

The arc of the western Alps in the light of geophysical data on deep crustal structure

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Abstract. Recently, two international deep seismic campaigns in the western and central Alps (Etude Continentale et Océanique par Reflexion et Refraction Sismique - Progetto Strategico Crosta Profonda, ECORS-CROP [Roure *et al.* 1996b]; and National Research Program 20, NRP-20 [Pfiffner *et al.* 1997]) have been completed. Here we present a synoptic interpretation of the wealth of geophysical data about deep crustal structure of the Alps collected during the past 40 years, including the two above mentioned geophysical projects. The tectonic reinterpretation of the ECORS-CROP high-resolution seismic transect is based on an evaluation of the deep crustal structure by integrating new and literature data on surface geology. Combined with previously published interpretations regarding the central Alps [Schmid *et al.*, 1996], this study reveals substantial differences in geometry and kinematics between transects across the western and central Alps, respectively. At depth the transition between the central and western Alps is marked by the western limits of an Adriatic lower crustal wedge-shaped structure and the northeastern limits of a similar structure made of European lower crustal material. At the surface it coincides with a corridor of dextral strike-slip along the Tonale and Simplon lines. In addition, the thickness of the seismogenic zone shows a remarkable variation from over 40 km beneath the Penninic realm of the western Alps to less than 20 km in the central Alps. The formation of the western Alpine arc was initiated during convergence and collision before 35 Myr ago, when the Adriatic micro-plate, moving northward with respect to the European foreland, caused sinistral transpression in the western Alps while the central and eastern Alps underwent head-on convergence and collision. During the post-collisional stage, i.e., after 35 Myr ago, the arcuate shape of the western Alps was accentuated by WNW-directed movement and anticlockwise rotation of the Adriatic microplate, decoupled from the central and eastern Alps along the Tonale-Simplon dextral shear zone. This led to wedging of lower crustal slices both in the western Alps and in the central Alps. The new tectonic interpretation of the ECORS-CROP transect allows a kinematic model to be established for crustal shortening in the western Alps during the past 35 Myr. The Ivrea mantle plays the role of a backstop in our tectonic model. We distinguish three episodes of post-collisional

crustal shortening along the ECORS-CROP transect: From 35 to 30 Ma the Briançonnais basement was back-thrusted over the Gran Paradiso units, leading to 30 km of crustal shortening (first episode). In the early Miocene, movements concentrated mainly along the Penninic frontal thrust and resulted in about 60 km shortening (second episode). Post-12 Ma shortening within the external massifs is associated with folding in the Jura mountains when the crust was shortened by an additional 30 km (third episode).

1. Introduction

Important progress has recently been made regarding two large-scale geophysical-geological transects situated in the Swiss-Italian central Alps and in the northern part of the French-Italian western Alpine arc. Published final reports summarize the achievements of the Swiss National Research Project 20 (NRP-20) [Pfiffner *et al.*, 1997] and of the French-Italian Etude Continentale et Océanique par Reflexion et Refraction Sismique - Progetto Strategico Crosta Profonda (ECORS-CROP) geophysical campaign in the western Alps [Roure *et al.*, 1996b] involving high-resolution deep seismic sounding. Schmid *et al.* [1996, 1997] and Escher *et al.* [1997] summarize the geodynamical evolution along two major transects in the eastern and western part of the Swiss-Italian Alps, respectively. A good summary of the geophysical data along the ECORS-CROP transect is available in the work of Roure *et al.* [1996b]. The former geodynamic models and the latter seismic data, however, still await a synoptic interpretation taking also into account all additional NRP-20 data and the wealth of geophysical data collected during the past 40 years [e.g., Kissling, 1993].

Direct comparison of the three transects (see Figure 1 for locations) shows that the style of deformation and, in particular, dating of individual tectonic and metamorphic events differ substantially. It is not clear yet which of these differences are real and which are due to different interpretations. A three-dimensional interpretation of NRP-20 results, also including the results of the ECORS-CROP traverse, is still lacking. A first step in this direction was made by Valasek and Mueller [1997] with regard to the major seismic crustal features imaged by NRP-20 data. This synthesis, however, does not consider additional geophysical data in the western Alps, it does not properly take into account the Ivrea geophysical body, and it does not include the shallow geological features. Older pioneering work by Steck [1984,

