SEISMIC STRUCTURE OF THE EUROPEAN CRUST
AND UPPER MANTLE BASED ON ADJOINT TOMOGRAPHY

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A DISSERTATION
PRESENTED TO THE FACULTY
OF PRINCETON UNIVERSITY
IN CANDIDACY FOR THE DEGREE
OF DOCTOR OF PHILOSOPHY

RECOMMENDED FOR ACCEPTANCE
BY THE DEPARTMENT OF
GEOSCIENCES
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NOVEMBER 2013
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Abstract

We use adjoint tomography to estimate three-dimensional variations in seismic parameters within the crust and upper mantle beneath Europe and the North Atlantic Ocean. Spectral-element and adjoint methods are used to numerically calculate synthetic seismograms and sensitivity kernels in three-dimensional Earth models. Combined with gradient-based optimization algorithms, e.g., preconditioned conjugate-gradient and L-BFGS methods, we iteratively update seismic models of Earth’s interior. A three-stage inversion strategy is designed to estimate variations in elastic wavespeeds, anelastic attenuation and radial & azimuthal anisotropy. In stage one, frequency-dependent phase differences between observed and simulated seismograms are used to determine a new radially anisotropic wavespeed model for the European crust and upper mantle, namely EU$_{30}$. Long-wavelength structures in EU$_{30}$ compare favorably with previous body- and surface-wave tomographic models. Some hitherto unidentified features naturally emerge from the smooth starting model. In stage two, frequency-dependent amplitude differences combined with remaining phase anomalies are used to simultaneously constrain elastic and anelastic structures. A new anelastic model, named EU$_{50}$, is constructed in this stage. We observe several notable features, such as enhanced attenuation within the mantle transition zone beneath the North Atlantic Ocean. In the first two stages, long-period surface waves and short-period body waves in three-component seismograms are combined to simultaneously constrain shallow and deep structures. In stage three, frequency-dependent phase and amplitude anomalies of three-component surface waves are used to construct a radially and azimuthally anisotropic model EU$_{60}$. We find that the direction of the fast axis is closely tied to the tectonic evolution in this region, such as extension along the North Atlantic Ridge, trench retreat in the Mediterranean, and counterclockwise rotation of the Anatolian Plate. Radial peak-to-peak anisotropic strength profiles identify distinct brittle-ductile transitions in lithospheric strength beneath oceans and continents, in agreement with observations in mineral physics experiments.
Acknowledgements

I was lucky that my advisor Jeroen Tromp gave me an option to transfer from Caltech to Princeton five years ago. Jeroen is a seismologist with unique talent on both theories and applications. Over the past five years, he helped me on both mathematic derivations and dealing with actual seismic observations. I got a lot of supports and suggestions from him when I met difficulties on my research. Sometimes it took me a while to figure out the issues and I was always impressed by the visionary of Jeroen. His enthusiasm and optimism gave me courage to improve and modify my original thoughts and approaches. It was impressive to see the efficiency of Jeroen on research. I would like to say it was great to work with him over the past five years.

I am very grateful to my committee members: Frederik Simons, Thomas Duffy, Allan Rubin, Blair Schoene and Jessica Irving. They helped me a lot during these five years. Especially before my general examine in the second year, I discussed my preliminary research projects with them. They were very nice and gave me valuable suggestions. I am also grateful to visiting professors Albert Tarantola and Michal Slawinski. They taught inverse theory, anisotropy and elasticity for a semester at Princeton. It was very great to attend their lectures and I believe they will be very helpful for my academic career in the future.

I would like to acknowledge my colleagues in seismology group: Yang Luo, Tarje Nissen-Meyer, Daniel Peter, Christina Morency, Ebru Bozdağ, Sharavan Hanasoge, Hom Nath Gharti, Ryan Modrak, Wenjie Lei, James Smith, Mattieu Lefebvre and Herurisa Rusmanugroho. It was very interesting to be exposed to a wide range of research topics in our Thursday group meetings, including tomography and migrations, normal mode, helioseismology, etc. It was great to discuss basic ideas, techniques details and see the progresses of each group member during these meetings.

I would also like to take this chance to acknowledge all previous members in our group: Dimitri Komatitsch, Qinya Liu, Carl Tape, Min Chen, Alessia Maggi, Vala Hjörleifsdóttir
Anne Sieminski... Without their important and fundamental works for the last decades on spectral-element method, adjoint method, FLEXWIN and many other tools, it is impossible for me to finish this thesis. Suggestions from Dimitri, Qinya, Carl and Min helped me a lot when I started this work three years ago.

It was great to have a lot of visitors in our group: Huub Douma, Irene Molinari, Federica Magnoni, Matthias Meschede, Rafael Abreu... They came from different universities and research institutes, and had quite different research backgrounds and interests. It was very interesting to discuss their research projects and see their progresses during the group meetings. I also appreciate helps from Irene for setting up her European crustal model: EPcrust, and upper mantle model: EPmantle.

I want to thank geophysics students in the third floor: Enning Wang, Yanhua Yuan, Jessica Hawthorne, Yajun Peng, Yue Tian, Yajing Liu, Pathikrit Bhattcharya, Max Werner. It was great to have Wednesday geophysics group meeting and I was happy to see progresses on research projects, such as earthquakes source, gravity, etc.

I give my best wish to our administrative assistants: Sheryl Robas, Mary Rose Russo, Theresa Autino, David Luet, Geogette Chalker, Mark Dalton and Nicole Leszczuk. They helped me a lot so that I am able to focus on my research during these years. Staffs at PICSciE helped me a lot to resolve issues about Sesame and Tiger clusters. Without their helps, it was impossible to finish a project requiring more than 10 million CPU hours.

I thank my former advisor at Peking University, Xiaofei Chen. He introduced me to seismology and helped me smoothly transfer from geology to geophysics. I also thank Wei Zhang at Peking University, who helped me a lot on numerical methods, especially finite different method. I appreciate discussions with my previous classmates at Peking University: Qiming Liu, Xiaolei Song, Ganpan Ke and Zhao Zheng.

I thank my family, my sister Hejia Zhu, my parents Zhongyuan Zhu and Shuying Shi and my beloved wife, Fujun Wei. It is their patient, consideration and love give me confidence and courage to pursue my Ph.D.
To my sister, parents and wife.
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Chapter 1

Introduction

Harnessing high-performance computers and accurate numerical methods to better constrain physical properties of Earth’s interior has become a very active research area in seismology. In this thesis, we apply adjoint method in both regional and exploration seismology. One of our targets is to improve three-dimensional image quality in elastic wavespeeds, anelastic attenuation, radial & azimuthal anisotropy and impedance contrasts. These results can be used to infer variations in temperature, water content, chemical composition as well as ancient and present deformation within Earth’s interior.

Chapter 2–Chapter 6 present adjoint tomography of the crust and upper mantle structures beneath Europe and the North Atlantic Ocean. In Chapter 2, we describe the dataset and inversion strategy used in this study. Distributions of earthquakes and seismographic stations, three-dimensional starting model and an initial source inversion are presented in this chapter. A three-stage inversion strategy is designed to iteratively determine elastic, anelastic, anisotropic heterogeneities in the crust and upper mantle beneath Europe and the North Atlantic Ocean. In Chapter 3, we present three-dimensional variations in elastic wavespeeds, anelastic attenuation and radial & azimuthal anisotropy in model EU60. These images are used to investigate tectonic activity, the distribution of water and the deformation state and history. Previous body- and surface-wave tomographic models are compared
with our model.

Chapter 4–Chapter 6 discuss elastic wavespeed structures in model EU\(_{30}\), anelastic attenuation in model EU\(_{50}\) and anisotropy in model EU\(_{60}\), respectively. In Chapter 4, we use the elastic wavespeed images in model EU\(_{30}\) to investigate detailed structures across the Iapetus and Tornquist Suture Zones. In Chapter 5, anelastic attenuation in model EU\(_{50}\) is discussed in detail. We observe enhanced attenuation within the mantle transition zone beneath the North Atlantic Ocean, which might imply a water-enriched transition zone beneath this region. In Chapter 6, we present azimuthally anisotropy in model EU\(_{60}\). These images are used to infer the tectonic deformation state and history beneath the European continent and the North Atlantic Ocean. In addition, radial peak-to-peak anisotropic strength profiles are used to identify brittle-ductile transitions in lithospheric strength beneath oceans and continents.

In Chapter 7, we apply adjoint methods in reverse-time and time-lapse migration. The two-dimensional SEG/EAGE salt dome model is used as an example. We find that sensitivity kernels with respect to impedance contrasts are much more suitable for seismic imaging compared to density kernels, which are equivalent to the conventional “imaging principle” in exploration seismology. We advocate the use of wavespeed kernels for iteratively updating volumetric wavespeed structures, while impedance kernels may be used to locate impedance contrasts within Earth’s interior.
Chapter 2

Seismic structure of the European crust and upper mantle based on adjoint tomography - I. Data set & inversion strategy

Note


2.1 Summary

We use adjoint tomography to iteratively determine seismic models of the crust and upper mantle beneath the European continent and the North Atlantic Ocean. Three-component seismograms from 190 earthquakes recorded by 745 seismographic stations are employed in the inversion. Crustal model EPcrust combined with mantle model S362ANI com-
prise the 3D starting model, EU\textsubscript{00}. Before the structural inversion, earthquake source parameters, e.g., centroid moment tensors and locations, are reinverted based on global 3D Green’s functions and Fréchet derivatives. This study consists of three stages. In stage one, frequency-dependent phase differences between observed and simulated seismograms are used to constrain radially anisotropic wavespeed variations. A new elastic model, named EU\textsubscript{30}, is constructed based on thirty preconditioned conjugate gradient iterations. In stage two, frequency-dependent phase and amplitude measurements are combined to simultaneously constrain elastic wavespeeds and anelastic attenuation. Elastic model EU\textsubscript{30} is chosen as the starting model for the second stage. Twenty additional iterations are performed to determine a new anelastic model, named EU\textsubscript{50}. In these two stages, long-period surface waves and short-period body waves are combined to simultaneously constrain shallow and deep structures. In stage three, frequency-dependent phase and amplitude anomalies of three-component surface waves are used to simultaneously constrain radial and azimuthal anisotropy. After ten additional iterations, we obtain a new anisotropic model, named EU\textsubscript{60}. Improvements in misfits and histograms in both phase and amplitude help us to validate this three-stage inversion strategy.

### 2.2 Introduction

Owing to advances in high-performance computing and numerical methods (e.g., the spectral-element method; Komatitsch & Vilotte, 1998; Komatitsch & Tromp, 1999), seismologists are now able to accurately simulate wave propagation in complex 3D Earth models, ranging from local (Komatitsch \textit{et al.}, 2004; Peter \textit{et al.}, 2011) to global scales (Komatitsch & Tromp, 2002a,b). Synthetic seismograms based on the spectral-element method (SEM) and the latest 3D Earth models, e.g., S362ANI (Kustowski \textit{et al.}, 2008a) or S40RTS (Ritsema \textit{et al.}, 2011), can match observed seismograms quite well at longer periods ( $> 60$ s) (Komatitsch \textit{et al.}, 2002; Tromp \textit{et al.}, 2010). However, at shorter periods there are remaining
differences between observed and simulated seismograms due to unmodeled source complexity and 3D heterogeneity. How to use these remaining differences to improve 3D Earth models is currently a very active research area in seismology (Chen et al., 2007; Tape et al., 2009, 2010; Fichtner et al., 2009, 2010; Lekic & Romanowicz, 2011; Zhu et al., 2012).

Adjoint methods, first introduced in seismology by Lailly (1983) and Tarantola (1984), enable us to numerically compute the gradient of a misfit function in complex 3D Earth models based on the interaction between two wavefields: one for a reference Earth model and a second obtained by using time-reversed differences between data and synthetics for the reference model at all receivers as simultaneous sources. This method is widely used to perform “full waveform inversion” in exploration seismology (Gauthier et al., 1986; Mora, 1987; Luo & Schuster, 1991; Pratt et al., 1998; Brossier et al., 2009; Virieux & Operto, 2009). Tromp et al. (2005) showed that the adjoint method is closely related to finite-frequency “banana-doughnut” theory, which has started to replace classical ray-based tomography in recent years (Marquering et al., 1998, 1999; Dahlen et al., 2000; Hung et al., 2000; Montelli et al., 2004). Liu & Tromp (2006, 2008) applied the SEM and adjoint methods to numerically compute 3D sensitivity kernels on both local and global scales. In combination with gradient-based optimization algorithms –e.g., a preconditioned conjugate gradient approach (Fletcher & Reeves, 1964) or the Broyden-Fletcher-Goldfarb-Shanno (BFGS) (Broyden, 1970; Fletcher, 1970; Goldfarb, 1970; Shanno, 1970) quasi-Newton algorithm, in particular its limited-memory version (L-BFGS) (Matthies & Strang, 1979; Nocedal, 1980)– adjoint methods can be employed to iteratively improve images of Earth’s interior by progressively minimizing discrepancies between observed and simulated seismograms (Akçelik et al., 2002, 2003; Tape et al., 2007). “Adjoint tomography”, a tomographic procedure based on the adjoint method, has been successively used to constrain crustal structure in southern California (Tape et al., 2009, 2010), as well as the upper mantle structure of Australia (Fichtner et al., 2009, 2010), Europe (Zhu et al., 2012; Fichtner et al., 2013b) and the North Atlantic (Rickers et al., 2013).
Seismic heterogeneities investigated in this study involve 3D variations in elasticity, anelasticity and anisotropy. Over the past several decades, most tomographic studies have focused on mapping lateral heterogeneities in (transversely) isotropic elastic wavespeeds based on traveltimes of body waves, dispersion of surface waves or splitting of free oscillations. However, the propagation of seismic waves is influenced by other physical properties besides isotropic elastic heterogeneity, such as attenuation and anisotropy. Anelastic attenuation leads to physical dispersion and dissipation, which affect both the phase and amplitude of seismic waveforms (Liu et al., 1976). In contrast to elastic wavespeed tomography, progress in attenuation tomography has been relatively slow (Dalton et al., 2008), and there are significant discrepancies between anelastic models determined by different groups (Romanowicz, 1995; Billien et al., 2000; Gung & Romanowicz, 2004; Lawrence & Wysession, 2006; Dalton et al., 2008; Wiens et al., 2008). Waveform amplitudes, which are usually employed to constrain attenuation, are relatively difficult to extract and are affected by a host of other factors besides intrinsic attenuation, such as earthquake magnitude, radiation pattern, elastic focusing and defocusing as well as scattering (Ruan & Zhou, 2010, 2012). Therefore, in order to investigate attenuation it is preferable to simultaneously invert for elastic wavespeeds and anelastic attenuation, using frequency-dependent phase and amplitude information.

Anisotropy is another important factor which affects the propagation of seismic waves. The constituent minerals of the crust and upper mantle, such as mica, amphibole and olivine, are highly anisotropic in terms of seismic wavespeeds. The fast “a” axes of these minerals are aligned with the directions of flow or principal extension depending on the state of deformation (Zhang & Karato, 1995). Thus, seismic wavespeeds vary with direction and polarization because of this lattice preferred orientation (LPO) (Ekström & Dziewonski, 1998; Simons et al., 2002; Gung et al., 2003; Marone & Romanowicz, 2007; Yuan & Romanowicz, 2010). Mapping these variations constrains the deformation state and history within Earth’s interior (Park & Levin, 2002). Shear-wave splitting measure-
ments have been widely employed to extract azimuthal anisotropy in terms of splitting times and fast propagation directions (Silver, 1996). However, owing to the relatively poor depth resolution, it is difficult to infer the vertical distribution of deformation and motion. This ambiguity has generated a long-lasting debate about the origin of seismic anisotropy within the uppermost mantle (Silver & Chan, 1991; Vinnik et al., 1992; Silver, 1996). Surface-wave tomography provides an important complementary tool—with better depth resolution—for mapping azimuthal anisotropy within the crust and upper mantle (Simons et al., 2002; Debayle et al., 2005; Marone & Romanowicz, 2007; Yuan & Romanowicz, 2010; Lin et al., 2011; Endrun et al., 2011).

The purpose of this study is to construct a reference seismic model for the crust and upper mantle beneath Europe and the North Atlantic by fully exploiting three-component seismic waveforms and utilizing modern numerical simulations. This is one of the main goals of the Initial Training Network in computational seismology named QUEST (quest-itn.org), funded within the 7th Framework People Programme by the European Commission. This paper mainly describes the data set and inversion strategy used in this study. A companion paper (Zhu et al., 2013c) (hereafter referred to as paper II) focuses on a discussion of the resulting 3D elastic, anelastic and anisotropic models. In Section 2.3, we discuss the distribution of earthquakes and seismographic stations used in the inversion. 3D starting model EU00 is presented in Section 2.4. An initial source inversion, aimed at correcting any biases in routine global CMT solutions, is discussed in Section 2.5. A three-stage inversion strategy for constraining elastic, anelastic and anisotropic heterogeneities is discussed in Section 2.6, Section 2.8 and Section 2.9, respectively. In Section 2.10, we discuss the behavior of various misfit functions and histograms of phase and amplitude anomalies.
2.3 Database

In this study, 190 earthquakes are used to illuminate the region of interest. These earthquakes are evenly distributed along the North Atlantic Ridge and the Mediterranean-Himalayan Belt (Figure 2.1a). Most occurred between 1996 and 2011, with magnitudes ranging from 4.5 to 6.5 (Figures 2.1e and b). The majority of events are shallower than 30 km (Figure 2.1c). Initial source parameters, e.g., origin times, locations and moment tensor solutions, are collected from the global Centroid-Moment-Tensor (CMT) website (globalcmt.org). Before the structural inversion, we perform source inversions for all 190 earthquakes (Section 2.5). For this purpose, good azimuthal coverage is crucial. Three-component seismic waveforms recorded by 239 global seismographic stations (from networks operated by IRIS/IDA, IRIS/USGS, GEOFON and GEOSCOPE) are collected from the Incorporated Research Institutions for Seismology (iris.edu).

In the structural inversion, seismic waveforms recorded by 338 seismographic stations of 40 European networks are collected from the Observatories and Research Facilities for European Seismology (orfeus-eu.org). In addition, several IRIS/PASSCAL arrays are included in the database. For example, the HOTSPOT array (Foulger et al., 2001; Allen et al., 2002a,b) is incorporated in order to illuminate subsurface structures beneath Iceland and the North Atlantic Ridge. We also assimilate some seismographic stations from the Kandilli Observatory (koeri.boun.edu.tr) in order to better constrain structures beneath the Anatolian Plate. In total, three-component seismic waveforms recorded by 745 seismographic stations are collected (Figure 2.2a). The instrument response is removed from the raw seismic data to obtain band-passed ground displacement, which is subsequently rotated into vertical, radial and transverse components.

Ray coverage of our dataset is very good for the European continent and the North Atlantic Ocean, especially for the Alpine-Mediterranean area. However, coverage of northern Africa and the western Urals is relatively poor (Figure 2.2b). Epicentral distances in our database range from $0^\circ$ to $70^\circ$ (Figures 2.3b and c). P wave raypaths for a typical shallow
Figure 2.1: 190 earthquakes used in this study. (a) Distribution of earthquakes. The blue quadrilateral denotes the SEM simulation region. Colors of beach balls indicate earthquake depths. (b)–(e) Histograms of earthquake moment magnitudes, depths, half durations and origin times.
earthquake in PREM (Dziewonski & Anderson, 1981) are shown in Figure 2.3a. When the epicentral distance is greater than $20^\circ$, seismic rays turn below the transition zone and have steep incidence angles, providing good resolution of the entire upper mantle in the region of interest.

2.4 3D starting model $\mathbf{E} \mathbf{U}_{00}$

Modern numerical techniques in combination with high-performance computing enable us to accurately and effectively calculate synthetic seismograms and misfit gradients in 3D heterogeneous Earth models. Therefore, instead of performing tomographic inversion in 1D spherical symmetric Earth models based on approximate, asymptotic methods, we are able to use a 3D model as the starting model, and iteratively improve it using gradient-based optimization techniques. In this study, EPcrust and S362ANI are chosen as initial crust and mantle models, respectively, which are combined to form starting model $\mathbf{E} \mathbf{U}_{00}$. 
Figure 2.3: Raypaths for a typical shallow earthquake. (a) Raypaths of P waves with epicentral distances ranging from $11^\circ$ to $60^\circ$ in PREM (Dziewonski & Anderson, 1981). (b) Propagation distances for event 201101290655A. Seismographic stations which recorded this event are labeled by red triangles. (c) Same as (b) for event 201005251009A.
2.4.1 3D crustal model: EPcrust

The crust is a highly heterogeneous region of the Earth, which can strongly affect seismic wave propagation. In tomography, mantle images can be severely distorted if 3D crustal structure is not taken into account properly (Waldhauser et al., 2002). Previous tomographic studies have frequently relied on “crustal corrections” to remove crustal effects before imaging mantle structure (Lekic et al., 2010; Panning et al., 2010). Since SEM and adjoint methods enable us to incorporate 3D crustal models in both forward and gradient calculations, no additional crustal corrections are required in our inversion. In addition, crust and upper mantle structures are updated simultaneously according to the behavior of 3D sensitivity kernels, which helps us to reduce tradeoffs between crust and mantle heterogeneity. There are a variety of crustal models for the European continent, ranging from global to local scales (Bassin et al., 2000; Tesauro et al., 2008; Molinari & Morelli, 2011). Considering our simulation region, we choose EPcrust (Molinari & Morelli, 2011) as our starting crustal model. EPcrust is a 3D crustal model with $0.5^\circ \times 0.5^\circ$ resolution, covering the entire European continent as well as Greenland and the North Atlantic Ocean. EPcrust is described in terms of three layers: sediments, upper and lower crust. It provides the following seismic parameters: the Moho, sedimentary basement, mass density, as well as isotropic compressional and shear wavespeeds. Since EPcrust incorporates several local-scale European crustal studies, in some areas it involves much more refined structures than Crust2.0, a global crustal model with $2^\circ \times 2^\circ$ resolution (Bassin et al., 2000).

2.4.2 SEM mesh

The Moho is one of the most important first-order discontinuities within Earth’s interior. From a numerical perspective, for weak-form implementations of the equation of motion—employed in the SEM— it is very important to design a spectral-element mesh which honors such a first-order discontinuity (Komatitsch & Vilotte, 1998; Komatitsch & Tromp, 1999). In our study region, Moho depth varies from 7–10 km beneath the North Atlantic
Figure 2.4: Spectral-element mesh for crustal model EPcrust (Molinari & Morelli, 2011). (a) Moho depths of EPcrust. (b) Histogram of Moho depths. (c) Areas where the Moho is honored by the spectral-element mesh. Red regions indicate the crust is captured by two or three layers of spectral elements. Blue regions indicate the crust is captured by one layer of spectral elements. White regions indicate the Moho is not honored by the mesh. (d) Spectral-element mesh and isotropic shear wavespeed perturbations in the starting model along cross section A–A’ in (c).
Ocean to greater than 50 km beneath the East European Craton (Figure 2.4a). It is challenging to honor such a dramatically varying discontinuity based on a hexahedral mesh. Following Tromp et al. (2010), the spectral-element mesh is stretched to honor the Moho discontinuity when it is shallower than 15 km and between 25 to 45 km, dominating the distribution of Moho depth variations in the area of interest (Figure 2.4b). The goal of this stretching is to employ one layer of spectral-elements to capture oceanic crust and two or three layers of spectral-elements to represent continental crust. The stretched mesh provides an adequate grid sampling for the entire crust, as illustrated in Figure 2.4d in vertical cross section A–A’. In Figure 2.4c, the areas in which the Moho is honored are colored blue (ocean) and red (continent), whereas white regions denote areas where the Moho runs through spectral elements.

The simulation domain has lateral dimensions of $65^\circ \times 65^\circ$, ranging from northern Africa to the North Pole, and from Greenland to the Urals (Figure 2.1a). One “cubed sphere” chunk of a global mesh (Komatitsch & Tromp, 2002a) is used in this study. The total number of elements is 4,692,600, and the horizontal element size is approximately 42 km on the free surface, resulting in an average spacing of $\sim 10$ km between Gauss-Lobatto-Legendre interpolation points. The minimum period resolved by a forward calculation is $\sim 12$ s. Using 100 cores on a Dell Intel Nehalem PC cluster, it takes approximately 1 hour for a forward simulation (a 30 minutes record) and 2 hours for a gradient calculation.

### 2.4.3 3D mantle model: S362ANI

S362ANI (Kustowski et al., 2008a), a global radially anisotropic shear wavespeed model, is chosen as the 3D starting mantle model. It provides six model parameters: mass density ($\rho$), two compressional wavespeeds ($\alpha_v$ and $\alpha_h$), two shear wavespeeds ($\beta_v$ and $\beta_h$), and a dimensionless parameter ($\eta$). Instead of describing 3D perturbations with respect to PREM (Dziewonski & Anderson, 1981) or IASP91 (Kennett & Engdahl, 1991), S362ANI has its own 1D reference model, namely STW105 (Figure 2.5a). In contrast to PREM,
Figure 2.5: Shear wavespeeds and attenuation for starting mantle model S362ANI (Kustowski et al., 2008a). (a) 1D wavespeed model STW105. Black and red lines denote speeds of horizontally traveling and vertically ($\beta_v$) and horizontally ($\beta_h$) polarized shear waves, respectively. (b) 1D shear attenuation $Q$ model. (c) & (d) Horizontal cross sections of relative perturbations in $\beta_v$ and $\beta_h$ at 75 km depth in model S362ANI.
radial anisotropy in STW105 extends to depths in excess of 300 km. Considering the complexity of the area of interest, in our inversion radial anisotropy is allowed from the bottom of the crust to the bottom of the transition zone, i.e., to depths of 660 km. The 1D $Q$ model is shown in Figure 2.5b, which involves a fairly strongly attenuating asthenosphere between 80 km and 220 km. This 1D attenuation model is fixed in the elastic and anisotropic inversions (Sections 2.6 and 2.9), and iteratively updated in the anelastic inversion (Section 2.8). Figures 2.5c and d show relative perturbations in $\beta_v$ and $\beta_h$ from S362ANI at a depth of 75 km. Since S362ANI is a global model, it only describes large-scale lateral variations, such as the old and cold East European Craton as well as the relatively young Mediterranean Sea and Anatolian Plate.

2.5 Source inversion

Hjörleifsdóttir & Ekström (2010) estimated uncertainties associated with routine global CMT solutions. One of their conclusions is that the CMT procedure tends to locate sources deeper than their actual depths; usually this depth bias ranges from 5 km to 8 km. They attributed this bias mainly to 3D crustal heterogeneity, which is only approximately considered in the CMT algorithm. This systematic depth bias necessitates a source parameters inversion prior to the structural inversion.

