Upper mantle structure beneath the Alpine orogen from high-resolution teleseismic tomography

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[1] To understand the evolution of the Alpine orogen, knowledge of the actual structure of the lithosphere-asthenosphere system is important. We perform high-resolution teleseismic tomography with manually picked $P$ wave arrival times from seismograms recorded in the greater Alpine region. The resulting data set consists of 4199 relative $P$ wave arrivals and 499 absolute $P$ wave arrivals from 76 teleseismic events, corrected for the contribution of the Alpine crust to the travel times. The three-dimensional (3-D) crustal model established from controlled-source seismology data for that purpose represents the large-scale Alpine crustal structure. Absolute $P$ wave arrival times are used to compute an initial reference model for the inversion. Tests with synthetic data document that the combination of nonlinear inversion, high-quality teleseismic data, and usage of an a priori 3-D crustal model allows a reliable resolution of cells at 50 km $\times$ 50 km $\times$ 30 km. Hence structures as small as two cells can be resolved in the upper mantle. Our tomographic images illuminate the structure of the uppermost mantle to depth of 400 km. Along strike of the Alps, the inversion reveals a high-velocity structure that dips toward the SE beneath the Adriatic microplate in the western and central Alps. In the eastern Alps we observe a northeastward dipping feature, subducting beneath the European plate. We interpret this feature in the western and central Alps as subducted, mainly continental European lower lithosphere. For the east, we propose that parts of the Vardar oceanic basin were subducted toward the NE, forcing continental Adriatic lower lithosphere to subduct northeastward beneath the European plate.

INDEX TERMS: 7203 Seismology: Body wave propagation; 7218 Seismology: Lithosphere and upper mantle; 8180 Tectonophysics: Tomography;

KEYWORDS: crust and upper mantle, seismic tomography, Alpine orogen, body waves


1. Introduction

[2] Because of extensive cooperative near-vertical and wide-angle reflection and refraction studies (i.e., ECORS-CROP [Roure et al., 1990]; EGT [Blundell et al., 1992]; NRP20 [Pfiffner et al., 1997]; TRANSALP [Transalp Working Group, 2001]; for earlier studies, see references in these publications) knowledge of the crustal structure of the Alpine orogen has markedly improved over the past decades (for a tectonic map, see Figure 1). However, very little can be said about the lower lithosphere, beneath the Moho, from these projects. Surface wave studies [Panza et al., 1980] and delay time tomography studies [Spakman, 1991; Spakman et al., 1993] have been carried out on a bigger scale to resolve the large-scale structure of the mantle. However, to better understand the evolution of the Alpine orogen, much more detailed information on deeper lithosphere and upper mantle structure is required. On the basis of a review of active and passive seismic studies on the deep structure of the Alps, Kissling [1993] concluded that there is lithospheric slab material beneath the Po plain that is probably connected to European continental lithosphere. This finding is in general agreement with subsequent tomographic studies of Morelli and Piromallo [1999], Bijwaard and Spakman [2000], and Bijwaard et al. [1998]. Their upper mantle models show a band of relative high velocity that follows the Alpine arc and extends to a depth of $\sim 300$ km. Regional-scale tomographic studies in the western Alps [Solarino et al., 1996] show an approximately NW-SE trending high-velocity anomaly, which Solarino et al. interpreted as remnants of subduction responsible for the Alpine orogeny.

[3] Although it is commonly understood that the European lithosphere is subducting beneath the African promontory [Stampfli et al., 1998], the amount of subducted lithospheric material (continental and oceanic) is still a matter of debate. Estimates of the cumulative large-scale convergence in the Alps since Cretaceous times are as high as 1100 km but have to be interpreted with extreme caution. The width of the different oceans (Alpine Tethys, Valais, Meliata) involved in the subduction process is not known nor the amount which can be budgeted by escape tectonics or by crustal thickening [Regenauer-Lieb, 1995]. Hence only
additional detailed and qualitative imaging of slab geometry and orientation will provide key information for geodynamic model reconstructions of the neoalpine orogeny in space and time. Resolution of previous tomographic models is insufficient to address details such as the proposed detached Valais ocean [Marchant and Stampfli, 1997] or tears in the lithospheric slab [Wortel and Spakman, 2000].

