Warren et al., 2013). Ambient seismic noise generally includes a variety of energy sources. These range from primary and secondary microseisms, with peak energy at ∼15 and ∼7.5 s, respectively, caused by pelagic storms and coastal reflections (since Wiechert, 1904), to Earth's long-period 'hum', caused by atmosphere-ocean-seafloor coupling (Rhie and Romanowicz, 2004).

Ambient noise tomography is a practical and effective tool for studying regions of low seismic activity, such as Greenland. Most of Greenland's lithosphere is a Precambrian shield. The collision that occurred ∼1860–1840 Ma between the Rae craton to the north and the overriding North Atlantic craton to the south created a complex orogeny system (Fig. 1) (Kalsbeek et al., 1987; van Gool et al., 2002; St-Onge et al., 2009; Garde and Hollis, 2010). The roughly NW-SE striking Nagssugtoqidian belt constitutes the southern margin of this orogeny, while the N-S directed Rinkian belt represents continuation of the system further north (Fig. 1) (St-Onge et al., 2009; Garde and Hollis, 2010).

The Greenland craton is surrounded by complex passive margins both to the west and east. The passive margin to the west was created during the opening of the Labrador Sea ∼61 Ma, when the North American plate separated from Greenland (Chalmers and Pulvertaft, 2001). The passive margin to the east is associated with continental breakup and opening of the North Atlantic Ocean ∼56-55 Ma (Storey et al., 2004; Larsen et al., 2014).

Between ∼70 and ∼40 Ma the Greenland craton passed over the Iceland-Jan Mayen plume (e.g. Lawver and Müller, 1994). The proposed plume tracks based on different assumptions on numbers of hotspots and plate motion data are mainly in agreement over the past ∼50 to ∼60 Ma, but the reconstructed paths prior to that time greatly diverge. The envelope of previously proposed paths is shown in Fig. 1 (Braun et al., 2007), with tracks suggested by Lawver and Müller, 1994 and Forsyth et al., (1986) delineating the southern and northern end members.

A set of unusual observations in eastern and north-central Greenland has been associated with this passing plume. First, both receiver functions (Kumar et al., 2005) and tomographic images of the uppermost mantle (Darbyshire et al., 2004; Darbyshire et al., 2018) indicate substantial variations in depth of the lithosphere-asthenosphere boundary under Greenland, ranging from ∼120 km to the west to ∼70 km to the east. The observed lithospheric thinning to the east has been attributed to thermal erosion due to an underlying plume ∼40 Ma (Kumar et al., 2005). Second, both gravity study (Braun et al., 2007) and receiver functions (Dahl-Jensen et al., 2003) inferred a 5–10 km thinner crust to the north, which was also attributed to thermal erosion

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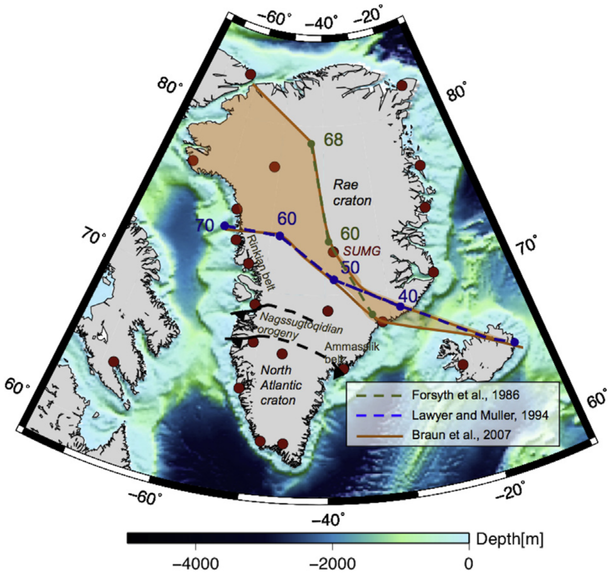

Fig. 1. Principal tectonostratigraphic units and orogeny belts discussed in the text. Black lines represent complex suture between Rae and North Atlantic cratons (after St-Onge et al., 2009). Dark red circles are stations used in the study with marked SUMG station in central Greenland. Shaded area represents the envelope of previously proposed plume paths (Braun et al., 2007), with tracks proposed by Forsyth et al., (1986) and Lawver and Müller, 1994 delineating northern and southern end members (age is given in Ma).

by the plume ~70 Ma (Braun et al., 2007). Third, the area in north-central Greenland, covered by an old and thick ice sheet (MacGregor et al., 2015), is associated with elevated heat flow ($> 70\,\mathrm{mW/m^2}$) (Fox-Maule et al., 2009) and faster ice sheet basal melting (Fahnestock et al., 2001; Rogozhina et al., 2012). This area coincides with the proximity of the plume ~60 Ma (Forsyth et al., 1986; Rogozhina et al., 2016). Based on a coupled ice–lithosphere model, Petrunin et al., (2013) showed that the observed high geothermal flux must have involved input of heat from the deep Earth and speculated that anomalously thin lithosphere in this area, thermally eroded by the plume, could have caused increased temperatures within the lithosphere. Another modeling study simulated periodic ice-sheet loading scenarios along an east-west transect across central Greenland and surmised that the modeled long-term changes in ice-sheet size in north-central Greenland may have been enhanced by possible plume-related magmatism (Stevens et al., 2016).