We use the source inversion algorithm of Liu et al. (2004). Targeted least-squares waveform differences between observed, $d$, and simulated, $s$, seismograms are used to define the non-dimensional misfit function

$$\chi = \sum_{c=1}^{N_c} w_c \sum_{m=1}^{N_m} w_m \frac{\int [d_m(t) - s_m(t - \Delta t, m_0)]^2 dt}{\int [d_m(t)]^2 dt},$$

(2.1)

where $N_c$ denotes the number of categories that define the misfit. In this study, three-component body waves with periods between 30 s and 80 s and three-component surface waves with periods between 80 s and 120 s are combined to constrain the source param-
Figure 2.6: Depth changes for the 190 earthquakes used in this study after source inversion. (a) Map view of the depth changes for all events. (b) Distribution of global seismographic stations employed in the source inversion (see Section 2.3). (c) Depth comparison between global CMT solutions and reinverted sources; only events with depths shallower than 50 km are shown. (d) Histogram of depth changes, indicating a general shallowing compared to CMT depths.

eters, i.e., $N_c = 6$. The quantity $w_c$ represents a weighting term associated with each category, which is equal to the reciprocal of the number of measurements in each category, i.e., $N_m$. The quantity $w_m$ refers to a weighting factor for each measurement $m$, which is related to azimuth, distance and component (for details see Liu et al., 2004). Cross-correlation traveltime differences between observed and synthetic seismograms, $\Delta t$, are used to correct simulated seismograms. These traveltime anomalies are attributed to unmodelled lateral heterogeneities, which are the target of the structural inversion. The initial source model, $m_0$, is the global CMT solution. Waveform differences are weighted by the integrated data power within each measurement window.
The SEM is used to calculate synthetic seismograms and Fréchet derivatives in the 3D global model. Source parameters may include latitude, longitude, depth and the six components of the moment tensor. Fréchet derivatives with respect to these source parameters are calculated based on a finite-difference approximation. In order to obtain good azimuthal coverage, we use global datasets (Section 2.3 and Figure 2.6b) and global simulations based on crustal model Crust2.0 and mantle model S362ANI. Thus, the source inversion for each event requires ten global forward calculations and each simulation takes approximately 5 hours on 150 cores for a 100 minutes record.

The source inversion procedure routinely provides results for three combinations of model parameters: moment tensor only, moment tensor plus depth, and moment tensor plus depth, longitude and latitude. In addition, each combination involves either zero-trace or double-couple constraints. Therefore, in total six scenarios are considered in the source inversion procedure. When depth, longitude and latitude are included in the inversion, the problem becomes nonlinear and should be solved iteratively (Liu et al., 2004). In this study, source parameters are updated only once because the changes are small compared to structural effects on phase and amplitude, and the computational cost is high. However, between the elastic and anelastic inversions of stages I & II, we perform a grid search for the scalar moment and centroid time of each earthquake to limit structural tradeoffs with the source. Variance reduction is calculated for each scenario by comparing waveform differences between data and synthetics with original and updated source parameters. In principle, the solution with the largest variance reduction is chosen as the new source model. In some specific areas, e.g., along the North Atlantic Ridge, the double-couple constraint is always preferred.

After carefully selecting the best result from the six scenarios, we obtain a new source model for each event. Figure 2.6 presents depth differences between updated source models and original global CMT solutions. Most new depths are shallower than the global CMT solutions, and the average depth difference is approximately 3–8 km (Figures 2.6c and d),
in agreement with the conclusions of Hjörleifsdóttir & Ekström (2010). In Figure 2.6a, new
depths of earthquakes beneath the North Atlantic Ridge are 3–5 km shallower than global
CMT solutions, which is attributed to a bias due a uniform PREM crust (which is too thick
underneath the oceans) used in the global CMT algorithm.

In the following three sections, we discuss inversion strategies for determining 3D vari-
ations in elastic (Section 2.6), anelastic (Section 2.8) and anisotropic (Section 2.9) hetero-
geneities.

2.6 Stage I: elastic inversion

As discussed in the Introduction, our goal is to constrain elastic, anelastic and anisotropic
heterogeneities in a three-stage inversion. In this section we discuss the Stage I elastic
inversion.

2.6.1 Misfit function

In Stage I, we only use phase differences between observed and simulated seismograms to
constrain elastic wavespeeds. The total misfit at this stage is

\[ \chi_I = \chi_\phi, \quad (2.2) \]

where \( \chi_\phi \) refers to the phase misfit. Three-component short-period body waves and three-
component long-period surface waves are combined to simultaneously constrain deep and
shallow structures. Therefore, the total phase misfit consists of six categories: P-SV
body waves on vertical and radial components, SH body waves on transverse components,
Rayleigh waves on vertical and radial components and Love waves on transverse compo-
nents. FLEXWIN (Maggi et al., 2009), an automated time window selection tool, is used
to select suitable measurement windows. A multitaper technique (Laske & Masters, 1996b;
Zhou et al., 2004) is used to measure frequency-dependent phase differences between data and synthetics. Thus, the phase misfit may be expressed as

\[ \chi_\phi = \sum_{c=1}^{N_c} w_c \sum_{m=1}^{N_m} \int w_m \left[ \frac{\Delta \tau_m(\omega)}{\sigma_m(\omega)} \right]^2 d\omega , \]  

(2.3)

where \( N_c \) denotes the number of categories, i.e., \( N_c = 6 \), \( w_c \) is a weighting term associated with each category, and \( \omega \) denotes angular frequency. Since the misfit values in each category are balanced in this study, we set \( w_c \) to the reciprocal of the number of measurements in each category, i.e., \( N_m \). The quantity \( w_m \) is a weighting term associated with each measurement. The quantities \( \Delta \tau_m(\omega) \) and \( \sigma_m(\omega) \) are angular frequency-dependent phase differences and associated uncertainties for multitaper measurement \( m \).

### 2.6.2 Model parameters

General anisotropic materials are described by a fourth-order elastic tensor \( c_{ijkl} \), which involves 21 independent elements. For an anisotropic material with a radial symmetric axis, the number of independent model parameters is reduced to five: \( A, C, L, N \) and \( F \), the so-called Love parameters (Love, 1927). In seismic tomography, we prefer to use wavespeeds as model parameters rather than the Love parameters, because traveltime measurements are more sensitive to wavespeeds. Therefore, radially anisotropic Earth models, such as PREM, are usually described in terms of six parameters, as explained in Section 2.4.3: \( \rho, \alpha_v, \alpha_h, \beta_v, \beta_h \) and \( \eta \). The relationships between wavespeeds and Love parameters are given in eqn. (2.4)
\[ \alpha_h = \sqrt{\frac{A}{\rho}} , \]
\[ \alpha_v = \sqrt{\frac{C}{\rho}} , \]
\[ \beta_h = \sqrt{\frac{N}{\rho}} , \]
\[ \beta_v = \sqrt{\frac{L}{\rho}} , \]
\[ \eta = \frac{F}{A - 2L} . \] (2.4)

In our inversion, we assume that the bulk modulus, \( \kappa \), remains isotropic, and that radial anisotropy is solely due to shear anisotropy, i.e., the parameters \( L \) and \( N \). In this case, we may use the isotropic bulk sound wavespeed, \( c = \sqrt{\kappa/\rho} \), together with the two shear wavespeeds, \( \beta_v = \sqrt{L/\rho} \) and \( \beta_h = \sqrt{N/\rho} \). Thus, rather than considering four wavespeeds (\( \alpha_v, \alpha_h, \beta_v \) and \( \beta_h \)), we consider just three (\( c, \beta_v \) and \( \beta_h \)), such that \( \alpha_v^2 = c^2 + \beta_v^2 \) and \( \alpha_h^2 = c^2 + \beta_h^2 \).

In general, traveltimes of seismic waves are much more sensitive to wavespeeds than mass density, \( \rho \), which is usually constrained by free oscillation and the Earth’s moments of inertia. In this study, an empirical relationship between relative perturbation in mass density and isotropic shear wavespeed is used to update mass density (Montagner & Anderson, 1989), namely

\[ \delta \ln \rho = 0.33 \delta \ln \beta , \] (2.5)

where \( \beta \) refers to the Voigt average of the radially anisotropic shear wavespeeds (Babuska & Cara, 1991)

\[ \beta = \sqrt{\frac{2\beta_v^2 + \beta_h^2}{3}} . \] (2.6)

In summary, in the elastic inversion we consider four model parameters: the isotropic
bulk sound wavespeed \( (c) \), the waves speeds of horizontally traveling and vertically and horizontally polarized shear waves \( (\beta_v \text{ and } \beta_h) \), and the dimensionless radial anisotropic parameter \( (\eta) \). Perturbations in total misfit are expressed as a volume integral over relative perturbations in these four model parameters:

\[
\delta \chi_I = \int_V K_c \delta \ln c + K_{\beta_v} \delta \ln \beta_v + K_{\beta_h} \delta \ln \beta_h + K_\eta \delta \ln \eta \, dV ,
\]  

(2.7)

where \( V \) denotes the Earth’s volume. The quantities \( K_c, K_{\beta v}, K_{\beta h} \text{ and } K_\eta \) are sensitivity kernels with respect to relative perturbations in the radially anisotropic model parameters. These four kernels can be derived from primary kernels based on relationships given in eqn. (2.9)

\[
\begin{align*}
K_{\alpha h} &= 2 A K_A + 2 A \eta K_F , \\
K_{\alpha v} &= 2 C K_C , \\
K_{\beta h} &= 2 N K_N , \\
K_{\beta v} &= 2 L K_L - 4 L \eta K_F , \\
K_\eta &= F K_F .
\end{align*}
\]  

(2.8)

If the isotropic bulk sound wavespeed, \( c \), is chosen as a model parameter instead of compressional wavespeed, the relationships (2.8) may be rewritten as

\[
\begin{align*}
K_c &= \frac{c^2}{\alpha_h^2} K_{\alpha h} + \frac{c^2}{\alpha_v^2} K_{\alpha v} , \\
K_{\beta h}' &= K_{\beta h} + \frac{4}{3} \frac{\beta_h^2}{\alpha_h^2} K_{\alpha h} , \\
K_{\beta v}' &= K_{\beta v} + \frac{4}{3} \frac{\beta_v^2}{\alpha_v^2} K_{\alpha v} , \\
K_\eta &= F K_F .
\end{align*}
\]  

(2.9)
2.6.3 Radially anisotropy kernels

In order to investigate the behavior of sensitivity kernels for the radially anisotropic shear wavespeeds $\beta_v$ and $\beta_h$, we perform adjoint simulations for two scenarios: one involving Rayleigh waves and the other involving Love waves. Figures 2.7c and d show sensitivity kernels for $\beta_v$ and $\beta_h$ based on Rayleigh waves measured on vertical and radial components (Figure 2.7b). The magnitude of the sensitivity kernel $K_{\beta_v}$ is much larger than $K_{\beta_h}$. This behavior is reversed when only transverse-component Love waves are back-projected (Figures 2.7e and f). This is because Rayleigh waves are more sensitive to the speed of vertically polarized and horizontally traveling shear waves, i.e., $\beta_v$, while Love waves are more sensitive to the speed of horizontally polarized and horizontally traveling shear waves, i.e., $\beta_h$. The complementary behavior of the $\beta_v$ and $\beta_h$ sensitivity kernels allows us to map radial anisotropy within the upper mantle and resolve the well-known “Rayleigh/Love discrepancy” in surface-wave tomography (Shapiro et al., 2004; Moschetti et al., 2010).

2.6.4 Frequency-band selection

In this study, we use a simple multiscale strategy to reduce the nonlinearity of our problem. In the first several iterations, we start by fitting long-period signals, e.g., body waves with periods between 15 s and 50 s and surface waves with periods between 50 s and 150 s. As the model improves and the misfit between data and synthetics diminishes, we gradually decrease the short-period corner of the bandpass filter. For instance, the short-period corner of long-period surface-wave measurements is gradually reduced from 50 s to 25 s. This strategy allows us to first resolve large-scale features based on long-period signals, and gradually map smaller-scale features based on shorter-period data. This multi-frequency procedure is also used in the anelastic (Section 2.8) and anisotropic inversions (Section 2.9).
Figure 2.7: Sensitivity kernels for radially anisotropic model parameters based on measurements of Rayleigh and Love waves. (a) Locations of event (200907010930A) and station (OUL.FN). (b) Comparisons between three-component data (black) and synthetics (red). From top to bottom are shown the vertical, radial and transverse components. Blue rectangles denote windows selected by FLEXWIN (Maggi et al., 2009). (c) & (d) Sensitivity kernels for $\beta_v$ (c) and $\beta_h$ (d) at 75 km depth based on measurements on the vertical and radial components in (b), i.e., Rayleigh waves. (e) & (f) Sensitivity kernels for $\beta_v$ (e) and $\beta_h$ (f) based on measurements on the transverse component in (b), i.e., Love waves.
2.6.5 Kernel preconditioning

For an inverse problem, if the Hessian—that is, the second derivative of the misfit function—is available, model updates are equal to the dot product between the (generalized) inverse of the Hessian and the negative gradient (Tarantola, 2005; Tape et al., 2007), as dictated by the Newton method. However, it is usually prohibitively expensive to compute and store the full Hessian (Chen et al., 2007). Therefore, a variety of preconditioners has been designed to approximate the Hessian or its diagonal terms (Pratt et al., 1998). In adjoint tomography, preconditioning is an important procedure since it is infeasible to access the Hessian, which would involve the very expensive calculation of “banana-doughnut” kernels for every single measurement at every iteration. Here, we employ a preconditioner which involves the vector dot product and convolution of the forward and adjoint accelerations:

\[ P(x) = \frac{1}{\int \partial_t^2 s(x,t) \cdot \partial_t^2 s^\dagger(x,T-t) \, dt} , \quad (2.10) \]

where \( s \) and \( s^\dagger \) are the forward and adjoint displacement wavefields, respectively. Both are readily accessible during an adjoint calculation, and therefore there are no additional costs associated with the calculation of this preconditioner. It has a small magnitude in the vicinity of sources and receivers, allowing us to reduce relatively large magnitudes of sensitivity kernels near earthquakes and stations. In order to avoid division by a very small value, a water level is applied in the denominator of (2.10). Event kernels for individual earthquake are first summed to construct misfit kernels, which are then multiplied by the preconditioner (2.10) to obtain the preconditioned misfit gradient.

2.6.6 Kernel smoothing

Regularization is generally used to stabilize an inverse problem, especially if it is ill-posed (Aster et al., 2005). In adjoint tomography, we do not explicitly incorporate a regularization term in the misfit function (Section 2.6.1). Instead, a 3D Gaussian function
is used to smooth preconditioned misfit gradients, which may be regarded as a regularization procedure (Tape et al., 2010). In classical global or regional tomography, seismic models are often parameterized by a spline function in the radial direction and spherical harmonics in the lateral directions. In this study, a 3D Gaussian with different half widths in the radial and azimuthal directions is employed to smooth the preconditioned gradients. For every Gauss-Lobatto-Legendre (GLL) point \( r \), the kernels may be smoothed based on

\[
K(r) = \frac{1}{W(r)} \int_V K(r') \exp\left[-\left(\frac{\Delta^2}{2\sigma_\Delta^2}\right)\right] \exp\left[-\left(\frac{(r-r')^2}{2\sigma_r^2}\right)\right] \, \mathrm{d}^3r',
\]

(2.11)

where \( \Delta \) denotes the azimuthal distance between points \( r \) and \( r' \), \( r = ||r||, r' = ||r'|| \), and where \( W(r) \) is the normalization factor

\[
W(r) = \int_V \exp\left[-\left(\frac{\Delta^2}{2\sigma_\Delta^2}\right)\right] \exp\left[-\left(\frac{(r-r')^2}{2\sigma_r^2}\right)\right] \, \mathrm{d}^3r'.
\]

(2.12)

The quantities \( \sigma_\Delta \) and \( \sigma_r \) are the half widths of the Gaussian function in the azimuthal and radial directions, respectively. They are chosen based on the wavelengths of structure which can be resolved by the current frequency bands. These two widths are reduced as the model improves. For instance, \( \sigma_\Delta \) is gradually decreased from 100 km to 50 km, reflecting the incorporation of shorter-period surface-wave measurements. Figure 2.8 illustrates sensitivity kernels for \( \beta_v \) and \( \beta_h \) before (a and b) and after (c and d) preconditioning and smoothing. Based on these two operations, we remove small-scale features which cannot be resolved by the current data, especially in the vicinity of sources and receivers.

### 2.6.7 Conjugate gradient method

Gradient-based optimization approaches are used to iteratively update model parameters based on preconditioned and smoothed misfit gradients (Tromp et al., 2005; Tape et al., 2007). In the first iteration, the steepest descent method is used to update the model parameters, i.e., search directions are equal to negative misfit gradients. In subsequent
Figure 2.8: Comparisons of sensitivity kernels before and after preconditioning and smoothing. (a) & (b) Sensitivity kernels for $\beta_v$ (a) and $\beta_h$ (b) after summing all 190 event kernels. (c) & (d) Sensitivity kernels for $\beta_v$ (c) and $\beta_h$ (d) after preconditioning and smoothing.
iterations, a conjugate gradient method is used to compute search directions (Fletcher & Reeves, 1964), which are equal to combinations of current misfit gradients and previous search directions:
\[ d_i = -g_i + \beta d_{i-1}, \]
where
\[ \beta = \frac{g_i^T \cdot (g_i - g_{i-1})}{g_{i-1}^T \cdot g_{i-1}}. \] (2.13)

The search direction is denoted by \(d\), and \(g\) refers to the misfit gradient. There are alternative formulas to compute \(\beta\) (Tarantola, 2005). In this study, the Polak-Ribi`ere formula is employed (Tromp et al., 2005; Tape et al., 2007, 2010). One advantage of this formula is that \(\beta\) can be reset to zero when it is negative. Thus, the algorithm is able to automatically forget previous search directions and restart as a new steepest descent search.

We have also experimented with quasi-Newton optimization methods, specifically the limited-memory version of the Broyden-Fletcher-Goldfarb-Shanno algorithm (L-BFGS) (Matthies & Strang, 1979; Nocedal, 1980). These methods give comparable results and convergence rates (Luo et al., 2013).

### 2.6.8 Line search and model update

A model update is obtained based on the expression
\[ \ln \frac{m_{i+1}}{m_i} = \alpha d_i, \] (2.14)
where \(\alpha\) denotes a step length in the \(i\)th search direction \(d_i\). Tape et al. (2007) used quadratic and cubic interpolations to determine the optimal step length along the search direction. In this study, using a representative subset of earthquakes, we generate several test models by choosing different values of \(\alpha\), compute synthetic seismograms and evaluate the misfit function for each test model. The model with the minimum misfit value is selected.
Figure 2.9: Line search to determine the step length in the first iteration. (a) Distribution of 16 earthquakes used in the line search. Yellow triangles denote seismographic stations. (b) Evolution of total misfit for the five test models with different $\alpha$ values ranging from 0.01 to 0.05. (c)–(e) Evolution of misfit for P-SV body waves on vertical (c) and radial (d) components, and SH body waves on transverse components (e). (f)–(h) Evolution of misfit for Rayleigh waves on vertical (f) and radial (g) components, and Love waves on transverse components (h). A 3% update is chosen in the first iteration.
Figure 2.10: Adjoint tomography workflow for the Stage I elastic inversion. Computational requirements for the source inversion as well as for forward and adjoint calculations are indicated.

as the new model. We usually test five $\alpha$ values ranging from 0.01 to 0.05, corresponding to maximum 1–5% model perturbations along the search direction. The behavior of the total misfit, as well as its behavior in the six subcategories (discussed in Section 2.6.1), are monitored to determine the best test model. Figure 2.9 shows how we select the value of $\alpha$ in the first iteration based on this strategy. Since it is very expensive to perform forward calculations for all 190 earthquakes, we use a representative subset of 16 earthquakes (Figure 2.9a). Therefore, for each line search, we perform another $16 \times 5$ forward calculations.

The inversion strategy and computational costs of adjoint tomography for elastic iterations are summarized in Figure 2.10. Intensive computations are required for the source inversions, forward and adjoint calculations. In comparison, pre- and post-processing re-
quires a limited amount of storage and computation. A new elastic wavespeed model for the crust and upper mantle beneath Europe and the North Atlantic, named EU$_{30}$, is constructed based on 30 preconditioned conjugate gradient iterations, which required more than 17,100 wavefields simulations and 2.3 million central processing unit hours (Zhu et al., 2012).

2.7 **Source correction**

Before the Stage II anelastic inversion, we perform a grid search for the scalar moment, $M_0$, and centroid time, $t_0$, of each earthquake to minimize tradeoffs between structure and source. For every event, we define a two-dimensional misfit function of corrections in origin time, $\Delta t_0$, and relative scalar moment, $\Delta \ln M_0$. Phase and amplitude anomalies between observed seismograms, $d(t)$, and corrected synthetic seismograms, $(1+\Delta \ln M_0) s(t+\Delta t_0)$, are combined in the misfit function. A simple grid search is used to determine a pair of corrections which minimizes the misfit. The search dimensions are $-5 \text{s} < \Delta t_0 < 5 \text{s}$ and $-0.8 \text{<} \Delta \ln M_0 \text{<} 0.8$. An example of the source correction for event 200905241617A is shown in Figure 2.11. The mean values of both phase and amplitude histograms are moderately improved after the source corrections. However, there are no significant improvements in their standard deviations, which must be reduced by considering 3D anelastic heterogeneity.

2.8 **Stage II: anelastic inversion**

After 30 iterations, we have significantly reduced the phase anomalies between observed and simulated seismograms. However, there are no obvious improvements in the amplitude anomalies (see Section 2.10.2). Anelastic attenuation is an important factor affecting the amplitudes of waveforms. Therefore, at this stage we combine amplitude anomalies with remaining phase differences to simultaneously constrain elastic and anelastic heterogeneities. Anelasticity involves physical dispersion and attenuation, and therefore we
Figure 2.11: Example of a source correction for event 200905241617A. (a) Locations of event (indicated by the beach ball) and stations (indicated by yellow triangles) used in the source correction. (b) 2D misfit function as a function of centroid time, $\Delta t_0$, and relative scalar moment, $\Delta \ln M_0$. The black star denotes the pair of parameters used to correct the source. (c) & (d) Histograms of phase anomalies (c) and amplitude anomalies (d) before (red) and after (blue) the source correction.
fit phase and amplitude anomalies simultaneously to limit tradeoff between elasticity and anelasticity. Model EU$_{30}$ from the previous elastic inversion is chosen as the starting model for the current stage.

The same inversion strategy as described in Section 2.6 and Figure 2.10 is used to simultaneously constrain elastic and anelastic heterogeneity. Twenty additional preconditioned conjugate gradient iterations are performed to construct a new anelastic model of Europe and the North Atlantic, namely EU$_{50}$, which required more than 18,050 wavefields simulations and 2.5 million central processing unit hours.

### 2.8.1 Misfit function

The total misfit function of Stage II is defined as

\[
\chi_{\text{II}} = w_\phi \chi^\phi + w_A \chi^A ,
\]

(2.15)

where \(\chi^\phi\) and \(\chi^A\) refer to phase and amplitude misfits, and \(w_\phi\) and \(w_A\) are weighting factors associated with these two misfits, which are chosen to balance the relative contributions of phase and amplitude.

As in Section 2.6.1, time windows are selected with FLEXWIN (Maggi et al., 2009), and in these windows multitaper phase and amplitude anomaly measurements are made (Laske & Masters, 1996b; Zhou et al., 2004). Phase and amplitude misfits \(\chi^\phi\) and \(\chi^A\) are calculated based on the expressions

\[
\chi^\phi = \sum_{c=1}^{N_c} w_c \sum_{m=1}^{N_m} \int w_m \left( \frac{\Delta \tau_m(\omega)}{\sigma^\phi_m(\omega)} \right)^2 d\omega ,
\]

(2.16)

\[
\chi^A = \sum_{c=1}^{N_c} w_c \sum_{m=1}^{N_m} \int w_m \left( \frac{\Delta \ln A_m(\omega)}{\sigma^A_m(\omega)} \right)^2 d\omega ,
\]

(2.17)

where \(\Delta \ln A_m(\omega)\) and \(\sigma^A_m(\omega)\) are frequency-dependent amplitude differences and associ-
ated uncertainties for multitaper measurement $m$. The other parameters are the same as defined in Section 2.6.1. Both short-period body waves and long-period surface waves in three-component seismograms are combined in the misfit functions to simultaneously constrain deep and shallow structures. Thus, the number of categories $N_c$ in eqns. (2.16) and (2.17) equals six, i.e., P-SV body waves on vertical and radial components, SH body waves on transverse components, Rayleigh waves on vertical and radial components and Love waves on transverse components.

Based on the definition of the total misfit function in eqn. (2.15), the corresponding adjoint sources are

$$f^\dagger(x, t) = f^\dagger_\phi(x, t) + f^\dagger_A(x, t),$$

where $f^\dagger_\phi(x, t)$ represents the adjoint source for the phase misfit defined in eqn. (2.16) and $f^\dagger_A(x, t)$ denotes the contribution related to the amplitude misfit defined in eqn. (2.17).

### 2.8.2 Model parameters

The model parameters at the current stage include elastic and anelastic contributions. For the elastic part, we use the same radially anisotropic model parameters as defined in Section 2.6.2, including the wavespeeds of horizontally traveling and vertically and horizontally polarized shear waves $\beta_v$ and $\beta_h$, the isotropic bulk sound wavespeed $c$, and the dimensionless parameter $\eta$. As before, mass density is scaled to the isotropic shear wavespeed via eqn. (2.5).

The inverse quality factor $Q^{-1}$ is used to quantify anelasticity. Since the magnitude of bulk attenuation $Q_\kappa^{-1}$ is usually negligible compared to shear attenuation $Q_\mu^{-1}$ (Dalton et al., 2008), only shear attenuation is considered in this study. For brevity, in the following expressions, we use $Q^{-1}$ rather than $Q_\mu^{-1}$ to denote the inverse shear quality factor.

