

Sn velocities in western and eastern North America

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Abstract. We have investigated the use of the new delay time data set of *Engdahl et al.* [1997] for the construction of tomographic images of Sn velocity in North America. The delays show a significant difference in average delay between paths located west and east of 105°W. This difference is larger when paths longer than 10° are included, indicating a positive gradient for the S velocity in eastern North America. Discarding outliers, two separate regions in the western and eastern United States and Canada were imaged. Low velocities were found under the Colorado Plateau, extending westwards towards the Gulf of California. In the eastern U.S. the Michigan Basin stands out with low Sn velocity, and the Grenville province is characterized by a positive anomaly.

Introduction

The knowledge of the velocity of shear waves in the lithosphere is important for our understanding of the tectonic regime that influences the evolution of a continental region. This velocity is a strong function of temperature, more than composition, and may decrease dramatically in the presence of volatiles or melt [Nolet and Zielhuis, 1994].

Information about the S velocity in the lithosphere is obtained from surface waves, as well as from local studies of Sn [Beghoul et al., 1994]. However, with the exception of the study by Bannister et al. [1991], who successfully imaged Sn velocity variations of the order of ±1% in Fennoscandia, body wave tomography studies have been limited to compressional velocities [e.g. Hearn et al., 1991; Hearn and Rosca, 1994, in the western US].

Method

In this paper we denote by Sn those S arrivals that travel in the lid and have a turning point in the lithosphere. Though strictly speaking Sn is a headwave, the distinction between Sn and shallow S waves is blurred in practice. In fact, recent work on the sensitivity of seismic rays by Marquering et al. [1998] shows how over-simplified our intuitive notions based on ray theory are. While the true headwaves Pn and Sn are often too small in amplitude to be observable,

shear wave energy arrives at regional distances in the form of a complex wavetrain composed of S and its multiple reflections at the bottom of the

Moho, with a velocity approximately equal to the classically defined Sn velocity, i.e. the velocity just beneath the crust-mantle interface. Catalogues tend to mix the notation S and Sn indiscriminately [Van der Hilst, pers. comm., 1997]. Through careful selection (see section ‘Data’) we ensure that no contamination with crustal phases nor sublithospheric S can occur.

With an accuracy sufficient for this study, the travel time of the true headwave Sn satisfies an equation with three time terms:

$$T_i = \sum_k L_{ik} v_k^{-1} + \delta t_m^{source} + \delta t_n^{stat} \quad (1)$$

where L_{ik} is the length of ray i in cell k and v_k the average S velocity in cell k ; m denotes the index of the source, and n the index of the station belonging to ray i . Since earthquake depths vary, we solve for an unknown δt_m^{source} for every event. However, instead of also allowing for a station correction for every single station, we allowed for a ‘crustal’ delay for every cell and constrain all δt_n^{stat} in the same cell (k) to be equal to the same unknown delay δt_k for which we solve. In our inversions, we have parametrized v in cells of 3° latitude. Both the magnitude as well as the roughness of the solutions were damped, following strategies outlined in Nolet [1987].

The neglect of additional terms in (1) to take account of deeper propagation of S waves causes an error, which depends on the velocity gradient. If the S velocity is constant in the lithosphere (or decreases with depth), the error in the inferred velocity is negligible (0.3% at 15°). A positive velocity gradient however introduces an error, which is reduced by virtue of Fermat’s Principle: at 600 km, close to the average path length of 5.9° for the west, S and Sn arrival times differ about 1 s for a (large) gradient of 0.0033 s⁻¹, leading to an error in the inferred Sn velocity of less than 1%, in case the reported phase is actually S, but this error increases with Δ and is about 2% at 10°.

