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Seismic anisotropy tomography: New insight into subduction dynamics

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ABSTRACT

Body-wave and surface-wave tomography, receiver-function imaging, and shear-wave splitting measurements have shown that seismic anisotropy and heterogeneity coexist in all parts of subduction zones, providing important constraints on the mantle flow and subduction dynamics. P-wave anisotropy tomography is a new but powerful tool for mapping three-dimensional variations of azimuthal and radial seismic anisotropy in the crust and mantle. P-wave azimuthal-anisotropy tomography has been applied widely to the Circum-Pacific subduction zones, Mainland China and North America, whereas P-wave radial-anisotropy tomography was applied to only a few areas including Northeast Japan, Southwest Japan and North China Craton. These studies have revealed complex anisotropy in the crust and mantle lithosphere associated with the surface geology and tectonics, anisotropy reflecting subduction-driven corner flow in the mantle wedge, frozen-in fossil anisotropy in the subducting slabs formed at the mid-ocean ridge, as well as olivine fabric transitions due to changes in water content, stress and temperature. Shear-wave splitting tomography methods have been also proposed, but their applications are still limited and preliminary. There is a discrepancy between the surface-wave and body-wave tomographic models in radial anisotropy of the mantle wedge beneath Japan, which is a puzzle but an intriguing topic for future studies.

Contents

1. Introduction

During the past three decades, a great number of seismological studies have revealed that seismic velocity and attenuation heterogeneity, as well as seismic anisotropy, exist widely in the Earth's interior. The

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structural heterogeneities in the crust and mantle have been mapped by using seismic tomography and receiver-function methods, whereas seismic anisotropy has been detected by shear-wave splitting (SWS) measurements, apparent discrepancy between Love and Rayleigh waves, and azimuthal variations of Pn-wave velocity (e.g., Zhao, 2012; Long, 2013; Zhao, 2015a for recent reviews). However, in the conventional tomographic studies, in particular, body-wave tomography, the Earth is usually assumed to be isotropic to the propagation of seismic waves. As pointed out by Anderson (1989), this assumption is made for mathematical convenience, and the fact that a large body of seismic data can be satisfactorily modeled with this assumption does not prove that the Earth is isotropic. Seismic anisotropy can cause the largest variations in seismic velocity, which can be even greater than those caused by changes in temperature, composition or mineralogy. Hence, anisotropy is a first-order effect (Anderson, 1989).

Seismic anisotropy is a very useful and important physical parameter, because it can provide a wealth of new information regarding dynamic processes in the crust and mantle (e.g., Cara, 2002; Fouch and Rondenay, 2006; Wang and Zhao, 2008; Long, 2013; Wang and Zhao, 2013; Koulakov et al., 2015). The major causes of seismic anisotropy are lattice-preferred orientation (LPO) and shape-preferred orientation (SPO) of the materials constituting the Earth. In the crust, orientations of local and regional structure and tectonics can cause seismic anisotropy, such as fault systems (e.g., Zhang and Schwartz, 1994; Bokelmann, 1995; Huang and Zhao, 2013; Huang et al., 2014; Koulakov et al., 2015). In the mantle, seismic anisotropy may reflect convection flows and is usually interpreted by LPO of olivine crystals (e.g., Karato and Wu, 1993; Fouch and Rondenay, 2006; Wang and Zhao, 2008; Long, 2013; Wang and Zhao, 2013). However, the relationship between anisotropy and mantle flows is uncertain in some cases, e.g., when abundant fluids and melts exist, the anisotropy may be orthogonal to the flow direction. The petrologic basis of such behaviors has been investigated in laboratory experiments (e.g., Jung and Karato, 2001; Karato et al., 2008).

Both body-wave and surface-wave data can be used to study seismic anisotropy. The body-wave methods include SWS, receiver functions, and P-wave travel-time inversion. Fouch and Rondenay (2006) made a detailed review of the methods for studying seismic anisotropy, as well as their advantages and limitations. In the past three decades, many researchers have attempted to use P-wave travel-time data to study anisotropy tomography (e.g., Babuska et al., 1984; Hearn, 1984; Hirahara and Ishikawa, 1984; Hirahara, 1988; Babuska and Cara, 1991; Mochizuki, 1995; Gresillaud and Cara, 1996; Hearn, 1996; Plomerova et al., 1996; Mochizuki, 1997; Lees and Wu, 1999; Wu and Lees, 1999; Bokelmann, 2002; Eberhart-Phillips and Henderson, 2004; Ishise and Oda, 2005, 2008; Wang and Zhao, 2008; Koulakov et al., 2009; Eken et al., 2010; Plomerova et al., 2011; Tian and Zhao, 2012a; Huang and Zhao, 2013; Wang and Zhao, 2013; Wei et al., 2013; Koulakov et al., 2015; Menke, 2015; Wei et al., 2015). However, reliable and geologically reasonable results have been obtained only in recent years, thanks to the availability of abundant high-quality arrival-time data recorded by dense seismic arrays of permanent and portable stations at local and regional scales. The Pn-wave tomography can only estimate two-dimensional (2-D) P-wave velocity (Vp) variations and azimuthal anisotropy in the uppermost mantle directly beneath the Moho discontinuity (e.g., Hearn, 1996), whereas P-wave anisotropy tomography can determine three-dimensional (3-D) distribution of Vp anisotropy in the crust and mantle beneath a seismic network.