Thus, the five model parameters considered in this stage are $c$, $\beta_v$, $\beta_h$, $\eta$ and $Q^{-1}$. The
Stage II misfit variation $\delta \chi_{II}$ may be expressed as

$$\delta \chi_{II} = \int_V K_c \delta \ln c + K_{\beta_v} \delta \ln \beta_v + K_{\beta_h} \delta \ln \beta_h + K_\eta \delta \ln \eta + K_{Q^{-1}} \delta Q^{-1} \, dV , \quad (2.19)$$

where $K_c$, $K_{\beta_v}$, $K_{\beta_h}$, $K_\eta$ and $K_{Q^{-1}}$ are sensitivity kernels for the elastic and anelastic model parameters. Absolute perturbations in $Q^{-1}$ are used in (2.19) in order to balance the relative contributions of elastic and anelastic gradients.

### 2.8.3 Anelastic kernels

Liu et al. (1976) demonstrated that for an absorption band solid the complex, frequency-dependent shear modulus may be expressed as

$$\mu(\omega) = \mu(\omega_0)[1 + (2/\pi) Q^{-1} \ln(|\omega|/\omega_0) + i \text{sgn}(\omega) Q^{-1}] , \quad (2.20)$$

where $\omega_0$ denotes a reference angular frequency. The second and third contributions on the right side of eqn. (2.20) denote the effects of physical dispersion and dissipation due to anelasticity, respectively.

Upon perturbing both sides of eqn. (2.20) we obtain a relationship between $\delta \mu$ and $\delta Q^{-1}$, namely

$$\delta \mu(\omega) = \mu(\omega_0)[(2/\pi) \ln(|\omega|/\omega_0) + i \text{sgn}(\omega)] \delta Q^{-1} . \quad (2.21)$$

Tromp et al. (2005) showed that the same expression for calculating the shear modulus sensitivity kernel, i.e., $K_\mu$, may be employed to determine the shear attenuation sensitivity kernel:

$$K_{Q^{-1}} = - \int_0^T 2\mu(x) D^\dagger(x, T - t) : D(x, t) \, dt , \quad (2.22)$$

where $D$ and $D^\dagger$ denote the forward and adjoint traceless strain deviators, respectively.
Figure 2.12: Adjoint sources used to calculate sensitivity kernels for elastic and anelastic model parameters. (a) Locations of event (200911030525A) and station (OBN.II). (b) Comparisons of observed (black) and simulated (red) vertical-component seismograms. The blue rectangle denotes a measurement window selected by FLEXWIN (Maggi et al., 2009). (c) Adjoint sources based on phase measurements for elastic (black) and anelastic (red) adjoint calculations. (d) Amplitude spectrums for elastic (black) and anelastic (red) adjoint sources shown in (c). (e) & (f) Same as (c) & (d), except the adjoint sources are constructed based on amplitude measurements.
Figure 2.13: Sensitivity kernels for elastic and anelastic model parameters based on phase and amplitude measurements. (a) Sensitivity kernel for $\beta_v$ based on phase measurements. The corresponding adjoint sources are shown in Figure 2.12c. (b) Sensitivity kernel for shear attenuation $Q^{-1}$ based on phase measurements. (c) & (d) Sensitivity kernels for $\beta_v$ (c) and $Q^{-1}$ (d) based on amplitude measurements. The corresponding adjoint sources are shown in Figure 2.12e.
this case, however, the adjoint wavefield used to calculate the shear attenuation kernel is determined by the anelastic adjoint source

$$\tilde{f}_i^\dagger = \frac{1}{2\pi} \int_{-\infty}^{+\infty} \left[ \frac{2}{\pi} \ln\left(\frac{\omega}{\omega_0} - i \text{sgn}(\omega)\right) \right]^* f_i^\dagger(x, \omega) \exp(i\omega t) \, d\omega ,$$

(2.23)

where $f^\dagger$ denotes the elastic adjoint source. Therefore, two separate adjoint simulations are required to simultaneously determine the elastic and anelastic kernels.

In Figure 2.12, we use a simple example to illustrate the behavior of the sensitivity kernels for elastic wavespeeds and $Q^{-1}$ using phase and amplitude measurements. The locations of the earthquake (200911030525A) and station (OBN.II) are shown in Figure 2.12a. Figure 2.12b compares vertical-component observed and simulated seismograms. Only Rayleigh-wave measurements are considered in this example. When frequency-dependent phase measurements are used, we obtain the elastic $(f^\dagger)$ and anelastic $(\tilde{f}^\dagger)$ adjoint sources shown in Figures 2.12c and d. Sensitivity kernels for elastic shear wavespeed $\beta_v$ and $Q^{-1}$ are displayed in Figures 2.13a and b. Adjoint sources and sensitivity kernels for amplitude measurements are shown in Figures 2.12e and f and Figures 2.13c and d. In this case, the magnitudes of the amplitude kernels are larger than the phase kernels due to the large amplitude anomaly, as shown in Figure 2.12b. The sign of the amplitude kernels indicates an increase in wavespeed and attenuation is required to reduce the amplitudes of the synthetic seismograms to match the observations.

## 2.9 Stage III: anisotropy inversion

As described in Section 2.2, anisotropy is another important factor affecting the propagation of seismic waves. In this stage, we use long-period surface waves to map both radial and azimuthal anisotropy in the crust and upper mantle beneath Europe and the North Atlantic.
2.9.1 Misfit function

Similar to eqn. (2.15), phase and amplitude anomalies are combined in the Stage III total misfit:

\[ \chi_{\text{III}} = w_\phi \chi^\phi + w_A \chi^A, \]  

(2.24)

where \( \chi^\phi \) and \( \chi^A \) are frequency-dependent phase and amplitude misfits as defined in eqns. (2.16) and (2.17), respectively. However, only three categories are used in this stage, namely, Rayleigh waves on vertical and radial components and Love waves on transverse components.

2.9.2 Model parameters

Smith & Dahlen (1973) and Montagner & Nataf (1986) showed that in a weakly anisotropic medium the phase speed of surface waves, \( c \), is a function of both angular frequency, \( \omega \), and azimuth, \( \theta \), and may be expressed as the Fourier series

\[ c(\omega, \theta) = c_0(\omega) + c_1(\omega) \cos(2\theta) + c_2(\omega) \sin(2\theta) + c_3(\omega) \cos(4\theta) + c_4(\omega) \sin(4\theta), \]  

(2.25)

where radial anisotropy is captured by \( c_0 \), which involves combinations of mass density \( \rho \) and the five Love parameters: \( A, C, L, N \) and \( F \), as described in Section 2.6.2. The functions \( c_1 \) and \( c_2 \) are \( 2\theta \)-dependent components, and are combinations of \( G_{c,s}, H_{c,s} \) and \( B_{c,s} \) in surface-wave tomography, where subscripts \( c \) and \( s \) denote cosine and sine dependence, respectively. The functions \( c_3 \) and \( c_4 \) are \( 4\theta \)-dependent components, and are combinations of \( E_{c,s} \). Thus, it requires 13 model parameters to describe surface-wave anisotropy. However, in practical applications one cannot resolve all 13 parameters. Most surface-wave tomographic studies focus on mapping lateral variations in two radially anisotropic parameters, namely, \( L \) and \( N \), and two azimuthally anisotropic parameters, namely, \( G_c \) and \( G_s \). We adopt the same strategy at this stage to map radial and azimuthal anisotropy in the crust.
and upper mantle.

The perturbation in the total misfit may be expressed as

\[
\delta \chi_{III} = \int_V K_L \delta L + K_N \delta N + K_{G_c} \delta G_c + K_{G_s} \delta G_s \, dV ,
\]

\[
= \int_V K_{\beta_v} \delta \ln \beta_v + K_{\beta_h} \delta \ln \beta_h + K_{G'_c} \delta G'_c + K_{G'_s} \delta G'_s \, dV .
\] (2.26)

where \( K_L, K_N, K_{G_c} \) and \( K_{G_s} \) are sensitivity kernels for the four model parameters \( L, N, G_c \) and \( G_s \), whereas \( K_{\beta_v}, K_{\beta_h}, K_{G'_c} \) and \( K_{G'_s} \) are sensitivity kernels for the dimensionless model parameters \( \delta \ln \beta_v, \delta \ln \beta_h, G'_c \) and \( G'_s \). The dimensionless parameters \( G'_c \) and \( G'_s \) are defined as

\[
G'_c = G_c / (\rho \beta_0^2) ,
\]

\[
G'_s = G_s / (\rho \beta_0^2) .
\] (2.27)

(2.28)

where \( \beta_0 \) denotes the isotropic shear wavespeed in the 1D reference model. Relationships between these kernels and the primary kernels may be found in eqns. (2.29) and (2.30)

\[
K_L = K_{c_{44}} + K_{c_{55}} ,
\]

\[
K_N = K_{c_{66}} - 2K_{c_{12}} ,
\]

\[
K_{G_c} = K_{c_{55}} - K_{c_{44}} ,
\]

\[
K_{G_s} = -K_{c_{45}} .
\] (2.29)

Sensitivity kernels with respect to dimensionless model parameters \( \delta \ln \beta_v, \delta \ln \beta_h, G'_c \) and
\( G' \) may be expressed as

\[
\begin{align*}
K_{\beta_c} &= 2L K_L - 4L \eta K_F , \\
K_{\beta_h} &= 2N K_N , \\
K_{G'_c} &= \rho \beta^2_0 K_{G_c} , \\
K_{G'_s} &= \rho \beta^2_0 K_{G_s} ,
\end{align*}
\]

(2.30)

where \( \eta = F/(A - 2L) \).

We use the same preconditioned conjugate gradient approach to iteratively update these four anisotropic model parameters. Isotropic shear wavespeed \( \beta \) and the radially anisotropic model parameter \( \xi \) may be derived based on the updated \( L \) and \( N \) Love parameters as follows

\[
\beta = \sqrt{(2L + N)/3 \rho} ,
\]

(2.31)

\[
\xi = N/L .
\]

(2.32)

The direction of the fast anisotropic axis, \( \zeta \), and the strength of the azimuthal anisotropy, \( G_0 \), may be calculated based on \( G_c \) and \( G_s \) via

\[
G_0 = \sqrt{G_s^2 + G_c^2} ,
\]

(2.33)

\[
\zeta = \frac{1}{2} \arctan(G_s/G_c) .
\]

(2.34)

Ten additional preconditioned conjugate gradient iterations are performed to construct a new anisotropic model of Europe and the North Atlantic, named EU_{60}, which required more than 5,700 wavefields simulations and 0.8 million central processing unit hours.
2.10 Improvements in misfits and histograms

2.10.1 Misfit function evolution

Figure 2.14 shows the evolution of the phase misfit during the three-stage inversion. During the first 50 iterations, the total misfit involves six categories (see Sections 2.6.1 and 2.8.1): P-SV body waves on vertical and radial components, SH body waves on transverse components, Rayleigh waves on vertical and radial components and Love waves on transverse components. As described in Section 2.6.4, the short-period corner of the surface-wave bandpass filter is gradually reduced from 50 s to 25 s to progressively resolve smaller-scale structures. The period range of body waves is changed from 15–50 s to 15–40 s at iteration 4 and fixed in subsequent iterations. The total misfit —as well as the misfit in each of the six subcategories— is gradually reduced over the 50 iterations, except for several slight increases when the short-period corner is reduced. For example, at iteration 4 the total misfit increases from 2.84 to 3.19 due to the incorporation of 40 s surface waves. Thus, it is important to recognize that the “misfit function” is a moving target that changes every several iterations. The increase in misfit at iteration 18 is due to the assimilation of several datasets recorded by stations from temporary IRIS/PASSCAL experiments and the Kandilli Observatory. Over the last 10 iterations, only surface waves are employed to constrain radial and azimuthal anisotropy in the crust and upper mantle (Section 2.9.1). Therefore, we only monitor the behavior of the misfit for three-component surface waves.

Amplitude measurements are used in the inversion after 30 elastic iterations. In Figure 2.15, we monitor the behavior of the amplitude misfits from iteration 30 to 60. Misfits in the first 20 anelastic iterations involve six categories, i.e., three-component body- and surface-wave seismograms. Over the last 10 anisotropic iterations, there are only three categories, i.e., just three-component surface waves. The overall amplitude misfit and the contributions from each of its subcategories are gradually reduced over the 30 iterations.
Figure 2.14: Evolution of phase misfits during the three-stage inversion. (a) Evolution of the total phase misfit, where blue, red and black dots label elastic, anelastic and anisotropic iterations, respectively. (b)–(d) Evolution of the phase misfit for P-SV body waves on vertical (b) and radial (c) components, and SH body waves on transverse components (d). (e)–(g) Evolution of the phase misfits for Rayleigh waves on vertical (e) and radial (f) components, and Love waves on transverse components (g). A simple multiscale strategy is used in the iterations, i.e., the short-period corner of the bandpass filter is gradually reduced as the models improve, allowing us to steadily resolve smaller-scale structures (see Section 2.6.4). Note that the “misfit function” is a moving target that changes every several iterations. The important observation is that after each change, the new misfit is gradually reduced overall and in all six categories.
Figure 2.15: Evolution of amplitude misfits during the anelastic and anisotropic inversions, Stages II & III. (a) Evolution of the total amplitude misfit from iteration 30 to 60. Red and black dots label the anelastic and anisotropic inversions, respectively. (b)–(d) Evolution of amplitude misfits for P-SV body waves on vertical (b) and radial (c) components, and SH body waves on transverse components (d). (e)–(g) Evolution of amplitude misfits for Rayleigh waves on vertical (e) and radial (f) components, and Love waves on transverse components (g).
2.10.2 Comparisons of histograms

In this section we compare phase and amplitude histograms for starting model $EU_{00}$ and final model $EU_{60}$. Phase histograms in all six categories are summarized in Figure 2.16. Compared with starting model $EU_{00}$, both mean values and standard deviations of phase histograms for the final model are significantly improved, e.g., for Rayleigh waves on the vertical component the mean value is reduced from $-1.36$ to $0.02$ and the standard deviation is reduced from $3.93$ to $2.40$. These improvements demonstrate that synthetic seismograms based on the final model $EU_{60}$ are able to simultaneously match observed short-period body waves and long-period surface waves.

In Figure 2.17, we compare amplitude histograms for the starting and final models. Similar to the phase histograms shown in Figure 2.16, amplitude histograms in all six categories are improved after the last 30 iterations. For instance, for Rayleigh waves on the vertical component the mean value and standard deviation are reduced from $0.07$ to $0.05$ and $0.36$ to $0.29$, respectively. In contrast, we compare amplitude histograms for starting model $EU_{00}$ and model $EU_{30}$ after the elastic inversion in Figure 2.18. There are no significant improvements in the amplitudes after 30 elastic iterations, because only phase measurements are used to constrain elastic wavespeeds in Stage I (Section 2.6). Amplitudes for some specific paths may be improved due to focusing and defocusing effects, however, in order to significantly reduce amplitude anomalies, shear attenuation $Q^{-1}$ has to be considered in the structural inversion.

2.11 Conclusions

In this paper, we use adjoint tomography to image the crust and upper mantle beneath Europe and the North Atlantic. A three-stage inversion strategy is designed to determine 3D variations in elastic, anelastic and anisotropic model parameters, as summarized in Figure 2.19. In Stage I, only phase information is used to image elastic wavespeeds. Af-
Figure 2.16: Comparisons of phase histograms between starting model EU$_{00}$ (red) and final model EU$_{60}$ (blue) for 15–40 s body waves and 25–100 s surface waves. (a)–(c) Comparisons of phase histograms for P-SV body waves on vertical (a) and radial (b) components, SH body waves on transverse components (c). (d)–(f) Comparisons of phase histograms for Rayleigh waves on vertical (d) and radial (e) components, Love waves on transverse components (f).

Figure 2.17: Same as Figure 2.16 except for amplitudes.
After 30 preconditioned conjugate gradient iterations, we determine a new elastic model, named $\text{EU}_{30}$, which is simultaneously constrained by three-component short-period body waves and long-period surface waves. In Stage II, we combine phase and amplitude differences between observed and simulated seismograms to simultaneously constrain elastic wavespeeds and anelastic attenuation. A new anelastic model, namely $\text{EU}_{50}$, is constructed based on 20 additional iterations. In Stage III, remaining phase and amplitude anomalies for three-component surface waves are used to constrain radial and azimuthal anisotropy, culminating in anisotropic model $\text{EU}_{60}$. Gradual reductions in misfit and significant improvements in phase and amplitude histograms help us to validate our inversion strategy. A companion paper (Zhu et al., 2013c) provides a tectonic interpretation of the 3D variations in elastic, anelastic and anisotropic heterogeneity determined in this three-stage inversion.
Figure 2.19: Three-stage inversion workflow. Computational requirements for each stage are indicated.
2.12 Acknowledgments

We acknowledge IRIS (iris.edu), ORFEUS (orfeus-eu.org) and the Kandilli Observatory (koeri.boun.edu.tr) for providing the data used in this study. We thank Daniel Stich for providing Spanish data, which help us to illuminate the Iberian peninsula. Numerical simulations for this article were performed on a Dell cluster built and maintained by the Princeton Institute for Computational Science & Engineering (PICSciE). Data and synthetics processing was accomplished based on the Seismic Analysis Code (SAC; Goldstein et al. (2002)). All maps and cross sections were made with the Generic Mapping Tool (GMT; Wessel & Smith (1991)). The open source spectral-element software package SPECFEM3D_GLOBE and the seismic measurement software package FLEXWIN used for this article are freely available for download via the Computational Infrastructure for Geodynamics (CIG; geodynamics.org). This research is supported by NSF grants 1063057 and 1112906.
Chapter 3

Seismic structure of the European crust and upper mantle based on adjoint tomography -II. Model interpretation

Note


3.1 Summary

We present model EU$_{60}$, a new crust and upper mantle model of the European continent and the North Atlantic Ocean. It is constructed based on adjoint tomography and includes 3D variations in elastic wavespeeds, anelastic attenuation, and radial & azimuthal
anisotropy. The study area was shaped by complicated tectonics driven by collisions among the Eurasian, African, Anatolian and Arabian Plates, and divergence between the Eurasian and North American Plates. Small-scale subduction and extension accompanied these plate-scale motions, creating complicated structures in the crust and upper mantle. Long-wavelength elastic wavespeed variations in model EU60 compare favorably with previous body- and surface-wave tomographic models. Some hitherto unidentified features, such as the Adria microplate, naturally emerge from the smooth starting model. Subducting slabs, slab detachments, ancient suture zones, continental rifts and back-arc basins are well-resolved in model EU60. We find an anti-correlation between shear wavespeed and anelastic attenuation at depths < 100 km. At greater depths, this anti-correlation becomes relatively weak, in agreement with previous global attenuation studies. Furthermore, enhanced attenuation is observed within the mantle transition zone beneath the North Atlantic Ocean. Consistent with typical radial anisotropy in 1D reference models, the European continent is dominated by features with a radially anisotropic parameter \( \xi > 1 \), indicating predominantly horizontal flow within the upper mantle. In addition, subduction zones, such as the Apennines and Hellenic arcs, are characterized by vertical flow with \( \xi < 1 \) at depths greater than 150 km. We find that the direction of the fast anisotropic axis is closely tied to the tectonic evolution of the region. Averaged radial peak-to-peak anisotropic strength profiles identify distinct brittle-ductile deformation in lithospheric strength beneath oceans and continents. Isotropic shear wavespeed and azimuthal anisotropy are used to explore radial rheology beneath, for example, the Alps, Aegean, Anatolia, North Atlantic and Iceland. Finally, we use the “point-spread function” to assess image quality and analyze tradeoffs between different model parameters.
3.2 Introduction

The European continent has been shaped by elaborate tectonic activities since the Archaean, such as continent-continent collisions, continental rifting, subduction, and back-arc extension (Artemieva et al., 2006). These processes not only modified surface geological features, but also created complicated subsurface structures, making the European continent a fascinating place for tomographic studies.

Seismic tomography is an important geophysical tool for imaging Earth’s interior (Woodhouse & Dziewonski, 1984; Van der Hilst et al., 1997; Romanowicz, 2003; Montelli et al., 2004). Such images are used to investigate tectonic evolution, simulate mantle convection, and determine earthquake locations. Moreover, lateral variations in temperature, water content, and composition may be inferred based on combined images of elastic wavespeeds, anelastic attenuation and radial & azimuthal anisotropy. Over the past several decades, body-wave tomography based on P and S wave arrival times has been used to construct 3D compressional and shear wavespeed models of the European continent (Spakman, 1986, 1990, 1991; Spakman et al., 1993; Piromallo & Morelli, 2003; Lippitsch et al., 2003; Amaru, 2007; Koulakov et al., 2009; Mitterbauer et al., 2011). In spite of good constraints on deep mantle structure, teleseismic arrival times have relatively poor resolutions at shallow depths because of steeply incident rays and an uneven distribution of earthquakes and stations.

In contrast, surface-wave tomography based on the dispersion of Rayleigh and Love waves provides relatively good lateral resolution on variations in shear wavespeed at shallow depths. Global-scale surface-wave tomographic studies constrain long-wavelength structures in Europe (Shapiro & Ritzwoller, 2002; Kustowski et al., 2008b; Boschi et al., 2009), and regional-scale surface-wave tomography has been employed to determine 3D variations in smaller-scale shear wavespeed variations (Boschi et al., 2004; Pasyanos, 2005; Weidle & Maupin, 2008; Schivardi & Morelli, 2009, 2011). Due to the limited depth sensitivity of surface waves, deep structures in these models are poorly constrained.
compared to shallower features. In order to simultaneously constrain shallow and deep structures, it is preferable to jointly invert body and surface waves. Partitioned waveform inversion was developed to fit both shear body waves and surface waveforms (Nolet, 1990). It has been successfully employed to construct 3D shear wavespeed models of Europe (Zielhuis & Nolet, 1994a; Marone et al., 2004; Schmid et al., 2008; Chang et al., 2010a,b). Finally, ambient noise tomography also provides good constraints on crust and uppermost mantle structure beneath the European continent (Yang et al., 2007).

Due to significant advances in numerical techniques and high-performance computing, seismologists are now able to routinely calculate accurate synthetic seismograms in general 3D Earth models (Komatitsch & Tromp, 1999; Peter et al., 2011). Adjoint tomography was developed to combine synthetic seismograms with high-quality observed seismographic records to estimate 3D variations in seismic model parameters based on massive data assimilation (Tromp et al., 2005; Tape et al., 2007; Liu & Tromp, 2006, 2008). This approach bridges gaps between traditional body- and surface-wave tomography by fully exploiting information in complete seismic records (Zhu et al., 2012). This type of tomographic method has been successfully used to image crustal structures in southern California (Tape et al., 2009, 2010), and upper mantle structures beneath Australia (Fichtner et al., 2009, 2010), Europe (Zhu et al., 2012; Fichtner et al., 2013b) and the North Atlantic (Rickers et al., 2013).

Complementing elastic wavespeeds, 3D variations in attenuation and anisotropy provide additional constraints on physical properties of Earth materials. Contrary to tomographic investigations of elastic wavespeeds, there is little agreement among different attenuation studies, even on a global scale (Romanowicz, 1995; Billien et al., 2000; Gung & Romanowicz, 2004; Lawrence & Wysession, 2006; Dalton et al., 2008). Unfortunately, for the European continent there are few regional or local models of attenuation.

Shear-wave splitting measurements are widely used to investigate azimuthal anisotropy in Europe (Walker et al., 2005; Diaz et al., 2010). Such measurements represent an in-
tegral effect of azimuthal anisotropy over the entire crust and upper mantle. The limited depth resolution of shear-wave splitting measurements has led to a long-standing argument about the origin of seismic anisotropy (Silver & Chan, 1991; Vinnik et al., 1992; Silver, 1996). Constraints on European azimuthal anisotropy either come from large-scale inversions (Pilidou et al., 2004; Debayle et al., 2005; Ekström, 2011) or small regional-scale studies, e.g., the Alps (Fry et al., 2010) and the Aegean (Endrun et al., 2011).

In this paper, we present a new seismic model of the crust and upper mantle beneath Europe and the North Atlantic, named EU$_{60}$. It incorporates 3D variations in elastic wavespeeds, anelastic attenuation, and radial & azimuthal anisotropy. A three-stage inversion strategy for constructing this model is presented in Zhu et al. (2013b) (hereafter referred to as paper I). The main tectonic structures of the European continent are described in Section 3.3. In Sections 3.4–3.6, various horizontal cross sections are used to illustrate 3D variations in elastic wavespeeds, anelastic attenuation, and radial & azimuthal anisotropy, and several previous surface- and body-wave tomographic models are compared with model EU$_{60}$. Detailed images beneath several notable regions –such as the Alps, the Aegean and Anatolian Plates, the North Atlantic, and Iceland– are discussed in Section 3.7. Finally, the “point-spread function” (Fichtner & Trampert, 2012) is used in Section 3.8 to analyze resolution in the tomographic images and tradeoffs between different model parameters.

### 3.3 Tectonic setting

The northeastern part of the European continent is dominated by the East European Craton (EEC), which is composed of the Baltic Shield, the Ukrainian Shield and the East European Platform (Figure 3.1). This region is of Archaean-Proterozoic age and has been stable over a long geological time. The Tornquist-Tessseyre Suture Zone (TTSZ) (Zielhuis & Nolet, 1994a) separates this Precambrian craton from the Phanerozoic parts of Europe. During
the Paleozoic, central and western Europe were mainly shaped by the Caledonides and
Variscides Orogenies, which are currently distributed along the western coast of Scandi-
navia, the British Isles, Germany, and France (see Figure 3.1).

Mesozoic and Cenozoic tectonic activities were primarily driven by the convergency
of the Eurasian and African-Arabian Plates. This broad-scale compression resulted in sub-
duction of the Tethys ocean. Numerous mountain belts with arcuate shapes, such as the
Maghrebides, Apennines, Alps, Dinarides, and Hellenides, were created under a tectonic
background of convergence. Back-arc extension accompanied by the subduction of oceanic
lithosphere occurred in the Algero-Provençal Sea, Tyrrhenian Sea, Pannonian Basin, and
Aegean Sea. Slab roll-back and trench retreat are important mechanisms for the devel-
opment of these arcuate mountain belts and back-arc basins (Wortel & Spakman, 2000).
For instance, the Calabrian arc migrated to its current position due to a roll-back of slabs
that started approximately 30 million years ago, resulting in the opening of the Algero-
Provençal Basin and the Tyrrhenian Sea, as well as the rotation of the Corsica and Sardinia
blocks.

The European Cenozoic Rift System (ECRS) (Ziegler, 1992) is responsible for exten-
sion and volcanism in western Europe during the Cenozoic. This rift system extended from
northern Africa to the North Sea, and created several massifs, grabens, and hotspots, e.g.,
the Massif Central, Rhine Graben, and Eifel Hotspot (Goes et al., 1999).

