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New insights into the structural elements of the upper mantle beneath the contiguous United States from *S*-to-*P* converted seismic waves

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SUMMARY

The *S*-receiver function (SRF) technique is an effective tool to study seismic discontinuities in the upper mantle such as the mid-lithospheric discontinuity (MLD) and the lithosphere–asthenosphere boundary (LAB). This technique uses deconvolution and aligns traces along the maximum of the deconvolved *SV* signal. Both of these steps lead to acausal signals, which may cause interference with real signals from below the Moho. Here we go back to the origin of the SRF method and process *S*-to-*P* converted waves using *S*-onset times as the reference time and waveform summation without any filter like deconvolution or bandpass. We apply this ‘causal’ SRF (C-SRF) method to data of the USArray and obtain partially different results in comparison with previous studies using the traditional acausal SRF method. The new method does not confirm the existence of an MLD beneath large regions of the cratonic US. The shallow LAB in the western US is, however, confirmed with the new method. The elimination of the MLD signal below much of the cratonic US reveals lower amplitude but highly significant phases that previously had been overwhelmed by the apparent MLD signals. Along the northern part of the area with data coverage we see relics of Archean or younger northwest directed low-angle subduction below the entire Superior Craton. In the cratonic part of the US we see indications of the cratonic LAB near 200 km depth. In the Gulf Coast of the southern US, we image relics of southeast directed shallow subduction, likely of mid-Palaeozoic age.

Key words: Body wave; Dynamics of lithosphere and mantle; Cratons; North America.

1 INTRODUCTION

The structure and history of tectonic plates are essential research topics in geosciences. Seismology has long contributed significantly to this research. The analysis of converted seismic waves is a highly effective seismological technique to study material discontinuities in the crust and upper mantle. Jordan & Frazer (1975) and Faber & Müller (1980, 1984) studied *S*-to-*P* (*Sp*) wave conversions recorded on analogue teleseismic records. Faber & Müller (1980, 1984) used the reflectivity method (Kind & Müller 1975) to model converted waveforms from the crust–mantle boundary and the upper-mantle transition zone. Burdick & Langston (1977) modelled *P*-to-*S* (*Ps*) converted waveforms from crustal discontinuities also with theoretical seismograms. Vinnik (1977) stacked long period *Ps* converted waves using convolution while accounting for the distance-dependent slowness of the converted phases. Langston (1979) introduced deconvolution in the time domain processing of converted waves. Deconvolution is a source equalization procedure that also reduces the effects of the source side structure and of the deeper mantle structure. However, it may also treat parts of the receiver side response as source signal. In the case of *Sp* conversions

the *SV* component is deconvolved from the *P* component. Possible errors caused by the choice of the deconvolution parameters are considered to be small. The basic observational quantity in the converted-wave technique is the differential traveltime between the input signal and the converted signal. These times were determined in the early applications of this method by reading the onset times of both phases directly from the records. In the receiver function method the differential time is measured between the maximum of the deconvolution spike and the maximum (or minimum) of the deconvolved converted signal. However, deconvolution processing, as well as the use of other acausal filters, may produce acausal zero-phase wavelets with significant sidelobes on both sides of the wavelets that will interfere with real converted signals. The particular problem of sidelobes when using deconvolution is discussed in several papers (Li *et al.* 2007; Kumar *et al.* 2012; Hansen *et al.* 2015; Liu & Gao 2018). Krueger *et al.* (2019) adopt a method that minimizes deconvolution sidelobes and conclude that mid-lithospheric discontinuity (MLD) depths and amplitudes are highly variable, and MLDs are less common and lower amplitude in the centres of cratons. Lekic & Fischer (2017) compare different deconvolution methods and illustrate the main limitations of this processing tool.

We return here to the original usage of the *Sp* conversion technique by picking the *SV* arrival time as the reference time for stacking rather than using the deconvolution spike as reference time. The wavelet produced by this approach does not lead to any acausality and is similar to a minimum phase wavelet, therefore, it doesn't contain sidelobes prior to the signal. This applies for the *SV* signal and also for the *S*-to-*p* converted signal from the Moho. Our approach is not entirely new and there are notable examples of processing converted waves without deconvolution (Menke & Levin 2003; Bostock 2004; Kumar *et al.* 2010, 2012; Bodin *et al.* 2014; Eilon *et al.* 2018). The advantages of waveform summation techniques were reported by Shearer (1991) who produced impressive images of stacked long period record sections that clearly resolved otherwise weak seismic phases.

