Earth and Planetary Science Letters 402 (2014) 82–89

Contents lists available at ScienceDirect

Earth and Planetary Science Letters

www.elsevier.com/locate/epsl

A sharp cratonic lithosphere–asthenosphere boundary beneath the American Midwest and its relation to mantle flow

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ARTICLE INFO

Article history:
Accepted 12 November 2013
Available online 12 December 2013
Editor: P. Shearer

Keywords:
lithosphere–asthenosphere boundary
cratonic lithosphere
mid-lithosphere discontinuity
receiver functions
teleseismic converted waves

ABSTRACT

Beneath the American Midwest, S-to-P (Sp) converted wave imaging and multi-mode surface wave tomography identify a north-trending transition in seismic structure at 150–250 km depth. To the east of this American Midwest transition (AMT), the lithosphere–asthenosphere boundary (LAB) is imaged as a 1–2% Sp/Sv amplitude arrival at 200–240 km depth, consistent with the depth of negative shear velocity and azimuthal anisotropy gradients imaged by surface wave tomography. To the west of the AMT, Sp conversions are much shallower at 150–190 km depth and are much weaker ($<0.7\%$) or absent. Azimuthal anisotropy constrained by surface wave tomography also changes across the AMT, with stronger anisotropy to the east of the transition beneath the thicker lithospheric root. We suggest that the seismic changes across the AMT can be explained by considering the effects of asthenospheric flow beneath the leading edge of the thick lithospheric root. The mantle flow is dominantly driven by the drift of the North America plate. Locally higher flow velocities are expected where the asthenosphere is forced to flow beneath the thicker root. This mantle underflow could create a sharper seismic LAB east of the AMT via two effects. First, the local increase in flow velocities could steepen the thermal gradient at the base of the lithosphere, and hence the isotropic velocity contrast. Second, the increased strain rate along edge of the lithosphere could enhance the magnitude of azimuthal anisotropy. Our results suggest that seismically detectable LAB sharpness variations could be used to constrain geographic variations in coupling between plates and mantle convection. A secondary result is the image of a Mid-Lithospheric Discontinuity arrival at 80–110 km depth that is found primarily to the east of the AMT. This arrival is interpreted as produced by a layer of low-velocity metasomatic minerals that have accumulated since the >1.8 Ga creation of the lithosphere.

1. Introduction

The magnitude of stress coupling between mantle convection and the base of the plates is a first-order issue in assessing the force balances on plates, and hence the motion of the plate mosaic (Forsyth and Uyeda, 1975; Ghosh et al., 2013). A first-order control with regard to the coupling between tectonic plates and underlying mantle convection is the physical state property variations at the base of the lithosphere that produce a viscosity gradient that has a 2–3 order of magnitude decrease across the 1000–1300 °C thermal gradient (Sleep, 2011). This basal viscosity gradient is where the buoyantly driven flow of the mantle (Ghosh and Holt, 2012) and the relative motion of the plate must be accommodated via exchange of stress.

Generally increasing viscosity with depth predicts that the rheologically mobile boundary layer in the ∼1000–1300 °C interval will experience higher flow stresses as lithospheric thickness increases, potentially producing an approximately 250 km limit on lithospheric thickness for the present day convective regime (Cooper and Conrad, 2009). Additionally, mantle flow around undulations in lithospheric thickness and variations in relative velocity between the lithosphere and asthenosphere will modulate the shear strain rates in the basal thermal boundary layer that is capable of deforming viscously. Such flow induced shear strain variations could create variations in the strength of olivine LPO development that would be observed as velocity anisotropy detectable with seismology (Kaminski and Ribe, 2002; Ribe, 1989). Thus, seismic interrogation of physical property variations at the base of the lithosphere can help to constrain geographic variations in the degree of plate–mantle stress coupling.

Ideally, one seeks to seismically interrogate a region where a plate's basal physical properties are changing. Here we focus on such a region in the American Midwest, where the lithosphere thickens substantially and the North American plate motion is nearly perpendicular to the strike of the narrow area over which the lithosphere thickens. In this scenario, the sub-lithospheric

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0012-821X/$ – see front matter
http://dx.doi.org/10.1016/j.epsl.2013.11.018

mantle must flow down and below the thickened lithospheric root and hence the mantle flow must accelerate based on Bernoulli's principle given the reasonable assumption that the base of the low viscosity asthenospheric channel remains relatively constant.

In thin lithosphere areas (<130 km), the LAB velocity gradient can be adequately resolved via inversion of fundamental mode surface wave dispersion data (Lebedev et al., 2009; Pedersen et al., 2009). However, in thick lithosphere areas (>130 km), fundamental mode resolution at these depths is poor and higher-mode surface wave data is required to constrain velocity gradients associated with the base of the lithosphere (Yuan and Romanowicz, 2010). Even higher resolution of velocity gradients at the base of the plates is provided by Ps and Sp converted wave images that directly sense velocity gradients (Farra and Vinnik, 2000; Rychert et al., 2007) and do not require a regularized inversion to image structure. In thin lithosphere areas, Ps converted waves can isolate LAB arrivals with minimal interference with free-surface to Moho reverberations (Rychert et al., 2005). However, in thick lithosphere areas, Ps prospecting for LAB arrivals is problematic due to the overprinting of small amplitude LAB arrivals by free-surface to Moho reverberations (Julià, 2007).