In eastern Europe, due to the closure of the Tethys, the northward moving Arabian Plate
collided with the stable EEC and turned its trajectory westerly. This movement contributed
to the counter-clockwise rotation of the Anatolian Plate, accommodated by strike-slip mo-
tion along the North Anatolian Fault. The main tectonic structures of the European con-
tinent are illustrated in Figure 3.1. A detailed overview of the tectonic evolution of the
Alpine–Mediterranean region may be found in Dercourt (1986) and Dewey et al. (1989).
3.4 Elastic wavespeeds

3.4.1 Isotropic shear wavespeed

In Figure 3.3, we present lateral variations in relative isotropic shear wavespeed in model EU$_{60}$. Model STW105 (Kustowski et al., 2008a), a transversely isotropic model shown in Figure 3.2, is used as a 1D reference model. Long-wavelength structures at shallow depths are consistent with previous surface-wave models (Boschi et al., 2004; Shapiro & Ritzwoller, 2002; Kustowski et al., 2008b; Chang et al., 2010a; Schivardi & Morelli, 2011). The EEC is characterized by faster-than-average wavespeeds (> 4%) down to depths in excess of 250 km, representing a cold and old continental lithospheric lid. The TTSZ is imaged as a sharp boundary between the EEC and western and central Europe down to depths greater than 250 km (Zielhuis & Nolet, 1994a). The Alpine-Himalaya orogenic belt, starting from the western Mediterranean, continuing through the Pannonian Basin, and extending to Anatolia, is resolved as a slow wavespeed anomaly at depths shallower than 150 km.

Within the mantle transition zone, central Europe is dominated by fast wavespeed anomalies (> 3%), which are related to slab roll-back associated with the Apennines-Calabrian-Maghrebides, Carpathians-Vrancea-Adria and Hellenic-Cyprus arcs (Zhu et al., 2012). These fast wavespeed anomalies are in excellent agreement with results from body-wave traveltime tomography (Wortel & Spakman, 2000; Piromallo & Morelli, 2003). Since in our inversion short-period body waves and long-period surface waves are combined to simultaneously constrain deep and shallow structures, there is generally good agreement between model EU$_{60}$ and previous surface-wave inversions at shallow depths and body-wave images at greater depths (for details see Section 3.4.5).

Numerous short-wavelength model features are resolved in EU$_{60}$. Most of these are missing in large-scale surface-wave tomographic models because of their inherently limited lateral resolution. Fast wavespeed anomalies associated with the Central Graben and
Figure 3.2: 1D reference model STW105 (Kustowski et al., 2008a). It is used to calculate relative perturbations in elastic wavespeeds and anelastic attenuation in this paper. (a) Radially anisotropic shear wavespeeds. Black and red lines denote vertically ($\beta_v$) and horizontally ($\beta_h$) polarized shear wavespeeds, respectively. (b) 1D shear quality factor (Q).
Figure 3.3: Relative perturbations in isotropic shear wavespeed at various depths in model EU$_{60}$. Depths are denoted in the right top corner.
the Armorican Massif are well imaged down to depths in excess of 200 km. The Alps, Adriatic and Hellenides are imaged as a continuous belt with fast wavespeeds down to 200 km. At greater depths, this continuous belt disappears in most places, except the Hellenic arc, where the subducting slab is clearly traceable down into the lower mantle (see Figure 3.14). At depths greater than 250 km, a localized fast wavespeed anomaly related to the Calabrian arc is resolved. Within the mantle transition zone, there are two fast wavespeed anomalies. The one beneath the western Mediterranean is related to slab roll-back associated with the Apennines-Calabrian-Maghrebides arc. The second one beneath central Europe is related to subducting slabs associated with the Carpathians-Vrancea-Adria and Hellenic-Cyprus arcs (Zhu et al., 2012). Further aspects of these subducting slabs are discussed in Sections 3.7.1–3.7.3 based on vertical cross sections. Small-scale slow wavespeed anomalies related to the ECRS (Ziegler, 1992), such as the Massif Central, Eifel Hotspot, Bohemian Massif, and Central Slovakia Volcanic Field, are mapped at depths shallower than 300 km.

### 3.4.2 Radial anisotropy

Both vertically ($\beta_v$) and horizontally ($\beta_h$) polarized shear wavespeeds are considered in this study. Thus, radial anisotropy may be captured based on the parameter

$$\xi = (\beta_h/\beta_v)^2.$$  

Variations in the $\xi$ parameter may be used to infer vertical and horizontal flows in the crust and upper mantle (Ekström & Dziewonski, 1998; Gung et al., 2003; Shapiro et al., 2004; Moschetti et al., 2010). Figure 3.4 shows lateral variations in $\xi$ at various depths in model EU60. At depths shallower than 150 km, the horizontally polarized shear wavespeed ($\beta_h$) is generally faster than the vertically polarized shear wavespeed ($\beta_v$), i.e., $\xi > 1$, in agreement with reference model STW105 (Kustowski et al., 2008a) (see Figure 3.2). Large values of $\xi$ may indicate the presence of horizontal flow within the upper mantle. At greater depths
Figure 3.4: Radially anisotropic model parameter $\xi = (\beta_h/\beta_v)^2$ at various depths in model EU$_{60}$. A value of $\xi > 1$ indicates the presence of horizontal flow, while values $\xi < 1$ indicate vertical flow. Depths are denoted in the left bottom corner.

( >150 km), some regions, such as the Calabrian and Hellenic arcs, are characterized by $\beta_h < \beta_v$, which may indicate vertical flow induced by subducting slabs. The small values of $\xi$ at a depth of 220 km beneath the North Atlantic Ocean may be related to upwelling within the upper mantle driven by the divergency of the Eurasian and North American Plates.

3.4.3 Other model parameters

We simultaneously determine 3D variations in bulk sound wavespeed ($c$), and vertically ($\beta_v$) and horizontally ($\beta_h$) polarized shear wavespeeds (see details in paper I). In Figure 3.5, we compare lateral variations in different model parameters at a depth of 100 km. Long-wavelength features, such as the EEC, are imaged in both compressional ($\alpha$) and shear ($\beta$) wavespeeds. However, fast wavespeed anomalies beneath the Adria microplate and the Hellenic arc are much broader in the compressional wavespeed image. In contrast, slow anomalies related to the western Mediterranean and the Pannonian Basin are relatively weak in compressional wavespeeds. Based on the compressional and shear wavespeeds,
we are able to calculate the $V_p/V_s$ ratio. Figure 3.5 illustrates the relative perturbation of $V_p/V_s$ ratio at a depth of 100 km. There is a good correlation between regions with slow shear wavespeed and a high $V_p/V_s$ ratio, for instance, beneath the Alpine-Himalaya orogenic belt. High $V_p/V_s$ ratios might be indicative of the presence of partial melt in this regions (Zhang et al., 2004). In contrast, the EEC is characterized by low $V_p/V_s$ ratios at this depth.

3.4.4 Model evolution

As discussed in paper I, a three-stage inversion strategy is used to constrain 3D variations in elastic, anelastic and anisotropic model parameters. Global radially anisotropic model
S362ANI (Kustowski et al., 2008a) is chosen as the starting model. In Stage I, only frequency-dependent phase differences between observed and simulated seismograms are employed to map lateral variations in radially anisotropic elastic wavespeeds. Based on 30 preconditioned conjugate gradient iterations (Fletcher & Reeves, 1964), we construct a new elastic model, named EU$_{30}$. In Figure 3.6, we observe significant improvements in isotropic shear wavespeed from the starting model EU$_{00}$ to the new elastic model EU$_{30}$ at both shallow (100 km) and great (600 km) depths. Since the starting model is based on a global-scale study (Kustowski et al., 2008a), it only involves long-wavelength features, such as, slow wavespeed anomalies along the Alpine-Himalaya orogenic belt, fast wavespeed anomalies beneath the EEC, and relatively weak slab signatures at a depth of 600 km. In EU$_{30}$, numerous small-scale features, as described in Section 3.4.1, naturally emerge from the smooth starting model.

In Stage II, i.e., iterations 30 to 50, we combine frequency-dependent amplitude anomalies with remaining phase differences from Stage I to simultaneously estimate elastic wavespeeds and anelastic attenuation. Most long-wavelength elastic wavespeed features of EU$_{30}$ are preserved in model EU$_{50}$, except for slight modifications of some small-scale heterogeneities. For instance, slow anomalies beneath the western Mediterranean and Eifel Hotspot become more prominent after the anelastic inversion. At a depth of 600 km, fast wavespeed slab features beneath central Europe become stronger.

In Stage III, i.e., iterations 50 to 60, only surface waves phase and amplitude anomalies are employed to constrain variations in radial & azimuthal anisotropy at shallow depths. The change in $\beta_v$ and $\beta_h$ from EU$_{50}$ to EU$_{60}$ is relatively modest because of small updates in wavespeeds from iteration 50 to 60. No body waves are incorporated in the anisotropic inversion, therefore, there is no change in shear wavespeed at great depths, e.g., 600 km.
Figure 3.6: Evolution of isotropic shear wavespeed from starting model $EU_{00}$ to purely elastic model $EU_{30}$ (Stage I) to anelastic model $EU_{50}$ (Stage II) to final model $EU_{60}$ (Stage III) at depths of 100 km (left) and 600 km (right).
3.4.5 Comparisons with pervious tomographic images

In recent years, a variety of compressional and shear wavespeed models of the European upper mantle has been developed based on different datasets and approaches. In order to assess similarities and discrepancies between model EU$_{60}$ and these complementary studies, we compare elastic shear wavespeed structures in EU$_{60}$ with five previous surface-wave tomographic models, namely EPmantle (Schivardi & Morelli, 2011), Chang2010 (Chang et al., 2010a), S2.9EA (Kustowski et al., 2008b), LRSP30EU02 (Boschi et al., 2009) and CUSDT1.0 (Shapiro & Ritzwoller, 2002), as well as body-wave traveltime model LLNL-G3Dv3 (Simmons et al., 2012).

EPmantle (Schivardi & Morelli, 2011) was constructed based on fundamental mode Rayleigh and Love group wavespeeds; only regional earthquakes were used in this study. Chang et al. (2010a) combined regional S and Rayleigh waveforms, teleseismic arrival times, and Rayleigh group wavespeeds to constrain upper-mantle shear wavespeeds along the Tethyan margin. S2.9EA (Kustowski et al., 2008b) is a shear wavespeed model constructed based on surface-wave phase wavespeeds, long-period waveforms and body-wave traveltimes. LRSP30EU02 (Boschi et al., 2009) is a shear wavespeed model based on Rayleigh and Love fundamental-mode phase anomalies. Both S2.9EA and LRSP30EU02 are large-scale tomographic inversion with a finer parameterization beneath Eurasia (S2.9EA) and the Mediterranean (LRSP30EU02), respectively. CUSDT1.0 (Shapiro & Ritzwoller, 2002) is a global shear wavespeed model based on fundamental-mode phase and group wavespeeds. All of models are radially anisotropic.

In Figure 3.7, we compare relative perturbations in isotropic shear wavespeed at a depth of 100 km. At long wavelength, the level of agreement among these six models is very good. For instance, the EEC and the Alpine-Himalaya orogenic belt are imaged as fast and slow anomalies in all six models, respectively. Model EU$_{60}$ involves more short-wavelength features which are not well resolved by the other five models. For instance, the continuous belt with fast wavespeed connecting the Alps, Adriatic and Hellenides is
Figure 3.7: Comparison of relative perturbations in isotropic shear wavespeed at a depth of 100 km for six different tomographic models, namely, EU60 (this study), EPmantle (Schivardi & Morelli, 2011), Chang2010 (Chang et al., 2010a), S2.9EA (Kustowski et al., 2008b), LRSP30EU02 (Boschi et al., 2009), CUSDT1.0 (Shapiro & Ritzwoller, 2002). STW105 (Kustowski et al., 2008a) is used as a reference model to calculate relative perturbations.

not obvious in EPmantle and Chang2010, and is only resolved as a very smooth feature in S2.9EA, LRSP30EU02 and CUSDT1.0. A similar observation can be made for fast wavespeed anomalies associated with the Central Graben and the Armorican Massif. In addition, the amplitudes of slow anomalies beneath the western Mediterranean, Pannonian Basin and Anatolian Plate in EPmantle and Chang2010 are relatively large compared to the other four models.

In Figure 3.8, we compare radial anisotropy in these six models at a depth of 150 km. In contrast to the reasonable consensus in isotropic shear wavespeeds among the various models, the level of agreement in radial anisotropy is relatively poor. This discrepancy is due to different weights assigned to Rayleigh and Love waves in the various inversions. At long wavelengths, all models involve large values of $\xi$ beneath the European continent, indicat-
Figure 3.8: Same as Figure 3.7, except for a comparison of the radially anisotropic model parameter $\xi$ at a depth of 150 km.

Model $E_{U60}$ identifies small values of $\xi$ beneath the Apennines, Adria and Hellenic arcs, which are only weakly imaged in model LRSP30EU02.

Finally, in Figure 3.9 we compare model $E_{U60}$ with body-wave traveltime model LLNL-G3Dv3 (Simmons et al., 2012). There is good agreement between shear wavespeed signatures in $E_{U60}$ and compressional wavespeed features in LLNL-G3Dv3 at both shallow (100 km) and great (600 km) depths. The EEC is not well imaged in model LLNL-G3Dv3 due to vertically incident body-wave and a lack of stations and earthquakes in this region. Both models reveal strong fast wavespeed anomalies beneath central Europe within the mantle transition zone.
Figure 3.9: Comparison of EU60 with body-wave traveltime model LLNL-G3Dv3 (Simmons et al., 2012) at depths of 80 km (top) and 600 km (bottom). STW105 (Kustowski et al., 2008a) is used as a 1D reference model to calculate relative perturbations in isotropic shear wavespeed for EU60, while PREM (Dziewonski & Anderson, 1981) is used as a 1D reference model to calculate relative perturbations in compressional wavespeed for LLNL-G3Dv3 (Simmons et al., 2012).
3.5 Attenuation

Horizontal cross sections of relative perturbations in shear attenuation $Q^{-1}$ are displayed in Figure 3.10. The radial Q profile from STW105 (Kustowski et al., 2008a) is used as 1D reference model to calculate relative perturbations (Figure 3.2). At shallow depths, e.g., 100 km, there is a clear anti-correlation between elastic wavespeeds and anelastic attenuation (compare with Figure 3.3). For instance, the EEC is revealed as a region with low attenuation and fast wavespeed. Below 200 km, the anti-correlation between wavespeeds and attenuation becomes relatively weak, in agreement with observations of global surface-wave attenuation tomography (Romanowicz, 1995; Billien et al., 2000; Gung & Romanowicz, 2004; Lawrence & Wysession, 2006; Dalton et al., 2008).

As discussed in Section 3.4.1, within the mantle transition zone central Europe is dominated by several fast wavespeed anomalies related to subducting slabs in the Mediterranean-Alpine region. However, there is no obvious anti-correlation between anelastic attenuation and elastic wavespeeds at these depths (Figure 3.10).

In Figure 3.11, three vertical cross sections (N1–N3) are used to explore shear attenuation beneath the North Atlantic. The lithosphere generally exhibits weak attenuation whereas the asthenosphere is characterized by relatively high attenuation, and we recognize these characteristics in the cross sections. Perhaps surprisingly, enhanced attenuation is revealed within the mantle transition zone, but no significant accompanying reduction in shear wavespeed is observed. This feature might be related to the presence of water. Major minerals of the mantle transition zone, namely Wadsleyite and Ringwoodite, have larger water solubilities than minerals in the shallow upper mantle (Kohlstedt et al., 1996). These minerals might be reservoirs for significant amounts of water in Earth’s mantle (Bercovic & Karato, 2003; Karato, 2011). A detailed discussion of the implications of a water-enriched mantle transition zone may be found in Zhu et al. (2013a).
3.6 Anisotropy

Azimuthal anisotropy in EU60 between 50 km to 220 km is illustrated in Figure 3.12. It involves complex lateral and depth variations in the directions and amplitudes of the fast anisotropic axis beneath continental Europe and the North Atlantic. Because of the limited depth sensitivity of surface waves, the strength of azimuthal anisotropy decreases below 200 km. Above 200 km, anisotropic fabrics are well correlated with regional tectonic evolution. Along the North Atlantic Ridge (NAR), the fast axis runs parallel to the extensional direction of the ridge system within the upper mantle. Small-scale complexities, such as the convergency of the fast axis along the western coast of England, are intriguing and require further investigations involving other geophysical observables, such as gravity and electrical conductivity. The fast axis beneath the western Mediterranean follows the opening trajectories of the Algero-Provençal and Tyrrhenian Seas, suggesting trench retreat of the Apennines-Calabrian arc.

At shallow depths, e.g., 50–150 km, the EEC involves complex azimuthally anisotropic
Figure 3.11: Vertical cross section of shear attenuation $Q^{-1}$ beneath the North Atlantic Ocean. (a) Locations of the three vertical cross sections N1–N3. (b)–(d) Shear attenuation in the three vertical cross sections N1–N3. The dashed black lines in N1–N3 denote the 220 km, 410 km and 660 km discontinuities.
Figure 3.12: Azimuthal anisotropy at various depths in model EU$_{60}$. The direction and amplitude of the fast axis are given by the orientation and length of the yellow bar. Blue lines denote global plate boundaries (Bird, 2003).
patterns, which might be correlated with ancient continental rifts, as discussed in Zhu & Tromp (2013). At greater depths, relatively weak anisotropy is observed within the EEC. Around the EEC, the fast axis is well correlated with tectonic activities which are closely related to the accretion of the EEC since the Paleozoic. The fast axis along the western coast of Scandinavia follows the trend of the Caladonian Orogeny during the Paleozoic (see Figure 3.1). The TTSZ is delineated by the fast anisotropic direction throughout the upper mantle, separating the EEC and western and eastern Europe. To the southern border of the EEC, the fast direction indicates northward motion and counter-clockwise rotation of the Arabian Plate, reflecting collision between Eurasia and Arabia.

In Figure 3.13, we compare averaged radial peak-to-peak anisotropic strength profiles for several regions. For oceanic regions, such as the North Atlantic Ocean (profile 1) and the Mediterranean (profile 2), strong azimuthal anisotropy is observed at a depth of 100 km. Within the upper mantle, the anisotropic strength increases monotonically, and then steadily decreases with depth below its maximum value. The same feature is observed beneath the Pannonian Basin, except that the depth of maximum anisotropic strength is approximately 150 km (profile 3).

In contrast, beneath the Aegean and Anatolian Plates (profiles 4 and 5), two peaks in anisotropic strength are observed. The first peak indicates weak and ductile lower crust. The second peak reflects strong mantle flow within the lithosphere and asthenosphere. They are accommodated by a transition zone with relatively weak anisotropic strength. These profiles are consistent with mineral physics experiments involving distinct brittle-ductile deformation in lithospheric strength beneath oceans and continents (Kohlstedt et al., 1995).

Beneath the EEC and Ukrainian Shield (profiles 6 and 7), strong azimuthal anisotropy is observed within the lower crust. However, the lower lithosphere (depth of \( \sim 100 \) km) is observed to have relatively weak anisotropic strength, which might be indicative of a plunging axis of symmetry in the continental lithosphere (Debayle et al., 2005). The current model parameterization does not accommodate this type of dipping anisotropy (see
Figure 3.13: Comparisons of averaged radial peak-to-peak anisotropic strength profiles for different regions. (a) Locations of seven averaged profiles. Different colors refer to different regions. White lines denote global plate boundaries (Bird, 2003). (b) Averaged radial peak-to-peak anisotropic strength profiles for seven regions: 1, the North Atlantic Ocean; 2, the Mediterranean; 3, Pannonian Basin; 4, the Aegean Sea; 5, the Anatolian Plate; 6, the EEC; 7, the Ukrainian Shield. Black dashed lines denote a reference depth of 100 km.
3.7 Local structures

In this section, we highlight structures beneath several interesting geographic regions.

3.7.1 Apennines-Calabrian-Maghrebides arc

Slab roll-back and trench retreat (Elsasser, 1971) have been investigated based on laboratory experiments (e.g., Kincaid & Griffiths, 2003) and numerical simulations (e.g., Schellart et al., 2007). Wortel & Spakman (2000) invoked these tectonic mechanism to explain the arcuate shape of mountain belts in central Europe, e.g., the Alps and Carpathians. Based on P-wave traveltime tomography, Wortel & Spakman (2000) found slab detachments beneath the central Apennines and western Carpathians. We find the same features at three locations in this region: the central Apennines, Calabrian arc and northern Africa. As illustrated in vertical cross sections A-a, B-b and C-c of Figure 3.14, there are gaps between the shallow lithosphere and a deep fast anomaly beneath the central Apennines and northern Africa, in agreement with P-wave tomographic images determined by Wortel & Spakman (2000). In vertical cross section A-a, model EU$_{60}$ identifies another slab detachment beneath the Calabrian arc. Strong uplift, a down-dip compression stress state, and termination of roll-back in the Tyrrhenian indicate the presence of a slab detachment beneath the Calabrian arc (Wortel & Spakman, 2000). The subducting slab from this arc is stagnant within the mantle transition zone. Strong slow anomalies are identified beneath the Algero-Provençal Basin and the Tyrrhenian Sea at depths shallower than 200 km. They may be attributed to back-arc basins associated with retreat of the Apennines-Calabrian-Maghrebides arc. Compared with vertical cross sections from body-wave tomography (Wortel & Spakman, 2000; Piromallo & Morelli, 2003), the 660 km discontinuity in model EU$_{60}$ is imaged as a sharp boundary.
Figure 3.14: Relative perturbations in isotropic shear wavespeed in various vertical cross sections beneath the Mediterranean. A-a, B-b and C-c are vertical cross sections in the western Mediterranean, D-d and E-e are vertical cross sections in the Carpathians-Vrancea region, F-f is a vertical cross section in the Hellenic-Cyprus region. Black lines in the vertical cross sections denote the 220 km, 410 km and 660 km discontinuities.
3.7.2 Carpathians-Vrancea-Adria arc

Like the Apennines-Calabrian-Maghrebides arc discussed in the previous section, the Carpathians-Vrancea-Adria arc started roll-back 30 million years ago and caused the opening of the Pannonian Basin (Rosenbaum et al., 2002). In model EU$_{60}$, the western part of the Carpathians-Vrancea arc is completely detached from the surface (Figure 3.14 D-d), in agreement with observations by Wortel & Spakman (2000). However, in the eastern part, we find the signature of a slab subducted towards the southwest. In vertical cross section E-e of Figure 3.14, this fast anomaly meets the Adria slab at nearly 400 km and overlays the Hellenic slab, (for details, see Zhu et al., 2012). The Pannonian Basin is imaged as a slow wavespeed anomaly (\(\langle -4\%\)) down to a depth of 150 km. Slabs from the Carpathians-Vrancea-Adria arc and Hellenic arc lie flat at 600 km, defining the second large fast wavespeed anomaly beneath central Europe (Figure 3.3).

3.7.3 Hellenic-Cyprus arc

In contrast to the previous two arc systems, the Hellenic-Cyprus arc is relatively young and started roll-back approximately 15 million ago. Instead of being confined within the upper mantle, as the previous two slab systems, the Hellenic slab penetrates through the 660 km discontinuity into the lower mantle, as illustrated in vertical cross section F-f of Figure 3.14. This observation agrees with previous P-wave tomographic images (Papazachos & Nolet, 1997; Wortel & Spakman, 2000; Piromallo & Morelli, 2003; Widiyantoro et al., 2004). The Hellenic arc involves a prominent slow wavespeed structure (\(\langle -4\%\)) underneath the Aegean Sea at depths shallower than 100 km, which we attribute to the consequences of slab roll-back and back-arc extension.
3.7.4 Alps

Tomographic images of variations in elastic wavespeed and anisotropic fabric in the Alpine region are crucial for understanding the evolution of this orogeny (Lippitsch et al., 2003; Fry et al., 2010). In Figure 3.15, a continuous arcuate fast wavespeed anomaly is imaged along the strike of the Alps ranging from 90–240 km in model EU$_{60}$. At a depth of 90 km, the Po Basin and the Adria microplate are revealed as slow and fast anomalies, respectively. Beneath 180 km, the continuous fast anomaly of the Alps is separated into two pieces, the western and eastern Alps. They can be traced down to depths in excess of 350 km. Based on relative traveltime residuals, Babuska et al. (1990) showed the existence of two fast wavespeed anomalies beneath the western and eastern Alps, in agreement with our model EU$_{60}$. However, in the teleseismic traveltime tomographic model of Lippitsch et al. (2003), three separate fast wavespeed anomalies are observed beneath the western, central and eastern Alps, respectively.

In order to compare our images with the teleseismic P-wave model of Lippitsch et al. (2003), we make comparable vertical cross sections through the Alps (Figure 3.15, bottom row). The agreement between EU$_{60}$ and the P-wave images of (Lippitsch et al., 2003) is good, even though entirely different datasets and methods were employed in their construction.

Azimuthally anisotropic fabrics are superimposed on isotropic shear wavespeed variations in Figure 3.15. An orogenic-parallel fabric is imaged along the strike of the Alps at depths ranging from 90–180 km. The fast axis beneath the Po Basin and the Adria microplate suggests indentation of the Adria microplate (Fry et al., 2010). In addition, beneath the northern Apennines the fast direction runs parallel to imaged fast wavespeed anomalies at depths ranging from 120 km to 180 km.
Figure 3.15: Relative perturbations in isotropic shear wavespeed and azimuthal anisotropy beneath the Alpine region. Depths are indicated in the top left corner. Three vertical cross sections are shown in the bottom row, for an easy comparison with figures from Lippitsch et al. (2003). The yellow bars in the top panel denote the direction of the anisotropic fast axis.
3.7.5 Aegean and Anatolia

The southward retreat of the Hellenic arc and the westward motion of the Anatolian Plate are responsible for the tectonic evolution of the Aegean region (Endrun et al., 2011). Lateral variations in isotropic shear wavespeed and azimuthal anisotropy from the crust to the mantle transition zone are displayed in Figure 3.16. Within the lower crust, the Hellenides are imaged as a slow wavespeed anomaly, while the Aegean is mapped as a fast wavespeed anomaly, in agreement with surface-wave phase wavespeed images from Endrun et al. (2011). Beneath the Moho, the Hellenic arc is dominated by strong fast wavespeed anomalies down to depths in excess of 450 km, indicating northward subduction of the Hellenic slab. The Aegean is mapped as a slow wavespeed anomaly from the top of the upper mantle down to 200 km, suggesting the presence of a mantle wedge related to dehydration of the Hellenic slab.