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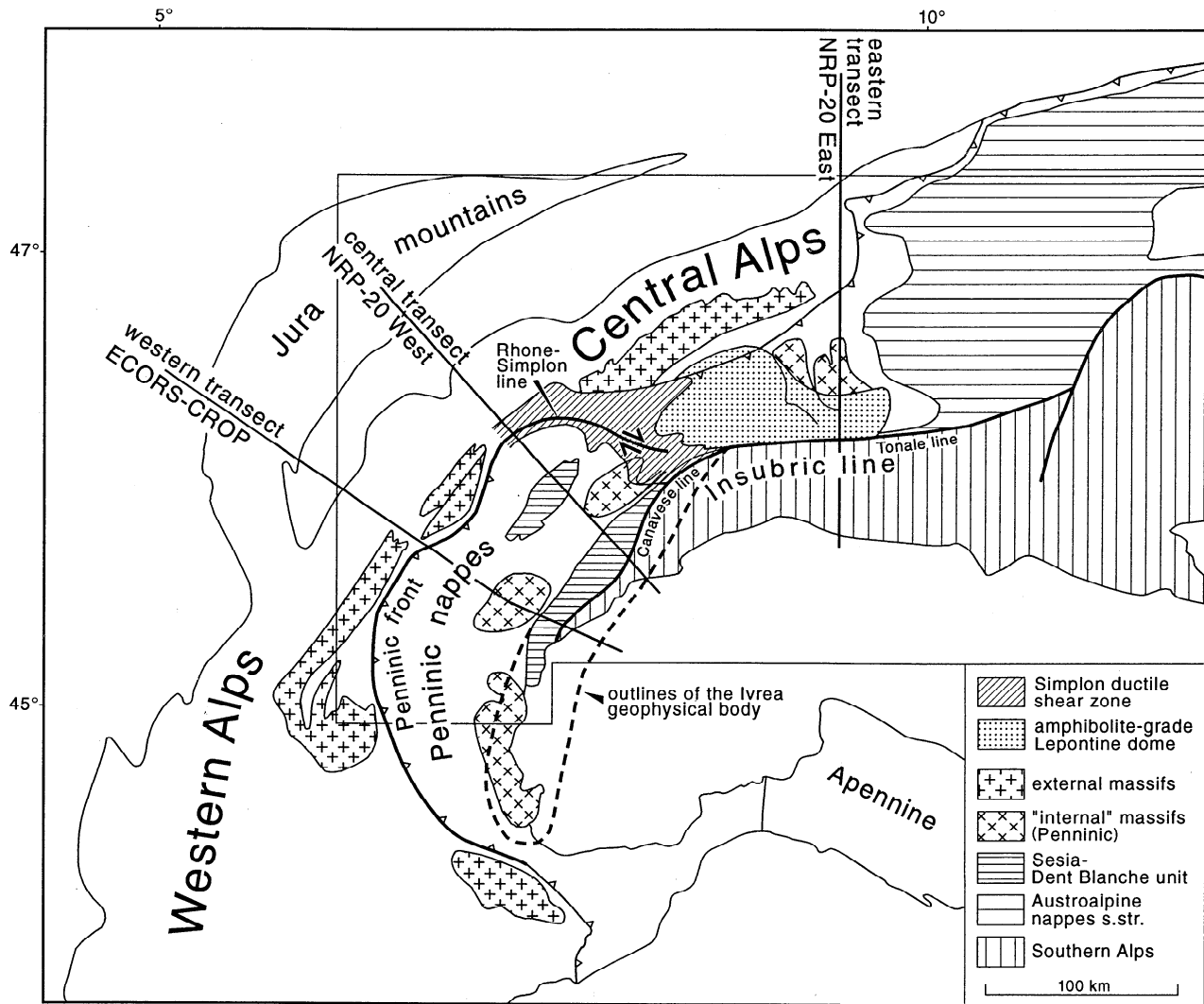


Figure 1. Sketch map of the Alps indicating locations of the three geophysical-geological transects, major tectonic features discussed in the text, and the region covered by the contour map depicted in Figure 5. ECORS-CROP, Etude Continentale et Océanique par Reflexion et Refraction Sismique - Progetto Strategico Crosta Profonda; NRP-20, National Research Project 20.

1990] recognized the importance of the Simplon ductile shear zone (Figure 1), which crosses the Alps in an oblique direction. This shear zone is intimately linked to the transition of the central to the western Alps and the formation of the Western Alpine arc, which is the main topic of the present contribution.

2. Tectonic Reinterpretation of the ECORS-CROP Profile of the Western Alps

2.1. Deep Structure

Previous interpretations of the deep structure along the ECORS-CROP profile heavily relied on results of a wide-angle seismic experiment consisting of five radial fans [Hirn *et al.*, 1989; Thouvenot *et al.*, 1990]. These data, together with results from gravity modeling [Bayer *et al.*, 1989, 1996], led

Nicolas *et al.* [1990a, b] to propose the existence of an intermediate Moho reflector at a depth of about 30 km beneath the Gran Paradiso massif (reflector II in the work of Nicolas *et al.* [1990b, Figure 4]), underlain by a mantle wedge which connects eastward with the mantle part of the Ivrea geophysical body. Nicolas *et al.* [1990a, Figure 7] proposed that this mantle wedge was sliced off from the European lithosphere during lithospheric-scale wedging. Alternatively, Roure *et al.* [1990, Figure 6] interpreted this mantle wedge below reflector II to directly connect with the mantle of the Adriatic lithosphere, indicating wedge-shaped indentation of Adriatic lithosphere into a split-apart within the European lithosphere, a model similar to interpretations along the NRP-20 East profile [Pfiffner, 1992; Schmid *et al.*, 1996]. More recently, Roure *et al.* [1996a, Figure 1b] proposed that the high-density material beneath reflector II, whose existence is indicated by gravity data [Bayer *et al.*, 1989, 1996], does not

represent mantle material at all but could be made up of duplexes derived from European lower crust. In fact, this third option offered by Roure *et al.* is in fairly close agreement with our reinterpretation presented below.