To better understand the link between deep and surface processes, we have imaged Greenland's lithospheric structure using ambient noise tomography. We provide important new constraints on lateral variations in the seismic velocity structure of Greenland's crust and the uppermost mantle. Reduced Rayleigh wave phase velocities in our maps coincide with NW-SE trending Iceland plume track, illuminating significant effects of the plume on the overpassing Greenland's lithosphere.

2. Data and methods

Seismic records for this study were obtained from the Greenland Ice Sheet Monitoring Network (GLISN; Dahl-Jensen et al., 2010; Clinton et al., 2014) (Fig. 1). All available vertical component seismograms from 21 broad-band stations in the period from December 2011 to July 2016 were processed following the procedure of Bensen et al., (2007). All data used in this study were extracted from seismic stations that included Guralp CMG-3T, Nanometrics Trillium, Quanterra 330 or Streckeisen STS-(1 and 2) sensors. Sample rates ranged from 20 to 100 samples/s. All waveform data are publically available and were extracted from the IRIS DMC. The seismograms were cut into daylong segments; records with less than 80 percent of the day were rejected. Since seismic stations differ in types of sensors and sampling rate, we removed daily trends, means and instrument response. The time series were decimated to 1 sample/s and bandpass filtered between 5 and 150 s. Discrete signals, such as earthquakes or explosions, as well as instrumental irregularities, were removed by applying temporal normalization using running-absolute-mean (Bensen et al., 2007). The waveforms were further converted into the frequency domain using Fast Fourier Transform (FFT) and spectrally whitened to broaden the band of the ambient noise signal and avoid the effects of monochromatic sources. Daylong seismic records were then cross-correlated and stacked.

The full cross-correlation waveform is a two-sided time function with positive and negative signals offset from zero by the same amount of time, corresponding to waves traveling in opposite directions between two stations (Suppl. Fig. 1). Asymmetry of positive and negative lags can be attributed to the inhomogeneous distribution of ambient noise sources. Inhomogeneity of ambient noise sources may be problematic for seismic analysis in Greenland, since previous studies detected a strong source of primary and secondary microseisms in the North Atlantic during winter (e.g. Kedar et al., 2008; Yang and Ritzwoller, 2008). Stacking of at least three years of data should avoid potential heterogeneities associated with seasonal source variations (Warren et al., 2013). In addition, Lin et al., (2008) demonstrated that travel times at two lags are almost the same in spite of different waveforms, which allows stacking of the two components of cross-correlations into a one-sided signal.

To illustrate the effect of source inhomogeneity we show a rose diagram of sources averaged over winter and summer months recorded at the central station ICESG (following Stehly et al., 2006, Fig. S4). The winter months show a strong orientation towards the North Atlantic, a known source of storms and microseismic activity (Kedar et al., 2008). The summer months are more heterogeneous, suggesting that it is important to include a wide temporal range to insure a broad source pattern. This observation may have implications for other studies that use seismic noise cross-correlation to investigate seasonal changes in ice sheet effects on Greenland (Mordret et al., 2016).

Following seismic stacking, Rayleigh wave phase velocity dispersion curves were measured for periods between 6 and 50 s using automated frequency-time analysis (FTAN) (Levshin and Ritzwoller, 2001). The analysis involves filtering of signals over narrow bandwidths centered at the frequency of interest. The measurements were made only for paths with $\mathrm{SNR} > 15$ and with interstation distance of at least 3 wavelengths. Due to poor path coverage, periods shorter than 8 s and longer than 40 s were removed prior to inversion analysis.

The measured Rayleigh wave phase velocity dispersion curves were inverted to obtain phase velocity maps of Greenland. The tomographic inversion was accomplished following the procedure of Barmin et al., (2001) for short (8–12 s), intermediate (15–25 s) and long (28–40 s) periods. The 2-D isotropic phase velocity model for each period is determined by minimizing a penalty function, which depends on three user defined regularization parameters: α, β, and σ. Parameters α and β control the damping and the averaging of data in regions with poor path coverage, while parameter σ defines the smoothing width. Inversion results using different values for parameters α 1:50:1000, β 1:50:1000, and σ 1:50:500 yielded similar main features, even though the amplitude of the features varied. The final results were obtained using $\alpha = 400$, $\beta = 100$, and $\sigma = 100$ km, which produced maps without apparent streaking and other spurious features. The average uncertainties of phase velocities obtained with this model are ~0.02 km/s for periods < 35 s, while uncertainties increase to 0.03 km/s at our longest periods (35 and 40 s).

The resolution matrix was used to estimate spatial resolution by finding the minimum distance at which two δ-like input functions can be resolved (Barmin et al., 2001) (Suppl. fig. 2). Since the station array is widespread and the distance between stations is large, the inversion was implemented on a $1^\circ \times 2^\circ$ grid. Resolution is highest in the central