The anisotropic fast direction is displayed at depths shallower than 200 km in Figure 3.16. Within the uppermost mantle, the Aegean is dominated by a northeast-southwest fast direction, correlated with the westward motion of the Anatolian Plate and the southward retreat of the Hellenic Trench. Although there is no obvious signature of distinct anisotropic directions between the lower crust and mantle lithosphere, as suggested by ambient noise Rayleigh-wave tomography (Endrun et al., 2011), radial peak-to-peak anisotropic strength profiles do identify a ductile lower crust and lithosphere beneath the Aegean (Zhu & Tromp, 2013).

Detailed images of isotropic shear wavespeed and azimuthal anisotropy beneath the Anatolian Plate are also presented in Figure 3.16. Several distinct slow wavespeed anomalies are well resolved in the interior of the Anatolian Plate, which may be traced from the crust down to 200 km. They might be related to volcanic fields in Anatolia. Strong slow wavespeed anomalies are imaged along the trace of the North Anatolian Fault down to a depth of 250 km. The images are similar to those of Fichtner et al. (2013a), except the amplitude of our slow anomalies is relatively smaller. Anisotropic fabrics from the crust
Figure 3.16: Isotropic shear wavespeed and azimuthal anisotropy at various depths beneath the Aegean and Anatolian regions. The first row shows absolute shear wavespeeds, while the remaining rows show relative perturbations in isotropic shear wavespeed with respect to STW105 (Kustowski et al., 2008a). White lines denote global plate boundaries (Bird, 2003). The yellow bars in the first and second rows denote anisotropic fast axis directions.
to the upper mantle reflect counter-clockwise rotation of the Anatolian Plate. Especially within the lower crust, it seems that this fabric follows the trace of the North Anatolian Fault.

Below 300 km, the Anatolian region is dominated by strong fast wavespeed anomalies, which can be traced to the Hellenic arc and Caucasus. There is good agreement between our model and P-wave images derived from teleseismic body waves (Biryol et al., 2011). It is interesting to observe a gap between fast anomalies from the Hellenic arc and the Caucasus at a depth of 450 km, which might indicate the presence of a slab detachment. Between 550 km and 650 km, fast anomalies beneath the Anatolian Plate become relatively weak compared to the subducting slab beneath the Hellenic arc.

3.7.6 North Atlantic Ridge and Iceland

In Figure 3.17, we present upper mantle structures beneath the North Atlantic Ocean. Only horizontally polarized shear wavespeeds ($\beta_h$) are shown to facilitate a comparison with images determined by Rickers et al. (2013). The NAR is resolved as a strong slow anomaly down to depths in excess of 200 km. However, the amplitude of the slow anomalies is relatively weak compared to the images of Rickers et al. (2013). The slow wavespeed anomaly becomes broader from 60 km to 200 km, and involves an elongated feature away from the ridge, e.g., between 120–160 km, which might be responsible for post-rift uplift in western Scandinavia and the British Isles (Rickers et al., 2013). Both Iceland and the Jan-Mayen Fracture zone are imaged as strong slow anomalies at depths shallower than 200 km.

Over the past several decades, seismic tomography and receiver functions have been used to search for mantle plume signatures (Morgan, 1971) and to map the thickness of the transition zone beneath Iceland (Wolfe et al., 1997; Shen et al., 1998; Ritsema et al., 1999; Bijwaard & Spakman, 2000; Foulger et al., 2001; Allen et al., 2002a,b; Montelli et al., 2004). Here, we make several vertical and horizontal cross sections to explore upper
Figure 3.17: Relative perturbations in horizontally polarized shear wavespeed $\beta_h$ beneath the North Atlantic Ridge. Reference shear wavespeeds at each depth are indicated in the left top corner, depths in the left bottom corner. White lines denote global plate boundaries (Bird, 2003).
Figure 3.18: Relative perturbations in isotropic shear wavespeed beneath Iceland. Two vertical cross sections A-a and B-b are used to illustrate deep structures. The black bar in the vertical cross sections denotes the lateral extent of Iceland.

mantle structures beneath Iceland in EU60. Since only regional earthquakes are used in the inversion (see paper I), the dataset is insufficient to constrain lower mantle structures, and hence we do not discuss wavespeed anomalies at depths greater than 400 km beneath Iceland. We observe strong slow wavespeed anomalies along the NAR down to 200–250 km (Figure 3.18). At a depth of 400 km beneath Iceland, there is a slow wavespeed anomaly with $-2\%$ perturbations, which might be related to the Iceland upwelling within the upper mantle.

### 3.8 Resolution analysis

Because of the demanding computational requirements of adjoint tomography, it is very expensive to perform traditional “check-board” tests for assessing image quality, which require the same amount of computational resources as an actual structural inversion. In this study, the “point-spread function” (Fichtner & Trampert, 2011, 2012) is used to assess image quality in models EU30, EU50 and EU60, as well as for analyzing tradeoffs between
elastic, anelastic and anisotropic model parameters. A finite-difference approximation is used to calculate the action of the Hessian on a localized model perturbation:

\[ H \cdot \delta m \approx g(m + \delta m) - g(m) \quad . \]  

(3.2)

where \( H \) denotes the Hessian and \( \delta m \) refers to a localized model perturbation with respect to the current model \( m \). The misfit gradient \( g \) is evaluated for both models \( m \) and \( m + \delta m \). Based on the action of the Hessian on the model perturbation, \( H \cdot \delta m \), we are able to assess the curvature of the misfit function at a particular “point” in the model space, reflecting the degree of “blurring” of that point. Since we have to calculate the misfit gradient \( g(m + \delta m) \) for the perturbed model and we already have the gradient \( g(m) \) for the current model, the computational requirements for a single spot analysis are the same as one full iteration. To perform this test numerically, we employ a \(-1\%\) 3D Gaussian model perturbation with a \(~120 \text{ km}\) half width in Figures 3.19, 3.20 and 3.21 for models \( \text{EU}_{30} \), \( \text{EU}_{50} \), and \( \text{EU}_{60} \). Experiments with center differencing rather than forward differencing or with smaller (positive or negative) perturbations lead to comparable results and conclusions.

In Figure 3.19, \( \beta_v \) is perturbed with respect to elastic model \( \text{EU}_{30} \) at a depth of 480 km beneath eastern Europe, while the remaining three model parameters, \( \beta_h \), \( c \) and \( \eta \), are held fixed. Although there is some smearing, the main features of the 3D Gaussian perturbation are preserved in the “Hessian kernel”, \( H \cdot \delta m \), thereby confirming image quality at this location. By comparing \( H \cdot \delta m \) for \( \beta_v \) and \( \beta_h \), we conclude that there is limited tradeoff between the two shear wavespeeds in our tomographic images.

In Figure 3.20, we perturb anelastic attenuation \( Q^{-1} \) with respect to anelastic model \( \text{EU}_{50} \) at the same location as in Figure 3.19, while keeping the other parameters fixed. There is modest smearing, while the tradeoff between elastic and anelastic model parameters is relatively weak. The same behavior is observed when we perturb the azimuthally anisotropic parameter \( G_c \) with respect to the final anisotropic model \( \text{EU}_{60} \) in Figure 3.21.
Figure 3.19: Resolution analysis beneath eastern Europe after the Stage I elastic inversion. (a) and (c) Model perturbations in $\beta_v$ and $\beta_h$ with respect to elastic model EU$_{30}$ at a depth of 480 km. (b) and (d) Corresponding “point-spread function” for $\beta_v$ and $\beta_h$, respectively. The half width of the Gaussian is 120 km.

(the depth of the model perturbation is changed from 480 km to 120 km, because only surface waves are used for mapping anisotropic heterogeneity). Based on the results shown in Figures 3.19, 3.20 and 3.21, we conclude that features imaged in EU$_{30}$, EU$_{50}$ and EU$_{60}$ are robust, and that tradeoffs between different model parameters are relatively weak.

Finally, we use the “approximate Hessian” —a scalar field— to assess ray coverage in this study (Luo et al., 2013). Figure 3.22 illustrates the depth dependence of the approximate Hessian for starting model EU$_{00}$. Within the upper mantle, we have very good ray coverage for the entire European continent and North Atlantic Ocean. At greater depths, e.g., 600 km, our dataset is still able to illuminate the European continent very well because of the incorporation of body waves.
Figure 3.20: Resolution analysis beneath eastern Europe after the Stage II anelastic inversion. (a), (c) and (e) Model perturbations in $\beta_v$, $\beta_h$ and $Q^{-1}$ with respect to anelastic model EU$_{50}$ at a depth of 480 km. (b), (d) and (f) Corresponding “point-spread function” with respect to $\beta_v$, $\beta_h$ and $Q^{-1}$.
Figure 3.21: Resolution analysis beneath eastern Europe after the Stage III anisotropic inversion. (a), (c), (e) and (g) Model perturbations in $L$, $N$, $G_c$ and $G_s$ with respect to anisotropic model EU$_{60}$ at a depth of 120 km. (b), (d), (f) and (h) Corresponding “point-spread function” with respect to $L$, $N$, $G_c$ and $G_s$. 

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Figure 3.22: Approximate Hessian for starting model EU\textsubscript{00} at various depths. This 3D map may be used to assess ray coverage. White lines denote global plate boundaries (Bird, 2003). Depths are denoted on the bottom.
3.9 Conclusions

We used a three-stage adjoint tomography strategy to estimate 3D variations in elastic wavespeeds, anelastic attenuation, and radial & azimuthal anisotropy beneath the European continent and North Atlantic Ocean. Elastic wavespeed variations in the final model, named $EU_{60}$, confirm most long-wavelength features determined in previous body- and surface-wave tomographic studies. However, numerous hitherto unidentified, small-scale structures are gradually revealed in our iterative inversion. These features are related to subducting slabs, slab detachments, roll-back and back-arc basins. Our images show an anti-correlation between shear wavespeed and anelastic attenuation at shallow depths, where both parameters are correlated with surface tectonic provinces, such as the EEC and the western Mediterranean. At greater depths, this anti-correlation becomes relatively weak, in agreement with conclusions from previous global attenuation tomography. Enhanced attenuation is observed within the mantle transition zone beneath the North Atlantic Ocean, which might be indicative of the presence of water in this region. Most parts of the European continent are characterized by a radially anisotropic parameter $\xi > 1$, indicating the presence of horizontal flow within the upper mantle. Beneath subduction zones, such as the Apennines and the Hellenic arc, radially anisotropic features with $\xi < 1$ are resolved at depths greater than 150 km, indicating predominantly vertical flow. Azimuthally anisotropic fabrics in $EU_{60}$ are well correlated with the tectonic evolution of Europe over the past hundreds of millions of years, identifying extension along the NAR and counter-clockwise rotation of the Anatolian Plate. Averaged radial peak-to-peak anisotropic strength profiles reveal different regimes of brittle-ductile deformation in lithospheric strength beneath oceans and continents, in agreement mineral physical experiments. Simultaneously analyzing these elastic, anelastic and anisotropic variations will improve our understanding of variations in temperature, water content, chemical composition as well as ancient and present deformation within the crust and upper mantle.
3.10 Acknowledgments

We acknowledge IRIS (www.iris.edu), ORFEUS (www.orfeus-eu.org) and the Kandilli Observatory (www.koeri.boun.edu.tr) for providing the data used in this study. We thank Daniel Stich for providing Spanish data, which help to illuminate Iberia. Numerical simulations for this article were performed on a Dell cluster built and maintained by the Princeton Institute for Computational Science & Engineering (PICSciE). Time series analysis was accomplished based on the Seismic Analysis Code (SAC; Goldstein et al. (2002)). All maps and cross sections were made with the Generic Mapping Tool (GMT; Wessel & Smith (1991)). The open source spectral-element software package SPECFEM3D_GLOBE and the seismic measurement software package FLEXWIN used for this article are freely available for download via the Computational Infrastructure for Geodynamics (CIG; geodynamics.org). This research is supported by NSF grants 1063057 and 1112906.
Chapter 4

Seismic wavesspeed images across the Iapetus and Tornquist Suture Zone

Note

This chapter was published as a paper entitled “Seismic wavespeed images across the Iapetus and Tornquist Suture Zone” by H. Zhu, E. Bozdağ, D. Peter and J. Tromp in Geophysical Research Letters, 2012.

4.1 Summary

Closures of the Iapetus Ocean and the Tornquist Sea lead to the collision of the paleocontinents of Laurentia, Baltica and Eastern Avalonia during the Caledonian orogeny. It has been speculated that relicts of these two closures may be preserved within the crust or upper mantle. Over the past decades, numerous wide-angle seismic profiles were gathered in northwestern Europe to search for related subsurface features. Although active source studies revealed detailed crustal structures across the Iapetus and Tornquist suture zones, there are relatively few clear three-dimensional upper mantle images beneath this region. We use a new European crust and upper mantle model, EU$_{30}$, determined based on continental scale, nonlinear adjoint tomography, to explore upper mantle structures across these
two suture zones. Model EU$_{30}$ reveals two fast anomalies within the upper mantle: one
dips in a northwesterly direction down to approximately 400 km beneath the North Sea,
and the other dips in a southwesterly direction down to nearly 250 km across the Tornquist
Suture Zone. In addition, we observe a “gap” between the lithospheres of Laurentia and
Eastern Avalonia across the Iapetus Suture Zone beneath the central British Isles. These
seismic images suggest that heterogeneity related to the closures of the Iapetus Ocean and
the Tornquist Sea have been preserved within the upper mantle over hundreds of millions
of years.

4.2 Introduction

During the Neoproterozoic and Paleozoic eras (\(\sim 600–400\) Ma), the Iapetus Ocean — re-
garded as a predecessor of the Atlantic Ocean — was a relatively wide ocean separating
three paleocontinents, namely Laurentia, Baltica and Eastern Avalonia (Abramovitz et al.
, 1999). Between Baltica and Eastern Avalonia there was a relatively narrow waterway
named the Tornquist Sea (Cocks & Fortey, 1982). These three paleocontinents collided
during the Caledonian orogeny (\(\sim 440\) Ma) and formed Laurussia. This collision lead to
the closures of the Iapetus Ocean and the Tornquist Sea, and resulted in the Iapetus Suture
Zone (ISZ) and the Tornquist Suture Zone (TSZ) (Abramovitz et al. , 1999). In northwestern
Europe, the ISZ has been identified beneath Ireland, the British Isles, and the North
Sea. The TSZ, which is also known as the Trans-European Suture Zone, has mainly been
found beneath northern Germany and southern Scandinavia, consisting of the Sorgenfrei-
Tornquist Zone in the northwest and the Teisseyre-Tornquist Zone in the southeast (Shomali
et al. , 2002, 2006) (see Fig. 4.1).
Numerous wide-angle active source experiments were designed to investigate seismic structures beneath these two ancient suture zones, such as CSSP (Bott et al., 1985), BIRPS (Klemperer & Matthews, 1987), VARNET-96 (Landes et al., 2000), MONA LISA (Abramovitz et al., 1998, 1999; Abramovitz & Thybo, 2000; Lyngsie & Thybo, 2007), and the TOR array (Arlitt et al., 1999; Shomali et al., 2002; Gregersen et al., 2002; Shomali et al., 2006). Using reflection data from the BIRPS experiment, Klemperer & Matthews (1987) found prominent crustal reflectors dipping in a NNW direction across the ISZ beneath the central British Isles. The VARNET-96 and MONA LISA profiles were used to identify significant differences in crustal structure across the ISZ beneath Ireland, as well as across the TSZ beneath the eastern North Sea (Abramovitz et al., 1998, 1999; Abramovitz & Thybo, 2000). Based on an S waveform inversion, Zielhuis & Nolet (1994a) found a strong upper mantle contrast in shear wavespeed across the TSZ, which is difficult to observe from previous active source experiments. Shomali et al. (2002, 2006) used teleseismic body-wave tomography based on TOR array data to investigate 2D P and S wavespeed structures across the TSZ down to nearly 300 km depth. Their model reveals a
sharp increase in lithospheric thickness from northern Germany to southern Sweden. They also found a fast wavespeed body dipping in a southwesterly direction and extending down to approximately 200 km beneath northern Germany. The location of this fast wavespeed body coincides with the surface expression of the Elbe Lineament. Medhus et al. (2012) integrated several seismic datasets recorded in northwestern Europe to image upper mantle structure beneath the TSZ and southern Scandinavia. This traveltime tomographic study provided the first targeted upper mantle image beneath this area. However, their vertical resolution is insufficient to reveal detailed structures.

Over the past decade, numerous traveltime tomographic images of the European continent have been produced (e.g., Bijwaard & Spakman, 2000; Piromallo & Morelli, 2003; Lebedev & van der Hilst, 2008). However, to date, there have been relatively few studies focused on 3D upper mantle images across the ISZ and TSZ. Previous seismic studies were unable to provide such images either because they mainly focused on crustal structure based on active source experiments, or because they were only able to resolve 2D cross sections beneath a local, linear array.

Here, we explore upper mantle structures beneath the ISZ and the TSZ based on a new 3D European crust and upper mantle model, called EU30, derived based on a nonlinear iterative inversion (Zhu et al., 2012). Several horizontal and vertical cross sections of relative perturbations in vertically propagating and horizontally polarized shear wavespeed, $\delta \ln \beta_v$, are used to investigate upper mantle structure associated with the closures of the Iapetus Ocean and the Tornquist Sea.

### 4.3 Data and Method

EU30 is a new European crust and upper mantle model derived from continental-scale adjoint tomography (Zhu et al., 2012). Data from 190 earthquakes recorded by 745 seismographic stations were employed in this study. Three-component seismograms were fully
exploited in the inversion, resulting in 26,581 useable seismograms and more than 123,205 frequency-dependent traveltime measurements. In the inversion, short-period body waves and long-period surface waves are combined to constrain seismic structures.

Adjoint tomography is an iterative tomographic procedure which utilizes 3D forward simulations and gradient calculations (Tromp et al., 2005; Tape et al., 2007). Its benefits include: accurate calculation of synthetic seismograms in fully 3D models, numerical computation of Fréchet derivatives in 3D models, incorporation of finite-frequency effects (Dahlen et al., 2000; Hung et al., 2000), avoidance of “crustal corrections” for imaging the mantle, and full exploitation of three-component seismograms. In the inversion, a spectral-element method (Komatitsch & Tromp, 1999, 2002a,b) is used to compute 3D synthetic seismograms, and an adjoint method (Tarantola, 1984; Tromp et al., 2005; Liu & Tromp, 2006, 2008) is used to calculate 3D misfit gradients. Based on a preconditioned conjugate gradient method (Fletcher & Reeves, 1964), we are able to iteratively update the model and progressively minimize discrepancies between observed and simulated seismograms. Adjoint tomography has been successively used to image crustal structure in southern California (Tape et al., 2009, 2010), upper mantle structure beneath Australia (Fichtner et al., 2009, 2010), and crust and upper mantle structure beneath Europe (Zhu et al., 2012). In this article, we focus on 3D structures across the ISZ and TSZ in new European upper mantle model EU$_{30}$.

### 4.4 3D shear wavespeed structure

In Fig. 4.2, we present relative perturbations in $\beta_v$ for model EU$_{30}$ at various depths, ranging from 75 km to 325 km. Spherically symmetric Earth model STW105 (Kustowski et al., 2008a) is chosen as the reference model to calculate the relative perturbations. The Sorgenfrei-Tornquist Suture Zone and the Teisseyre-Tornquist Suture Zone (Shomali et al., 2002, 2006) comprise a striking tectonic feature which separates the Precambrian East
European Craton and Phanerozoic central and western Europe (Zielhuis & Nolet, 1994a,b; Artemieva et al., 2006). In EU$_{30}$, this prominent contrast extends from 75 km down to nearly 250 km. Beneath southern Scandinavia, in the depth range from 75 km to 225 km, we detect a sharp transition between fast and slow anomalies, which was previously discovered by Medhus et al. (2012). The Danish Basin and Northern German Basin are separated by a linear fast anomaly from 75 km to 225 km. At 75 km depth, there are several small-scale features beneath central and western Europe, such as a fast anomaly associated with the Armorican Massif, and slow anomalies related to the Eifel and Harz hotspots, the Bohemian Massif, and the Central Slovakia Volcanic Fields.

Beneath the North Sea, model EU$_{30}$ contains a strong fast anomaly down to depths of nearly 175 km. This fast and presumably cold anomaly may be the cause of the Central Graben. The absence of fast anomalies beneath southwestern Norway and Denmark at depths ranging from 75 km to 225 km may explain their elevated expression. Around 60°N and between 0°E and 5°E, we observe a fast anomaly at depths between 175 km and 225 km, which we interpret as a subsurface structure related to the Viking Graben. At 75 km depth around 58°N and between 7.5°E and 10°E, there is an elongated fast anomaly connecting fast anomalies beneath the North Sea and the Baltic Shield. This structure coincides with the locations of the Skagerrak and Oslo grabens. A relatively small and weak slow anomaly is found beneath the southern coast of Norway between depths of 275 km and 325 km, which may be a result of the rift of the Oslo Graben (Heeremans et al., 1996). In addition, an interesting NW-SE trending fast anomaly, extending from northern England towards northern Germany, is revealed in the depth range from 175 km down to nearly 325 km. We interpret this feature as the subsurface expression of the Elbe lineament (Shomali et al., 2002, 2006; Lyngsie et al., 2006; Lyngsie & Thybo, 2007).

In Fig. 4.3, three vertical cross sections are used to further illustrate upper mantle structures across the ISZ and TSZ. In vertical cross section A-a across the TSZ between northern Germany and southern Sweden, we find a fast anomaly dipping towards the southwest
Figure 4.2: Relative perturbations in the wavespeed of vertically propagating and horizontally polarized shear waves, $\delta \ln \beta_v$, for model EU$_{30}$ at various depths ranging from 75 km to 325 km. Spherically symmetric model STW105 (Kustowski et al., 2008a) is used as a reference model. Blue and red colors denote fast and slow wavespeed anomalies, respectively.
Figure 4.3: Relative perturbations in the wavespeed of vertically propagating and horizontally polarized shear waves, $\delta \ln \beta_v$, in three vertical cross sections across the ISZ and the TSZ. A. Map locations of three vertical cross sections; black bars denote locations of interest in (B)–(D). B. Relative wavespeed perturbations in vertical cross section a-A. C. Relative wavespeed perturbations in vertical cross section b-B. D. Relative wavespeed perturbations in vertical cross section c-C.
down to depths of nearly 200 km (see Fig. 4.3 B). The location of this fast anomaly coincides with the Elbe lineament on the surface (Shomali et al., 2002, 2006). In agreement with body-wave tomographic results based on TOR array data (Shomali et al., 2002, 2006), the lithosphere of EU\textsubscript{30} is relatively thin (∼100 km) beneath northern Germany. Within a very short horizontal distance, the lithosphere becomes much thicker (nearly 200 km–300 km) beneath southern Sweden. In the image from Shomali et al. (2002, 2006), the fast wavespeed body is detached from the surface. However, in our image, the SW dipping fast anomaly is continuous from the surface down to 200 km depth. Furthermore, our image for the A-a cross section is consistent with tomographic results from Amaru (2007). At ∼52°N, the fast anomaly within the transition zone beneath northern Germany may be a detachment from the fast lithospheric anomaly at 200 km depth. From approximately 250 km to 450 km, between 54°N and 58°N, there is an interesting fast anomaly laying flat at the 410 km discontinuity, which may be delaminated lithosphere beneath the thick and old Baltic Shield.

In vertical cross section B-b across the ISZ beneath the southern North Sea, we observe a prominent fast anomaly dipping towards the northwest down to nearly 400 km depth (see Fig. 4.3 C). Although the location of this anomaly is a little to the southeast of the geologically speculated location of the ISZ, the dip direction of the fast anomaly agrees with such speculations. In addition, the direction of dip of the fast anomaly in EU\textsubscript{30} is similar to the orientation of crustal reflectors imaged based on 2D active source experiments (Klemperer & Matthews, 1987; Abramovitz et al., 1999). Beneath the northern part of cross section B-b, EU\textsubscript{30} contains a nearly 120 km thick lithosphere. The slow anomaly in the southern part of the cross section is correlated to the Rhine Graben in western Germany and Netherlands.

In vertical cross section C-c across the central British Isles, we find a “gap” in lithosphere between the northern and southern British Isles (see Fig. 4.3 D). The location of this “gap” coincides with the geological exposure of the ISZ in the British Isles. The lithosphere of southern England is relatively thick, nearly 200–300 km. The thickest part
Figure 4.4: Resolution test for image quality across the TSZ. A and C are vertical and horizontal cross sections through a Gaussian perturbation in $\beta_v$ with respect to model $EU_{30}$. The dashed white line in C indicates the location of vertical cross section in A. B and D are vertical and horizontal cross sections of the point-spread function for $\beta_v$. E and G are vertical and horizontal cross sections for perturbations in $\beta_h$, which are absent, that is, we are only perturbing $\beta_v$. F and H are point-spread functions for $\beta_h$, illustrating limited tradeoff between $\beta_v$ and $\beta_h$. 
extends down to more than 300 km depth, for instance between 51°N and 53°N. Previous wide-angle reflection studies found a significant contrast in crustal structure across the ISZ (Abramovitz et al., 1998, 1999; Abramovitz & Thybo, 2000). Based on our image, the structural contrast between Laurentia and Eastern Avalonia extends through the crust and into the lithosphere. It is interesting to note that such a contrast can, apparently, be preserved within the upper mantle over hundreds of millions of years.

In order to assess the quality of these images, we perform a “point-spread function” test (Fichtner & Trampert, 2011, 2012). A Gaussian anomaly with a half width of 60 km and a 2% shear wavespeed reduction is used to perturb model EU_{30} beneath the TSZ, as illustrated in Fig. 4.4. Although there is some smearing across the TSZ, its main features are preserved and well-localized. By comparing the results for vertically and horizontally propagating and horizontally polarized shear wavespeeds, $\beta_v$ and $\beta_h$, respectively, we are confident that there is limited tradeoff between these two parameters. This resolution test confirms the quality of the images used in our paleotectonic interpretation of model EU_{30}.

4.5 Discussion and Conclusion

It is fascinating that we are still able to observe several prominent upper mantle wavespeed anomalies—related to the closure of the Iapetus Ocean and the Tornquist Sea—despite the fact that after the Caledonian orogeny the area was affected by a host of subsequent tectonic events, such as the Variscan and Alpine orogenies. Compared to previous seismic studies based on wide-angle active source experiments, model EU_{30} enables us to explore much deeper structures beneath the ISZ and TSZ. Furthermore, seismic studies based on dense local arrays, such as the TOR array (Shomali et al., 2002, 2006), are unable to constrain the 3D geometry of these two suture zones, and the current resolution of global tomography is insufficiently detailed.