The aim of this study is to verify, with a modern version of the original *Sp* conversion technique, previous observations of upper-mantle discontinuities like the lithosphere–asthenosphere boundary (LAB) and MLD using densely recorded data from the contiguous United States (US). There exists a large number of seismological studies of the LAB and MLD in the US, particularly those using *S*-receiver functions (SRFs; Ramesh *et al.* 2002; Rychert *et al.* 2005; Li *et al.* 2007; Rychert *et al.* 2007; Eaton *et al.* 2009; Hansen & Ducker 2009; Abt *et al.* 2010; Fischer *et al.* 2010; Yuan & Romanowicz 2010; Lekic *et al.* 2011; Kumar *et al.* 2012; Levander & Miller 2012; Hansen *et al.* 2013; Foster *et al.* 2014; Hopper *et al.* 2014; Lekic & Fischer 2014; Wirth & Long 2014; Fischer 2015; Hansen *et al.* 2015; Hopper & Fischer 2015; Karato *et al.* 2015; Kind *et al.* 2015; Calo *et al.* 2016; Cooper *et al.* 2017; Roy & Romanowicz 2017; Chen *et al.* 2018; Liu & Gao 2018). A generalized result of these observations is that the LAB is clearly measured in the tectonically active western US and at the eastern margin of the continent at 100 km depth or less. Surface wave tomography found in the cratonic North America a lithospheric thickness near 200 km (e.g. Priestley & McKenzie 2006; Priestley *et al.* 2019). However, receiver function observations of the cratonic LAB at the expected larger depths (200–240 km) are controversial. This controversy is important because the depth and width of the LAB are decisive parameters needed to understand the origin of the LAB beneath cratons, and particularly the role that temperature plays (Mancinelli *et al.* 2017). Equally controversial is the question of the existence, interpretation and areal distribution of the relatively recently observed MLD that has been reported beneath the cratonic US. The MLD is commonly reported at a depth of 80–100 km which is considerably shallower than the expected depth of the cratonic LAB. Several petrological and physical models of the LAB and MLD have been suggested based on these observations (Mierdel *et al.* 2007; Eaton *et al.* 2009; Karato *et al.* 2015; Rader *et al.* 2015; Selway *et al.* 2015; Cooper *et al.* 2017; Dasgupta 2018; Saha *et al.* 2018; Aulbach 2019).

2 DATA AND METHOD

We use open access data from the IRIS archive, mainly from the USArray project (Ekström *et al.* 1998), the USGS national network, other permanent networks, and some temporary networks. All networks used are listed in the Acknowledgments. Altogether we obtained data from about 2200 stations (Fig. 1). We downloaded data from events with magnitude > 5.6 within epicentral distances between 60° and 85° and requested seismic broad-band waveforms for a time window of 400 s before and after the theoretical *S* arrival time using the IASP91 global reference model (Kennett & Engdahl

1991). The software package Seismic Handler (Stammiller 1993) is used for the data processing. The three-component traces are rotated from the Z, N, E coordinate system using the theoretical backazimuth and incidence angles according to the IASP91 model into the local ray coordinate system *P*, *SV*, *SH* (also commonly called the L, Q and T components). We automatically pick the first arrival of the *SV* signal (Baer & Kradolfer 1987) using data with a signal-to-noise ratio greater than 6 and correcting for the sign of the onset. We also normalize the traces with the absolute maximum within the window of 10 s after the *S* onset on the *SV* component. To ensure high-quality data we chose only events for which the amplitudes on the *P* component are less than 30 per cent of the input *SV* signal on the *SV* component within the time window of -50 to -20 s before the *S* onset. The reason for applying this criterion is that very large signals on the *P* component in that window can only be noise. At the same time, if we lower this limit too much we have significantly less data and may not be able to obtain a good signal because the converted signals of interest are in most cases below the noise level in individual traces. The expected *S*-to-*P* converted LAB or MLD signals are only a few per cent of the incident signal. After applying these data quality criteria, up to 150,000 three-component records remained. To reduce the amount of data we resampled all traces to 10 samples per second by linear interpolation. No other filter is used since our intention is to do everything possible to avoid any acausality.

