Earth and Planetary Science Letters 297 (2010) 95–102

Contents lists available at ScienceDirect

Earth and Planetary Science Letters

journal homepage: www.elsevier.com/locate/epsl

Layered azimuthal anisotropy of Rayleigh wave phase velocities in the European Alpine lithosphere inferred from ambient noise

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ARTICLE INFO

Article history:
Received 11 December 2009
Received in revised form 1 June 2010
Accepted 5 June 2010
Available online 24 July 2010

Editor: R.D. van der Hilst

Keywords:
surface waves
anisotropy
seismic tomography
Alpine crust
crustal thickening

ABSTRACT

Passive seismic imaging of the earth is a rapidly developing field of study. Recent advances in noise cross correlation techniques allow imaging of isotropic surface-wave and shear-wave velocities in areas where earthquake numbers and distributions are insufficient to implement traditional earthquake based tomography. Furthermore, advances in the theory underpinning surface wave inversion have led to depth-dependent mapping of seismic anisotropy in the lithosphere and upper asthenosphere. We show that by merging these two rapidly advancing fields we can invert noise-based phase velocity measurements for azimuthally anisotropic phase speed, thereby providing a highly resolved image of layered azimuthal anisotropy in continental crust. We apply this new algorithm in the western Alps, an area of complex lithospheric structure. Alpine crustal thickening results from continental collision, indentation of the Ivrea mantle into the middle crust of the European plate, rollback of the European lithospheric mantle, and crustal slicing. We find that anisotropy beneath the central Alps is stratified in two layers — one with an orogen-parallel fast direction above ~30 km depth and another with a strong orogen-perpendicular fast direction at greater depth. Although our resolution is reduced outside the central Alps, we map orogen-parallel anisotropy in the crust of the northern Alpine foreland. We interpret the results in the central Alps as first-order evidence for a model of azimuthal anisotropy in which (1) near-vertical emplacement of crustal slices following detachment of the lithospheric mantle from the crust gives rise to orogen-parallel fast directions of wave propagation in the crust, and (2) dominantly horizontal tectonics in the thickened crustal root and uppermost mantle yield orogen-perpendicular fast directions at greater depth.

1. Introduction

Although the central Alps have likely been studied as long as any orogen in the world, there is still no consensus model explaining the geodynamic development of the Alpine lithosphere and uppermost asthenosphere. The Alpine orogeny resulted from the collision of the Adriatic lithosphere (promontory of the African plate) with the European plate in the Late Cretaceous (e.g. Coward and Dietrich, 1989). In the Oligocene, subducting lithosphere changed from oceanic to continental and slab breakoff likely occurred (Davies and von Blanckenburg, 1995). Much of our knowledge of the current state of the Alpine lithosphere comes from traditional body and surface wave seismic imaging which have shown high seismic velocities in the upper mantle below the Alps (Lippitsch et al., 2003; Piromallo and Morelli, 2003; Fry et al., 2008; Peter et al., 2008; Boschi et al., 2009). These high velocity anomalies have been interpreted as evidence for subducted slabs. Thickened crust in the Alps has also been imaged (Wortel and Spakman, 1993; Waldhauser et al., 1998; Lippitsch et al., 2003). Prior to our current study, tomography of the Alps has been limited to inversions for isotropic velocity structure. Seismic aniosotropy information is complementary to the isotropic velocity studies and can provide constraints on geodynamic models of the region.

In response to strain, crustal (e.g., amphibole) and mantle (e.g., olivine) minerals can develop some specific fabrics that result in seismic anisotropy. The direction of fast propagation is usually the direction of maximum deformation (Christensen, 1984; Nicolas and Christensen, 1987). Seismic anisotropy beneath continents has been mapped with high-resolution using shear-wave splitting (Silver, 1996; Savage, 1999; Fouch and Rondenay, 2006), but the origin and interpretation of this anisotropy is controversial. Some measurements agree with regional tectonic structure (e.g., in collisional regimes, a direction of fast propagation that is sub-parallel to the belt), suggesting a lithospheric origin, whereas others show large scale patterns that

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doi:10.1016/j.epsl.2010.06.008

correspond to the absolute plate motion, suggesting an asthenospheric origin. Surface waves sample different depths depending on their period. This trait provides the vertical resolution that is missing in shear-wave splitting measurements. Global and continental scale models based on surface-wave data have resolved different anisotropic patterns in the lithosphere and in the upper asthenosphere (Simons et al., 2002; Debayle et al., 2005; Marone and Romanowicz, 2007). Because of the lateral resolution of these models, relating lithospheric anisotropy to regional tectonics is difficult. However, by using data from dense arrays of seismic stations, isotropic and anisotropic anomalies can be mapped with good resolution at a regional scale (Pedersen et al., 2006; Yang and Forsyth, 2006; Zhang et al., 2007; Deschamps et al., 2008a; Eberhart-Phillips and Reyners, 2009). Layered anisotropic structure has been interpreted as resulting from both frozen and present sources, and relates to both past tectonics (Darbyshire and Lebedev, 2009) and present absolute plate motion (Deschamps et al., 2008b). Previous studies utilized earthquake sources, which are naturally spatially and temporally limited. Areas with inhomogeneous distributions of earthquakes such as the European Alps are difficult to image with these methods. Noise-based tomography does not use earthquake energy and consequently is not hindered by heterogeneous earthquake distributions. We therefore employ a method to invert ambient noise data for regional scale layered anisotropy.

