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Insights on the upper mantle beneath the Eastern Alps

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ABSTRACT

Analyses of Ps and Sp receiver functions from datasets collected by permanent and temporary seismic stations, image a seismic discontinuity, due to a negative velocity contrast across the entire Eastern Alps. The receiver functions show the presence of the discontinuity within the upper mantle with a resolution of tens of kilometers laterally. It is deeper (100–130 km) below the central portion of the Eastern Alps, and shallower (70–80 km) towards the Pannonian Basin and in the Central Alps. Comparison with previous studies renders it likely that the observed discontinuity coincides with the lithosphere–asthenosphere boundary (LAB) east of 15°E longitude, while it could be associated with a low velocity zone west of 15°E.

1. Introduction

The Alps are the result of long term convergence between the Eurasian and African plates, which began around 120 Ma ago (DeMets et al., 1994). Subduction initiated ∼80 Ma ago and plate collision started 35 Ma ago (Handy et al., 2010, and references therein) followed by uplift of the Alpine orogenic belt after 23 Ma (e.g. Schmid et al., 2004; Castellarin and Cantelli, 2000). As expected, such a long series of tectonic processes have led to the formation of a highly complex and heterogeneous structure of the crust (Hirn et al., 1980; Pfiffner, 1990; Ye et al., 1995; Bleibinhaus and Transalp Working Group, 2001; TRANSALP Working Group, 2001; TRANSALP Working Group et al., 2002) and the upper mantle (Hirn et al., 1984; Panza et al., 1980; Pfiffner et al., 1988; Kissling, 1993; Lippitsch et al., 2003; Koulakov et al., 2009; Bokelmann et al., 2013).

Tomographic models of the upper mantle beneath the convergent zone (Wortel and Spakman, 2000; Piromallo and Morelli, 2003; Giacomuzzi et al., 2011) have determined the current position of ancient suture zones by imaging high-velocity anomaly bodies running parallel to the Alpine chain axis that extend into the mantle transition zone. Regional tomographic models (Lippitsch et al., 2003; Mitterbauer et al., 2011) show that the positive velocity anomalies (ascribed as the Alpine slab) are interrupted along the Alpine chain, testifying the presence of fragmented subduction. Seismic models based on P-wave residuals (Babuška et al., 1998), MT and electromagnetic studies (cf. Jones et al., 2010; Korja, 2007), as well as geothermal (mostly steady-state) models (Artemieva et al., 2006, and references therein) describe an anomalously thin (60–80 km) lithosphere below the Pannonian Basin. Previous works show that the lithosphere is thickened beneath the Bohemian Massif to 120–140 km (Babuška and Plomerová, 2001; Heuer et al., 2007; Geissler et al., 2010, 2012; Plomerová et al., 2012), which is a large stable body of crystalline rock representing the eastern part of the European Variscan Orogen. The whole mantle in this area has been imaged by P receiver function studies of Kummerow et al. (2004) and Hetényi et al. (2009). However in Kummerow et al. (2004) the occurrence of the Moho multiples obscures possible conversions due to structures within the upper mantle, while Hetényi et al. (2009) primarily explore the mantle transition zone and irregular lateral variation of the 660 discontinuity due to accumulation of cold and denser material, described as subducting slabs actively impinging on the lower mantle. Receiver functions computed along longitude ∼12°E show the thickening of the European crust towards south until the central part of the Eastern Alps (Kummerow et al., 2004).

The surface expression of the arcuate Alpine belt can be divided into two distinct blocks; the arcuate western Alps, and the Eastern Alps, which extends towards the east to the Carpathians. There are many open questions regarding the eastern part of the Alpine belt. Is it linked at depth to the Carpathian arc or to the Dinaridic arc? Does a Pannonian plate exist and is there a triple junction between Europe, Adria, and Pannonia (Brückl et al., 2010)?

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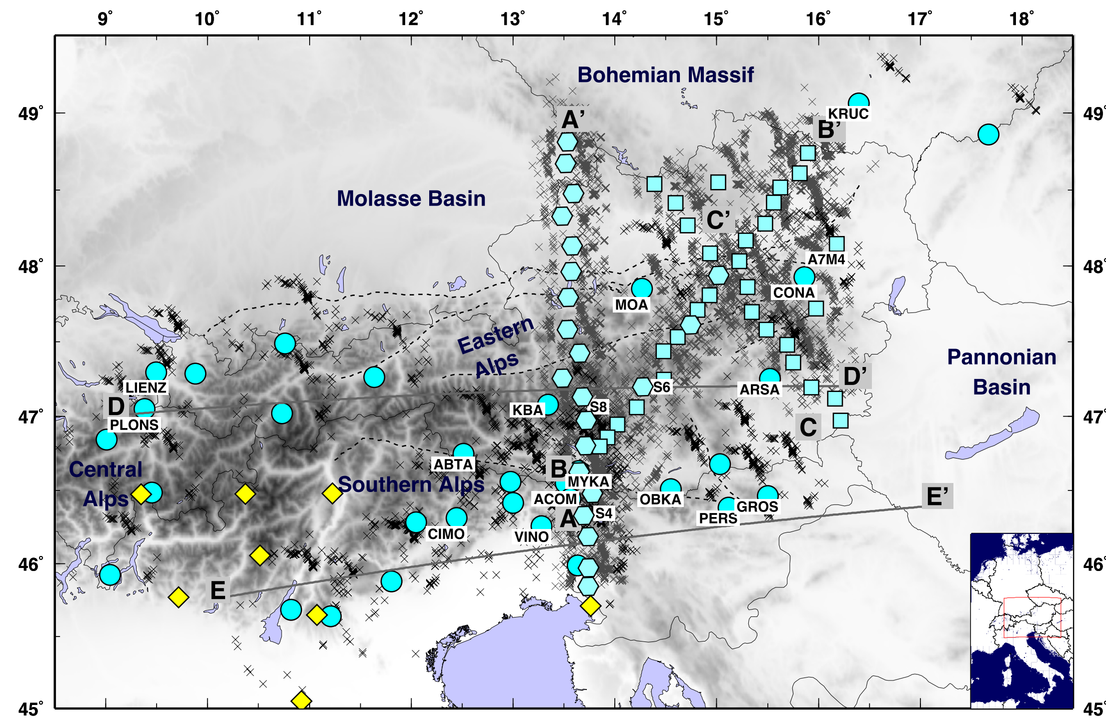