In Fig. 2 we use the raw data of 127 traces from station SADO of the Canadian National Seismograph Network in the region of the Great Lakes (Fig. 1) to show how our stacking method works. The *SV* component is displayed in Fig. 2(a) and the *P* component in Fig. 2(b). Both components are aligned along the automatically picked onset of the *SV* signal. The summation traces are shown on the top of the suites of traces. We applied bootstrap resampling analysis to estimate the uncertainty in the summation traces. For each component, a subset of 60 per cent of total traces is randomly selected to generate a sample stack. We repeated this process 100 times and then calculated the mean and standard deviation (σ) of the 100 generated sample stacks. The mean summation of each component is plotted along with a range of $\pm 2\sigma$ from the mean, which can be considered as a rough estimate of the uncertainty. Clear signals are visible on the *SV* components and the summation trace has a simple waveform (Fig. 2a). In contrast, no converted signal is visible on the individual *P* component traces (Fig. 2b). These signals are all below the noise level. However, a precursor signal is visible in the summation trace. The summation *SV* component (Fig. 2a) is similar to a minimum-phase wavelet with the positive maximum amplitude at ~ 1.5 s and a significant negative swing afterwards. We note that onset times of the converted signals on the *P* component are referenced to the onset of the wavelet on the *SV* component.

3 DATA AND COMPARISON WITH THEORETICAL SEISMOGRAMS

Here we discuss summation traces of unfiltered broad-band *SV* and *P* components from two large regions of the US and compare them with theoretical seismograms of models within these regions. These regions are indicated in Fig. 3. The western region covers most of the tectonically active western US and the central region is primarily within the cratonic central US. Ten thousand randomly chosen traces recorded with the USArray in both regions are summed in Fig. 3. All traces are lined up along the picked onset times of the *SV* signals and summed. The summation uncertainty is determined

Figure 1. Distribution of seismic broad-band stations used in this study on a simplified tectonic map of North America (Whitmeyer & Karlstrom 2007). The orange dashed line marks the boundary between the shallow (<100 km) LAB region in the west and the deep LAB region (~200 km) in the east according to our study. It agrees well with the tomographic results of Zhu *et al.* (2017). WyC denotes the Wyoming Craton. The blue dot (north of Lake Ontario) marks the location of a sample station, whose data are shown in Fig. 2. Inset shows locations of earthquakes that are used.

similar to Fig. 2. The SV and the Moho signals are clearly visible in both regions. The LAB signal is visible clearly above the noise level in the western region before the Moho signal on the P component. No comparable signal is visible in the data of the central region which could indicate the existence of the MLD. In the following we compare summation traces from both regions with theoretical seismograms of velocity models of these regions. About 17 000 traces have been summed in the central region and about 35 000 in the western region. The black traces are theoretical seismograms calculated with the reflectivity method (Kind 1985) which allows different structures at the source and receiver sides. In order to reduce effects of the source side we avoided the crust there and extended the mantle to the surface. The theoretical seismograms are computed for the distance range of 60° – 85° and summed like the observed data. The blue trace in the Fig. 4 is the source-time function used. It is a nearly perfect spike without any sidelobes. No response of a recording system is added to the theoretical seismograms. The broad-band recording systems have a flat response within the period band that we are using. The red traces in Fig. 4 are theoretical seismograms calculated using the crust and upper-mantle model of Massé (1973). This model is for the cratonic US and was obtained from a large controlled source experiment using wide-angle observations and it was considered to indicate the existence of a low velocity layer within the upper mantle. The resulting

theoretical signals contain much longer periods than the input signal which is due to the attenuation applied in the model. The Moho in the model is at 42 km depth, a high velocity zone between 77.5 and 93.5 km depth and a low velocity zone between 93.5 and 107 km depth. In Fig. 4 the averaged observed Moho signal is at about 4 s precursory time. No negative signal is observed (black trace) before the Moho on the P component which could be associated with an averaged signal from the MLD. The resulting theoretical seismograms are clearly more oscillatory than the observed data. However, we should keep in mind the observed traces are averaged over a very large region with significant variations of the crustal structure which could smooth the signals. The complicated signals of the theoretical seismograms before the SV onset are produced by reverberations within the multilayered upper mantle and crust. The crustal multiples after the SV onset time are also much more complicated than the observed data.