This paper presents the results of a high-resolution teleseismic tomography study in the greater Alpine region capable to resolve P wave velocity variations of 50 km length in the upper mantle to a depth of 400 km. In this study we use high-quality manually picked P wave travel time data and an a priori 3-D crustal model. Since the accuracy of imaged upper mantle structure in teleseismic tomography critically depends on precise knowledge of the 3-D crustal velocity structure beneath the receiver array, we correct the observed travel time residuals for the crustal effects with a method proposed by Waldhauser et al. [1998, 2002]. With this a priori crustal correction and high-quality teleseismic data, we are able to decrease the appropriate cell volume by a factor of ten relative to previous studies and obtain reliable images of smaller-scale upper mantle structures. Specific aspects of teleseismic tomography are described later in the text; for a general description of the technique see sections 3 and 4.

The goal of this study is to reveal the upper mantle structure beneath the Alpine arc to depths of 400 km, in particular, to image lithospheric slab geometries to provide a basis for further discussion of tectonics and recent evolution of the Alpine orogen.

2. Data

Several permanent seismic station networks are operated since the early 1980s in the greater Alpine region. Digital teleseismic recordings from these permanent networks as well as data from the temporary passive seismic network TRANSALP [TRANSALP Working Group, 2001] were collected and merged for this study (Figure 2). The temporary TRANSALP stations were arranged along a 40 km wide traverse from Munich to Venice, following an active seismic reflection survey. The array filled the gap between nearby regional station networks and consisted of broadband and short-period stations operated between May 1998 and April 1999 (Figure 2). Teleseismic P wave data...
from the following regional seismic networks have been included in this work: DISTER (Genoa), ING (Rome), LED (Freiburg), RENASS (Strasbourg), Rete Sismometrica del Friuli-Venezia Giulia (Trieste), SED (Zürich), URGS (Ljubljana), and ZAMG (Vienna). The database consists of digital mostly three-component seismograms of 76 selected teleseismic events (Figure 3a) recorded at more than 200 stations in the greater Alpine region. Selection criteria for the events were (1) at least 20 observations per event, (2) event magnitude larger than 4.5 and (3) epicentral distance between $20^\circ$ and $90^\circ$. For the picking of arrival times, we assigned different errors to every single pick (Figures 4a and 4b). This so-called “observational weighting scheme” [Arlitt, 1999] consistently reflects arrival time errors in phase correlation caused by scattering, dispersion, and noise effects. A conservative estimation of the data error for the 4199 relative $P$ wave travel time readings (Figure 4c) used in this study yields an average 0.36 s corresponding to a data variance (noise) of 0.13 s$^2$ of the data set. Data with obviously large timing/picking errors (picks with residuals of more than 5 s) and data with very low signal-to-noise ratio were excluded from the data set.

In order to obtain uniform resolution in the center region of the study area between $5^\circ$–$15^\circ$E longitude and $44^\circ$–$49^\circ$N latitude, excellent azimuthal coverage of the data is required. Figure 3b shows that the azimuthal coverage is complete, though not homogeneous with respect to the number of readings, since the southern sector shows lower coverage due to small number of events of magnitude greater than 4.5 in the African craton and the Atlantic. However, the present selection criteria are still fulfilled, and hence the areal coverage of stations (Figure 2) and azimuthal distribution of selected events (Figure 3b) provide a good ray geometry for the subsequent inversion procedure as it ensures ray crossing in most of the area under study. Nevertheless, one has to be aware that the uneven azimuthal distribution of picks can bias the resolution of the inversion process.