The reinterpretation depicted in Plate 1 is primarily based on migrated high-angle seismic reflection data from the ECORS-CROP profile, provided by *Sénéchal and Thouvenot* [1991], and differs from an earlier geological interpretation based on the same data [*Marchant*, 1993]. The migration is based on the principle of the common tangent of two spherical wave fronts. The migrated line drawings, which are particularly suitable for geological interpretation (Figure 2), are taken from *Thouvenot et al.* [1996]. Note that the migration depicted in Figure 2 follows a tortuous trace along the geophone line (see location map in Plate 1 of *Roure et al.* [1996b]). To arrive at the interpretation depicted in Plate 1, the key reflectors from Figure 2 have been orthogonally projected into a straight profile trace (coordinates indicated in Plate 1). For our interpretation we also used information from wide-angle reflection and refraction seismic experiments [*Ansorge*, 1968; *Hirn et al.*, 1989; *Thouvenot et al.*, 1990]. Rather than projecting the wide-angle reflection data on the ECORS-CROP profile [*Thouvenot et al.*, 1990], however, we followed the procedure outlined by *Kissling* [1993] and *Waldhauser et al.* [1998] to establish a 3-D crustal model with the experimentally determined seismic reflector elements located in those places where they were actually imaged. Note that this procedure sometimes results in large information gaps but also in more reliable seismic structural models in those areas covered by controlled source seismology profiling.

The depth of the European and Adriatic Moho is well constrained at the locations indicated in Plate 1 by 3-D interface modeling of two-dimensional seismic data [*Waldhauser et al.*, 1998]. The top of the European lower crust between points A and B in Plate 1 is placed at the top interface of a band of strong reflectors visible on the migration of reflection data [*Thouvenot et al.*, 1990] along the ECORS-CROP profile. The position of reflections visible around point F (Figure 2) is close to that of reflector II, detected by the wide-angle reflection experiment, when this reflector II is projected 30 km laterally onto the ECORS-CROP profile [*Hirn et al.*, 1989]. For reasons discussed later, we interpret these strong reflections around point F (and reflector II of the wide-angle experiment) to represent the interface between lower and upper European crust, rather than the top of a mantle slice, as proposed by *Nicolas et al.* [1990a, b]. Note, however, that these reflections around point F are situated slightly above the proposed top of a thin mantle slice, according to the gravity model by *Bayer et al.* [1989, 1996]. The outlines of this mantle slice with a density of 3200 kg m^{-3} are shown for comparison only (dashed lines in Plate 1). According to our interpretation, the area below point F and down to the European Moho is entirely made up of European lower crust. This results in a duplication in total thickness of European lower crust compared to that found in the foreland, i.e., below points A and B. The gravity effects of a thin very high density layer (i.e., the mantle slice by *Bayer et al.* [1989] and of a thick moderately high density layer are undistinguishable within the framework of the ECORS-CROP gravity profile, and hence, our interpretation is in accordance with the observed

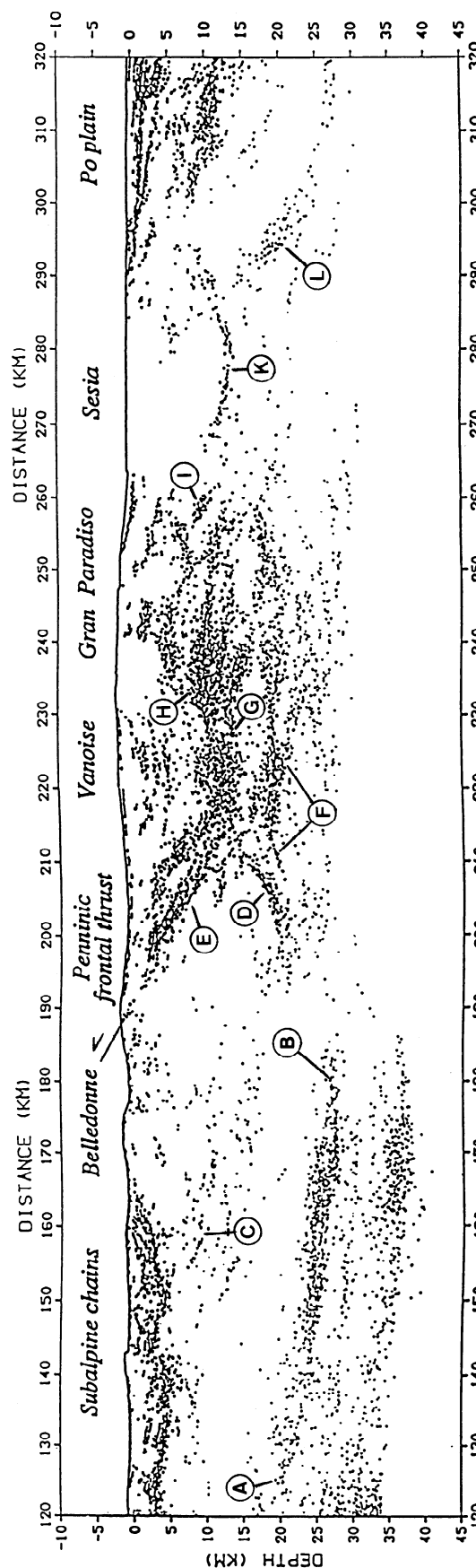


Figure 2. Migration by the common tangent method (CTM) of a part of the ECORS-CROP transect as published by *Thouvenot et al.* [1996]. Labels of some key reflectors allow for comparison with the geological interpretation of Plate 1.