Data

Propagation of high-frequency Sn is difficult or impossible in some areas with tectonic activity such as the central Basin and Range and the Colorado Plateau [Beghoul et al., 1993]. Elsewhere, the Sn arrival may be difficult to read and noise in reported Sn arrival times has caused am-

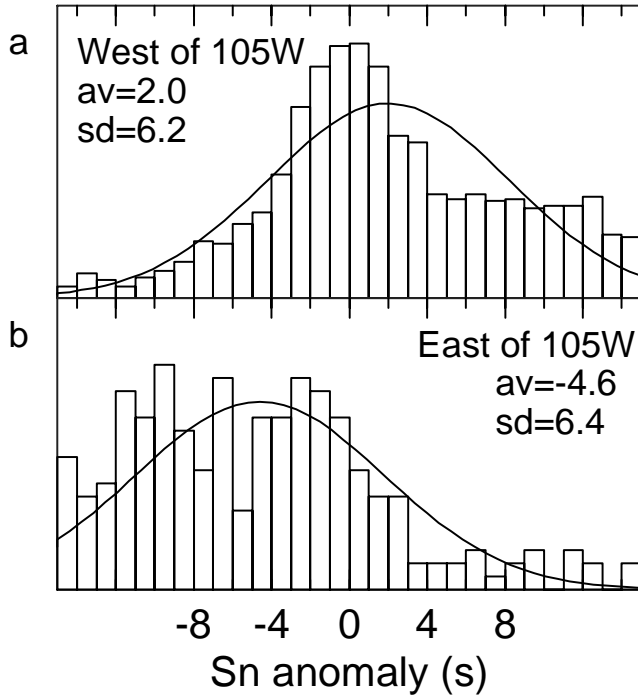


Figure 1. (a) Histogram of Sn delays for all events with $\Delta > 15^\circ$ west of 105°W . The best fitting Gaussian curve has an average delay (av) of +2.0 s, and a standard deviation (sd) of 6.2 s. Delays are w.r.t. model ak135, and taken from *Engdahl et al.*, [1997], who apply a cut-off at ± 15 s. (b) As in (a) but for all events east of 105°W . Average arrival time is -4.6 seconds (early). If paths with $\Delta > 10^\circ$ are excluded, the average decreases to -1.5 s, indicating a positive velocity gradient with depth (see text).

biguous results in inversions of ISC delays [*Zielhuis*, 1992]. However, even though it is small in amplitude, and therefore only intermittently observed, shear energy propagates even in the lithosphere of the western US, as evidenced by stacked records (K. Vogfjörð, pers. comm., 1998), or from several broadband seismograms for regional earthquakes that we inspected. We see therefore no reason in principle to question the reported observations. The fact that this study results in a coherent image for the western US gives further credibility to the reported data.

3460 Sn arrival times were selected from a new catalogue [*Engdahl et al.*, 1997] from 419 reporting stations across the North American Plate. Events reported by fewer than 10 stations or absolute delay times larger than 7.5 s were rejected. Station elevation and ellipticity corrections have been applied. As a first test for ‘signal’ in the data set, we plot the histogram for delays at epicentral distance $\Delta < 15^\circ$ from events west and east of 105°W , respectively, in Fig. 1. It shows a clear difference in the distribution of the delay times, with on average more than 6 s difference.

The distribution of the ray paths (Fig. 2) clearly shows two separate regions with raypath density. We shall analyze these separately.

Large distances ($\Delta > 10^\circ$) are found mostly in the east, where the average Δ is 8.6° (versus 5.9° in the west). When we restrict the epicentral distance to 10° , the average delays in Fig. 1 reduce to +1.8s and -1.5s, respectively. The large

change in the average for the eastern delays indicates the existence of a velocity gradient there, with the possibility of errors in the interpretation (see *Method*). We therefore decided to restrict the eastern data set to $\Delta < 10^\circ$. In both regions we adopted a minimum distance of 2° .

After winnowing, 107 ray paths from stations in the east and 865 for the west were left. The density of raypaths in the Basin and Range and the Colorado Plateau is much lower than in California, but not negligible. This is not necessarily in contradiction with the observations of *Beghoul et al.* [1993], since most of the reported Sn phases in this area are at $\Delta < 5^\circ$, whereas some may have been read from long-period instruments. Contamination with crustal phases Sg and Lg is minimized by our cut-off threshold of 7.5 s.