### 3. Analysis and Correction of Travel Time Residuals

In teleseismic tomography, travel times are inverted for velocity anomalies relative to an assumed reference velocity model, which in our study is a 1-D reference model computed from absolute $P$ wave travel times. Contributions to travel time residuals come from the lateral heterogeneity of the Earth’s mantle and crust, seismogram reading errors, station and instrument effects, and hypocenter mislocation. To minimize the effects arising from uncertainties in lower mantle structure and source effects, relative travel times are usually calculated by subtracting the weighted mean residual arrival time for each event from every station recording that event. In regional teleseismic tomography this approach is justified for plane waves only, an approximation not valid when crustal variations strongly distort teleseismic wave fronts [Waldhauser et al., 2002; Arlitt et al., 1999]. The linear approximation we apply to the travel time equation to solve the inverse tomographic problem (for a mathematical formulation, see Spakman et al. [1993]) is a valid approximation only, if the initial reference model for the inversion is a close approximation to the average seismic structure in the study area. Thus we adapt the method of a “minimum 1-D model” used in local earthquake tomography [Kissling et al., 1994] to generate a starting model close to the true upper mantle and crustal model. We use absolute travel times to determine this 1-D initial reference model. Relative travel times are calculated with respect to the weighted average for arrivals from one event, while absolute travel times denote the difference relative to travel times calculated for the IASP91 model [Kennett and Engdahl, 1991]. Hence, while for relative residuals the weighted average per event is removed, when inverting the absolute travel times one obtains first order information on the average (1-D) velocity-depth function below the study region. A natural starting point for the calculation of this 1-D initial reference model is the global
1-D model itself; the following determination of the reference model is then a trial and error process. The calculation of a 1-D initial reference model can lead to ambiguous results, several parameters that control the inversion need to be varied and the corresponding results have to be examined. We selected the best events in the data for absolute travel time picks, preferably recorded by the whole station network, and invert for velocity changes relative to the IASP91 model and repeat this procedure several times with new (updated) velocities. The determined “new” initial reference model (Figure 5) shows slightly higher velocities at depth of 150–210 km and is \( \frac{1}{C^2} \% \) slower in deeper layers compared to the 1-D global IASP91 model. Our a priori 3-D crustal-uppermost mantle model derived by CSS data (described below) includes velocity information down to 72 km depth (using the \( P_n \) phases in refraction profiling). The travel time effects of this independent 3-D model with respect to model IASP91 are corrected before the inversion. Hence the fact that above 100 km only the velocity gradient can be resolved (Figure 5) does not bias the inversion results. Thus the newly derived 1-D initial reference model represents a close approximation to the average velocity structure in the upper mantle in the well resolved center part of the study area. We do not violate the linearization assumption and the model provides additional valuable information to improve the constraints on the amplitude of velocity variations and to check the performance of our 3-D crustal model as described below.

As mentioned above, a 3-D crustal correction of the teleseismic travel time residuals is performed. In the Alpine region the effect of the crust may account for up to 50% of the observed travel time residuals [Kissling, 1993]. Because of almost exclusively near-vertical ray penetration of the crust in teleseismic tomography, these crustal anomalies may erroneously be mapped into the upper mantle by the inversion process. Thus we apply the method proposed by Waldhauser et al. [1998] and Arlitt et al. [1999] to correct for crustal contributions to teleseismic travel times.

We have built a 3-D crustal model from existing controlled source seismology (CSS) profiles [Waldhauser, 1996; Waldhauser et al., 1998]. This 3-D crustal model (Figure 6) comprises the large-scale Alpine crustal structure. In addition to the European, the Adriatic and the Ligurian crustal units, this crustal model includes specifically the Po...
basin [Bigi et al., 1983], the Molasse basin [Lohr, 1978], and the Ivrea body [Solarino et al., 1997]. The steeply dipping Ivrea body (Figures 1 and 6) is a structure of high-density, high seismic velocity, and high magnetic susceptibility associated with the Adriatic crust and located at the inner side of the western Alpine arc [Menard and Thouvenot, 1984].

[11] The effect of the 3-D crust on teleseismic travel time depends on the local structure itself, on the backazimuth, and on the epicentral distance of a specific event because the rays sample different crustal structures. Thus we calculate the time field for each single event (Figure 7) from the base of the 3-D crustal model to the surface with the finite difference (FD) scheme of Vidale [1990] modified by Hole and Zelt [1995]. In the study region, severe distortions of the wave front of more than 1 s occur for regions with deep and strongly heterogeneous velocity structure in the crust beneath the southern Alps, the Apennines, and the Po plain. With this approach we can separate and correct for the contribution of the 3-D crustal structure on teleseismic travel time for every single event-station configuration. Waldhauser et al. [2002] recently used this method and showed that the correction of near-surface structure prevents crustal effects from leaking into deeper layers. Result is a clearer image of the actual subcrustal structure since it is not masked by crustal anomalies.

[12] We selected four teleseismic events (Table 1 and Figure 3) to illustrate the effect of a 3-D crustal correction on travel time residuals for the mantle in the Alpine area (Figures 8a and 8b) and to demonstrate that the effects depend on back azimuth and distance (Figures 8c–8d). Comparing the pattern of travel time residuals before and after crustal correction (Figures 8a and 8b) for the event located in Alaska and recorded at the Alpine station network, we observe a more homogeneous distribution of travel time residuals after correction, since we subtract the effect of the strongly heterogeneous crustal velocity struc-

Figure 5. Initial 1-D global reference model (IASP91 [Kennett and Engdahl, 1991], dashed line) and final (solid line) 1-D reference model after inversion of absolute travel time residuals. Above 100 km only the velocity gradient can be resolved, no absolute velocities. We derive Vp down to 72 km by the a priori crustal model based on CSS data (Figure 6). Crosses denote depths of grid layers, where velocities are parameterized.

