Relative arrival-time upper-mantle tomography and the elusive background mean

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SUMMARY

The interpretation of seismic tomographic images of upper-mantle seismic wave speed structure is often a matter of considerable debate because the observations can usually be explained by a range of hypotheses, including variable temperature, composition, anisotropy, and the presence of partial melt. An additional problem, often overlooked in tomographic studies using relative as opposed to absolute arrival-times, is the issue of the resulting velocity model’s zero mean. In shield areas, for example, relative arrival-time analysis strips off a background mean velocity structure that is markedly fast compared to the global average. Conversely, in active areas, the background mean is often markedly slow compared to the global average. Appreciation of this issue is vital when interpreting seismic tomographic images: ‘high’ and ‘low’ velocity anomalies should not necessarily be interpreted, respectively, as ‘fast’ and ‘slow’ compared to ‘normal mantle’. This issue has been discussed in the seismological literature in detail over the years, yet subsequent tomography studies have still fallen into the trap of mis-interpreting their velocity models. I highlight here some recent examples of this and provide a simple strategy to address the problem using constraints from a recent global tomographic model, and insights from catalogues of absolute traveltime anomalies. Consultation of such absolute measures of seismic wave speed should be routine during regional tomographic studies, if only for the benefit of the broader Earth Science community, who readily follow the red = hot and slow, blue = cold and fast rule of thumb when interpreting the images for themselves.

Key words: Tomography; Body waves; Seismic tomography; Cratons; Hotspots.

1 INTRODUCTION

1.1 Overview

Constraining accurately the amplitudes of seismic wave speed anomalies is particularly desirable to seismologists since this information can subsequently be used to develop hypotheses concerning the physical state of the mantle (e.g. temperature, partial melt, composition and anisotropy; Karato 1993; Sobolev et al. 1996; Goes et al. 2000; Takei 2002; Artemieva et al. 2004; Anderson 2011). Seismic tomography is a commonly used method to achieve this, but tomographic images can be generated in a variety of ways, with each yielding fundamentally different information about subsurface seismic wave speed structure. Surface-wave methods, for example, can be used to constrain absolute seismic wave speed anomalies across a region with good depth resolution (e.g. Tanimoto & Anderson 1985; Zhang & Lay 1996; Ritzwoller et al. 2002; Pilidou et al. 2005; Priestley et al. 2008). However, they lack the high-quality lateral resolution afforded by higher-frequency body-wave studies, which are thus an extremely commonly used tool for the study of regional upper-mantle seismic wave speed structure (e.g. VanDecar 1991; Tilmann et al. 2001; Allen et al. 2002; Bastow et al. 2005; Rawlinson & Kennett 2008).

In the case of temporary seismograph networks, high levels of background noise, and the emergent (as opposed to impulsive) nature of many body-wave phase arrivals (Fig. 1), means it is often impossible to identify accurately the onset of P-wave or S-wave energy, and thus the absolute traveltime anomaly ($t_{abs}$) for a given station-earthquake pair. Many studies thus instead compute relative arrival-times (e.g. VanDecar 1991; Tilmann et al. 2001; Allen et al. 2002; Bastow et al. 2005; Rawlinson & Kennett 2008). These are times that have been normalized by the removal of a mean arrival-time for a given earthquake from each station of a regional network (e.g. Romanowicz 1979; Aki & Richards 1980).

In the relative arrival-time approach, seismic phases are picked on seismograms at the first identifiable peak or trough in the first cycle of coherent energy across a network (Fig. 1). The similarity of teleseismic waveforms then lends itself naturally to the use of multichannel cross-correlation techniques (e.g. VanDecar & Crosson 1990) to compute relative arrival-times to an extremely high degree of accuracy (Fig. 2). Thus, while absolute traveltime anomalies ($t_{abs}$), computed from the first arriving phase energy ($t_{fb}$) are...
Figure 1. Examples of vertical component $P$-wave seismograms; 40 s of data are shown. Panel (a) shows a noisy $P$ waveform recorded at three temporary seismograph stations in Ethiopia deployed during the Ethiopia Afar Geoscientific Lithospheric Experiment (EAGLE, see e.g. Bastow et al. 2011). Panel (b) shows the same waveforms after application of a zero-phase two-pole Butterworth bandpass filter with corner frequencies of 0.4 and 2 Hz. Panel (c) is an unfiltered example of the best quality data recorded at EAGLE stations. The waveforms in panel (d) are the filtered traces. The $P$ wave picks labelled on each figure are the expected traveltimes according to the IASP91 traveltime tables. The later arriving red picks are those that would typically be used during relative arrival-time analysis (see e.g. VanDecar & Crosson 1990). The first break of energy, shown by the green bars (the absolute traveltime, $t_{fb}$) is particularly difficult to identify in panels (a–c). More impulsive, high signal-to-noise ratio waveforms, such as panel (d), are relatively uncommon in short-duration temporary deployments; hence the popularity of the relative arrival-time approach.