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gravity (R. Bayer, oral communication, 1997). Unfortunately, the European lower crust loses its characteristic high-reflectivity pattern ESE of point B while the Moho wide-angle reflections derived from refraction data can be followed farther to the ESE. Consequently, the exact geometry of the thickening of the European lower crust below point F is not constrained by seismic data, except for Moho depth, and we are aware of the fact that this geometry might be considerably more complex than that depicted in Plate 1.

The results of 3-D tomography based on well-locatable seismic events [Solarino *et al.*, 1997] are crucial for the reinterpretation discussed above and presented in Plate 1. Plate 2 is a 2-D section through the 3-D tomographic model along the trace of the profile depicted in Plate 1. The NW boundary of the pronounced high-velocity perturbation, corresponding to the Ivrea geophysical body (a thick slab of predominantly mantle material [Kissling, 1984]), appears steeply east dipping to subvertical, not only in the section of Plate 2 but everywhere within the 3-D model of Solarino *et al.* [1997]. Plate 2 indicates that the high-velocity Ivrea body rather abruptly terminates toward the WNW, where an adjacent low-velocity perturbation is found at a depth of 20-35 km. Hence the volume of rocks beneath the reflector at point F, interpreted to be made up of European lower crust in Plate 1, is clearly separated from the Ivrea mantle material down to a depth of 45 km by low-velocity material interpreted to mark the suture between the European and Adriatic lithospheres (Subpenninic nappes and Valais and Piemont-Liguria sutures).

The strong group of reflectors between points G and H (Figure 2 and Plate 1), situated above the reflector at point F, directly connects with two bands of high reflectivity (at point E in Figure 2 and Plate 1). Updip projection to the surface shows that these reflections near point E correlate with the Valais suture zone and the Penninic frontal thrust at its base [Mugnier *et al.*, 1996; Marchant, 1993; Fügenschuh *et al.*, 1999]. This geometry indicates that the band of high reflectivity, interpreted to mark the Valais suture zone, flattens out eastward. Hence all material below reflectors at points E and G is inferred to be of more external origin with respect to the Valais suture, and therefore we assign it to the European lithosphere (Plate 1). This geological interpretation lends further support for attributing the volume of rocks below point F to the European lower crust.

The top of the Ivrea mantle rocks (corresponding to reflector III of the wide-angle experiment [Nicolas *et al.*, 1990a]) is well visible on the migrated section (reflector at point K in Figure 2 and Plate 1). The SE dipping band of strong reflectors at point L (Figure 2 and Plate 1), in combination with the location of the Adriatic Moho given by Waldhauser *et al.* [1998], is taken to represent the Adriatic lower crust, directly overlying the Ivrea mantle material. According to this interpretation, the NW edge of the Adriatic Moho is uplifted and may be directly correlated with the top of the Ivrea mantle slab, dissected by a group of back thrusts below the Po plain, which are drawn in Plate 1 according to the interpretation given by Roure *et al.* [1990].

One of the key features, particularly well visible in unmigrated line drawings of the ECORS-CROP seismic

reflection profile [Damotte *et al.*, 1990], is a group of NW dipping reflectors (point W in Figure 3c), also visible in the migrated data (point D in Figure 2). This group of reflectors is discordant with respect to several SE dipping events visible in Figure 3c: (1) the European Moho (point M in Figure 3c) and the top of the "autochthonous" European lower crust (point C in Figure 3c) situated below and (2) the Pennine frontal thrust and the Valais suture zone (point P in Figure 3c and equivalent to point E in Figure 2) situated above reflector group W, respectively. In Plate 1 this group of NW dipping reflectors (near point D in Plate 1) is taken to mark a foreland-dipping shear zone, or back thrust, above the tip of the allochthonous slice of European lower crust.

Although a discussion of the dynamical aspects of orogeny is beyond the scope of this contribution, focusing on geometrical and kinematic aspects, the interpretation of the deep structure depicted in Plate 1 suggests that the allochthonous slice of European lower crust might have acted as a flow-resistant wedge. The shear zone near point D in Plate 1 is interpreted to balance wedging of allochthonous European lower crust into the interface between lower and upper European crust, in analogy to a scheme proposed by Roure *et al.* [1990, Figure 7] for the ECORS-CROP section. Wedging of the lower crust with similar geometry was also postulated and discussed in detail with respect to the NRP-20 East transect [i.e., Schmid *et al.*, 1996]. This wedging would imply a flow-resistant feldspar-dominated rheology within the lower crust and decollement at its upper interface and near its base (Moho), respectively. Such a postulate is not unrealistic in the light of the concept of rheological stratification applied to a two-layer (quartz- and feldspar-dominated) crust [see Ranalli and Murphy, 1987; Handy, 1989]. Below we will make use of the term "lower crustal wedge" rather than "wedge-shaped geometry of lower crust" being aware that given our geometric-kinematic approach, rheological inferences remain speculative.