Figure 6. Southeast perspective view of newly compiled 3-D Alpine crustal P wave velocity model to 70 km depth, shown (a) in a cutaway display and (b) by south–north oriented cross sections. The Ligurian (ML), the Adriatic (MA), and the European (ME) crustal blocks with Moho depths are visible as well as the sedimentary basins and the Ivrea body. Perspective contours at the surface indicate national boundaries [after Waldhauser et al., 2002].
events indicate a rather steep high-velocity upper mantle structure located in the upper mantle beneath the eastern Alps, east of the temporary TRANSALP array.

4. Three-Dimensional Inversion of Teleseismic Travel Time Residuals

[13] Next we invert the crustal corrected travel time residuals to obtain a three-dimensional view of the upper mantle structure using a nonlinear inversion technique that is based on the original formulation known as the ACH technique [Aki et al., 1977; Iyer and Hirahara, 1993] with several significant modifications. The forward problem is solved by 3-D minimum travel time ray ray tracing [Steck and Prothero, 1991] using the simplex algorithm of Press et al. [1986]. The nonlinear inversion scheme iteratively inverts the residuals for velocity changes of an initial reference model, in our case the new 1-D reference model shown in Figure 5. Each iteration involves a complete one-step inversion, including both ray tracing and change of velocity distribution in the model. The solution for \( V_p \) is obtained using a weighted, damped least squares technique. To invert the linearized travel time equation, we used the singular value decomposition. The \( V_p \) values are estimated at nodes of a 3-D grid, where the grid node is the center of a grid cell. We parametrize the model by a three-dimensional grid with 50 km horizontal and 30 km vertical grid spacing in areas with reasonable ray coverage, and grid spacing of 50 × 50 × 50 km at the border and the base of the model (300–400 km). The gridding is adapted to the distribution of stations and ray paths [Kissling et al., 2001] and guarantees uniform and reasonable resolution in most areas under study in the depth range 70–400 km. In the inversion, velocities are held fixed (velocity changes are not allowed) for grid nodes not hit by any rays and for crustal nodes, as we already corrected for crustal effects. In total, our model consists of 2937 unknown \( V_p \) values and 4199 P observations which lead to an average over determination factor of 1.43.

[14] To quantify the influence of regularization parameters (number of iterations and damping factor) and to assess the resolution capability of our data set, we performed various tests with synthetic upper mantle models and analyzed the results of inversions. A critical parameter in the inversion of teleseismic data is the damping value. Depending strongly on source-receiver distribution and on grid spacing, damping does not only affect the final solution but also the absolute value of the resolution matrix. In the inversion process amplitude damping was applied, and based on a series of single iterations with varying damping values for travel times calculated for a synthetic upper mantle model (Figure 9), a damping value of 80 was chosen for the 3-D inversion. The synthetic upper mantle model mimics two slab structures

<table>
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<th>Event</th>
<th>Latitude, deg</th>
<th>Longitude, deg</th>
<th>Backazimuth, deg</th>
<th>Distance, deg</th>
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<tbody>
<tr>
<td>Central America</td>
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<td>-96.87</td>
<td>293</td>
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<td>-161.19</td>
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<td>155.03</td>
<td>21</td>
<td>79.782</td>
</tr>
<tr>
<td>Indonesia</td>
<td>1.28</td>
<td>100.32</td>
<td>59</td>
<td>89.636</td>
</tr>
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beneath the Alpine region with velocity changes of $\pm 4\%$ and $-4\%$ relative to the IASP91 model [Kennett and Engdahl, 1991]. The two structures have a horizontal extension of three cells, 150 km, respectively. Figure 9d shows that with a damping value of 80 the best fit between synthetic input model and recovered structure is achieved. The damping value of 80 is approved by plotting data variance against model (solution) variance (Figure 10a). It represents a sensible compromise between minimizing model variance and optimizing data fit (data variance). The 3-D inversion process was stopped after a maximum of four iterations, because adjustments to the model parameters became insignificant, compared to the errors of observed phases. The variance reduction that is obtained after four iterations amounts to $\approx 70\%$ (Figure 10b). Starting variance for the data set is the variance of the relative travel time difference after the crustal correction, the remaining 30\% data variance correspond to noise in the data.