Resolution

To estimate the available resolution, we used a ‘spike’ test. Synthetic data were calculated for the models shown by the contour lines in Fig. 3a and 4a, which show a +5% anomaly in S velocity. We added Gaussian errors with a standard deviation of 3 s and inverted using the same damping parameters as in the actual inversions. In Fig. 3-4 we have used linear interpolation between the centers of the $3^\circ \times 3^\circ$ cells to reduce the grid size to 1° for plotting purposes. As seen from Fig. 3a, the path coverage is sufficient for a reasonable resolution of the 3° cells for those areas covered by raypaths in the west. The resolution in the east (Fig. 4a) is marginal, and we shall not wish to attach too much significance to the results obtained here. The artificial data errors do not seem to give rise to strong artifacts in the west, except near the border of the region, but do seem to influence the northern part in the east.

Results and interpretation

Fig. 3b shows the inversion result for Sn in the west. The background velocity used is that of model ak135 [*Kennett et al.*, 1995], which has a crustal thickness of 35 km and a sub-Moho S velocity of 4.48 km/s. Though higher Sn velocities are common in oceanic areas or ancient shields, ak135 produces a better fit to continental S travel times [*Engdahl et al.*, 1997].

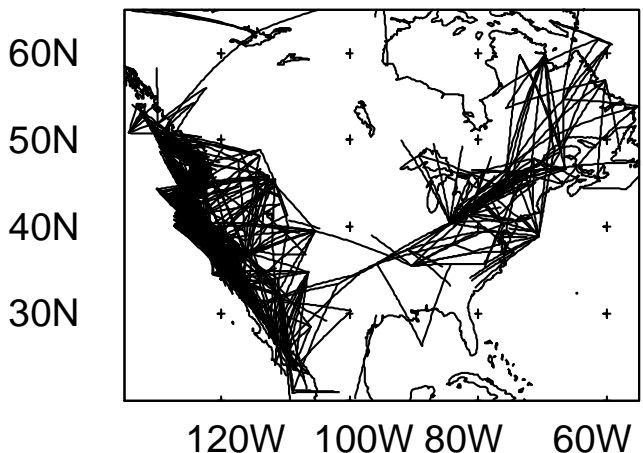


Figure 2. Rays paths ($2^\circ < \Delta < 15^\circ$) of Sn data with absolute delay less than 7.5 s.

As seen in Fig. 3b, lower velocities coincide with tectonic areas that are suspected to involve high temperatures in the upper mantle such as in NW Nevada/SE Oregon where underplating may have caused high heat flow resulting in lower velocities [Klemperer *et al.*, 1987]. This low is also visible in the P velocity map at 70 km depth of *Humphreys and Dueker* [1994]. A less pronounced velocity low is observed beneath the Mohave desert in southern California and extending towards the Walker Lane Belt along the California/Nevada border, where *Hearn et al.* [1994] infer the presence of partial melt from large negative Pn anomalies. The Sn anomaly does not exactly overlap with this Pn anomaly, but the 3° resolution precludes a more detailed comparison. The tomographic image shown in Fig. 3b does show a general characteristic of northeast-southwest trends, following the strike of the basins. The most notable of these starts with the weak negative anomaly under the Snake River Plain (42°N, 111°W) and connects to the negative velocities beneath southern California. The connection coincides with the St George Volcanic Trend [Humphreys and Dueker, 1994]. In the Pn image of *Hearn et al.* [1994] the Snake River Plane anomaly is instead connected to low velocities in eastern and central Oregon. The more neutral Sn velocity in northeastern Nevada seems to preclude a clear continuation for shear velocities in the western direction. This is a feature in the Sn map that differs from the Pn image.

Further south, we also observe low velocities under the Colorado Plateau, which extend westward towards the Gulf of California. Normal to high Sn velocities are found along the coastline in central and northern California, agree with high Pn found by *Hearn et al.* [1994], and are likely related to remnants of oceanic lithosphere from the Juan the Fuca Plate which is currently subducting beneath Oregon and Washington and which has been imaged by both P and S wave tomography at deeper levels [Humphreys and Dueker, 1994; Bostock and VanDecar, 1995; Van der Lee and Nolet, 1997a]. However, the ‘window’ inferred to exist in this slab, and which is visible in the P image at 70 km *Humphreys and Dueker* [1994] as well as in the S velocity image of *Van der Lee and Nolet* [1997a] at deeper levels, does not show up in