A remedy with regard to this Ps Moho reverberation interference is the use of converted Sp waves that has proven useful for imaging thick lithosphere LAB arrivals because Sp converted arrivals are not overprinted by Moho reverberations (Farra and Vinnik, 2000). The caveat to this advantage over Ps receiver functions is that S-wave receiver functions have reduced imaging power due to: less high-frequency bandwidth, fewer usable earthquake recordings due to epicentral distance range constraints, deconvolutional issues associated with the S and SKS branch cross-over at 84° (Yuan et al., 2006a), and, the requirement that SV polarized incident S-waves be used. Herein, a comparison of Sp and multi-mode surface wave images is performed to constrain lateral variations in isotropic and anisotropic seismic structure across the edge of North America's cratonic core in the American upper Midwest.

2. Data and methods

All EarthScope USArray seismic data recorded through November 2011 and most of the available PASSCAL broad-band data were processed into an Sp image that extends from the west coast to the longitude of Chicago with 41,731 binned Sp receiver functions. From this dataset, a subset within the American Midwest sampled by 6993 receiver functions is presented (Fig. 1). Our Sp receiver functions dataset consists of both the direct S-wave branch from 58°–84° epicentral distance and the SKS branch from 84°–110° epicentral distance. The percentage of direct-Sp and SKSp arrivals in our dataset is 64% and 36% respectively. Images made using only SKSp arrivals are noisy because the data sampling is reduced to 36% of the full dataset. With regard to P contamination of SKSp arrivals, we note that only pPPPP arrivals from >300 km depth event arrivals are predicted to arrive in our uppermost mantle time window of interest. Thus, given the rarity of this occurrence, we are not concerned about P-phase contamination of our SKSp dataset. To minimize cross-contamination between the SKS and S branches, extensive data culling and careful time windowing of the first-arriving S-wave packet was performed. The data culling consists of viewing up to 440 USArray three-component traces on multi-monitors and interactively processing the data. The first S-wave packet is windowed with an 8–12 s window and the SV-component relative travel-time residuals are measured and applied to line up the data so that waveform coherence may be assessed. Traces that are incoherent are deleted and events that have significant variations in the first-arriving S-wave packet waveshape are not used as that would violate our common source wavelet assumption in our deconvolution method. Rather than deconvolving the entire SV-component traces from the P-component traces, the first-S arrival is windowed and used in our source estimation procedure (Hansen and Dueker, 2009).

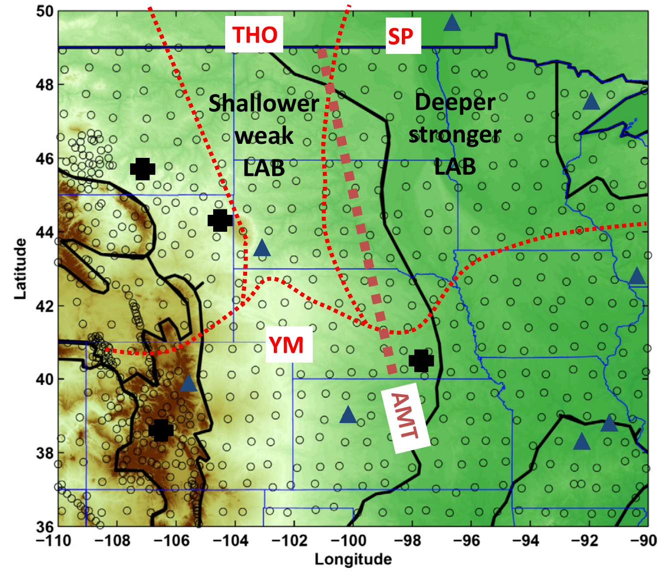

Fig. 1. Stations, topography, crustal province boundaries. The boundary between a deeper LAB and shallower weak to absent LAB signals is demarked as the American Midwest Transition (AMT) red dotted line. The crustal provinces boundaries are outlined as red dotted lines: circa. 2.6 Ga Superior Province (SP); 1.8–1.9 Ga Trans-Hudson orogen (THO); 1.8–1.6 Ga Yavapai–Mazatzal (YM). The black crosses show carbonatite intrusions locations that are consistent with the re-melting of an ancient metasomatic layer. The blue triangles are the permanent stations analyzed by Abt et al. (2010) discussed in text.

Our method to produce receiver functions (Hansen and Dueker, 2009) is significantly different with respect to signal station spectral deconvolution methods (Langston, 1977; Vinnik, 1977) that use the incident component as an estimate of the source wavelet. Instead, we assume for each event that a common source wavelet is incident at all stations (up to 440) and a least-square inverse is conducted to isolate the source wavelet spectra from the three-component receiver function spectra. Additionally, we do not assume that pure mode scattering (Sv–Sv) potential is zero as done in traditional receiver function analysis. The receiver function phase spectra are calculated by using the assumption that pure-mode scattering (SV to SV in this work) is effectively minimum phase for velocity models relevant to the Earth (Bostock, 2004) and the Kolmogorov minimum phase operator (Kolmogorov, 1939) is used in the phase reconstruction of the receiver functions (Mercier et al., 2006).

The Sp receiver function data were filtered using a suite of filter band-pass values and migrated into our image volume using a common-conversion point migration method (Dueker and Sheehan, 1997). The band-passes used were a high pass of 30 s with low-passes of 2 s, 3 s, 4 s, 6 s, and 8 s. The image volume is parameterized using 110 km wide image bins that are two km thick. In the upper 150 km, our migration velocity model consists of an isotropic shear velocity model (Shen et al., 2013) that is smoothly merged with the AK135 velocity model at 150–170 km depth to produce a model that extends to 300 km depth. A corresponding P-wave velocity model is derived by scaling this shear velocity model to a P-wave model using an assumed crust and mantle V_p/V_s of 1.76 and 1.81 respectively. Sp times are mapped to depth by extracting the S- and P-wave velocity functions along the incident and converted ray-paths and using these velocity functions in the one-dimensional Sp move out equation.