[15] The size of the inverse problem did not allow the computation of a formal resolution analysis. Thus resolution tests were performed with synthetic upper mantle models. As one example for resolution tests, we show the test with a characteristic model for the upper mantle structure of the Alpine region (Figure 11). A characteristic model contains size and amplitude of anomalies seen in the inversion results of the real data, but with different strike, shape and sign of the amplitude [Haslinger et al., 1999]. Velocity anomalies have been implemented in such a way, that it is possible to identify areas with reliable resolution and to test for potential horizontal and vertical velocity smearing. We designed a high-velocity slab structure accompanied by upwelling of faster material, a low-velocity slab structure, and different smaller anomalies to test for vertical and horizontal limits of resolution.

[16] The ray bending method of Steck and Prothero [1991] is used to calculate travel times relative to the IASP91 reference model for the characteristic upper mantle structure at the Alpine stations. Gaussian noise is added with a standard deviation in the size of the observational weight, which represents the pick uncertainty of a real data set. By comparing the results of the inversion to the input model, areas of reliable resolution can be identified (Figure 11). At 90 km depth the synthetic structure is poorly recovered due to a lack of crossing rays. In deeper layers velocity anomalies in the Alpine area are properly resolved, although we observe smearing effects between adjacent velocity structures with same polarity (indicated at depth 150 km), due to the azimuthal distribution of picked arrivals (Figure 3b). There is limited vertical resolution between 150 and 240 km depth because of the limited vertical resolution of teleseismic rays, which is a consequence of rays crossing the upper mantle with similar and almost

Figure 8. Travel time residuals (s) for events with hypocenter in Alaska (a) before and (b) after crustal correction. Circles represent fast and pluses represent slow travel times relative to the IASP91 model. Size of symbols scales with amplitude of the travel time delay/advance. Dashed box indicates enlarged area shown in Figures 8c–8f, with travel time residuals for event with hypocenter for (c) Central America, (d) Alaska, (e) Kamchatka, and (f) Indonesia after crustal correction. The migration pattern (see text) is indicated by dashed contours.
vertical angle of incidence (between $10^\circ$ and $40^\circ$). The images show vertical smearing from layers above (120–150 km) and below (270–240 km). An oblique profile (Figure 9) through a high-velocity slab structure shows the capability of the data set to resolve dipping structures. The amplitude recovery in general is decreased as a consequence of damping. In deeper layers (beneath 270 km), amplitude recovery is reasonably good. Additional resolution tests with different synthetic upper mantle structures, checkerboard, and spike and spike harmonic tests [Spakman and Nolet, 1988] were also performed. We also calculated the number of rays which intersect each cell to find areas with poor coverage. From these calculations, rectangular contours are drawn to mark areas of reliable resolution where the geometry of anomalies is restored correctly.

5. Presentation of Inversion Results

[17] To show the inversion results, we plot a series of horizontal depth slices (Figure 12) located at the respective inversion grid nodes. The 10 layers show the $P$ wave velocity heterogeneity in the upper mantle beneath the greater Alpine region obtained after four iterations with a damping value of 80. To obtain a smoother image, we apply a linear interpolation between the inversion cells. In general, model features can only be discussed if they are equal or larger than the resolvable size. Areas of good resolution are shown in full colors, whereas areas of poorer resolution are depicted in faded colors. Cells with fewer than two rays are masked grey. As we corrected the teleseismic travel times for crustal effects, we kept the $P$ wave velocity at crustal nodes (at surface and 30 km depth) fixed during the inversion process. We allow the node at 60 km depth to float (to change its velocity) to account for possibly uncorrected crustal contributions, and start the discussion of the newly derived upper mantle $P$ wave velocity model at a depth of 90 km.

[18] In the two uppermost layers (depths of 90 and 120 km), large areas beneath the Po plain are scarcely sampled by seismic rays and therefore not reasonably resolved, due to the lack of seismic stations in the Po basin. Nevertheless, there are indications for low velocity material in the illuminated surrounding regions down to 240 km depth. The most prominent feature in the layers from 90 to 210 km depth is a band of relative high-velocity following the structural strike of the Alps, with up to +5% higher $P$ wave velocity than 1-D initial reference model (Figure 5). At 120 km this high-velocity band seems to be discontinuous beneath the western Alps. The velocity anomaly is rather steep as it does not change horizontally. At a depth of 240 km we observe three distinctly separated high-velocity anomalies. While the central high-velocity structure is vanishing with depth, the high-velocity structures beneath the western and eastern Alps can still be seen at 270 km partly down to 300 km depth. Three high-velocity blocks appear in deeper layers, one in the area of the Apennines at 350 km depth and two beneath the Po plain and the northern Alpine foreland in 400 km depth. Beneath the Apennines and the Ligurian Sea, a chain of high velocities can be resolved at 120 and 150 km depth, which disappears again at deeper layers. In the layers down to 210 km depth, the
northern foreland of the Alpine chain is bordered by an extended low-velocity anomaly which is broken up into several segments at deeper layers. Because of insufficient ray crossing, the tomographic images get more diffuse at layers deeper than 400 km, so we exclude these layers from the presentation.