By exploring European crust and upper mantle model EU_{30} beneath northwestern Eu-
rope, we observe two fast anomalies across the ISZ and TSZ. One dips towards the NW and extends down to nearly 400 km beneath the North Sea, and the other dips towards the SW and extends down to nearly 200 km beneath the TSZ. These characteristics agree with geological speculation with regards to the closings of the Iapetus Ocean and the Tornquist Sea. A resolution test across the TSZ confirms the image quality of this feature. In addition, across the ISZ beneath the central British Isles, we find a “gap” between the lithospheres of Laurentia and Eastern Avalonia, which is consistent with a previously observed contrast in crustal structure across this suture zone. Understanding how seismic features capturing the closure of the Iapetus Ocean and Tornquist Sea are preserved within the upper mantle for such a long time requires detailed future geodynamic and tectonic modeling.

4.6 Acknowledgments

This research was supported by the US National Science Foundation under grants EAR-1112906 and OCI-1063057. We thank Jerry X. Mitrovica and an anonymous reviewer for constructive reviews which helped to improve an earlier version of the manuscript.
Chapter 5

Seismic attenuation beneath Europe and the North Atlantic: Implications for water in the mantle

Note

5.1 Summary

It is well known that anelasticity has significant effects on the propagation of seismic waves, as manifested by physical dispersion and dissipation. Investigations of anelasticity provide complementary constraints on the physical properties of Earth materials, but — contrary to imaging with elastic waves — progress in mapping Earth’s anelasticity has been relatively slow, and there is only limited agreement between different studies or methodologies. Here, within the framework of adjoint tomography, we use frequency-dependent phase and
amplitude anomalies between observed and simulated seismograms to simultaneously con-
strain upper mantle wavespeeds and attenuation beneath the European continent and the
North Atlantic Ocean. In the sea-floor spreading environment beneath the North Atlantic,
we find enhanced attenuation in the asthenosphere and within the mantle transition zone
(MTZ). In subduction zone settings, for example beneath the Hellenic arc, elevated atten-
uation is observed along the top of the subducting slab down to the MTZ. No prominent
reductions in wavespeeds are correlated with these distinct attenuation features, suggesting
that non-thermal effects may play an important role in these environments. A plausible
explanation invokes the transport of water into the deep Earth by relatively cold subducting
slabs, leading to a hydrated MTZ, as previously suggested by mineral physics and geo-
dynamics studies.

5.2 Introduction

Current knowledge of heterogeneities in the Earth’s mantle primarily comes from tomo-
graphic studies of elastic wavespeeds. Based on traveltimes of body waves, dispersion of
surface waves, and splitting of free oscillations, seismologists routinely estimate lateral
variations in elastic wavespeeds within the Earth’s interior (e.g., Woodhouse & Dziewon-
ski, 1984; Van der Hilst et al., 1997; Romanowicz, 2003; Montelli et al., 2004). However,
Earth materials exhibit anelasticity, an energy dissipation mechanism that manifests itself
in the form of physical dispersion and attenuation of seismic waves (Liu et al., 1976).
3D maps of lateral variations in anelastic attenuation provide complementary constraints
on variations in temperature, water content, partial melt, and composition (Karato, 2003).
For instance, guided by mineral physics experiments, Shito et al. (2006) combined tomo-
geraphic images of elastic wavespeeds and anelastic attenuation to estimate 3D variations
in temperature, water content and other parameters, such as major element chemistry and
melt fraction.
There is no consensus among seismologists on how the effects of attenuation should be quantified and measured. Contrary to wavespeed models, existing global models of attenuation exhibit limited agreement, e.g., Romanowicz (1995), Gung & Romanowicz (2004), Dalton et al. (2008). In contrast to body-wave traveltimes or surface-wave dispersion, which are solely governed by seismic wavespeeds, seismic wave amplitudes are affected by a host of competing factors besides anelastic attenuation, such as earthquake magnitudes and radiation patterns, elastic focusing and defocusing, and scattering (Ruan & Zhou, 2010, 2012).

We have developed a new tomographic technique, called “adjoint tomography” (Tromp et al., 2005; Tape et al., 2007; Liu & Tromp, 2008; Tape et al., 2009, 2010; Zhu et al., 2012), to simultaneously constrain elastic wavespeeds and anelastic attenuation. Frequency-dependent phase and amplitude differences between observed and simulated seismograms (Laske & Masters, 1996a; Ekström et al., 1997; Zhou et al., 2004) are simultaneously considered in the inversion in order to ensure a consistent treatment between anelastic attenuation and elastic focusing/defocusing (Billien et al., 2000). Synthetic seismograms are computed based on a spectral-element method (Komatitsch & Tromp, 1999; Peter et al., 2011), and Fréchet derivatives with respect to the model parameters are numerically calculated in a 3D background model based on adjoint methods (Lailly, 1983; Tarantola, 1984; Tromp et al., 2005; Liu & Tromp, 2008). Body and surface waves recorded in three-component seismograms are combined to simultaneously constrain deep and shallow upper mantle structures (Zhu et al., 2012). Based on a preconditioned conjugate gradient method (Fletcher & Reeves, 1964), we iteratively improve the elastic and anelastic models and gradually reduce the phase and amplitude differences.
5.3 Dataset and method

5.3.1 Starting model

Dataset used in this study can be found in Section 2.3. A new 3D crust and upper mantle model of Europe and the North Atlantic Ocean, EU30 (Zhu et al., 2012), is used as the starting elastic model. EU30 was constructed based on adjoint tomography (Tape et al., 2009, 2010; Zhu et al., 2012). Three-component body and surface waves were combined to constrain radially anisotropic shear wavespeeds throughout the European upper mantle. Thirty preconditioned conjugate gradient iterations (Fletcher & Reeves, 1964) were performed to minimize frequency-dependent phase differences between observed and simulated seismograms, requiring more than 17,100 wavefield simulations and 2.3 million central processing unit core hours. Fig. 5.1c and d illustrate relative perturbations in vertically and horizontally polarized shear wavespeeds in EU30 at a depth of 75 km. In addition, the 1D (radial) shear quality factor $Q_\mu$ profile from reference model STW105 (Kustowski et al., 2008a) is chosen as the starting anelastic model (Fig. 5.1b). Since the bulk quality factor $Q_\kappa$ is much larger than the shear quality factor $Q_\mu$ (Durek & Ekström, 1996), only shear attenuation is considered in this paper. For brevity, in the rest of this paper, we use the symbol $Q$, rather than $Q_\mu$, to denote the shear quality factor.

Choice of misfit function, model parameterization and source correction can be found in Section 2.8.1, Section 2.8.2 and Section 2.7, respectively. Figure 5.2 and Figure 5.3 illustrate improvements in amplitude and phase histograms after source corrections.

5.3.2 Improvements in misfits and histograms

Twenty preconditioned conjugate gradient iterations are performed to obtain a new crust and upper mantle model, named EU50, which required more than 18,050 wavefield simulations and 2.5 million central processing unit core hours. Both phase and amplitude misfits are significantly reduced after these iterations. In Figure 5.4, the overall amplitude misfit
Figure 5.1: Elastic and anelastic structure of starting model EU$_{30}$. a. 1D shear wavespeed profiles of STW105 (Kustowski et al., 2008a). Black and red lines denote vertically ($\beta_v$) and horizontally ($\beta_h$) polarized shear wavespeeds, respectively, which are used as a reference model to compute relative perturbations in c and d. b. 1D shear $Q$ profile of STW105. c. Relative perturbations in vertically polarized shear wavespeed $\beta_v$ in starting model EU$_{30}$ at 75 km depth. d. Relative perturbations in horizontally polarized shear wavespeed $\beta_h$ in EU$_{30}$ at 75 km depth. White lines in c and d denote plate boundaries (Bird, 2003).
Figure 5.2: Comparisons of amplitude histograms before and after source correction for the whole dataset. a-c. Amplitude histograms for P-SV waves on vertical (a) and radial (b) components, SH waves on transverse components (c). d-f. Amplitude histograms for Rayleigh waves on vertical (d) and radial (e) components, Love waves on transverse components (f). Red and blue histograms denote the amplitude fits for the starting model EU30 and the model after the source correction, respectively. 15–40 s body waves and 25–100 s surface waves are used. The source corrections slightly improve the mean values of the distributions, but they do not improve the standard deviations.

Figure 5.3: Same as Figure 5.2 except for comparisons of phase histograms before and after source correction.
as well as the contributions from the six body- and surface-wave categories are gradually reduced. In the structural inversion, we start with long-period measurements for surface waves, and short-period information is gradually incorporated, e.g., the short-period corner for surface waves is gradually reduced from 40 s to 25 s. This strategy allows us to first resolve large-scale features. With gradual improvements in models and simulated seismograms, shorter-period measurements are included to constrain smaller-scale features. Meanwhile, phase misfits are also gradually reduced during the iterations (see Figure 5.5).

Figure 5.6 compares histograms of amplitude anomalies for the starting model (after the source correction) and the new model, EU50. Both the average values and standard deviations of amplitude anomalies are considerably reduced in all six categories. As shown in Figure 5.7, the phase anomalies of the new model are also significantly reduced compared with the starting model.

## 5.4 Tomographic images and discussion

### 5.4.1 Horizontal cross sections

Fig. 5.8 presents horizontal cross sections of isotropic shear wavespeed ($\beta$) and shear attenuation ($Q^{-1}$) in model EU$_{50}$ at depths of 100 km and 600 km. Relative perturbations in shear wavespeed are calculated with respect to 1D reference model STW105 (Fig. 5.1a). At 100 km depth, the isotropic shear wavespeed image contains numerous interesting features correlated with well-known tectonic provinces (Figs. 5.8a and 3.1). For instance, we observe a sharp contrast associated with the Tornquist-Teisseyre Suture Zone, separating the Precambrian East European Platform from Phanerozoic central and eastern Europe. The Tyrrenhenian Sea and the Anatolian Plate are revealed as two prominent slower-than-average wavespeed anomalies. In central and eastern Europe, there are several notable slow wavespeed anomalies associated with the Massif Central, the Rhine Graben, the Eifel Hotspot, the Bohemian Massif as well as the Pannonian Basin. At this shallow depth, most
Figure 5.4: Evolution of amplitude misfits as defined in eqn.(3) in the main text. a. Evolution of the total amplitude misfit. b–d. Evolution of the amplitude misfit for P-SV body waves on vertical (b) and radial (c) components, and SH body waves on transverse components (d). e–g. Evolution of the amplitude misfit for Rayleigh surface waves on vertical (e) and radial (f) components, and Love surface waves on transverse components (g). The corner period of the surface-wave measurements is gradually reduced from 40 s to 25 s.
Figure 5.5: Same as Figure 5.4 except for the evolution of the phase misfits as defined in eqn.(2) in the main text.
Figure 5.6: Comparisons of amplitude histograms before and after the anelastic inversion. The anelastic model improves both the means and the standard deviations of the distributions. a–c. Amplitude histograms for P-SV waves on vertical (a) and radial (b) components, SH waves on transverse components (c). d–f. Amplitude histograms for Rayleigh waves on vertical (d) and radial (e) components, Love waves on transverse components (f). Red and blue histograms denote the amplitude fits for the starting model $\text{EU}_{30}$ (after source correction) and the new model $\text{EU}_{50}$, respectively.

Figure 5.7: Same as Figure 5.6 except for comparisons of phase histograms between the starting model (after source correction) and the new model.
regions show an interesting relation between shear wavespeed and attenuation (Fig. 5.8b). For example, the East European Platform is characterized by fast shear wavespeed and low attenuation while the Tyrrhenian Sea is characterized by slow shear wavespeed and high attenuation. These features suggest that the same factors, e.g., temperature, control both elastic wavespeeds and anelastic attenuation at shallow depths (Billien et al., 2000; Dalton et al., 2008). In addition, some areas, e.g., the Anatolian Plate, involve more complicated relations between shear wavespeed and attenuation.

At a depth of 600 km, central and eastern Europe are dominated by two prominent faster-than-average wavespeed anomalies (Fig. 5.8c). These features might be related to remnants of slab roll-back due to trench migration over the past 30 million years, a process responsible for the current Apennines-Calabrian-Maghrebides, Carpathian-Vrancea, and Hellenic-Cyprus arcs (Wortel & Spakman, 2000). Our isotropic shear wavespeed images are consistent with previous traveltime-based body-wave tomographic studies (Wortel & Spakman, 2000; Piromallo & Morelli, 2003). However, as illustrated in Fig. 5.8d, there are no strong lower-than-average attenuation anomalies which correspond to slab remnants at a depth of 600 km, indicating that different factors control elastic wavespeeds and anelastic attenuation at greater depths. Previous global-scale attenuation studies also demonstrated that correlations between attenuation and shear wavespeed decrease significantly below depths of 200–250 km (Billien et al., 2000; Gung & Romanowicz, 2004; Dalton et al., 2008).

5.4.2 The MTZ beneath the North Atlantic Ocean

Model EU50 contains several notable features in shear attenuation beneath the North Atlantic Ocean. A number of vertical cross sections perpendicular to the northern Mid-Atlantic Ridge are used to illustrate these features in Fig. 5.9. At shallow depths, the oceanic and continental lithosphere are characterized by low attenuation, e.g., in N3, N4, and N5. At depths ranging from 80 km to 200 km, we observe a prominent enhanced
Figure 5.8: Horizontal cross sections of isotropic shear wavespeed (left column) and shear attenuation (right column) for model EU\textsubscript{50}. a and c. Relative perturbations in isotropic shear wavespeed at depths of 100 km and 600 km, respectively. b and d. Shear attenuation at depths of 100 km and 600 km, respectively. White lines indicate plate boundaries (Bird, 2003).
attenuation layer associated with the oceanic asthenosphere. The most striking feature of shear attenuation in these vertical cross sections is enhanced attenuation within the MTZ beneath the North Atlantic Ocean, for instance in cross sections N2 and N4. Interestingly, there are no strong reductions in shear wavespeed related to the highly attenuating MTZ (see Figure 5.10), suggesting that non-thermal effects, such as water content, grain size, or compositional variations, may play important roles in this environment.

There is growing evidence for enhanced water content in the transition zone. The major minerals of this region, wadsleyite and ringwoodite, have water solubilities that are an order of magnitude larger than for minerals in the shallow upper mantle (Kohlstedt et al., 1996). Electrical conductivity measurements are consistent with higher water content in the transition zone compared to the rest of the upper mantle (Karato, 2011). The transition zone water filter hypothesis was proposed to explain discrepancies between geophysical and geochemical observations (Bercovic & Karato, 2003). A thin melt layer on top of the 410 km discontinuity, an important prediction of this hypothesis, was found beneath the western United States by both seismological and electromagnetic studies (Song et al., 2004; Coutier & Revenaugh, 2007; Tauzin et al., 2010), indicating that a hydrated MTZ might exist, at least on regional scales.

The presence of small quantities of water (∼0.1 wt%) is expected to have a modest effect on seismic wavespeeds but may have a strong effect on attenuation (Shito et al., 2006; Karato, 2011). Thus, elevated attenuation in the absence of marked reductions in shear wavespeeds in the vertical cross sections shown in Fig. 5.9 might be indicative of a water-enriched MTZ beneath the North Atlantic Ocean. Furthermore, a hydrous MTZ has direct impact on the character of upper mantle discontinuities (Meijde et al., 2003). A receiver function study in the northern North Sea observed a significant weakening of the 660 km discontinuity (Helffrich et al., 2003), which was interpreted to be a consequence of a hydrated MTZ beneath the North Atlantic Ocean, an interpretation that is consistent with our images. Using body waveforms, Fuji et al. (2010) found a low shear quality
Figure 5.9: Vertical cross sections of shear attenuation beneath the North Atlantic. a. Locations of vertical cross sections in b–g and Fig. 5.11. b–g. Shear attenuation for vertical cross sections N1–N6, perpendicular to the northern Mid-Atlantic Ridge. The dashed black lines in b–g denote the 220 km, 410 km and 660 km discontinuities.
Figure 5.10: Vertical cross sections of relative perturbations in isotropic shear wavespeed perpendicular to the North Atlantic Ridge (see Fig. 5 in the main text).
factor within the MTZ beneath the Northwestern Pacific, reminiscent of our vertical cross sections beneath the North Atlantic Ocean. Interestingly, 1D radial Q model QL10 (Durek & Ekström, 1996) shows a slightly reduced shear quality factor in the transition zone, whereas the more coarsely parameterized 1D model QL6 (Durek & Ekström, 1996) does not. We speculate that the water-enriched MTZ beneath the North Atlantic Ocean might be due to ancient subduction related to the closure of the Iapetus Ocean during the Caledonian Orogeny. This interpretation, involving water transport during subduction, is corroborated by our wavespeeds and attenuation images of present-day subduction in the Hellenic and Apennines-Maghrebides arcs, as we discuss next.

5.4.3 Subduction Zones

Over the past several decades, relationships between subducting slabs and Earth’s deep water cycle have been investigated based on observations from seismology, mineral physics, and geodynamics (Karato, 2003; Jacobsen & van der Lee, 2006; Maruyama & Okamoto, 2007; Iwamori, 2007; Zhao et al., 2009). However, questions related to how much and how deep subducting slabs can transport water down into the mantle remain largely unresolved (Karato, 2003). Here, we examine relationships between isotropic shear wavespeeds, compressional-to-shear wavespeed ratios (\(V_p/V_S\)), and attenuation based on model EU\textsubscript{50} for two subduction zones in central and eastern Europe: the Hellenic and Apennines-Maghrebides arcs (Fig. 5.11). As illustrated in Figs. 5.11a, b, d and e, these subducting slabs are characterized by strong fast shear wavespeeds and low \(V_p/V_S\) ratios. The Hellenic slab penetrates the 660 km discontinuity down into the lower mantle while slabs associated with the Apennines-Maghrebides arc are stagnant within the MTZ, in agreement with previous teleseismic body-wave studies (Wortel & Spakman, 2000; Piromallo & Morelli, 2003). At depths shallower than 150–200 km, we observe reductions in shear wavespeed and elevated \(V_p/V_S\) ratios overlying the subducting plate. These anomalies might be related to partial melt associated with dehydration of subducting oceanic litho-
Figure 5.11: Vertical cross sections of the Hellenic arc (top row) and the Apennines-Maghrebides arc (bottom row). a-c. Relative perturbations in isotropic shear wavespeed $\delta \ln \beta$, $V_P/V_S$ ratio, and shear attenuation ($Q^{-1}$) of the Hellenic subduction zone. The location of cross section A-a is shown in Fig. 5.9a. d–f. Structures of the Apennines-Maghrebides subduction zone. The location of cross section B-b is shown in Fig. 5.9a. White lines outline the seismically fast subducting slabs.

Several attenuation studies have been used to investigate anelastic structure in subduction zones, e.g., Stachnik et al. (2004), Wiens et al. (2008), Rychert et al. (2008). In Figs. 5.11c and f, we observe channels of enhanced attenuation along the top of the seismically fast slabs down to the MTZ beneath the Hellenic and Apennines-Maghrebides arcs. Similar to the highly attenuating MTZ beneath the North Atlantic Ocean, there are no obvious corresponding reductions in shear wavespeed (Figs. 5.11a and d). Petrological and geodynamical models show that a significant amount of water can be carried down into the deep upper mantle by slabs if their thermal gradients are low or subduction speeds fast (Maruyama & Okamoto, 2007; Iwamori, 2007). Minerals such as lawsonite, phengite, and phase A or E may be responsible for transport of water in the slab into the MTZ (Maruyama & Okamoto, 2007), where it can be incorporated into nominally anhy-
drous phases such as wadsleyite and ringwoodite. This high-attenuation channel overlying the subducting slab might be a complementary seismic indicator for the presence of highly water-soluble minerals in descending slabs.

Within the MTZ beneath these regions of subduction, portions of the slab are characterized by higher-than-average attenuation (Figs. 5.11c and f). Previous attenuation studies of subduction zones beneath Japan and south America demonstrated that shear $Q$-values of slabs within the MTZ might be as low as values within the asthenosphere (Sacks & Okada, 1974). Furthermore, global attenuation tomography, constrained by Rayleigh-wave phase and amplitude anomalies, has identified anomalies with high attenuation surrounding the Pacific. These high-attenuation features correlate with the locations of subducting slabs at MTZ depths (Billien et al., 2000). One plausible explanation for these higher-than-average attenuation anomalies is the presence of water transported down into the MTZ by subducting slabs. Such a scenario is consistent with speculation for a water-enriched MTZ beneath the Mediterranean region based on a receiver function study (Meijde et al., 2003). Enhanced attenuation extending into the top of the lower mantle (Fig. 5.11c) may reflect water (up to 0.1 wt%) carried into this region and incorporated in silicate perovskite (Lawrence & Wysession, 2006; Hernandez et al., 2013).

Electrical conductivity studies provide independent constraints on the amount and distribution of water within the deep mantle, because of its high sensitivity to the presence of hydrogen (Huang et al., 2005). A global 3D electromagnetic study has shown that high conductivity correlates well with cold subducting slabs around the circum-Pacific margin within the MTZ (Kelbert et al., 2009). Furthermore, an enhanced conductivity layer was imaged along the top of the subducting Nazca slab down to depths in excess of 250 km (Booker et al., 2004). These electromagnetic anomalies were attributed to the presence of water or water-induced partial melt within the deep Earth, in agreement with our interpretation of enhanced shear attenuation overlying subducting slabs as well as within the MTZ beneath the North Atlantic.
5.4.4 Resolution Tests

The “approximate Hessian”, $\tilde{H}$, is a good proxy for ray density and resolution. Fig. 5.12 illustrates the approximate Hessian in vertical cross sections perpendicular to the northern Mid-Atlantic Ridge. We have good ray coverage in these vertical cross sections, as discussed in Section 5.4.2.

The “point spread function” is used to assess image quality in the final model (Fichtner & Trampert, 2011, 2012). This is used to assess the curvature of the misfit function at a particular “point” in the model space, and reflects the degree of “blurring” of that point, thus, the nomenclature “point spread function”.

A spherical Gaussian function in shear attenuation is used to perturb model $EU_{50}$ beneath eastern Europe (Fig. 5.13) and the North Atlantic (Figure 5.14). By comparing the resulting images for $\beta_v$, $\beta_h$ and $Q^{-1}$, we conclude that there is weak tradeoff between elastic wavespeeds and anelastic attenuation. However, the test beneath eastern Europe (Fig. 5.13) demonstrates that we cannot resolve whether or not regions of high attenuation are partially located inside the Hellenic slab or entirely outside of it. In Figure 5.15, we calculate the “point spread function” by perturbing $\beta_v$ for model $EU_{50}$; again we see good resolution and limited tradeoff with anelasticity. We conclude that elastic and anelastic structures in the MTZ beneath eastern Europe and the North Atlantic are reasonably well resolved.

5.5 Conclusions

We use adjoint tomography to construct an anelastic upper mantle model of the European continent and the North Atlantic, named $EU_{50}$. Frequency-dependent phase and amplitude measurements are combined to simultaneously determine elastic wavespeeds and anelastic attenuation. Short-period body waves (15–40 s) and long-period surface waves (25–100 s) recorded in three-component seismograms are utilized to constrain deep and shallow upper mantle structures. Twenty preconditioned conjugate gradient iterations are performed, re-
Figure 5.12: Images of the approximate Hessian in vertical cross sections perpendicular to the northern Mid-Atlantic Ridge (see Fig. 5.9).
Figure 5.13: Resolution analysis beneath eastern Europe. (a), (c) and (e) are horizontal cross section of model perturbation in $\beta_v$, $\beta_h$ and $Q^{-1}$ with respect to the final model EU$_{50}$ at 480 km depth. (b), (d) and (f) are corresponding “point spread function” for $\beta_v$, $\beta_h$ and $Q^{-1}$. The half width of Gaussian perturbation is 120 km.
Figure 5.14: Resolution analysis beneath the North Atlantic. The same setting as Fig. 5.13, except the half width of Gaussian perturbation is 180 km.
Figure 5.15: Resolution analysis beneath the North Atlantic. The same setting as Fig. 5.14 except here we only perturb $\beta_v$ while keeping $\beta_h$ and $Q^{-1}$ fixed.
quiring 18,050 wavefields simulations and 2.5 million central processing unit core hours. Both phase and amplitude differences between observed and simulated seismograms are significantly reduced after these iterations. Resolution analyses show limited smearing in the tomographic images and weak tradeoffs between elastic and anelastic model parameters.

We find that regions with fast shear wavespeed anomalies correspond to regions with low attenuation at shallow depths, e.g., 100 km. However, at greater depths, this correspondence vanishes. Enhanced attenuation anomalies are found within the asthenosphere and the MTZ beneath the North Atlantic Ocean, as well as along the top of the subducting slabs down to the MTZ beneath the Hellenic arc. There are no strong shear wavespeed reductions associated with these high attenuation anomalies, suggesting a non-thermal origin. A plausible interpretation invokes the transport of water into the deep Earth by cold subducting slabs, leading to a water-enriched MTZ, in agreement with a previous hypothesis proposed based on mineral physics and geodynamics investigations.

## 5.6 Acknowledgments

This research was supported by the NSF under grants 1063057 and 1112906. We thank Shun-ichiro Karato and an anonymous reviewer for constructive reviews which helped to improve an earlier version of the manuscript. Numerical simulations were performed on a Dell cluster built and maintained by the Princeton Institute for Computational Science and Engineering (PICSciE). The open source spectral-element software package SPECFEM3D GLOBE and the seismic measurement software package FLEXWIN used in this study are freely available for download via the Computational Infrastructure for Geodynamics (CIG; geodynamics.org).
Chapter 6

Mapping tectonic deformation in the crust and upper mantle beneath Europe and the North Atlantic Ocean

Note

This chapter was published as a paper entitled “Mapping tectonic deformation in the crust and upper mantle beneath Europe and the North Atlantic Ocean” by H. Zhu and J. Tromp in *Science*, 2013.

6.1 Summary

We constructed a 3D azimuthally anisotropic model of Europe and the North Atlantic Ocean based on adjoint seismic tomography. Several features are well correlated with historical tectonic events in this region, for example, extension along the North Atlantic Ridge, trench retreat in the Mediterranean, and counter-clockwise rotation of the Anatolian Plate. Beneath northeastern Europe, the direction of the fast anisotropic axis follows trends of ancient rift systems older than 350 million years, suggesting “frozen-in” anisotropy related
to the formation of the craton. Local anisotropic strength profiles identify the brittle-ductile transitions in lithospheric strength. In continental regions, these profiles also identify the lower crust, characterized by ductile flow. The observed anisotropic fabric is generally consistent with the current surface strain rate measured by geodetic surveys.