We have added to Fig. 4 the convolved traces of the computed P and SV components with the observed SV component, which may represent the averaged source-time function. This improves the agreement of theoretical and observed seismograms (dashed red lines). The SV waveform contains the averaged source function, the Earth response (especially from near the receiver) and the response of the recording system. There are only slight changes before the SV onset time. The fit of the data is, however, considerably

Figure 2. (a) Normalized and sign corrected SV signals recorded at the station SADO of the Canadian National Seismograph Network in the Great Lakes area (blue dot in Fig. 1). Traces are aligned and summed along the automatically picked first arrivals. (b) P components of the same data using the same amplitude scale with the simultaneous time-shift with each corresponding SV component (summation trace is enlarged by a factor of 5). In each summation panel, the thick line denotes the mean from the bootstrap test (see the text for more details), and the two thin lines on either sides of the mean represent bootstrap mean $\pm 2\sigma$. Positive amplitudes within 2σ are shaded in blue, while negative amplitudes within 2σ are shaded in red. The precursor on the P component is the Moho conversion.

improved after the SV onset time. This means that considering the SV waveform as the average source-time function allows us to better explain the coda of the converted signals.

Next we compare the same data with theoretical seismograms of the IASP91 upper-mantle model (Kennett & Engdahl 1991) and the crustal model for the cratonic US from Dreiling *et al.* (2017; Fig. 5). The IASP91 model has no velocity discontinuity between the Moho and the 410 km discontinuity. In the Dreiling *et al.* (2017) model the Moho is at a depth of 40 km. This model reproduces the Moho signal well. There is no signal in front of the Moho, in agreement with the observed data. The crustal multiples differ somewhat. However, this may be not unexpected since the observed trace is the result of stacking of many records obtained in a large region, whereas the theoretical seismograms are obtained from one receiver side model.

In Fig. 6 we compare averaged receiver function data in the tectonically active western US with theoretical data computed with the crustal model based on McCarthy *et al.* (1991), which was derived for the southwestern US from controlled source wide-angle data. The Moho is at 30 km depth in this model and the LAB has a velocity reduction of 0.3 km/s at 90 km depth with a gradual velocity increase beneath. The fit of the Moho amplitude is also

relatively good. The larger width of the observed Moho signal in the western US is likely due to the greater variability of the crustal structure there. The observed trace has a negative swing from about -13 to -7 s and the theoretical signal of the negative discontinuity at 90 km appears at about 9 s. That means our SRF data confirm the existence of the LAB in the western US at a depth of less than 100 km.

We note that the fit of the theoretical and observed traces in the coda after zero time is not very good in either of the two components displayed in Figs 4–6. There are probably two reasons for this result. First, the source time functions may have generally a long period negative over-swing, and second, the near-receiver response may not be fitted very well. It is not trivial to separate these two effects. The instrument response is probably not the reason. Application of the Streckeisen STS-3 response to theoretical seismograms did not lead to significantly different traces. We did not want to remove the instrument response from the data to avoid any possible acausality due to signal processing. We did not study this question further because we concentrate here on the SV precursors. However the separation of source effects and near-receiver structure effects is worthy of further investigation.

Figure 3. (a) Two regions in the US from which causal S -receiver functions are compared, (b) about 10,000 summed traces from the western region and (c) about 10,000 summed traces from the central region. Bottom traces (P components) are five times amplified. Note the LAB observation in the western region.

4 MIGRATED CAUSAL SRF (C-SRF) PROFILES

Fig. 7 shows the location of a number of SRF profiles across the US computed with the new method. In Fig. 8, we compare a migrated SRF profile computed with the C-SRF method with results from the deconvolution method. The IASP91 model is used for the time-to-depth migration. The width of the profiles is indicated by two black lines on the map. All rays traveling within the profile width contribute to the resultant seismic image. We used a large profile width in order to obtain enough traces for a good signal-to-noise ratio in the image. Some additional profiles with narrower widths are shown in the Supporting Information (Figs S1–S3). The Moho, LAB and the 410 km discontinuities are marked in Fig. 8(a). The LAB in Fig. 8(a) is marked by a dashed line which is not drawn through the centre of the red signal (as is done when deconvolution is used) but marks approximately the onset of the LAB signal. In addition to the mentioned phases, an inclined positive converted phase below the Great Lakes is marked 'X'. In the SRF method using deconvolution (Fig. 8b) the MLD is also marked. The presence of a strong MLD in Fig. 8(b) is the most significant difference between the two methods. We note that the 410 km discontinuity is also observed in Fig. 8 and