Measuring SWS is a popular and useful method for studying (detecting) seismic anisotropy. A great number of researchers have used this method to study seismic anisotropy in many regions of the world, which have provided important information on mantle dynamics (e.g., Crampin, 1984; Silver, 1996; Savage, 1999; Huang et al., 2011a,b; Long, 2013). However, the SWS measurements have a poor depth resolution, and so their interpretations are usually not unique. This drawback can be overcome by Vp anisotropy tomography.

Many researchers have used surface-wave tomography to study seismic heterogeneity and anisotropy (e.g., Tanimoto and Anderson, 1984; Nishimura and Forsyth, 1988; Montagner and Tanimoto, 1991; Ritzwoller and Lavely, 1995; Nettles and Dziewonski, 2008; Yoshizawa et al., 2010; Montagner, 2011; Yuan et al., 2011). As compared with body-wave tomography, however, surface-wave tomography generally has a lower spatial resolution, and so it has been mainly applied to study shear-wave velocity (Vs) structure and anisotropy in the crust and upper mantle at a global scale or a large regional scale for oceanic or continental regions. The information on subduction dynamics provided by surface-wave tomography has been limited by its lower spatial resolution. Recently, Long (2013) made a detailed review of seismic anisotropy studies of subduction zones, whereas her review focused on the SWS measurements and receiver function studies.

In this article, we review the methods and applications of body-wave anisotropy tomography and discuss their implications for the structure and dynamics of subduction zones. Some related SWS measurements and high-resolution surface-wave tomography studies are also mentioned for better interpreting the results of body-wave anisotropy tomography.

2. P-wave anisotropy tomography: Methods

Twenty one independent elastic moduli are required to fully express an anisotropic medium, which is very hard to handle in both theory and practice. Fortunately, anisotropy with hexagonal symmetry is a proper approximation to the materials in the Earth's crust and mantle, which can reduce the number of physical parameters describing seismic anisotropy (e.g., Christensen, 1984; Park and Yu, 1993; Maupin and Park, 2007). To further simplify the problem, we can assume the hexagonal symmetry to be horizontal when the azimuthal anisotropy is concerned in SWS measurements (e.g., Crampin, 1984; Silver, 1996; Savage, 1999; Huang et al., 2011a,b; Long, 2013) and P-wave velocity studies (e.g., Hess, 1964; Backus, 1965; Raitt et al., 1969; Hearn, 1996; Eberhart-Phillips and Henderson, 2004; Wang and Zhao, 2008, 2013); whereas we can assume the hexagonal symmetry to be vertical when the radial anisotropy is concerned in the form of a Vsh/Vsv variation (Vsh and Vsv are the velocities of shear waves polarized horizontally and vertically, respectively) in surface-wave studies (e.g., Nettles and Dziewonski, 2008; Fichtner et al., 2010; Yuan et al., 2011) and in the form of a Vph/Vpv variation (Vph and Vpv are the velocities of P-waves propagating horizontally and vertically, respectively) in P-wave velocity studies (e.g., Ishise et al., 2012; Wang and Zhao, 2013; Wang et al., 2014). Here we introduce the recent tomographic methods for P-wave azimuthal and radial anisotropy, following Wang and Zhao (2008, 2013).

2.1. Azimuthal anisotropy tomography

For hexagonal anisotropy, P-wave slowness can be expressed as (Barclay et al., 1998):

$$S = S_0 + M\cos(2\theta), \tag{1}$$

where S is the total slowness (i.e., 1/V), S_0 is the average slowness (i.e., isotropic component), θ is the angle between the propagation vector and the symmetry axis (Fig. 1), and M is the parameter for anisotropy. In a weak anisotropic medium with a horizontal hexagonal symmetry axis (Fig. 1a), the P-wave slowness of a horizontal ray can be approximately expressed as (e.g., Backus, 1965; Raitt et al., 1969; Hearn, 1996; Eberhart-Phillips and Henderson, 2004; Wang and Zhao, 2008, 2013):

$$S(\phi) = S_0(1 + A_1\cos(2\phi) + B_1\sin(2\phi)), \tag{2}$$

where S is the total slowness, S_0 is the azimuthal average slowness, A_1 and B_1 are the azimuthal anisotropy parameters, and ϕ is the ray path