6.2 Introduction

Minerals that comprise the Earth’s crust and upper mantle, such as amphibole and olivine, are highly anisotropic in terms of seismic wavespeeds. Deformation induced by tectonic activity can align the fast axes of these minerals with the directions of flow or principle extension (Zhang & Karato, 1995). This process of lattice preferred orientation leads to a directional- and polarization-dependence of seismic wavespeeds. Seismic anisotropy has been observed within the crust, lithosphere and asthenosphere at various scales (Simons et al., 2002; Marone & Romanowicz, 2007; Yuan & Romanowicz, 2010; Lin et al., 2011; Endrun et al., 2011), providing important constraints on past and present deformation within Earth’s interior (Vinnik et al., 1992; Silver, 1996; Park & Levin, 2002).

The European continent has undergone extensive tectonic deformation since the Archaean, and was especially heavily reworked due to Meso-Cenozoic volcanism and subduction (Goes et al., 1999; Wortel & Spakman, 2000). Geodetic surveys have detected substantial deformation due to the convergence of the Eurasian, African and Arabian Plates (Kreemer et al., 2003). Most seismic studies in this area focus on mapping lateral heterogeneities in compressional (Wortel & Spakman, 2000) and shear wavespeeds (Schivardi & Morelli, 2011). Anisotropic images either come from investigations at the global scale (Debayle et al., 2005) or local studies limited to a relatively small region (Endrun et al., 2011).

Here, we present an azimuthally anisotropic model of Europe and the North Atlantic Ocean constructed based on adjoint tomography. We collected 26,581 three-component seismograms for 190 earthquakes recorded by 745 seismographic stations (see Section 2.3).
We used spectral-element (Komatitsch & Tromp, 1999) and adjoint methods (Tarantola, 1984; Tromp et al., 2005; Tape et al., 2009) to numerically calculate synthetic seismograms and Fréchet derivatives with respect to radial and azimuthal anisotropic model parameters (Smith & Dahlen, 1973; Montagner & Nataf, 1986) in fully 3D Earth models (see Section 2.9). Frequency-dependent phase and amplitude measurements of three-component surface waves with periods between 25 s to 100 s are used to constrain structure at depths shallower than 200–250 km (see Section 2.9.1). A new anisotropic model of Europe and the North Atlantic Ocean, namely EU₆₀, is obtained after ten preconditioned conjugate gradient iterations (Fletcher & Reeves, 1964).

### 6.3 Azimuthally anisotropic images

Both isotropic shear wavespeed and the anisotropic fast axis reveal numerous features correlated with tectonic activity in the region (Figure 6.1). Along the North Atlantic Ridge (NAR), the shear wavespeed is relatively slow ($< -3\%$) down to a depth of 250 km. The fast axis is perpendicular to the ridge system, in agreement with the extensional direction of this divergent plate boundary.

In the western Mediterranean, the east-west oriented fast direction runs parallel to the opening trajectory of the Algero-Provençal and Tyrrhenian Seas, induced by slab roll-back and migration of the Calabrian and Apennines arcs starting $\sim30$ Ma year ago (Wortel & Spakman, 2000). This strong azimuthal anisotropy correlates with a large shear wavespeed reduction ($\sim-4\%$) down to depths in excess of 200 km, suggesting intensive deformation and asthenospheric flow associated with trench retreat. Beneath central Europe, the fast direction shows a strong correlation with the arcuate-shape of the fast Alpine anomaly down to a depth of 200 km.

The old and stable East European Craton (EEC) is imaged as a fast shear wavespeed anomaly ($\sim +4\%$) down to depths in excess of 250 km (Zhu et al., 2012). Around its
Figure 6.1: Isotropic shear wavespeed and azimuthal anisotropy in model EU_{60}. Top and bottom panels illustrate horizontal cross sections of relative perturbations in isotropic shear wavespeed and azimuthal anisotropy at depths of 75 km and 125 km, respectively. Relative perturbations in shear wavespeed are calculated with respect to 1D reference model STW105 (Kustowski et al., 2008a). The direction and magnitude of the fast axis are given by the orientation and length of the yellow bars. White lines denote global plate boundaries (Bird, 2003).
border, there are striking azimuthally anisotropic fabrics associated with the accretion of this craton since the Paleozoic era. To the northwest, the Caledonian Orogeny along the western coast of Scandinavia is dominated by a nearly northeast-southwest fabric within the depth ranging from 75–150 km, which changes abruptly to a weak east-west trend beneath the Baltic Shield (BS).

On the western border of the EEC, there is a prominent contrast in isotropic shear wavespeed separating the Precambrian East European Platform (EEP) from Phanerozoic central and western Europe, namely the Tornquist-Teisseyre Suture Zone (TTSZ) (Zielhuis & Nolet, 1994a). The fast axis runs parallel to the trend of this suture zone from Denmark to the Black Sea. To the west of the TTSZ, the anisotropic fabric follows the strike of several remarkably slow wavespeed anomalies associated with the Bohemian Massif, the Central Slovakian Volcanic Field and the Pannonian Basin (Zhu et al., 2012). The anisotropic strength is relatively weak to the east of the TTSZ at depths greater than 125 km.

Along the southern border of the EEP, northward motion of the Arabian Plate encounters the stable EEC, where its trajectory turns from northerly to east-westerly and contributes to the counter-clockwise rotation of the Anatolian Plate (Figure 6.1). Furthermore, the Aegean is revealed as a strong slow anomaly and the fast direction changes from northeast-southwest in the north to north-south in the south, resulting from the westward motion of Anatolia and retreat of the Hellenic trench. Across the fast wavespeed anomaly of the Hellenic arc, the fast axis changes abruptly from trench-normal in the north to trench-parallel in the south, indicating retreat of the Hellenic trench since the Mid-Eocene (Endrun et al., 2011).

Although the shear wavespeed structure of the EEC is relatively simple compared to central and western Europe, it exhibits lateral and depth variations in the direction and strength of anisotropy at depths shallower than 200 km (Figure 6.1), in agreement with investigations of anisotropic layering within the continental lithosphere beneath North America (Marone & Romanowicz, 2007; Yuan & Romanowicz, 2010; Lin et al., 2011). Tectonic
Figure 6.2: Azimuthal anisotropy within the EEC at a depth of 50 km. Yellow bars represent the direction and strength of the fast axis. Green lines denote tectonic boundaries modified from Artemieva et al. (2006). Cyan lines show the locations of ancient rift systems and suture zones modified from Artemieva et al. (2006). TTSZ: Tornquist-Teisseyre Suture Zone, CRRS: Central Russia Rift System, PR: Pachelma Rift, PDDR: Pripyat-Dnieper-Donets Rift. The age of each rift event is given according to Artemieva et al. (2006).
boundaries, such as the Paleozoic Orogenies and the TTSZ, as well as some ancient rift systems are superimposed on the anisotropic field within the BS and EEP at 50 km depth (Figure 6.2). It is interesting to observe that the anisotropic fabric follows the trends of these ancient rift systems, such as the Pachelma Rift, the Central Russia Rift System and the Pripyat–Dnieper–Donets Rift (Artemieva et al., 2006), which are at least 350 million years old. This azimuthally anisotropic pattern might originate from a “frozen-in” fabric in the lower crust during Precambrian rifting (Silver, 1996), which is preserved owing to relatively low temperatures and a lack of subsequent tectonic activity within the interior of the EEC.

6.4 1D anisotropic strength profiles

We compared 1D radial profiles of isotropic shear wavespeeds and peak-to-peak anisotropic strength for oceanic and continental lithospheres (Figure 6.3). For typical oceanic regions (profiles 1–3), strong shear wavespeed reductions are resolved beneath the oceanic lithosphere (from 70 km to nearly 200 km), indicating the regime of the ductile asthenosphere. Anisotropy within the crust and lithosphere is relatively weak compared to the asthenosphere. The peak-to-peak anisotropic strength first increases monotonically with depth, exceeding more than 1%, and then steadily decreases with depth below its maximum value. By comparing profiles 1–3, it appears that the depth of maximum anisotropy increases with the thickness of the oceanic plate. Substantial anisotropy underlying a weakly anisotropic oceanic lithosphere may reflect a slow spreading rate and strong shear at the base of the Atlantic Plate, an asthenosphere drag mode in contrast to the plate drag mode of the Pacific Ocean. The depth of maximum anisotropic strength is nearly 150 km beneath the Pannonian Basin, as demonstrated in profile 4, reflecting substantial deformation beneath this strong slow anomaly (Figure 6.1). The anisotropic strength profiles for oceanic lithosphere are consistent with mineral physics experiments focused on the transition from brittle to
Figure 6.3: 1D profiles of isotropic shear wavespeed and peak-to-peak anisotropic strength at locations 1–8. A. Simplified tectonic map of the European continent modified from Artemieva et al. (2006). Colors denote the age of each tectonic province. Green lines refer to global plate boundaries (Bird, 2003). B. 1D profiles of isotropic shear wavespeed $V_s$ (upper panel) and peak-to-peak anisotropic strength (bottom panel) at locations 1–8 in A. Red dashed lines in the upper panel represent wavespeed profiles from 1D reference model STW105 (Kustowski et al., 2008a). Gray boxes indicate crustal thickness from model EPCrust (Molinari & Morelli, 2011). Black dashed lines in the bottom panel denote a reference depth of 100 km.
ductile deformation (Kohlstedt et al., 1995), with weak anisotropy in the brittle oceanic crust, and lithosphere underlain by a ductile region that exhibits substantial anisotropy (Ekström & Dziewonski, 1998; Gung et al., 2003).

In contrast, two peaks in anisotropic strength are observed beneath the Aegean and Anatolian in profiles 5 & 6, one within the lower crust and a second at nearly 100 km depth. These features agree with laboratory investigations of lithospheric strength beneath continents, namely, a “sandwich” model with brittle-ductile-brittle-ductile transitions (Kohlstedt et al., 1995). The brittle upper crust is revealed as a regime with relatively weak anisotropy. The first peak indicates ductile flow within the lower crust, which may be related to the “frozen-in” fabric induced by palaeo-extensions in the Miocene (Endrun et al., 2011). The strong secondary peak is related to current mantle flow associated with the retreat of the Hellenic trench and the counter-clockwise rotation of the Anatolian Plate (Figure 6.1). The strain rate in this region is on the order of $10^{-7}$ yr$^{-1}$ (Kreemer et al., 2003), which is sufficient to align the fast axis of olivine with the direction of asthenospheric flow (Zhang & Karato, 1995). These two peaks are accommodated by a transition zone with relatively modest anisotropy.

In profiles 7 & 8, the BS and EEP exhibit large positive shear wavespeed anomalies down to depths in excess of 250 km, corresponding to the fast wavespeed lid underlying continents. The strength profiles involve strong azimuthal anisotropy within the lower crust, in agreement with the location of the ductile lower crust, where the orientation of the fast axis aligns with ancient rift systems (Figure 6.2). The anisotropic strength within the lithosphere of the EEC is relatively weak compared to the oceanic lithosphere (profiles 1–3) and the continental lithosphere beneath Anatolia (profile 6), which may be indicative of a plunging axis of symmetry owing to relatively small shear underlying this old and stable continent (Debayle et al., 2005). Such dipping anisotropy is not accommodated by our current model parameterization.
6.5 Comparison with geodetic survey

Geodetic surveys provide direct constraints on deformation at Earth’s surface, which may be compared with deformation inferred from surface-wave azimuthal anisotropy. Contours of peak-to-peak anisotropic strength and the second invariant of the strain rate field (Kreemer et al., 2003) exhibit intensive deformation along the NAR, the western Mediterranean, the Aegean Sea and the Anatolian Plate (Figure 6.4 A, B). The direction of the fast axis from surface-wave tomography and the extensional component of the current strain-rate field beneath the NAR and the Aegean Sea are in good agreement (Figure 6.4 C, D). For instance, both run perpendicular to the NAR and reflect counter-clockwise rotation of Anatolian lithosphere. Moreover, the transition from trench-normal to trench-parallel patterns beneath the Hellenic arc is observed in both fields (Figure 6.4D). These comparisons imply it is appropriate to use these surface-wave azimuthal anisotropic images (both strength and direction) to infer deformation state and history.

6.6 Acknowledgments

This research is supported by NSF grants 1063057 and 1112906. The observed seismic waveforms used in this study are collected from the Incorporated Research Institutions for Seismology (IRIS; http://www.iris.edu), Observations and Research Facilities for European Seismology (http://www.orfeus-eu.org), and the Kandilli Observatory (http://www.koeri.boun.edu.tr). The open source spectral-element software package SPECFEM3D_GLOBE and the seismic measurement software package FLEXWIN used in this study are freely available for download via the Computational Infrastructure for Geodynamics (CIG; geodynamics.org).
Figure 6.4: Comparisons between surface-wave azimuthal anisotropy and current strain-rate field (Kreemer et al., 2003). A. Contours of peak-to-peak anisotropic strength at a depth of 75 km. B. Contours of the second invariant of the strain-rate field measured at Earth’s surface based on geodetic surveys (Kreemer et al., 2003). C and D. Comparisons between azimuthal anisotropy at 75 km depth and the extensional component of strain rate beneath the NAR (C) and the Anatolian/Aegean (D). Yellow and red bars represent the fast axis and the extensional component of the current strain-rate field, respectively. Blue lines in A and C denote global plate boundaries (Bird, 2003).
Chapter 7

Elastic imaging and time-lapse migration based upon adjoint methods

Note


7.1 Summary

We draw connections between imaging in exploration seismology, adjoint methods, and emerging finite-frequency tomography. All these techniques rely on spatial and temporal constructive interference between observed and simulated waveforms to map locations of structural anomalies. Modern numerical methods and computers have facilitated the accurate and efficient simulation of 3D acoustic, (an)elastic, and poroelastic wave propagation. We use a 2D cross section of the SEG/EAGE salt model to illustrate how such waveform simulations may be harnessed to improve on- and off-shore seismic imaging strategies and capabilities. We demonstrate that the density sensitivity kernel in adjoint tomography is closely related to the imaging principle in exploration seismology, and that
in elastic modeling the impedance kernel is actually a better diagnostic tool for reflector identification. The shear and compressional wavespeed sensitivity kernels in adjoint tomography are closely related to finite-frequency banana-donut kernels, and these kernels are well-suited for mapping larger-scale structure, i.e., for transmission tomography. We further substantiate these ideas by addressing problems in subsalt time-lapse migration.

7.2 Introduction

Mapping oil & gas reservoirs is frequently based upon the imaging principle introduced by Claerbout (1971), which states that reflectors exist at positions where first-arriving up- and down-going waves coincide in time. Imaging can provide constraints on the spatial distribution of reflection coefficients, but it cannot determine the amplitudes of reflectors (Zhang et al., 2003).

Traditionally, the imaging principle has been associated with acoustic seismic migration, i.e., the subsurface is treated as a fluid. For elastic media, the divergence and curl operators have been used in an attempt to separate compressional and shear wavefields, applying a modified imaging principle to the two wave types (Xie & Wu, 2005; Yan & Sava, 2008). Ray-based migration techniques (e.g., Kirchhoff migration) involve high-frequency asymptotic solutions to the wave equation. Due to multipathing, ray methods tend to break down in complex models involving sharp impedance contrasts. Reverse-time migration (McMechan, 1982, 1983; Baysal et al., 1983), which is based upon the full acoustic or elastic wave equation, was initially proposed in the 1980s, but, thanks to modern computers, is recently gaining popularity due to its ability to image complex structures.

Iterative inversion schemes based upon waveforms were advocated by Lailly (1983) and Tarantola (1984), who showed that the gradient of a least-squares waveform misfit function with respect to a set of model parameters may be constructed based upon the interaction between the wavefield for a reference model and a wavefield obtained by using time-reversed
differences between data and synthetics at all receivers as simultaneous sources. This approach requires only two numerical simulations per shot, and these simulations are based upon the same solver. More generally, Chavent (1974) and, in particular, Talagrand & Courtier (1987) introduced the now widely used ‘adjoint-state method’ for the calculation of the gradient of a generic misfit function. A recent review of geophysical applications of the adjoint-state method may be found in Plessix (2006).

Lailly (1983) and Tarantola (1984) recognized that the gradient of a least-squares waveform misfit function is equivalent to the images of seismic migration. Gauthier et al. (1986) and Mora (1987) numerically implemented these ideas for nonlinear inversion of multi-offset seismic reflection data, and an iterative, frequency-domain waveform minimization procedure was developed and implemented by Pratt et al. (1998).

In global & regional seismology, classical ray-based tomographic inversions are being abandoned in favor of inversions based upon finite-frequency sensitivity kernels (Marquering et al., 1998, 1999; Dahlen et al., 2000; Hung et al., 2000). These kernels, sometimes whimsically referred to as ‘banana-donut’ kernels due to their characteristics for direct arrivals in global Earth models (Dahlen et al., 2000), recognize the finite-frequency nature of seismic observables and accommodate wavefront healing. Tomographic inversions based upon finite-frequency kernels are rapidly becoming the standard in global and regional seismology (Montelli et al., 2004; Zhou et al., 2005; Zhao et al., 2005; Chen et al., 2007). The importance of these kernels for exploration geophysics was anticipated by Luo & Schuster (1991) and Woodward (1992), and they have recently been implemented for seismic migration and velocity analysis (Sava & Biondi, 2004a,b; Xie & Yang, 2008).

Tromp et al. (2005) demonstrated that adjoint methods, popular in the atmospheric and oceanographic sciences and rapidly gaining interest in geodynamics (Bunge et al., 2003), are closely related to both finite-frequency tomography and time-reversal imaging. In this article we further explore these connections in the context of exploration seismology, and more particularly in imaging and time-lapse migration.
Isotropic elastic models may be characterized in terms of three model parameters, e.g., density, shear modulus & bulk modulus, or density, shear wavespeed & compressional wavespeed. The choice of model parameterization has important implications for imaging and tomography (e.g., Tarantola, 1986; Forgues & Lambaré, 1997). Following Luo et al. (2009), we further explore the observation that from an imaging perspective, e.g., in reverse-time migration, the ‘impedance’ kernel, which naturally arises when the model is parameterized in terms of density, shear wavespeed & compressional wavespeed, is the best indicator of reflectors. In contrast, adjoint wavespeed kernels, which are closely related to banana-donut finite-frequency sensitivity kernels, are optimally suited for updating shear and compressional wavespeeds. We further explore, quantify, and develop these ideas in the context of on- and off-shore reverse-time and time-lapse migration.

The outline of this article is as follows. We begin with a review of adjoint methods and discuss choices of model parameterization. Next, we draw physical connections between adjoint kernels, finite-frequency sensitivity kernels, and various imaging principles. Finally, we present acoustic-elastic numerical examples of on- and off-shore seismic imaging and time-lapse migration based upon a 2D cross-section of the 3D SEG/EAGE salt-dome model.

7.3 Adjoint Method

Following Tarantola (1984) and Tromp et al. (2005), we use a least-squares waveform ‘misfit’ or ‘objective’ function to measure the goodness of fit between data and corresponding synthetics. In marine acquisition, one records pressure $p$, and for each shot we use the misfit function

$$
\chi(m) = \frac{1}{2} \sum_{r=1}^{N} \int_0^T \omega_r \| p_r^{\text{syn}}(m, t) - p_r^{\text{obs}}(t) \|^2 \, dt,
$$

(7.1)

where the sum is over all $r = 1, \ldots, N$ hydrophones, $T$ denotes the record length, and $\omega_r$ is a weighting & windowing function. The observations are denoted by $p_r^{\text{obs}}$ and the cor-
responding synthetics are identified as $p_r^{\text{syn}}$. The model vector $\mathbf{m}$ may be defined in terms of various combinations of parameters that characterize the model, e.g., density $\rho$, shear modulus $\mu$, and bulk modulus $\kappa$, or density $\rho$, shear wavespeed $\beta = \sqrt{\mu/\rho}$, and compressional wavespeed $\alpha = \sqrt{(\kappa + \frac{4}{3}\mu)/\rho}$. For numerical reasons, we solve the acoustic wave equation using a displacement potential $\phi$ defined by $s = \nabla \phi$, where $s$ denotes displacement (Komatitsch et al., 2005). The potential $\phi$ is related to pressure by $p = -\rho \partial^2 \phi$.

On shore one records ground velocity or displacement, and for each shot we use the misfit function

$$\chi(\mathbf{m}) = \frac{1}{2} \sum_{r=1}^{N} \int_0^T \omega_r ||s_r^{\text{syn}}(\mathbf{m}, t) - s_r^{\text{obs}}(t)||^2 \, dt. \tag{7.2}$$

The observations are denoted by $s_r^{\text{obs}}$ and the corresponding synthetics by $s_r^{\text{syn}}$.

In an isotropic medium, if we choose density $\rho$, shear modulus $\mu$, and bulk modulus $\kappa$ as our model parameters, changes in the misfit function $\delta \chi$, be it equation 7.1 or 7.2, may be written in terms of relative changes in density, $\delta \ln \rho = \delta \rho/\rho$, shear modulus, $\delta \ln \mu = \delta \mu/\mu$, and bulk modulus, $\delta \ln \kappa = \delta \kappa/\kappa$, as (Tromp et al., 2005)

$$\delta \chi = \int_S \left( K_\rho \delta \ln \rho + K_\mu \delta \ln \mu + K_\kappa \delta \ln \kappa \right) \, d^3 x, \tag{7.3}$$

where $\int_S \, d^3 x$ denotes integration over the solid (elastic) domain. The waveform misfit sensitivity kernels $K_\rho$, $K_\mu$, & $K_\kappa$ are the Fréchet derivatives in the elastic region:

$$K_\rho(x) = \int_0^T \rho(x) \partial_t s^\dagger(x, t) \cdot \partial_t s(x, t) \, dt, \tag{7.4}$$

$$K_\mu(x) = -\int_0^T 2\mu(x) D^\dagger(x, t) : D(x, t) \, dt, \tag{7.5}$$

$$K_\kappa(x) = -\int_0^T \kappa(x) \nabla \cdot s^\dagger(x, t) \nabla \cdot s(x, t) \, dt. \tag{7.6}$$

Here $s^\dagger$ denotes the ‘adjoint’ wavefield in the solid, and $D = \frac{1}{2} \left[ \nabla s + (\nabla s)^T \right] - \frac{1}{3} (\nabla \cdot s) \, I$ and $D^\dagger$ denote the traceless strain deviator and its adjoint. The field $s^\dagger$ is the solution of the
adjoint wave equation (Tromp et al., 2005). The regular and adjoint wave equations are the same, with the exception of the source term. As a consequence, the same solver that is used to determine the regular wavefield may be used to determine the adjoint wavefield, simply by changing the source term. In marine acquisition, the adjoint source term for each shot is determined by time-reversed differences between the synthetic and observed pressure:

\[ f^\dagger(x,t) = \sum_{r=1}^{N} \omega_r [p_r^{\text{syn}}(T - t) - p_r^{\text{obs}}(T - t)] \delta(x - x_r), \]  

(7.7)

where \( x_r \) denotes the location of hydrophone \( r \). Note how all the hydrophones become simultaneous sources. Similarly, the adjoint source term for each shot during a land survey is given by

\[ f^\dagger(x,t) = \sum_{r=1}^{N} \omega_r [s_r^{\text{syn}}(T - t) - s_r^{\text{obs}}(T - t)] \delta(x - x_r), \]  

(7.8)

where \( x_r \) denotes the location of geophone \( r \). In this case all the geophones become simultaneous sources. Note from equations 7.7 and 7.8 how the contribution of each shot to the misfit function and its derivative may be calculated one at a time. Consequently, the cost of image construction scales linearly with the number of shots, but is independent of the number of receivers, components, or measurement windows.

To construct the finite-frequency sensitivity kernels 7.4–7.6, one needs simultaneous access to the regular wavefield \( s \) in forward time and the adjoint wavefield \( s^\dagger \) in reverse time. A brute-force strategy would be to first calculate and save the regular wavefield as a function of space and time. During a subsequent adjoint simulation, the regular wavefield may be read back from disk and combined with the adjoint wavefields to obtain the kernels. The storage requirements for this approach are generally prohibitive, especially for realistic 3D simulations. Several methods have been proposed to tackle this problem in reverse-time migration, e.g., an optimal checkpointing algorithm based upon interpolation (Symes, 2007). Our strategy is to reconstruct the regular wavefields on the fly during the calculation of the adjoint wavefields, based upon the last snapshot of the forward sim-
ulation and storage of the forward wavefield absorbed on the artificial boundaries of the domain. This doubles the CPU and memory requirements compared to a regular forward simulation (for details see Liu & Tromp, 2006). In this article, both the forward and adjoint wavefields are calculated based upon a spectral-element method (Komatitsch & Tromp, 1999). The advantages of this method include a natural accommodation of the free surface boundary condition, the ability to honor sharp first- and second-order discontinuities, coupled acoustic-elastic-poroelastic simulations based upon domain decomposition, and straightforward incorporation of full anisotropy as well as 3D attenuation.

7.3.1 Model parameterization

The primary density, shear, and bulk kernels are given by equations 7.4–7.6, and correspond to choosing \( \rho \), \( \mu \), and \( \kappa \) as independent model parameters. There are other, seismologically more sensible, parameterizations one could choose. For example, one could ask for kernels related to a parameterization in terms of density, \( \rho \), shear wavespeed, \( \beta \), and compressional wavespeed, \( \alpha \). The related kernels may be expressed in terms of the primary kernels as

\[
K'_\rho = K_\rho + K_\mu + K_\kappa ,
\]

\[
K_\beta = 2 \left( K_\mu - \frac{4}{3} \frac{\mu}{\kappa} K_\kappa \right) ,
\]

\[
K_\alpha = 2 \left( \frac{\kappa + \frac{4}{3} \mu}{\kappa} \right) K_\kappa .
\]

The kernels \( K_\beta \) and \( K_\alpha \) represent sensitivity to shear and compressional wavespeed perturbations, respectively. When measuring traveltime or phase delays between observed and predicted arrivals, e.g., in finite-frequency shear- or compressional-wave tomography, wavespeed kernels are a natural choice.