2.2. Near-Surface Structures

The geology of the NW foreland (Jura and chaînes subalpines) in Plate 1 is drawn according to that of Guellec *et al.* [1990] and Mugnier *et al.* [1996]. The suspected frontal thrust of the Belledonne massif (reflector group C in Figure 2) can only take up part of the total shortening of 55-60 km indicated by profile balancing (some 30 km in the Jura cover sequence plus around 28 km in the chaînes subalpines according to Guellec *et al.* [1990]). Hence much of the shortening within this Mesozoic cover has to be taken up in the internal Belledonne and the Mont Blanc massifs, including the Penninic frontal thrust (see discussion by Mugnier *et al.*). Shortening within the NW foreland occurred during an early Miocene phase (post-25 Ma, probably during the Aquitanian) in the case of the frontal thrust of the chaînes subalpines but during a later phase (late Miocene to Pliocene) in the case of the Jura mountains (stratigraphic constraints discussed by Guellec *et al.* [1990] and Mugnier *et al.* [1990, 1996]).

The Valaisan domain, which we consider a separate and more external oceanic suture zone with respect to the Piemont-Liguria ocean [Frisch, 1979; Stampfli, 1993; Froitzheim *et*

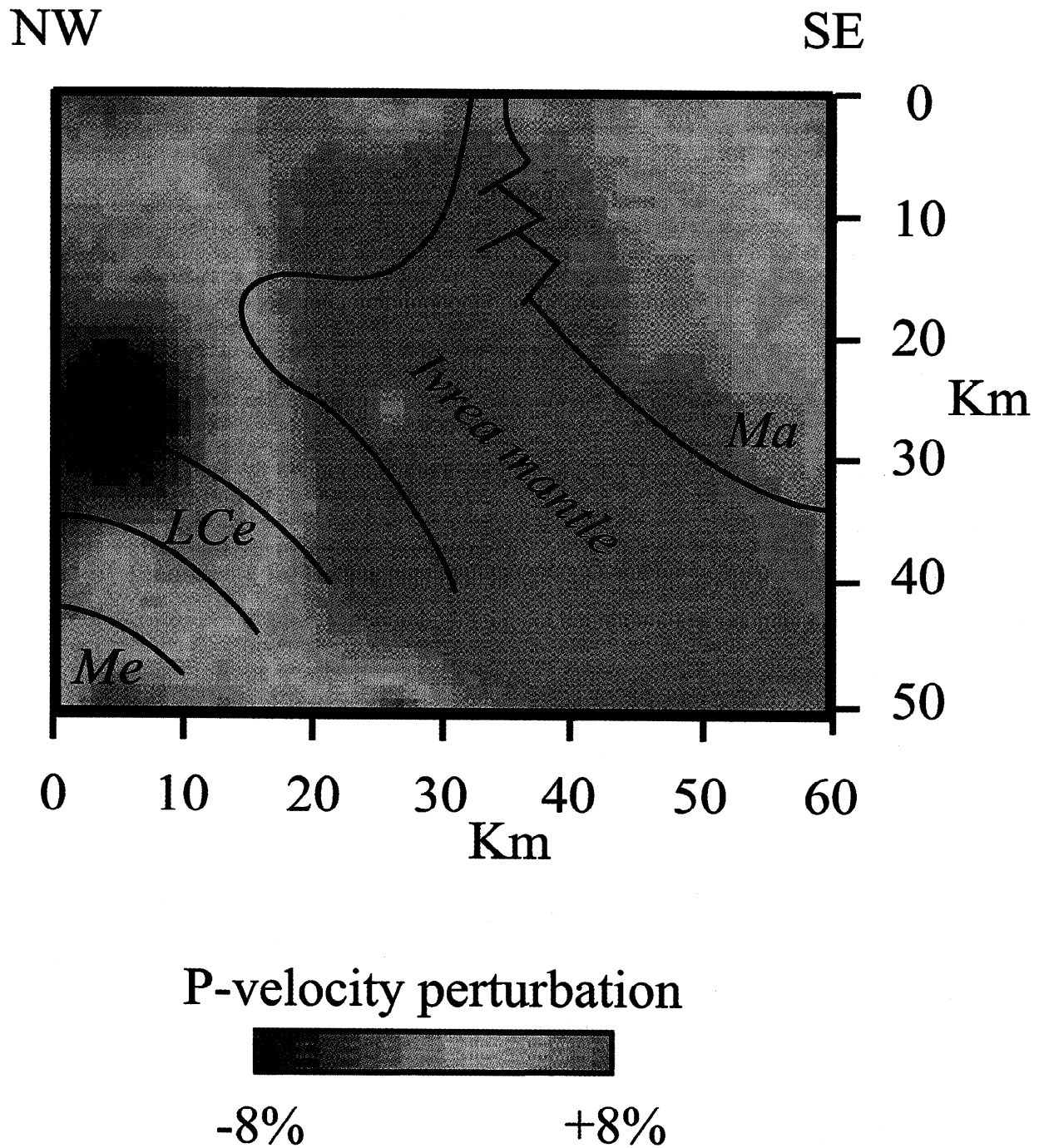