1.2 Relative arrival-time tomography and implications for $\delta V = 0$ per cent

The first 3-D inversion method for retrieving velocity structure from the traveltimes of seismic body waves recorded by a seismic array (the NORSAR array in southern Norway) was developed in the 1970’s by Aki, Christofferson and Husebye (Aki et al. 1976, 1977) and is known simply as the ‘ACH’ method. ACH uses relative-arrival times for two principal reasons. First, relative arrival-times do not require identification of the first breaking energy on a seismogram—instead, the first clear peak or trough of energy can be picked and cross correlation techniques used to align the traces accurately for a given earthquake recorded across a network (Fig. 2). Secondly, relative arrival-times isolate the lateral variations in traveltimes due to variations in the background mean velocity structure for a region. This problem has been well documented before (e.g. Aki et al. 1977; Evans & Achauer 1993; Lévêque & Masson 1999; Foulger 2010), yet unfamiliarity with the
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Figure 2. Bandpass filtered seismic traces. Traces have been aligned with respect to a cross-correlation optimized pick (solid red line) of the initial packet of P-wave energy. All traces have been bandpass filtered with a zero phase Butterworth filter with corner frequencies of 0.4–2 Hz. The vertical dashed lines indicate the 3 s window of data used in the cross-correlation procedure.

to mantle structures directly beneath a seismograph network (the depth range of this region is governed by the aperture of the network and can be up to several hundreds of kilometres; e.g. Evans & Achauer 1993). Because the rays for a given earthquake sample essentially the same Earth structure at large distances from the regional network, uncertainties in hypocentre information (source mis-locations and origin time errors) and the effects of un-resolvable source-side Earth structure are removed from the traveltime data set. This reduces the opportunity for un-warranted structure to be mapped into the region of interest.

In the years since ACH, body-wave traveltime inversion ray-theory techniques have been developed that adopt more sophisticated parametrization schemes (smoothed splines as opposed to blocks), ray geometries (curved as opposed to straight ray-paths) and regularization schemes (smoothing and flattening as well as amplitude damping; e.g. VanDecar et al. 1995; Allen et al. 2002; Rawlinson & Kennett 2008). As with the ACH method, the relative arrival-time data are inverted for a velocity structure assuming some starting 1-D velocity model (based, for example, on the IASP91 traveltime tables of Kennett & Engdahl 1991) to compute initial ray-paths through the Earth.

The use of relative arrival-time data has important consequences for the resulting tomographic images. First, the zero mean of the resulting velocity model will rarely be the same as the global mean (e.g. Aki et al. 1977; Lévêque & Masson 1999). Some studies use additional seismological constraints such as surface-wave data to help constrain the background starting model (e.g. Allen et al. 2002; Rawlinson & Fishwick 2011), but usually this information is not incorporated formally into the inversion and ‘standard’ Earth 1–D models are thus assumed instead. Relatively low and high wave speed areas in the final tomographic model (often shown as red and blue, respectively) cannot always, therefore, be interpreted as genuinely ‘fast’ and ‘slow’, respectively, compared to ‘normal’ mantle. A second problem with ACH-type inversions is that they can only estimate velocity contrasts in the horizontal direction; vertical velocity variations cannot, in a strict mathematical sense, be resolved (e.g. Lévêque & Masson 1999). If an anomalously fast or slow wave speed layer characterizes a given depth range beneath a regional network, its influence on the final inversion will be entirely removed during the computation of the relative arrival-times from the subvertical rays. This is illustrated in Fig. 3, where a network-wide low velocity zone at 200 km is demonstrated to be irretrievable using relative-arrival time tomography. A large, flat, fast wave speed slab would also be invisible during relative arrival-time analysis.