In seismic migration, the main goal is to locate interfaces with strong impedance contrasts. Therefore, compressional impedance \( Z_\alpha = \rho \alpha \), shear wavespeed \( \beta \), and compres-
sional wavespeed $\alpha$ seem to be another natural choice of parameters. The related kernels are

$$
K_{Z_\alpha} = K_\rho + K_\mu + K_\kappa,
$$
(7.12)

$$
K_\beta = 2 \left( K_\mu - \frac{4}{3} \mu K_\kappa \right),
$$
(7.13)

$$
K_\alpha = \left( \frac{\kappa + \frac{8}{3} \mu}{\kappa} \right) K_\kappa - K_\rho - K_\mu.
$$
(7.14)

Alternatively, one could choose shear impedance $Z_\beta = \rho \beta$, shear wavespeed $\beta$, and compressional wavespeed $\alpha$ as the independent parameters. In that case the kernels are

$$
K_{Z_\beta} = K_\rho + K_\mu + K_\kappa,
$$
(7.15)

$$
K_\beta = K_\mu - K_\rho - \left( \frac{\kappa + \frac{8}{3} \mu}{\kappa} \right) K_\kappa,
$$
(7.16)

$$
K_\alpha = 2 \left( \frac{\kappa + \frac{4}{3} \mu}{\kappa} \right) K_\kappa.
$$
(7.17)

Finally, we note that shear and compressional wavespeeds $\beta$ and $\alpha$ are mutually dependent, and a better model representation is based upon density, $\rho$, shear wavespeed, $\beta$, and bulk-sound speed, $\Phi = \sqrt{\kappa/\rho}$. In fact, to obtain proper Gaussian statistics, Tarantola (2005, Section 1.1.1) demonstrates that one should employ the logarithm of these quantities as the model parameters. The related kernels are

$$
K'_\rho = K_\rho + K_\kappa + K_\mu,
$$
(7.18)

$$
K_\beta = 2 K_\mu,
$$
(7.19)

$$
K_\Phi = 2 K_\kappa.
$$
(7.20)

It is important to recognize that regardless of whether we choose (1) $\rho$, $\beta$ & $\alpha$, (2) $Z_\alpha$, $\beta$ & $\alpha$, (3) $Z_\beta$, $\beta$ & $\alpha$, or (4) $\rho$, $\beta$ & $\Phi$, as our model parameters, the impedance kernels
$K_{Z\alpha}$ and $K_{Z\beta}$ are strictly equivalent to $K'_{\rho}$:

$$K'_{\rho} = K_{Z\alpha} = K_{Z\beta},$$

(7.21)

Therefore, we use $K'_{\rho}$ to denote the impedance kernel in the remainder of this article, and we will work with the kernels 7.9–7.11 as our preferred finite-frequency sensitivity kernels.

### 7.3.2 Choice of Misfit

The waveform misfit functions 7.1 and 7.2 are just one possible measure of the goodness of fit between observations and synthetics. One can also construct misfit functions by using cross-correlation traveltime measurements:

$$\chi(m) = \frac{1}{2} \sum_{r=1}^{N} \left| |T_{r}^{\text{syn}}(m) - T_{r}^{\text{obs}}| \right|^2,$$

(7.22)

where $T_{r}^{\text{syn}}(m)$ and $T_{r}^{\text{obs}}$ are the predicted and observed traveltimes of a chosen seismic phase. As demonstrated by Tromp et al. (2005), for a given pick, these misfit functions correspond to the marine or land adjoint sources

$$f^{\dagger}(x, t) = \sum_{r=1}^{N} \frac{w_r}{N_r} \delta T_r \partial_{t p_{r}^{\text{syn}}(T - t)} \delta(x - x_r),$$

(7.23)

$$f^{\dagger}(x, t) = \sum_{r=1}^{N} \frac{w_r}{N_r} \delta T_r \partial_{s_{r}^{\text{syn}}(T - t)} \delta(x - x_r),$$

(7.24)

where $\delta T_r = T_{r}^{\text{syn}} - T_{r}^{\text{obs}}$ denotes the cross-correlation traveltime anomaly at receiver $r$, and $w_r$ the cross-correlation time window containing the phase of interest. In marine acquisition the normalization factor $N_r$ is given by

$$N_r = \int_{0}^{T} w_r \left| \partial_{t p_{r}^{\text{syn}}(t)} \right|^2 \, dt,$$

(7.25)
whereas in land acquisition

\[ N_r = \int_0^T w_r \left| \partial_t s_{r}^{\text{syn}}(t) \right|^2 \, dt. \]  

(7.26)

The kernel expressions 7.9–7.11 for the misfit function 7.22 remain the same, but in this case the adjoint wavefield is generated by the traveltime adjoint source 7.23 or 7.24, and the resulting kernels will reflect this choice.

The normalization factor \( N_r \), which arises naturally in cross-correlation traveltime tomography, ensures that the resulting traveltime sensitivity kernels are independent of the seismic scalar moment, i.e., they do not depend on earthquake magnitude or on the amplitude of the controlled source. For example, in earthquake seismology the forward wavefield \( s \) is proportional to the seismic moment \( M \): \( \partial_t s \sim M \). If we want the sensitivity kernels 7.9–7.11 to be independent of source magnitude, the adjoint wavefield must be proportional to \( 1/M \): \( s^\dagger \sim 1/M \). In waveform tomography, the weighting factor \( \omega_r \) in the waveform misfit functions 7.1 and 7.2 should be chosen such that the related sensitivity kernels are also independent of source magnitude.

This may be accomplished by choosing

\[ \omega_r = 1/ \int_0^T \left| p_{r}^{\text{obs}}(t) \right|^2 \, dt \]  

(7.27)

for measurements in the acoustic medium, and

\[ \omega_r = 1/ \int_0^T \left| s_{r}^{\text{obs}}(t) \right|^2 \, dt \]  

(7.28)

in the elastic medium, i.e., by normalizing the waveform misfit with the power in the data, thereby rendering the misfit function dimensionless. The quality of the images obtained using waveform adjoint tomography with and without normalized adjoint sources is discussed in detail in Luo et al. (2009). Their results demonstrate that using normalized
adjoint sources yields better images. For this reason, in what follows when we refer to waveform adjoint tomography it is implied that we use normalized adjoint sources.

7.4 Drawing Connections

7.4.1 Banana-donut and adjoint kernels

Dahlen et al. (2000) relate the cross-correlation traveltime anomaly $\delta T_r$ for a given source and a receiver $r$ to relative wavespeed perturbations $\delta \ln v = \delta v/v$ via a finite-frequency ‘banana-donut’ kernel $\overline{K}_r$:

$$\delta T_r = \int \overline{K}_r \delta \ln v \, d^3x.$$  \hfill (7.29)

Note how in finite-frequency tomography a wave senses relative wavespeed perturbations $\delta \ln v$ through the banana-donut kernel $\overline{K}_r$. This relationship generalizes the ray expression

$$\delta T_r = -\int_{\text{ray}_r} v^{-2} \delta v \, dl,$$  \hfill (7.30)

which attributes the traveltime anomaly to wavespeed perturbations along the corresponding geometrical ray. At longer periods, the banana-donut kernels $\overline{K}_r$ are fat, recognizing that longer period waves sense farther away from the geometrical ray path, whereas at shorter periods the kernels are skinny, sensing close to the ray. Banana-donut kernels account for wavefront healing, recognizing that a traveltime anomaly ‘heals’ as the wave propagates further. These kernels are currently revolutionizing global seismic tomography (Montelli et al., 2004).

One may calculate a banana-donut kernel based upon an adjoint method by using the traveltime adjoint source 7.23 or 7.24 for a single pick at a single receiver (Liu & Tromp, 2006). Consequently, for a given source, we may write the cross-correlation traveltime
adjoint sensitivity kernel $K$ as a weighted sum of individual banana-donut kernels $\mathcal{K}_r$:

$$K = \sum_{r=1}^{N} \delta T_r \mathcal{K}_r,$$

(7.31)

where the weights are the corresponding traveltime anomalies $\delta T_r$ (Tromp et al., 2005).

### 7.4.2 Imaging and adjoint kernels

Mathematically, the imaging principle may be expressed as

$$I(x) = \frac{1}{P(x)} \int_{0}^{T} u(x, t)d(x, t) dt,$$

(7.32)

where $u$ and $d$ denote the vertical-component up- and down-going wavefields, respectively, $T$ denotes the length of the time window of interest, and the normalization factor $P$ is given by $P = \int_{0}^{T} d^2(t) dt$ (Chattopadhyay & McMechan, 2008). This normalization factor plays the same role as the weighting factor $\omega_r$ in the waveform misfit functions 7.1 and 7.2.

The connection between the density adjoint kernel, $K_\rho$, and the imaging principle 7.32 may be established as follows. In seismic migration, the downgoing velocity wavefield $d$ is related to the forward displacement wavefield $s$ by $d(t) = \partial_t s(t)$, and the upgoing velocity wavefield $u$ is related to the adjoint wavefield $s^\dagger$ by $u(t) = -\partial_t s^\dagger(T - t)$; it is convenient and conventional to use backward time, $T - t$, in the adjoint wavefield. Upon substituting these relationships into the imaging principle 7.32 and comparing the result to the density kernel 7.4, we find that

$$I(x) \sim K_\rho(x),$$

(7.33)

i.e., the imaging principle corresponds to the density finite-frequency Fréchet derivative.

To account for multicomponent data in elastic media, several modified imaging principles have recently been introduced. For example, for P-to-P and S-to-S reflections the
imaging conditions may be expressed as (Xie & Wu, 2005; Yan & Sava, 2008)

\[ I_{PP}(x) = \int_0^T [\nabla \cdot u(x,t)] [\nabla \cdot d(x,t)] \, dt, \] (7.34)

\[ I_{SS}(x) = \int_0^T [\nabla \times u(x,t)] \cdot [\nabla \times d(x,t)] \, dt, \] (7.35)

where \( \nabla \cdot u \) and \( \nabla \cdot d \) correspond to the compressional-wave components of up- and down-going wavefields, and \( \nabla \times u \) and \( \nabla \times d \) refer to the shear-wave components. Upon comparing these expressions to equation 7.6, we see that the modified imaging principle for P-to-P reflections corresponds to our bulk modulus kernel \( K_{\kappa} \):

\[ I_{PP}(x) \sim K_{\kappa}(x). \] (7.36)

Kiyashchenko et al. (2007) advocate a modified imaging principle that is reminiscent of the sum of our density and bulk kernels, i.e., the acoustic impedance kernel.

### 7.5 Numerical Examples

#### 7.5.1 SEG/EAGE salt-dome model

To illustrate the imaging capabilities of the adjoint method, we design two experiments representative of land and marine surveys. We use a modified version of a 2D cross-section of the 3D SEG/EAGE salt-dome model (Aminzadeh et al., 1997). In the original SEG/EAGE model, a high wavespeed salt body and several spike reflectors are superimposed on a smooth background wavespeed model. In our models, we retain the salt body and divide the background model into several distinct regions, each with constant material properties. To obtain an elastic version of the acoustic SEG/EAGE model, we assume that the shear wavespeed \( \beta \) may be obtained from the compressional wavespeed \( \alpha \) based upon the relationship \( \beta = \alpha / \sqrt{3} \), and the density, \( \rho \), is calculated based upon the empirical relationship
proposed by Brocher (2005). For marine surveys, we add an acoustic layer on top of the elastic domain, with constant compressional wavespeed and density.

Figure 7.1a shows the compressional wavespeed distribution in the 2D cross section of the SEG/EAGE model used for on-shore simulations. Figures 7.1b and c show the two types of marine surveys we will consider: a flat sea floor and a model with irregular bathymetry. These models are used as ‘true’ models to generate the ‘data’ \( p_r^{\text{obs}}(t) \) or \( s_r^{\text{obs}}(t) \). It is well known that the images of prestack depth migration are very sensitive to the accuracy of the initial smooth wavespeed model. In least-squares waveform inverse problems, the initial model should be reasonably accurate, otherwise gradient-based inversion strategies may get trapped into some local minimum rather than converging to the global minimum of the misfit function. For our purposes, smooth initial models are obtained by convolving the true models shown in Figures 7.1a–c with a 2D Gaussian. Considering the shortest wavelength in our numerical simulations, we use a Gaussian with a 100 m half width to smooth the true models and obtain the background models displayed in Figures 7.1d-f. For the models used in the marine survey shown in Figures 7.1b–c, we do not smooth the bathymetry. The smooth models are used to generate synthetic waveforms \( p_r^{\text{syn}}(m, t) \) or \( s_r^{\text{syn}}(m, t) \).

The open-source 2D spectral-element software package SPECFEM2D (Komatitsch & Tromp, 1999, geodynamics.org) is used for our forward and adjoint simulations. Mesh generation is an important aspect of a spectral-element simulation. Ideally, the mesh should honor all first- and second-order discontinuities in the geological model, resolve at least five points per shortest wavelength, and satisfy the Courant stability condition. Unstructured quadrilateral spectral-element meshes which satisfy these requirements are generated using CUBIT (Sandia National Laboratory, cubit.sandia.gov), and are superimposed on the models displayed in Figures 7.1a–c. In the land survey, Figure 7.1a, there are 10,003 elements and 161,017 global grid points. In the marine survey, due to the slow compressional wavespeed in the water layer, both the local grid spacing and the time
Figure 7.1: Compressional wavespeed distribution for modified 2D cross-sections of the 3D SEG/EAGE model used for land and marine experiments. (a) On-shore experiment: modified elastic model used as the ‘true’ model to generate ‘data’, $s_{\text{obs}}^{\text{true}}$. (b) Off-shore experiment I: modified acoustic-elastic model used to mimic marine experiments. A 1 km deep water layer is added on top of the elastic model shown in (a). The ocean bottom is flat. (c) Off-shore experiment II: same model as (b) but this time the bathymetry is irregular. The spectral-element mesh is superimposed on panels (a)–(c). The mesh honors all discontinuities and satisfies appropriate stability and accuracy conditions. Panels (d)–(f) display smooth models obtained by convolving the true models (a)–(c), with a Gaussian with a 100 m half width. These smooth models are used to generate synthetics. The two green triangles identify regions used for 4D seismic imaging.
step have to be decreased dramatically to satisfy the stability and accuracy requirements. In our spectral-element modeling we use domain decomposition between the fluid and the solid, honoring the continuity of traction and the normal component of displacement across the interface (Komatitsch et al., 2000, 2005). The flat sea floor marine model, Figure 7.1b, has 48,744 elements and 781,653 global grid points, while the model with irregular bathymetry, Figure 7.1c, contains 48,971 elements and 785,289 global grid points. We use the same spectral-element meshes for both the ‘true’ and smooth initial models to obtain ‘data’ and synthetics.

In this section we use surveys involving 51 shots to illuminate the on-shore model (Figure 7.1a) and the off-shore models (Figures 7.1b and c). Each shot illuminates a portion of the subsurface structure, resulting in one ‘shot’ kernel. Stacking all 51 shot kernels yields a complete image for the entire model, a procedure reminiscent of prestack depth migration. Because the number of simulations is independent of the number of receivers, we only need 102 forward simulations to generate ‘data’ and synthetics. After producing the adjoint sources, another 51 adjoint simulations are performed to obtain the sensitivity kernels. For the land survey, we use two recording geometries: a wide-angle deployment of 601 geophones with a maximum offset of 14 km, and a narrow-angle deployment of 101 geophones with a maximum offset of 2.35 km. Explosive point sources are directly deployed on the free surface. For the marine survey, 201 hydrophones move following the shots at a maximum offset of 2.35 km. Both sources and receivers are deployed 10 m below the water surface. A Gaussian source-time function with a 10 Hz center frequency is used for all simulations. The influence of source frequency content on image quality is discussed in Luo et al. (2009). They show that the lower the frequency content the blurrier the images. Each receiver in a common-shot gather records seismograms for a duration of 6 s. Because free surface boundary conditions are applied in all simulations, both surface and direct waves are muted in the pre-processing.
7.5.2 Reverse-time migration

By backprojecting the power-normalized waveform differences between synthetics and data in the smooth model shown in Figure 7.1d, we obtain the density, shear modulus, and bulk modulus kernels as well as the impedance, shear wavespeed, and compressional wavespeed kernels displayed in Figure 7.2 (wide-angle deployment, 601 geophones) and Figure 7.3 (narrow-angle deployment, 101 geophones). As previously discussed, the density kernel, \( K_\rho \), is equivalent to migrated images based upon Claerbout’s imaging principle. It is well known that reverse-time migration images suffer from low spatial frequency artifacts if full or two-way wave equations are used (Yoon et al., 2004). Those artifacts are regarded as the effects of diving waves, head waves, and backscattered waves, which are associated with sharp impedance contrasts. Several methods have been proposed to eliminate or attenuate these artifacts, for example, high-pass spatial filtering (Guitton et al., 2007) or iterative least-squares migration (Mulder & Plessix, 2004). These low-frequency, volumetric patterns are very clearly seen in our adjoint kernels, except in the impedance kernel \( K'_\rho \). When \( K_\rho, K_\mu, \) and \( K_\kappa \) are summed to produce the impedance kernel, \( K'_\rho \), the low-frequency artifacts cancel each other, while true reflectors are enhanced. Thus, the main structural features can be identified more clearly in the impedance kernel. It is for this reason that we advocate the use of the impedance kernel, \( K'_\rho \), instead of the density kernel, \( K_\rho \), to image reflectors. Notice that subsalt regions are still very difficult to image because of defocusing and multipathing. The high-frequency artifacts near the surface are attributed to interactions between body and surface waves. The shear modulus kernel, which does not arise in purely acoustic simulations, plays an important role in the construction of the impedance kernel. We also note that the low-frequency volumetric ‘banana-donut’ patterns in the wavespeed kernels, which are regarded as artifacts in reverse-time migration, can be used to update the wavespeeds of the initial model in an iterative inversion scheme. In fact, we anticipate that subsalt imaging can be significantly improved by harnessing impedance and wavespeed kernels in a joint iterative inversion scheme, starting with specific low-
Figure 7.2: Finite-frequency sensitivity kernels for the 2D modified elastic SEG/EAGE salt model shown in Figure 7.1a, generated by backprojecting the normalized differences between ‘data’ and synthetics from 601 geophones in the smooth model displayed in Figure 7.1d. This wide-angle experiment has a maximum offset of 14 km. (a) Density kernel $K_\rho$, (b) shear modulus kernel $K_\mu$, (c) bulk modulus kernel $K_\kappa$, (d) impedance kernel $K'_\rho$, (e) shear wavespeed kernel $K_\beta$, (f) compressional wavespeed kernel $K_\alpha$. All six kernels are displayed on the same scale. All of these sensitivity kernels exhibit low-frequency artifacts, except for the impedance kernel (d), defined as the sum of the density kernel (a), shear modulus kernel (b) and bulk modulus kernel (c).
Figure 7.3: Same as Figure 7.2, except in this case we use 101 geophones, using narrower angle reflections with a maximum offset of 2.35 km. (a) Density kernel $K_\rho$, (b) shear modulus kernel $K_\mu$, (c) bulk modulus kernel $K_\kappa$, (d) impedance kernel $K'_\rho$, (e) shear wavespeed kernel $K_\beta$, (f) compressional wavespeed kernel $K_\alpha$. Figure courtesy of Luo et al. (2009).
Figure 7.4: Marine survey: (a) Impedance kernel for a model with a flat ocean floor obtained by migration in the smooth model displayed in Figure 7.1e. (b) Impedance kernel for a model with irregular bathymetry obtained by migration in the smooth model displayed in Figure 7.1f. Notice that artificial reflectors are generated, which are attributed to multiples in the water layer.

frequency arrivals and steadily increasing the frequency content and number of arrivals.

Figure 7.4 shows the impedance kernels for the two marine surveys, with flat and irregular bathymetry. The initial smooth models are obtained by using a Gaussian with a 100 m half width. Due to the sharp contrast between the water layer and the underlying solid, strong multiples appear in the wavefields, which generate several artificial reflectors, but at the same time enhance the actual reflectors. Upon comparing Figure 7.4a and b, we observe that irregular bathymetry mutes artificial reflectors while not affecting the true reflectors.

7.5.3 Time-lapse migration

Our next application mimics a time-lapse experiment in which the properties of particular regions of the model change between two subsequent surveys. Time-lapse migration is also referred to as 4D seismic imaging, since it involves not only the spatial distribution (3D) of reflection coefficients, but also temporal changes (the fourth dimension) due to the flow of fluids. Time-lapse migration is an important tool for carbon sequestration and imaging
monitoring induced fluid injections in reservoirs (Landro et al., 1999; MacBeth et al., 2005). Therefore it is a critical asset for managing and forecasting production.

In our experiments, from one survey to the next, we change the wavespeeds of triangular regions on the deep left flank of the salt dome and below the salt dome by −20%, in an attempt to mimic the (possibly induced) migration of fluids (green triangles in Figure 7.1). We backproject the differences between sets of seismograms recorded during the two subsequent surveys to illuminate the regions that are responsible for the observed waveform differences.

For a land survey, Figure 7.5 shows the primary kernels $K_\rho$, $K_\mu$, and $K_\kappa$ as well as the impedance and wavespeed kernels $K'_\rho$, $K_\beta$, and $K_\alpha$ for changes occurring in the triangle on the left flank of the salt body. We see again how the impedance kernel $K'_\rho$ highlights the affected region, while the other kernels show volumetric changes. The top and bottom boundaries of the triangle are clearly identified. The right boundary of the triangle is missing, since waves reflected off that boundary are poorly recorded by geophones at the surface. A vertical seismic profile (VSP) can help to enhance spatial coverage (Dillon, 1988), leading to a better image of the right boundary. Notice that we produce a false image of a portion of the bottom of the salt body below the target region. This artifact is due to a mismatch between data and synthetics induced by the wavespeeds in the target region. These artifacts can be eliminated by restricting the seismogram used for imaging by excluding later portions of the records, which contain reflections from the bottom of the salt body. These observations underscore the need to develop a multi-scale, top-down inversion strategy that harnesses the low-frequency wavefield first, before increasing the frequency content to steadily enhance the quality of the model (e.g., Bunks et al., 1995; Pratt, 1999).
Figure 7.5: Primary kernels and impedance and wave speeds kernels for time-lapse migration. (a) density kernel $K_\rho$, (b) shear modulus kernel $K_\mu$, (c) bulk modulus kernel $K_\kappa$, (d) impedance kernel $K_\rho'$, (e) shear wavespeed kernel $K_\beta$, (f) compressional wavespeed kernel $K_\alpha$. All six kernels are displayed on the same scale.
Figure 7.6: Impedance kernels for time-lapse migration in different models. (a) land survey for the triangle on the left flank, (b) land survey for the triangle below the salt body, (c) marine survey for the triangle on the left flank, with flat ocean floor, (d) marine survey for the triangle below the salt body, with flat ocean floor, (e) marine survey for the triangle on the left flank, with irregular ocean floor, (f) marine survey for the triangle below the salt body, with irregular ocean floor.
The wavespeed sensitivity kernels $K_\beta$ and $K_\alpha$ exhibit more diffusive volumetric sensitivity. The reason for the diffuse nature of the wavespeed kernels may be understood in terms of expression 7.31, which states that the adjoint wavespeed sensitivity kernels are weighted sums of banana-donut kernels, with weights determined by the traveltime delay. Our shot-receiver geometry illuminates the target region from above, resulting in a sum of more-or-less vertical kernels connecting sources and receivers to the target. When the anomaly is slow, the resulting shear and compressional wavespeed kernels are red. Again, strings of geophones in wells would help to concentrate the wavespeed kernels more tightly in the region of interest.

Figure 7.6 compares impedance kernels in two marine settings. To investigate bathymetric effects, we consider flat and irregular bathymetry of the sea floor. In both marine surveys, the impedance kernel contains the correct triangular regions, together with false images generated by multiple reflections in the water layer. In both scenarios, the image of the target regions is not affected by the existence of the water layer. We can eliminate the false images by using only early arrivals in the waveform differences, or by removing multiples based upon standard pre-processing migration techniques (Backus, 1959; Verschuur et al., 1992).

7.6 Discussion

We have illustrated the capabilities of the adjoint method in imaging elastic and coupled acoustic-elastic models, which mimic land and marine acquisition experiments, respectively. On-shore, we are able to illuminate deeper structures (e.g., the lower boundary of a salt body) because of the large aperture of the recording geometry. Off-shore, our approach introduces some artifacts, which are attributed to reverberations in the water layer. Irregular bathymetry reduces these artifacts.

We have demonstrated that the density kernel which arises in adjoint tomography closely
corresponds to Claerbout’s imaging principle. Our numerical results show that the impedance kernel, defined as the sum of the density, shear modulus, and bulk modulus kernels, is more suitable from an imaging perspective than the density kernel. It is well known that reverse-time migration based upon the full wave equation suffers from low spatial frequency artifacts in the presence of strong impedance contrasts; these artifacts are frequently eliminated based upon high-pass spatial filtering. Unlike the density kernel, the impedance kernel does not contain these low-frequency patterns and yields clearly delineated reflectors. Impedance kernels may also be successfully used to highlight wavespeed anomalies induced by migrating fluids, i.e., in the context of 4D seismic imaging.

The volumetric shear and compressional wavespeed sensitivity kernels which arise in adjoint tomography correspond to the finite-frequency ‘banana-donut’ sensitivity kernels which are transforming global and regional seismic tomography. Banana-donut kernels are the finite-frequency extension of geometrical rays, and are well-suited for constraining large-scale wavespeed variations. They form a perfect complement to the impedance kernel, which highlights reflectors.

### 7.7 Conclusion

We have shown that adjoint methods, finite-frequency sensitivity kernels, and imaging principles commonly invoked in seismic migration are closely related. In particular, adjoint methods may be used to migrate on- and off-shore seismic data. Due to the high accuracy of the spectral-element method, using domain decomposition between the fluid and the solid, we are able to image in coupled acoustic-elastic models with irregular bathymetry. Besides identifying reflectors based upon the impedance kernel, shear and compressional wavespeed sensitivity kernels may be used to iteratively update model parameters in a tomographic inversion scheme. The inversion strategy should involve gradually increasing the frequency content of data and synthetics to avoid entrapment in local minima, as well as
a ‘top-down’ approach in which early arrivals affected by shallow structures are fit before later arriving signals associated with deeper structures. The emphasis should be on matching the phase of observed signals, i.e., on fitting targeted, frequency-dependent traveltime anomalies. Waveform tomography is largely controlled by amplitude differences, which are notoriously difficult to fit in seismology. Traveltime or phase, on the other hand, is a robust measure of misfit that has been used very effectively to constrain Earth structure.

### 7.8 Acknowledgments

We thank Assistant Editor Tamas Nemeth and two anonymous reviewers for comments & suggestions which helped to improve the manuscript.
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