Plate 2. Vertical cross section along the SE part of the ECORS-CROP transect (Plate 1) through the 3-D P wave velocity field in NW Italy obtained by seismic tomography [Solarino *et al.*, 1997]. The outlines of some key structures, taken from the geological interpretation given in Plate 1, are superimposed for comparison. Ma, Adriatic Moho; Me, European Moho; LCe, European lower crust. Note that the structural model presented in Plate 1 and schematically depicted here has been derived by using controlled source seismic, geological, and gravimetric information. Hence Plate 2 allows comparison of our proposed structural model with independent tomographic information. With regard to the maximal resolution capabilities of the seismic tomography model of about 10 km [Solarino *et al.*, 1997], the two models correspond very well.

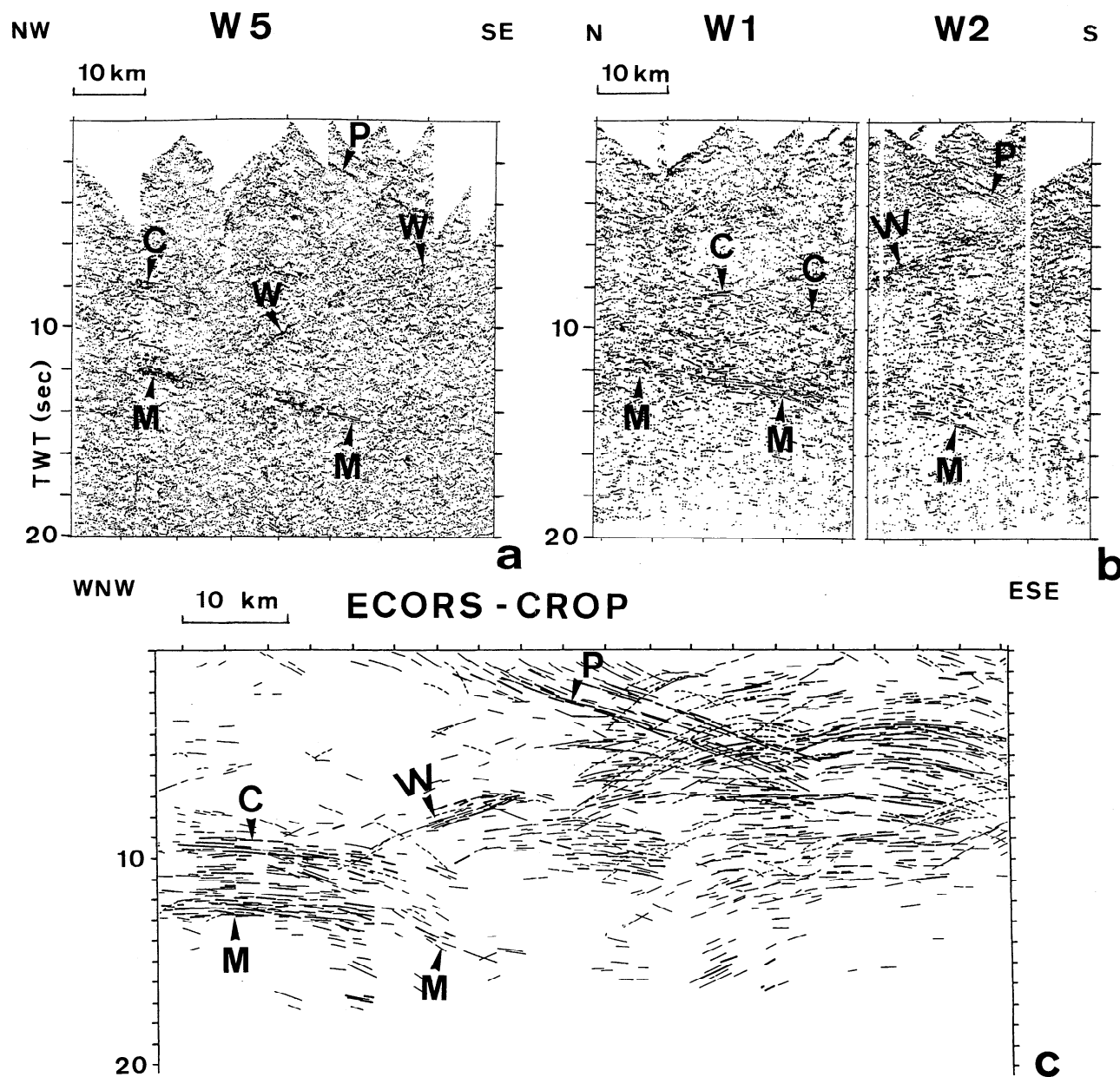


Figure 3. Comparison of unmigrated seismic reflection profiles whose locations are indicated in Figure 5. The reflectors discussed in the text and contoured in Figures 5 and 6 are the following: M: European Moho; C: top European lower crust (equals point B in Figure 2); W, top European lower crustal wedge (equals point D in Figure 2); and P, Penninic front (equals point E in Figure 2). (a) NRP-20 West line W5 (from Plate 12-9 of Steck *et al.*, [1997]), (b) NRP-20 West lines W1 and W2 (from Plates 12-1 and 12-3, respectively, of Steck *et al.*, [1997]), (c) ECORS-CROP profile [from Damotte *et al.*, 1990]. TWT, two-way travel time.