Fig. 4 highlights the difficulties in comparing tomographic images from different tomographic studies. In Fig. 4(b) a slice at 150 km depth through the global tomographic model of Ritsema et al. (2010) can readily be used to identify seismically fast shields in blue, which contrast with the seismically slow hotspots. Fig. 4(a) shows a slice at the same depth through the regional P-wave relative arrival-time tomographic model of Frederiksen et al. (2007) in the Canadian shield. The same study area is characterized in the global model by markedly fast wave speed anomalies (Fig. 4b) so red areas in Fig. 4(a) are, in reality, probably fast compared to the global mean. This assertion is corroborated by the study of van der Lee & Frederiksen (2005), whose NA04 velocity model for the North American continent indicates fast velocity structure for the entire region resolved in Fig. 4(a) (except, perhaps, for the region south of ~45°N where a southward transition to normal and then
Figure 3. A synthetic velocity model consisting simply of a $\delta V_p = -4$ per cent low velocity zone at 200-km-depth beneath Ethiopia. Tomographic inversion (using the method of VanDecar et al. 1995) of relative arrival-time residuals through this model results in a final velocity model with no structure. The influence of the layer is lost entirely during the computation of the traveltime data set. Even if absolute traveltimes were used instead, the subvertical rays would still be completely unable to resolve the layer. Station–earthquake pairs used in the inversion are the same as in the study of Bastow et al. (2008). A Gaussian residual time error component with a standard deviation of 0.02 s was added to the theoretical $P$-wave traveltimes to simulate the uncertainties often encountered in observed data. Areas of poor ray coverage are black.

slower structure occurs). Frederiksen et al. (2007) are thus entirely correct to refer in their manuscript to red regions as ‘low’, not ‘slow’ velocity anomalies. The same correct terminology is used by Sol et al. (2002) in their tomographic study of Western Superior upper-mantle seismic structure—a study area even closer to the heart of the Canadian shield.

Figs 4(b) and (c) highlight a similarly extreme example of the zero mean problem in Ethiopia, East Africa. While the global model identifies Ethiopia as a region of markedly low seismic wave speed, the depth slice through the regional model of Bastow et al. (2008) highlights both high and low wave speed anomalies, precisely as is expected in a relative arrival-time tomographic model. Bastow et al. (2005, 2008) noted that International Seismological Catalogue (ISC) traveltime data for permanent station AAE (Addis Ababa) in Ethiopia show that mean $P$-wave traveltime delays there (4.6 ± 0.15 s with respect to the IASP91 traveltime tables) are amongst the latest worldwide, with the implication that the mantle beneath the region may be amongst the slowest on Earth (an observation also made in the 1980s by surface wave studies such as Nakanishi & Anderson (1984)). The zero mean for Figs 4(b)
and (c) are thus probably as different as is possible anywhere on Earth. Even high wave speed anomalies presented by Bastow et al. (2008) are likely to be markedly slow compared to normal mantle.

2 RECENT MANIFESTATIONS OF THE ZERO MEAN PROBLEM

In a recent study using broadband seismic and data from Ireland, O’Donnell et al. (2011) suggest that compositional contrasts and small-scale convection likely dominate their upper-mantle relative arrival-time seismic tomographic images, with no requirement for elevated temperatures beneath the region. This was in contrast to earlier interpretations of regional tomographic images by Wawerzinek et al. (2008) and Arrowsmith et al. (2005) who cited low wave speed anomalies as evidence for hot (up to ~200 °C), partially molten Iceland plume material ponding beneath lithospheric thin spots. These interpretations are markedly different, and the discordance stems entirely from the studies’ different assumptions about the region’s zero mean velocity structure. While Arrowsmith et al. (2005) and Wawerzinek et al. (2008) interpreted low wave speed anomalies as slow compared to the global average, O’Donnell et al. (2011) cited evidence from a variety of measures of absolute velocity to propose a fast background model for theirs. Compilations of mean traveltime delays indicate that teleseismic P-waves arrive early, not late beneath Ireland and the UK (Poupinet 1979; Poupinet et al. 2003; Amaru et al. 2008), global tomographic studies (Megnin & Romanowicz 2000; Montelli et al. 2006; Li et al. 2008) and regional surface-wave studies (Pilidou et al. 2004, 2005) all indicate that fast P- and S-wave velocity anomalies characterize the upper-mantle beneath the region. Although new constraints on seismic wave speed beneath the UK/Ireland may reveal genuinely slow velocity structure, the published literature on the region presently suggests otherwise. The proposed ~200 °C thermal anomaly for the UK/Irish mantle would, in any case, plot towards the high end of the global temperature range of large igneous provinces and hotspots as determined from petrology (e.g. Rooney et al. 2012), which is somewhat unlikely given the paucity of recent volcanism and high heat flow in the region (e.g. Brock 1989).

3 ESTIMATING THE BACKGROUND MEAN

Some studies use constraints on absolute seismic wave speed structure, for example from surface-wave data, to help constrain their starting velocity model before tomographic inversion of relative arrival-times (e.g. Allen et al. 2002; Rawlinson & Fishwick 2011), but usually this information is not utilized and ‘standard’ Earth 1–D models are thus assumed instead. In these cases, in the absence of an over-determined inverse problem, and a data set of accurate error-free self-consistent absolute traveltimes, it is almost inconceivable that any tomographic model will successfully retrieve the true amplitudes of wave speed anomalies. However, global tomographic models (e.g. Megnin & Romanowicz 2000; Grand 2002; Ritsema et al. 2010) that use large data sets of absolute traveltimes and other closer measures of absolute velocity anomaly (e.g. surface-wave dispersion) can still be used to help place constraints on the background mean velocity structure of a region, before interpretation of a regional ‘relative’ tomographic image. Databases of mean traveltime anomalies recorded at long-running permanent seismograph networks can also be used to approximate the background velocity structure of a region (e.g. Poupinet 1979; Poupinet et al. 2003; Amaru et al. 2008).