marked '410'. In addition a negative signal marked LVL is observed above the 410 km discontinuity. Similar signals are also observed, for example by Vinnik & Farra (2007). The origin of such signals is controversial but a commonly discussed interpretation is that it may be due to partial melt caused by dehydration of the underlying mantle transition zone (Bercovici & Karato 2003). The LVL signal above the 410 is not the focus of our present study since it possibly could be the product of the coda of the teleseismic source signal (see observed SV signals in Fig. 3). Additional studies are needed to determine if this signal is due to the source or the structure below the receiver. Therefore we do not discuss the negative signal above the 410 further here. There is also a positive amplitude arrival at ~ 275 km depth in Fig. 8(a). Such signals are visible in nearly all images. They could be an indication of the 'Lehmann' discontinuity (considered as bottom of the asthenosphere). However, we defer further discussion of the discontinuities below the LAB in a separate paper. We should emphasize that the difference between the two methods is the use of deconvolution and the different definition of the reference time for the SV component. The LAB signal in the western US is observed with both methods. Below the cratonic central and eastern North America, a wide range of weak negative amplitudes can be observed in Fig. 8(a) at depths of 150–200 km

Figure 4. Comparison of observed and theoretical seismograms. The recorded traces (black, 17 000 traces) are summed records from USArray stations within the central region marked in Fig. 3. The red traces are theoretical seismograms computed with the reflectivity method (Kind 1985) using the Massé (1973) crustal and upper-mantle model. The blue trace is the input signal for the computation of theoretical seismograms. Zero time is the picked onset time of the SV signal which is used for lining up all traces for summation. The dashed red traces are the convolution of the computed P and SV components (continuous red lines) with the observed SV waveform (black).

Figure 5. Same as Fig. 4 but using the IASP91 mantle model and the Dreiling *et al.* (2017) crustal model for the computation of the theoretical seismograms. The Moho is visible but the LAB is not visible.

Figure 6. Same as Fig. 4 but for the tectonically active western US with 35 000 traces summed). The crust and mantle model modified from McCarthy *et al.* (1991) is used for the computation of the theoretical seismograms. The LAB depth is put at 90 km. Moho and LAB signals are clearly visible in the data and the theoretical seismograms.

Figure 7. Location of migrated *S*-receiver function profiles displayed in Figs 8–12.

that may be the cratonic LAB (marked LABc). The signal marked X is very clear in the new causal SRF image but is completely missing in the deconvolution image. In the deconvolution image the signal marked X is overwhelmed by the apparent MLD. We used a time-domain Wiener spiking filter for deconvolution. The computation is done using the Toeplitz matrix made from the autocorrelation of a specified time window over the *SV* waveform. Before evaluation the diagonal elements of the Toeplitz matrix are multiplied by a pre-whitening factor. There are many versions of the deconvolution

method, however the observation of the MLD in the cratonic US is common to all these methods.

5 RESULTS AND DISCUSSION

In the following we discuss more profiles through the US obtained with the new C-SRF method. Two east–west profiles in the northern and central US (Fig. 9) show the same significant features as in Fig. 8 (Moho, LAB and 410). Some phases extend in both pro-

Figure 8. Comparison of migrated southwest to northeast profiles across the US obtained by the C-SRF method (a) and the S-receiver function method (b). The width of the profiles is marked on the maps by the black lines. The maps above the profiles are oriented parallel to the profiles. All rays between the two black lines have contributed to the image. Positive signals (velocity increase with depth) are marked blue. Negative signals (velocity decrease with depth) are marked red. The amplitude scale (amplitude ratio of converted signal to SV signal) is given on the right. Typical phases like Moho, LAB and 410 are marked. The LAB, for example, is marked by a dashed line at the deepest part (not the maximum) of the red signal which is due to the use of onset times in our method. In panel (a) is also marked a scattered cratonic LAB (LABc) and previously unknown inclined discontinuity marked X. In panel (b) is marked the MLD. The most important difference between the two methods is that in the C-SRF method the MLD has disappeared, and now a new signal marked X can be imaged. The X signal is interpreted as relict of Archean or younger low-angle subducted plate below the Superior Craton.