al., 1996], is thrust onto the allochthonous cover of the Mont Blanc massif (the Ultra-Dauphinois) along the most spectacular feature depicted by reflection seismics: the Penninic frontal thrust (reflectors near point E in Plate 1). Details within the Valaisan domain (Plate 1) are drawn according to a recent structural analysis by Fügenschuh *et al.* [1999]. Updip projection of the two bands of high reflectivity (near point E in Figure 2) indicates that the two bands correspond to the Ultra-Dauphinois cover below and the

Versoyen unit (prasinities, interbedded with shales) above the Pennine frontal thrust (point P in Figure 3). Hence this late thrust, including the Versoyen unit, can be followed down to a depth of almost 15 km before it flattens out according to our interpretation of the deep structure. The exact age of the Penninic frontal thrust is poorly constrained. It certainly postdates the deposition of Priabonian-age nummulitic limestones of the Ultra-Dauphinois (i.e., it postdates 36 Ma). Since it appears that this thrust is indeed kinematically related

to some of the shortening within the chaînes subalpines [Mugnier *et al.*, 1990], its activity probably does not start before the early Miocene (25 Ma). Note that we use the term "Penninic frontal thrust" (see Figure 1, brief "Penninic front") to denote a late stage WNW directed thrust which postdates earlier (i.e., latest Eocene) top north stacking within the Valais domain [Fügenshuh *et al.*, 1999], linked to decollement and thrusting of more externally located Penninic cover nappes, such as, for example, the Préalpes Romands [Mosar *et al.*, 1996; Bagnoud *et al.*, 1998].

The large-scale structure within the more internal Penninic units is drawn (Plate 1) according to available geologic maps [e.g., Bigi *et al.*, 1990; Debelmas, 1979; Debelmas *et al.*, 1991] and literature data [e.g., Ballèvre and Merle, 1993; Baudin, 1987; Butler and Freeman, 1996; Caby, 1996; Elter, 1972; Fabre, 1961; Gouffon, 1993; Vearncombe, 1985; Wheeler and Butler, 1993]. A group of reflectors near point I (Figure 2 and Plate 1) is tentatively taken to mark the base of the Gran Paradiso thrust sheet. The NW dipping foliations, prominent within outcrops of the back-folded and back-thrusted Rutor and Vanoise-Mont Pourri basement units [Caby, 1996; Butler and Freeman, 1996], are too shallow to be depicted in the seismic reflection profile (Figure 2). A maximum age of 34 Ma has been inferred for one of these back thrusts (Entrelor shear zone) by radiometric dating [Freeman *et al.*, 1997]. The minimum age of this greenschist facies back thrust is indirectly constrained by zircon fission track cooling ages (30 Ma) from the Gran Paradiso massif in its footwall [Hurford *et al.*, 1991], indicating that temperatures fell below those required for mylonitization under greenschist facies conditions before 30 Myr ago.

The region to the SE of the Gran Paradiso unit has been drawn according to the interpretation given by Roure *et al.* [1990]. According to these authors, SE directed thrusting, detected in the subsurface of the Po plain, is of early Neogene age (around 25-22 Ma). Thrusting affects Oligocene to Aquitanian deposits (Gonfolite group) and is sealed by a Burdigalian unconformity. Post-Oligocene verticalization of the Ivrea zone, related to this back thrusting, is compatible with paleomagnetic evidence from Oligocene dikes intruding the western margin of the Ivrea zone [Schmid *et al.*, 1989].

2.3. Summary of Timing Constraints

The constraints available suggest that all the major structures characterizing the large-scale geometry of the profile depicted in Plate 1 formed during Oligocene and Neogene times (i.e., after 35 Ma). Hence most of the deformation visible in the NW-SE to WNW-WSE oriented ECORS-CROP section postdates early Tertiary collision and nappe stacking which led to at least 350 km N-S directed convergence in the eastern central Alps between 65 and 35 Ma [Schmid *et al.*, 1996, 1997]. Tectonic activity related to Tertiary collision certainly also affected the western Alps, but since this earlier nappe stacking is suspected to be associated with top-north displacements [Fügenshuh *et al.*, 1999], it cannot be properly imaged in the ECORS-CROP section.

A first postcollisional episode imaged in the ECORS-CROP section can be dated between 34 and 30 Ma. It led to back thrusting of the originally more external Vanoise-Mont Pourri basement over the more internal Gran Paradiso and

Grivola units. Probably, this back thrusting is kinematically linked and contemporaneous with a forethrust at the base of the Gran Paradiso thrust sheet (near point I in Plate 1). Following Butler and Freeman [1996], we interpret this linked fault system to have accommodated the WNW directed final emplacement of the Gran Paradiso thrust sheet.