Fig. 1. Map of the Eastern Alps, including seismic station locations. Light blue circles show broadband permanent stations from different national networks, hexagons for broadband temporary stations, and squares for short period temporary stations. Yellow diamonds show stations used in Miller and Piana Agostinetti (2012). Gray crosses are the piercing points at 100 km depth for PRF. Black crosses are piercing points at 100 km for SRF. The inset shows the study area location in Central Europe. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Is it possible to distinguish the effects of the indentation and of the lateral extrusion (e.g. Ratschbacher et al., 1991a)? In this paper we investigate the structure of the shallow portion of the upper mantle, down to 150 km, beneath the central and eastern Alps, to contribute additional information that may help addressing some of these outstanding questions, in particular by looking at the occurrence of a decrease in seismic velocities. Sharp velocity reductions in the upper mantle have been suggested to be due to the boundary that separates the lithosphere from the asthenosphere (LAB) (Barrell, 1914; Eaton et al., 2009) or to a mid-lithospheric discontinuity (MLD) globally located in a depth interval below 100 ± 20 km (Thybo and Perchuć, 1997; Thybo, 2006; Abt et al., 2010; Lekic and Romanowicz, 2011). A distinct velocity drop was globally identified by P receiver functions between 70 ± 4 km in oceanic environments, and down to 95 ± 4 km beneath Precambrian shields and platforms (Rychert and Shearer, 2009); possible explanation for these observations has been proposed by Karato (2012) that considering the amount of water content, the geothermal gradient and the grain boundary sliding, identifies a decrease in seismic velocities in the depth range between 70 (oceans) and 150 (continents) km. In this work we present P and S receiver function results that detect a shallow negative velocity contrast in the mantle below the Eastern Alps and its depth variations that occur over length scales of several tens of kilometers. A negative velocity contrast implies the occurrence of a seismic discontinuity below which the S-velocity decreases. The results for the two different methodologies are strikingly similar.

2. Data and methods

2.1. P receiver functions

More than 8000 waveforms, from 536 teleseismic events with Mw ≥ 5.5 occurred at epicentral distances between 30° and 100°, were used to compute the P receiver functions. Waveforms from these events were recorded at 56 three-components stations (Fig. 1), 53 belonging to the ALPASS temporary network (Mitterbauer et al., 2011) which were deployed between July 2005 and April 2006, and 3 stations from the Carpathian Basin Project (CBP) temporary network that were deployed from May 2006 to June 2007 (Dando et al., 2011).

The P receiver function (PRF) method isolates the effect of mode conversions generated at velocity discontinuities at depth beneath a seismic station. The technique has been employed primarily for determining the depth of the crust–mantle boundary (Moho) but has also been widely used for imaging other discontinuities such as the LAB or the mantle transition zone (e.g. Rychert and Shearer, 2009; Hetényi et al., 2009). PRFs are the result of P-to-S (Ps) waves generated by the conversion of the incoming teleseismic P-wave into an S-wave by the passage through a seismic interface at depth (Langston, 1979; Ammon, 1991). The presence of several velocity jumps at depth causes the presence of several Ps converted phases together with their multiples (such as PpPs, PsPs + PpSs phases); consequently a more complicated structure results into a more phase-populated receiver function. In this study, P receiver functions have been calculated in the RTZ reference system with a frequency domain algorithm using multitaper correlation estimates (Park and Levin, 2000) with a frequency cut off of 0.2 Hz in order to image the deep lithospheric structure. This method provides an estimate of PRF uncertainty in the frequency domain, using the pre-event noise spectrum for frequency-dependent damping. The multitaper spectrum estimates are leakage resistant, so low-amplitude portions of the P-wave spectrum can contribute usefully to the PRF estimate. This enables PRFs from different seismic events to be combined in a weighted-average PRF estimation according to the inverse of their variance. The weighted average PRFs are obtained by binning events in 10° bins for both epicentral distance and from backazimuth (Fig. 2). For each PRF we calculated the mean and standard deviation (σ) as in Abt et al. (2010). We consider the well-resolved portions of the re-