6. Interpretation of Tomographic Results

[19] To visualize better the variation in P wave velocity beneath the Alpine arc, we plot three vertical transects cutting the upper mantle along the western (Figure 13a), central (Figure 13b), and the eastern Alps (Figure 13c). Since we do not invert for crustal structures Moho topography (red) is extracted from the 3-D crustal P wave velocity model is superimposed instead. We define the lithosphere-asthenosphere boundary (LAB) as the depth, where the vertical gradient of P wave velocity is lower or equal to zero indicating the asthenosphere. In three sections we tentatively outline the extent of the lithospheric slabs by dotted lines (Figures 13a–13c).

[20] Figures 13a displays a section through the upper mantle below the western Alps. The profile is roughly oriented WE perpendicular to the strike of the Alps crossing the western Alpine arc and the Po plain. The location of the geophysical Ivrea body is schematically included in the crustal layers. The upper part of the profile shows, that the main feature of the European lower lithosphere is resolved and the LAB can be drawn at about 100 km, slightly deepening toward the inner part of the western Alpine arc. At larger depths, a well developed high-velocity anomaly, corresponding to the high-velocity anomaly beneath the Alpine arc (Figures 12a–12j), can be traced down to about 320 km depth with a dip in ESE direction. In the west, this high-velocity volume is bordered by a low-velocity zone, which we can follow down to $\sim$350 km. Two high-velocity anomalies (A and B) appear in the middle of the cross section at 350 km. Low-velocity material lies beneath the Po plain (10°–12.5°E longitude), which we can trace down to 210 km.

[21] We will first concentrate on the striking high-velocity feature between 150 and 320 km depth and its connection to the European lower lithosphere. To find the origin of this high-velocity structure we need additional information from kinematic reconstructions. Cross section AA’ (Figure 13a), is located closely to the ECORS-CROP geophysical transect, with detailed information on crustal structure. Schmid et al. [1996] and Schmid and Kissling [2000] derived from active seismic data that European lower lithosphere is subducted toward the E-SE beneath the Adriatic microplates in the western Alps. Kinematic reconstructions in the western Alps along the ECORS-CROP transect imply an estimated total amount of $\sim$124 km postcollisional crustal shortening [Schmid and Kissling, 2000]. Apart from subducted European crustal lower lithosphere we have to expect previously subducted oceanic lithosphere beneath the western Alps. Before the actual collision in the late Eocene (40 Myr), at least two oceanic basins (Piemont-Ligurian and Valais) were subducted beneath the Adriatic microcontinent in the wider Alpine orogen.

[22] In our cross section located slightly west of the ECORS-CROP, at least 60 km of European lower lithosphere seem to be subducted beneath the Adriatic microplate, as we can trace it to the ESE beneath (Figure 13a). Since we expect a total amount of $\sim$124 km European lower lithosphere in the upper mantle [Schmid and Kissling, 2000], we conclude that parts of the high-velocity material between 160 and 320 km depth are of European lower lithospheric origin. The length extension of the high-velocity structure ($\sim$180 km) indicates that either more European lower lithosphere has been subducted beneath the Adriatic microplate than estimated or the feature consists not only of European lower lithosphere. Since the Valais
ocean was subducted at a later stage in the Alpine orogeny, we consider that remnants of subducted Valais oceanic lithosphere could be also included in the high-velocity feature.

[23] A clear gap appears in the high-velocity structure between 110 and 150 km depth, which could indicate slab detachment. An explanation for slab detachment at this location could be that the Ivrea body, which acted as a buttress during the collision of the Adriatic and European plate, created opposing buoyancy and forced a tear apart of the subducting slab. Slab detachment for the western Alps was proposed earlier by Sue et al. [1999] to explain extensional earthquakes in the center of the western Alpine arc. The pronounced shallow low-velocity anomaly in the west (between 6° and 7°E longitude) underneath the LAB may be reinforced by asthenospheric upwelling, resulting from the slab detachment process.