Figure 9. Two depth migrated east-west profiles obtained with the C-SRF method. The same phases are marked as in Fig. 8(a). Both the signal X and the cratonic LAB (LABc) are significantly weaker in the southern profile (panel b).

files over more than 5000 km. The signal marked LAB (Fig. 9a) corresponds to the previously determined shallow LAB west of the Rocky Mountains front (e.g. Abt *et al.* 2010; Hansen *et al.* 2015). Significantly, this red signal (a velocity decrease) does not continue east of the Rocky Mountains front (Fig. 9). However, a negative

converted signal observed by the SRF at a similar depth in the mid-continental US is frequently interpreted as the MLD (see references in Section 1). There are more recent reports on the MLD. Hopper & Fischer (2018) used the SRF method and reported MLD signals in the entire contiguous US, however with weaker amplitudes in

Figure 10. Two profiles in the southern US: (a) east–west profile and (b) northwest to southeast profile parallel to the US– border. The same phases as in Fig. 8 are visible. In addition, a structure ‘Y’ is marked indicating an east to southeast dipping positive structure from the Rocky Mountains Front to the Gulf Coast, marked by a black line. A zone of more scattered negative energy is visible beneath the positive ‘Y’ structure.

Figure 11. Two north–south profiles. (a) Profile showing the LAB in the southern part of the profile. The LAB ends below the Wyoming Craton (‘WyC’). (b) A second image of the ‘Y’ structure and other features are marked.

the mid-continental region. Liu & Gao (2018) observed a strong negative velocity gradient at less than 100 km depth in the contiguous entire US using the SRF method. Chichester *et al.* (2018) also observed the MLD with the SRF method in the Midcontinent Rift. Similar observations of the MLD have been made in Europe by Knappmeyer-Endrun *et al.* (2017). However, Krueger *et al.* (2019) report that MLD observations are less common in the centres of several cratons using a deconvolution method especially designed to suppress sidelobes.

In the central part of the upper mantle (Fig. 9a) we have labelled a very scattered band of red signals as ‘LABc’ (cratonic LAB). These weaker signals do not correspond to sharp discontinuities like the Moho or the LAB in the western US. They indicate rather diffuse scattered zones. Hopper & Fischer (2018) obtained a few observations of the deep cratonic LAB (LABc) beneath the central and eastern US with the SRF method. Knappmeyer-Endrun *et al.* (2017) report similar results in the East European Craton. Thus, the cratonic LAB appears to be a broad (>50 km) transition

Figure 12. Two more north–south images with larger profile width. Profile (a) is west of the Rocky Mountains Front and profile (b) is east of it. The two profiles show a great contrast in the structure imaged. In profile (a) most of the upper mantle is nearly empty of discontinuities, except just below the Moho and from 250 km to the 410 discontinuity. Profile (b) has much more structure in the central part of the upper mantle. The cratonic LABc is clearly visible. Another significant feature of this image is the lack of a MLD along the entire profile (b).

zone that produces only weak S-to-P converted signals (Kind *et al.* 2012).

There is a prominent west-dipping structure below the Moho in Fig. 9(a) (blue, marked by an X and a black line) that is centred beneath the southern edge of Archean Superior Craton. The black line is marked at the deepest part of the blue structure because in the C-SRF method times and depths are determined at the onset, not maximum of the signal. The X signal reaches a depth of about 170 km where it intersects the cratonic LAB. The inclination is much less than it appears in Fig. 8(a) due to vertical exaggeration; the true inclination is about 5°. We infer that this west-dipping feature is a palaeosubduction zone, either with a similar age as the Superior Craton (late Archean, 2.7–2.6 Ga) or perhaps a mid-Proterozoic age that is similar to the adjacent accreted terrains. There are many other relics of Precambrian subduction in the North American cratons. Bostock (1998) measured PRFs in the north-west Canadian shield and suggested that the presence of prominent upper-mantle boundaries indicates ‘cratonic assembly through shallow subduction, imbrication and underplating’. Cook *et al.* (1999) present upper-mantle seismic reflection images, also from northwest Canada, that are consistent with shallow-dip subduction now frozen into the upper mantle. Seismic tomography shows an increase of the thickness of the craton in this region without resolving the slab structure (Clouzet *et al.* 2018). Nettles & Dziewonski (2008) observed the west inclined boundary of the cratonic lithosphere using surface wave tomography (see their fig. 8) at nearly the same location as our profile in Fig. 9. The profile in Fig. 9(b) is located south of the profile in Fig. 9(a) and it shows nearly the same structural elements (LAB, LABc and 410) but the dipping structure X is nearly lost. Here the dipping structure is located not beneath Archean lithosphere, but beneath the mid-Proterozoic accreted terrains. A similar

dipping structure below the lithosphere of the Bohemian Massif in Europe has been found by Kind *et al.* (2017).