2. Methods

It has been shown that the surface-wave portion of the Green's Function between two receivers can be estimated by stacking the cross correlation functions (CF) of ambient noise recorded at each of the stations (Lobkis and Weaver, 2001; Snieder, 2004; Shapiro et al., 2005; Sanchez-Sesma et al., 2006). Stacking cross correlations from each station pair statistically requires that incoherent energy propagating off the great-circle path between the stations interferes destructively; coherent energy that propagates directly between the receivers is isolated. Furthermore, taking the time derivative of the CF yields the time domain empirical Green's Function (EGF) from which reliable phase velocity estimates can be made (Sabra et al., 2005; Yao et al., 2006). However, in an ambient field resulting from a homogenous distribution of sources, extracting the phase velocities from the CF is possible if an appropriate phase correction is applied. In a homogeneous isotropic media, the phase of the CF resulting from an isotropic energy distribution is advanced from the GF by $\pi/2$ (Yao and van der Hilst, 2009). In the presence of a unidirectional energy source, the phase is still advanced, but to a lesser degree. In our study area, energy is well distributed when averaged over a year of cross correlations. In fact, at the most important periods of our study (~20 s), energy is dominated by a very wide distribution (~180°) owing to a large region of oceanic disturbances in the northern Atlantic Ocean during the northern hemisphere winter (Stehly et al., 2006). This energy distribution would therefore imply that at all inter-station azimuths, a phase correction of $\pi/2$ is appropriate to measure Rayleigh wave phase velocities. To confirm the validity of our phase correction, we have compared measurements from traditional two-station analysis to measurements made from the cross correlations along an azimuth parallel to the average trace of the central Alps.

In a directionally biased ambient wavefield, estimates of phase velocity from noise cross correlations will be azimuthally biased (Yao and van der Hilst, 2009). However, in their case study of central Asia, Yao and van der Hilst (2009) show that azimuthal bias is small and point out that spatial smoothing serves to further reduce the effects of azimuthally dependent velocity estimates on inversions for anisotropy. In the present study, we present first-order results for azimuthal anisotropy by classifying the anisotropy as either orogen-parallel or orogen-perpendicular. Small-scale variations in fast directions are not considered robust and are consequently not interpreted.

2.1. Cross correlation and phase velocity measurements

We apply a noise cross correlation and stacking procedure similar to the method of Stehly et al. (2009). Continuous data from 29 broadband (STS-2) stations in the SDSNet and 7 broadband (STS-2) stations from TomoCH have been correlated on a day-by-day basis. In this method, we first cut seismic recordings to 1-day files. The signals are then decimated to 1 Hz and the trend is removed by subtracting the best-fit line from the signal and the signals are tapered. We deconvolve the instrument responses, leaving a record of ambient velocity recorded at the site. The data are then spectrally whitened in a frequency band between 0.008 and 0.33 Hz (3 to 125 second periods). The whitening lessens the effects of spectral peaks that result from earthquakes and dominant oceanic microseismic peaks. We perform the time-domain cross correlation for the vertical component of each day of recording for each station pair. Because the particle motion of Rayleigh waves is recorded on the radial and vertical components, by looking at the vertical component, we are isolating the Rayleigh wave signal. The cross correlation functions are then stacked for each station pair. With a sufficient number of stacked correlations, the stacked-function becomes stable in our frequencies of interest. Stehly et al. (2006) found a stable CF after stacking just one year of data.

We then determine the phase information as a function of instantaneous frequency from the cross correlation function and use these data to determine the corrected phase dispersion curve with a modified 2-station method of Meier et al. (2004). Phase velocities are manually chosen to negate multiple-cycle ambiguities by comparing the obtained velocity to a background earth model and 'picking' the appropriate cycle number. Between 185 and 332 phase velocity observations were used for each inversion we present (Fig. 1).

2.2. Inversion of dispersion curves

For each inter-station path, the measured dispersion curve represents the average Rayleigh-wave phase velocity along the path as a function of period (or frequency). Following the same inversion procedure as in Deschamps et al. (2008a), we use a perturbative approach to invert our catalogue of phase velocity measurements for the isotropic and anisotropic components of phase velocity at discrete periods of observation, between 6 and 60 s, using a ray-theoretical approximation with an explicitly defined ray sensitivity width of 20 km. Deschamps et al. (2008a) noted that the inverted phase velocity model is not sensitive to the chosen ray sensitivity width. Similarly, we performed inversions for ray sensitivity widths between 10 and 100 km, but did not find significant differences in the phase velocity model.

At any given frequency, phase velocity variations in a laterally heterogeneous earth can be written as (Smith and Dahlen, 1973)

$$\delta C(\varphi,\theta) = \delta C_{iso}(\varphi,\theta) + \delta C_{2\psi}(\varphi,\theta) + \delta C_{4\psi}(\varphi,\theta) \tag{1}$$

where ψ is the azimuth of anisotropy. We discretize the anisotropic contributions of wave speed as

$$\delta C_{2\psi}(\varphi,\theta) = A_{2\psi}\cos(2\psi) + B_{2\psi}\sin(2\psi) \tag{2}$$
$$\delta C_{4\psi}(\varphi,\theta) = A_{4\psi}\cos(4\psi) + B_{4\psi}\sin(4\psi).$$

The anisotropic coefficients contain information relating to both the amplitude (Λ) and direction (Θ) of the anisotropy, as defined by:

$$\begin{cases} \Lambda_{2\Psi} = \sqrt{A_{2\Psi}^2 + B_{2\Psi}^2} \\ \Theta_{2\Psi} = \dfrac{1}{2}\arctan\left(\dfrac{B_{2\Psi}}{A_{2\Psi}}\right) \end{cases} \text{and} \begin{cases} \Lambda_{4\Psi} = \sqrt{A_{4\Psi}^2 + B_{4\Psi}^2} \\ \Theta_{4\Psi} = \dfrac{1}{4}\arctan\left(\dfrac{B_{4\Psi}}{A_{4\Psi}}\right) \end{cases} \tag{3}$$