[24] The low-velocity anomaly beneath the Po plain has been imaged in several tomographic studies before [Spakman et al., 1993; Solarino et al., 1996; Bijwaard and Spakman, 2000]. As it is located near to the border region of our study area, the geometry and amplitude of the anomaly remain uncertain. An additional study encompassing the region farther south is therefore necessary [see Di Stefano et al., 2003]. We do not assign the high-velocity feature below 320 km depth (A) (Figure 13a) to the lithospheric slab structure, as this could also be a leakage effect from deeper structures (below the studied upper mantle model volume which extends to 400 km only). Furthermore, this feature shows the opposite dip direction. The origin of the second high-velocity structure at this depth (B) is also unclear.

[25] Figure 13b shows a section across the central Alps, near the transition to the eastern Alps. The LAB beneath the European plate can be drawn at ~80 km in the NW, deepening toward the center of the orogen. The striking high-velocity feature seen in the horizontal depth slices (Figure 12) is now clearly connected to European lower continental lithosphere. The structure dips to the SE and reaches a depth of 240 km. Similar to the profile through the western Alps (Figure 13a), two high-velocity anomalies of unknown origin appear at larger depths (A and B). To the NW at around 49°N latitude, a low-velocity feature is resolved down to 130 km. The SE sector reveals low velocities at shallow depths beneath the Po plain.

[26] Differences in deep structure between central and western Alps were proposed by Schmid and Kissling [2000]

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Figure 11. Horizontal depth sections of synthetic (characteristic) model used to compute synthetic travel times and the recovered structure after the inversion with four iterations. For better differentiation, positive anomalies are marked with plus. Dashed black lines denote areas with reasonable resolution. Arrows indicate selected areas with vertical smearing at 120 km and horizontal smearing at 150 and 240 km depth.
Figure 12. Horizontal depth sections of final 3-D $V_p$ model at (a) 90 km, (b) 120 km, (c) 150 km, (d) 180 km, (e) 210 km, (f) 240 km, (g) 270 km, (h) 300 km, (i) 350 km and (j) 400 km depth. Depth locations of slices correspond to inversion grid nodes, linear interpolation is applied between the inversion cells to smoothen the image. Velocity variations are plotted relative to the 1-D initial reference model (Figure 5). Areas with no resolution (no hits) are left gray, and areas of critical resolution are displayed in faded colors.
on the basis of the significant differences between the crustal structures and their evolution. In the western Alps (along the ECORS-CROP profile) the geological-geophysical cross sections of Schmid and Kissling [2000] reveal a European lower crustal wedge. Contrary, in the central Alps (along the NRP20/EGT transect) the Adriatic lower crust is indenting European lithosphere, building an Adriatic lower crustal wedge. Under the western Alps the Ivrea body acted as a buttress during the collision of Adriatic and European plates, preventing indentation from SE. From kinematic reconstructions the estimated amount of postcollisional crustal shortening for the central Alps is with 164 km [Schmid et al., 1996] larger than in the western Alps. From our tomographic image (Figure 13b) we estimate that ~160 km lower lithospheric material has subducted beneath the Adriatic microcontinent. Thus the entire high-velocity anomaly seems to be European lower lithosphere. There is no evidence, that parts of a subducted Valais oceanic lithosphere are attached to European lower lithosphere along this profile (Figure 13b). The tomographic image suggests slab detachment between European lower lithosphere and subducted oceanic lithosphere, as already proposed by von Blanckenburg and Davies [1995]. They explained magmatic intrusions along the Periadriatic lineament with slab detachment in the Alpine region. The shallow low-velocity feature in the NW part of the section correlates with the Rhinegraben, a continental rift system with shallow LAB [Fanza et al., 1980; Prodehl et al., 1995].

Figure 13c shows a cross section of the upper mantle structure across the eastern Alps. Whereas the high-velocity feature in the central Alps (Figure 13b) was clearly connected to the European lower lithosphere, it now seems to be tied to the Adriatic lower lithosphere in the eastern Alps dipping toward north to ~230 km depth. The LAB beneath the Adriatic microplate is seen at ~100 km.