Two profiles in the southern US, an east–west profile (Fig. 10a) and a northwest to southeast profile along the US-Mexican border (Fig. 10b) show very clear features. The Moho can be seen continuously on both profiles, and the LAB is clearly seen west of the Rocky Mountains Front. A deeper LAB is also visible east of the Rocky Mountains Front in both Figs 10(a) and (b). A significant structure directly below the Moho near the Gulf Coast is marked as Y in Figs 10(a) and (b). This region is the southern margin of the former supercontinent Laurentia (Hoffman 1988; Whitmeyer & Karlstrom 2007). Feature ‘Y’ can be interpreted as mid-Palaeozoic subducted oceanic plate that marks the convergence zone between Laurentia and Gondwana, which resulted in the formation of the supercontinent Pangea. Evidence for this ancient subduction zone includes a structurally complex accretionary tectonic feature (Thomas 1977; Walper 1977). The Y structure is underlain by a low velocity feature in both profiles in Fig. 10. Evanzia *et al.* (2014) observed a high velocity body underneath the Gulf Coast plain and interpreted it as delaminated lower crust. Ainsworth *et al.* (2014) observed in an SRF study a complicated upper-mantle structure with the LAB in about 100 km depth and a positive discontinuity near 200 km depth but did not detect an inclined structures as we have resolved.

Two north–south profiles (Fig. 11) provide an additional perspective on the structure of the crust, LAB and upper mantle. In the southern part of Fig. 11(a) we see the LAB above 100 km depth west of the Rocky Mountains Front. There is no indication below the Wyoming craton of a similar negative discontinuity. However, we do see indications of the cratonic LABc near 200 km depth. Fig. 11(b) shows another clear image of the south dipping Y structure, as seen in Fig. 10(b).

Figure 13. Comparison of summation traces within boxes of different sizes in five regions. The numbers at the coloured lines refer to the numbers of summed traces within the boxes of the same colour. Panel (b) shows that large differences in the LAB depth in relatively small scales are possible. See the text for discussion.

Two wider north–south profiles provide a more broad scale view of the main features of lithosphere and upper mantle of the contiguous US (Fig. 12). Fig. 12(a) is west of the Rocky Mountains Front. In this region we see the LAB near 100 km depth. There is no LAB near 100 km depth observed in the cratonic US profile in Fig. 12(b). However, we see indications of the cratonic LABc at a depth of 200 km. Also the structure X from Figs 9(a) and 11(b) is visible again, although its maximum depth is only 150 km. In Supporting Information Fig. S4, we show summed time-domain traces from a region within Fig. 12(b) that indicates a LAB depth near 200 km there.

To summarize the imaging of the LAB in Figs 8–12, we can draw a boundary between the tectonically active western US, where the LAB is shallower than 100 km, and the cratonic US, where the LAB depth reaches as deep as more than 200 km. We indicated this LAB boundary on the map in Fig. 1. The boundary roughly follows the Rocky Mountain Front in most of the regions from north to south. However, it makes a curve around the Wyoming Craton (see also

Fig. 9a). Across the boundary depicted in Fig. 1 the LAB depth changes abruptly.

The width of the profiles we used in Figs 8–12 is many hundred kilometres, which means that we may smooth out over smaller scale structure within the upper mantle. Hansen *et al.* (2015) and Hopper & Fischer (2018) presented maps of negative velocity gradients (NVGs) near 100 km depth which show much smaller scale lateral variations. Motivated by those prior observations we compare in Fig. 13 summed traces within boxes with variable sizes in several regions. The boxes are marked in the map in Fig. 13. The smallest boxes (red) have the dimensions 2° latitude $\times$ 2° longitude, the medium size boxes are 3° latitude $\times$ 4° longitude (black) and the largest boxes are 4° latitude $\times$ 6° longitude, respectively. These box sizes refer to the location of theoretical Sp piercing points at 100 km depth using the IASP91 global model. A map of the distribution of the piercing points is given in Supporting Information Fig. S5. All traces have been moveout corrected for an epicentral distance of 67° . The Moho is a strong signal in all boxes. In box A in southern Texas