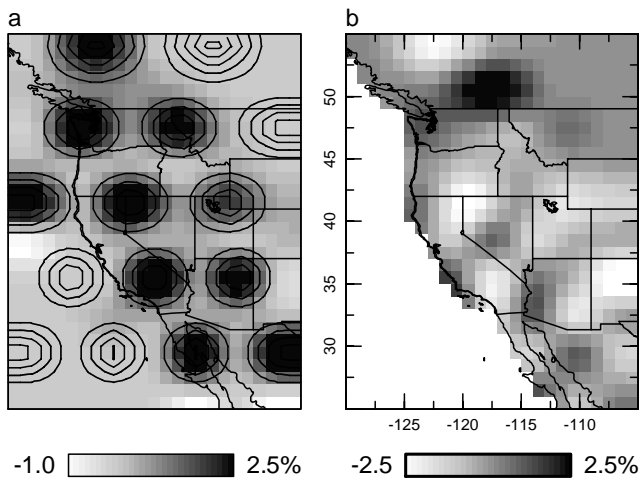


Figure 3. (a) ‘Spike’ resolution test. The contour intervals denote the 5% anomalies of the input model. The greyscale represent the solution found by inverting synthetic data with random errors ($\sigma = 3$ s) added. (b) Sn velocity anomalies.

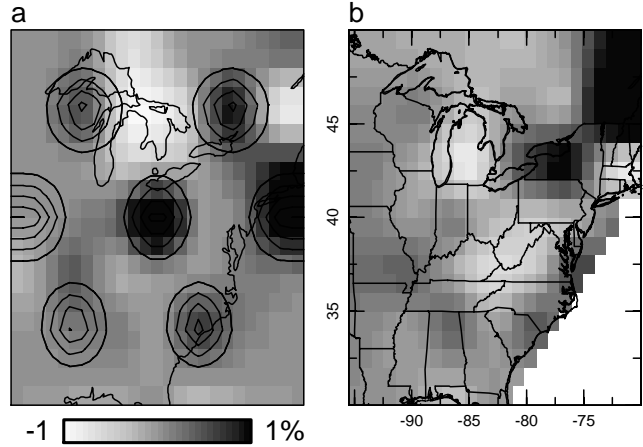


Figure 4. As Fig. 3, for eastern North America

the Sn map, which instead shows a positive anomaly in the Sn velocity at about 34°N, 121°W.

In contrast to the west, the image of the eastern area has a more uniform upper mantle. The most notable area of low velocity appears to correlate with the Michigan Basin, but this must be interpreted with care since the image is sensitive to data errors in this region. At deeper levels it is absent in the S wave image of *Van der Lee and Nolet* [1997b]. The Michigan Basin is an interior cratonic sag basin [Leighton *et al.*, 1990], with its subsidence caused by flexure under a load of unknown origin. A diapiric intrusion has been postulated to cause such a load through replacement of crustal material by heavier mantle rock [Haxby *et al.*, 1976]. However, a direct influence of high temperature to cause the negative anomaly is not likely, since diapiric penetration of asthenospheric material into the lower crust occurred too long ago. The last subsidence of the Michigan Basin occurred about 300 my ago, and may have coincided with the thermal decay of a presumed diapir. Yet, a local enrichment of the upper mantle in the area through such mantle plume could very well explain the lower Sn velocity.

The Saint Lawrence Seaway and the New England areas show up as higher Sn velocities. *Van der Lee and Nolet* [1997b] show a sharp transition from the oceanic to the cratonic upper mantle, which is located at the edge of the Grenville province. However, both the oceanic and the continental sub-Moho S velocity is high, and this is consistent with the extent of the higher Sn velocity observed in this study.

Conclusions

Our Sn velocity tomographic inversions are the first successful attempt to map Sn variations under the north American continent and do show broad correlations to known tectonics, as well as with P velocity maps from *Hearn et al.* [1991, 1994] and *Humphreys and Dueker* [1994]. This, together with the resolution analysis, attests to the usefulness of the new delay data compiled by *Engdahl et al.* [1997]. Among the interesting features that do not directly correlate with known tomographic images for compressional velocity are the high Sn above the slab window in California, a slight velocity high in the northernmost Basin and Range province, as well as a low Sn under the Michigan Basin. In

view of the exploratory nature of this study, it is desirable to substantiate these findings and above all to improve the resolution locally.

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