Fig. 5 shows vertically averaged S-wave velocity anomalies in the depth range 0–410 km for permanent seismograph stations (including the GeoScope and GSN networks) presently operating around the world (Fig. 6). For permanent stations AAE and FURI in Ethiopia (Fig. 6), mean velocity anomalies are amongst the slowest on Earth (negative anomalies), emphasizing the differences in δV_s between the study of Bastow et al. (2008) and Ritsema et al. (2010), as discussed earlier. The converse is true for the Canadian studies of Sol et al. (2002) and Frederiksen et al. (2007): stations such as FRB and FFC on the Canadian shield (Fig. 6) are characterized by some of the largest amplitude fast wave speed anomalies on Earth.

Figure 4. A comparison of global and regional tomographic models. Panel (A) Slice at 150-km-depth through the P-wave relative arrival-time tomographic model of Frederiksen et al. (2007) in Canada, computed from the inversion of relative arrival-time residuals. Panel (B) Slice at 150-km-depth through the global tomographic model of Ritsema et al. (2010). White lines are plate boundaries. Panel (C) Slice at 150-km-depth through the P-wave relative arrival-time tomographic model of Bastow et al. (2008) in Ethiopia, computed from the inversion of relative arrival-time residuals. Dark lines are mid-Miocene border faults that define the Ethiopian Rift. Areas of poor ray coverage are black.
Figure 5. Panel (a) Histogram of mean shear wave velocity anomalies through the upper 410 km of the global tomographic model of Ritsema et al. (2010). Early arrivals, indicative of fast mantle seismic wave speeds are positive anomalies; late arrivals, indicative of slow mantle seismic wave speeds are negative anomalies. Stations OBN and AAE, shown in Fig. 6 are characterized by the fastest and slowest upper mantle wave speeds on Earth.

Figure 6. Locations of permanent seismograph stations for which mean upper-410 km S-wave velocity anomalies are presented in Fig. 5. Stations are part of permanent networks including Geoscope and GSN.
In the case of the aforementioned UK/Ireland studies, mean wave speed anomalies presented in Fig. 5, indicate fast velocity structure beneath permanent station ESK ($\delta V_s \approx 1$ per cent) in Scotland, which supports the conclusion of O’Donnell et al. (2011) that elevated temperatures and partial melt are not required to explain tomographic images there.

Global absolute velocity tomographic models, and traveltime catalogues, of course, provide only ball-park figures to help guide interpretation of the relative arrival-time images. Archean keels surrounded by slow asthenospheric mantle (e.g. Tanzania) are also genuinely fast wave speed anomalies (Ritsema et al. 1998) so the vertically averaged anomalies shown in Fig. 5 cannot always be used to benchmark precisely a regional ‘relative’ study. Future body-wave tomographic inversions in places like the UK/Ireland could usefully replace their ‘global’ starting model with one constrained a priori from absolute measures of velocity such as surface wave dispersion (e.g. Allen et al. 2002; Rawlinson & Fishwick 2011, for Iceland and Australia, respectively). On the other hand, studies of relative arrival-times alone remain valuable additions to the literature and should not be discouraged. Tomographers should, though, be abundantly clear what their low and high velocity anomalies mean in the context of the global average (see e.g. Schimmel et al. 2003; Bastow et al. 2005, 2008; O’Donnell et al. 2011). While the seismological community may be familiar with the difference between low and slow, and high and fast anomalies, the broader Earth Science community, who are often keen to interpret tomographic images for themselves, are almost certainly not.

4 CONCLUSIONS

Seismic tomographic studies that use relative, not absolute, arrival-time data to constrain seismic wave speeds yield models with a zero mean that corresponds to the average velocity structure of the region. There are very few places on Earth, however, where this zero velocity contour will coincide precisely with the $\delta V = 0$ per cent global average. Low wave speed anomalies should thus not necessarily be interpreted as ‘slow’, and high velocities should not necessarily be interpreted as ‘fast’ compared to the global mean. Brief consultation of absolute traveltime catalogues and global tomographic models can help constrain the elusive $\delta V = 0$ per cent contours in the relative arrival-time studies and thus aid interpretation of the results. Perhaps most importantly, such practice could help avoid confusion amongst the broader readership of the tomographic literature.

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