we see a very complicated signal below the Moho marked Y. This is the same signal evident in Fig. 10 that is interpreted as the remnant of a fossil subduction zone. Boxes B to E together form an east–west profile through the northern part of the US. The LAB signal in the box B is remarkable. The two larger boxes (blue and black traces) have very similar LAB signals, whereas the LAB in the smallest box (red trace) is more than two seconds earlier (corresponding to an about 20 km larger depth). This is an indication of strong lateral heterogeneities in this region. If the size of the box is enlarged the additional traces overwhelm the signals from the smallest (red) box. In boxes C and E there is no indication of an NVG below the Moho regardless of the box size. In box E we see a positive velocity gradient (PVG) near -10 s that is marked ‘X’ because it is the same signal as observed in Figs 8(a), 9(a), 11(b) and 12(b). In box E the signal X is in relatively good agreement in the two smaller boxes (red and black) and it is not visible in the largest (blue) box. This is again an indication of small-scale heterogeneities in this region where data from a larger summation area overwhelm the signals from smaller areas. In box C we see also a positive signal marked PVG which is only weakly visible in the profile of Fig. 11(a). In the mid-continental box D there might be weak indications of an NVG, possibly the MLD. In addition, near 20 s we see good indication of an NVG that could be interpreted as the cratonic LABc.

6 CONCLUSIONS

There are two essential conclusions which could be drawn from our analysis, one methodological and one geological. The methodological conclusion is more fundamental since without proper observations no reliable geological conclusions can be drawn and too much room is left for speculations. The receiver function method has been used for about four decades and is considered to be an established method. We think that the results presented here can contribute to re-evaluation and improvement of this method. It will certainly remain a powerful method but its strengths and limitations must be explored further. An important goal is to find reliable methods to recover signals that are well below the noise level in single traces, as well as methods for the treatment of lateral heterogeneity within the local region where such small signals must be summed. Closely related to this problem is the question of determining the correct amplitudes of such summed signals. The new C-SRF method presented here appears to have the potential to improve the reliability of the converted wave method by removing some processing artefacts.

The main geological result is that the evidence for low-velocity mid-lithospheric discontinuities is elusive below the cratonic US. In the mid-cratonic area of the US it is probably not a general feature, but may exist locally. We did find local indications of the cratonic LAB at a depth of ~ 200 km. These observations cannot be generalized to other cratons since, for example, Shen *et al.* (2019) also found with our new technique the LAB near 80 km depth below the China cratons and no indication for a deeper LAB. Significantly, we have detected some previously unknown geological structures within the North American mantle lithosphere. These include relicts of a Precambrian subduction zone below the Superior Craton and a Palaeozoic subduction zone below the Gulf Coast of the southeastern USA.

SUPPORTING INFORMATION

Supplementary data are available at [GJI](https://doi.org/10.1093/gji/ggaa000) online.

Figure S1. Copy of Fig. 7(a) from main paper with 800 km profile width for comparison with narrower profiles in the following two figures. The two black lines in the map mark the profile width.

Figure S2. Same profile as in Fig. S1, however with 400 km profile width.

Figure S3. Same profile as in Fig. S1, however with 200 km profile width. The main observed structures, the western LAB, the dipping interface X and the cratonic LABc, are visible in all three profiles, which are very different in their width. The signal-to-noise ratio is highest in the 800 km wide profile. That means that the marked structures are really large-scale structures, which, however, do not rule out that more dense observations might see more complicated local structures. However, we think that the improvement of the signal-to-noise ratio with increasing width in the above profile is due to the increasing number of data in a relatively homogeneous region. Otherwise an increase of data from laterally inhomogeneous regions could perhaps even reduce the signal-to-noise ratio.

Figure S4. Time-domain observation of the cratonic LAB below a region within the central region in Fig. 3 (main text) where the deep LAB is observed in Fig. 11(b). 1600 traces are summed.

Figure S5. Theoretical Sp conversion points at 100 km depth of about 150 600 seismic records used and computed with the IASP91 global reference model.

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Observatories Initiative, PB-Plate Boundary Observatory Borehole Seismic Network, PE-Penn State Network, PN-PEPP-Indiana, RV-Regional Alberta Observatory for Earthquake Studies Network, SB-UC Santa Barbara Engineering Seismology Network, SC-New Mexico Tech, SN-Southern Great Basin Network, TA-USArray Transportable Array (NSF EarthScope Project), TS-New Mexico Tech Seismic Network, TX-Texas Seismological Network, UO-University of Oregon Regional Network, US-United States National Seismic Network, UU-University of Utah Regional Network, UW-Pacific Northwest Seismic Network, WU-The Southern Ontario Seismic Network, WY-Yellowstone Wyoming Seismic Network, Y5-Canadian Rockies and Alberta Network.

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