Because of the opposite dip direction of high-velocity material we imaged in the eastern Alps, and the connection to the Adriatic, instead of European, lower lithosphere we state that the three tomographic images show lower lithospheric subducting beneath the Adriatic microcontinent beneath the western and central Alps, and lithosphere subducting northeastward beneath the European plate in the eastern Alps. We identify the high-velocity volume beneath the central part of the orogen as European lower lithosphere. This observation corroborates the general concept of different stages in the evolution of the Alpine orogen [Marchant and Stampfli, 1997; Stampfli et al., 1998; Schmid and Kissling, 2000]. For the eastern Alps our tomographic images reveal a different upper mantle structure. Thus we assume that a different ocean was subducted, forcing Adriatic continental lower lithosphere to subduct northeastward beneath the Austro-Alpine. Recent plate kinematic reconstructions for the Alpine regions [Channell and Kozur, 1997; Stampfli et al., 1998; Wortmann et al., 2001] show the Varad ocean south, respectively southeast of the Austro-Alpine. We assume that parts of this oceanic basin were subducted toward the north beneath the Austro-Alpine. With this hypothesis we can explain the location of the northward subducting continental slab beneath the European plate in the eastern Alps.

We consider that the subducted high-velocity material we find beneath the eastern Alps is similar to the western and central Alps of continental origin and hence was subducted after collision. In the eastern Alps we lack sufficient geological-geophysical information. Valuable information on convergence rates and kinematic reconstructions is hardly available for the eastern part of the Alpine orogen. Considering the rotation of the Adriatic microcontinent after collision, we must assume that the amount of subducted continental lithosphere increases toward the east. In the Friuli region the amount of crustal shortening for the last 20 Mio years is estimated to ~100 km [Nussbaum, 2000]. More conclusive data can be expected from the active reflection and refraction survey along the TRANSALP traverse [TRANSALP Working Group, 2001]. From our tomographic images we estimate 210 km (Figure 13c) lithospheric material beneath the European plate. These findings suggest crustal shortening of about 200 km after collision at 40 Myr), which is possible if we consider the proposed 100 km for the last 20 Myr by Nussbaum [2000]. Thus we identify the high-velocity material connected to Adriatic lower lithosphere and subducting toward the northeast as Adriatic lower lithosphere.

### 7. Comparison With Other Studies

Babuska et al. [1990] compiled a map for the lithosphere-asthenosphere boundary in the Alps on the basis of relative travel time residuals. In this map the existence of two high P wave velocity features beneath the western and eastern part of the Alpine orogen separated by an asthenospheric upwelling (relatively low P wave velocities) is proposed. The locations of these two anomalies coincide with high-velocity structures in our tomographic images. Belts of high velocity beneath the Alpine arc are also shown by model EUR89B [Spakman et al., 1993], model PM0.5 [Morelli and Piromallo, 1999], and model BSE2000 [Bijwaard and Spakman, 2000]. Because of limited resolution, these mantle models do not reveal the different slab geometries. Our findings are in good agreement with the earlier ones, but show significantly more detailed structures. Comparison with the BSE2000 3-D global P wave velocity model [Bijwaard and Spakman, 2000] show that the upper...
mantle region beneath the Alpine orogen is underlain by massive high-velocity anomalies below 400 km depth, which seem to be remnants of earlier subducted oceans (e.g., Meliatta and Alpine Thetys). Thus we have to consider that the deeper parts in our regional upper mantle image could be biased by leakage effects of deeper sitting structures, as commented earlier (Figure 13). Hence further studies should consider a correction of deeper mantle structures, similar to the crustal corrections we performed in this study.

8. Conclusion

[31] The inversion of crustal corrected teleseismic travel time residuals recorded in the greater Alpine region reveals significant lateral P wave velocity variations in the upper mantle beneath the Alpine orogen. Various synthetic resolution tests document that the two pronounced high-velocity anomalies located beneath the Alps are separated and dip in oblique direction. In the western and central Alps the high-velocity anomaly is connected to European lower lithosphere and dipping toward the southeast beneath the Adriatic microplate to depths of ~300 km. The high-velocity anomaly in the eastern Alps, with a clear connection to Adriatic lower lithosphere, dips to the northeast and is subducted beneath the European plate. We interpret these structures as continental European lower lithosphere in the western and central part of the orogen, and as Adriatic lower lithosphere in the eastern part.

[32] Previous tomographic studies did not have the resolution capabilities to resolve the slab geometries in detail. By applying crustal correction to teleseismic travel time residuals and performing a nonlinear inversion with a 3-D ray tracer, we are able to resolve the upper mantle structure beneath the Alpine orogen in much more detail. This enables us to identify two different collisional lithospheric slabs, which suggest two separate earlier subduction zones beneath the Alps.

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