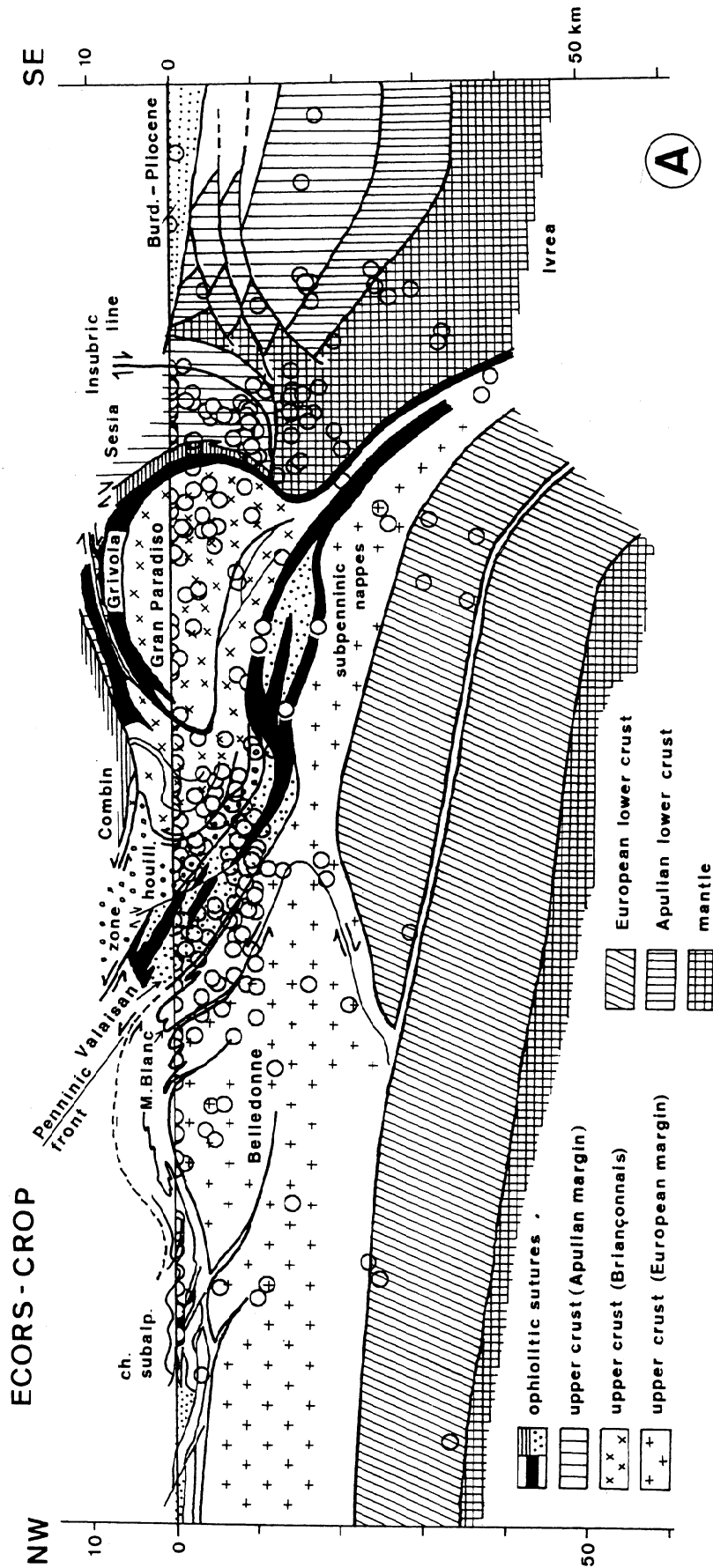
A second episode during the early Miocene is again associated with simultaneously active fore-thrusts and back thrusts (Penninic front and thrusting below the Po plain, respectively). A third episode beginning at about 12 Ma is related to thrusting in the external Belledonne massif and shortening within the Jura mountains; that is, the thrust front jumped by some 80 km into the foreland [Burkhard and Sommaruga, 1998].

The timing of some of the inferred deep structures beneath the Mont Blanc and Gran Paradiso massifs (back thrust near point D and duplication of European lower crust below point F in Plate 1) remains speculative. Since we assume that detachment of the European lower crustal wedge is contemporaneous with back thrusting near point D and forethrusting along the Penninic frontal thrust (see later discussion of the kinematic reconstruction in section 6 and Figure 7), these features are most likely to be associated with the second episode of postcollisional shortening during the Miocene.

3. Comparison With NRP-20 Transects Through the Central Alps

The locations of all the Alpine transects discussed in this section are indicated in Figures 1 and 5. In a first step we compare the ECORS-CROP profile (briefly referred to as the "western" transect) with the NRP-20 East transect ("eastern" transect [e.g., Pfiffner and Hitz, 1997; Schmid *et al.*, 1996]), situated east of the central Alpine Lepontine metamorphic dome and part of the "European Geotraverse" [Blundell *et al.*, 1992; Pfiffner, 1992]. The major features common to these two transects, schematically sketched in Figures 4a and 4c, are (1) ESE to south directed subduction of the European lithosphere, (2) a gap between European and Adriatic (or Apulian) Moho, and (3) the presence of wedge-shaped bodies of lower crust, largely decoupled from the piling up and refolding of thin flakes of upper crustal material (the Alpine nappes). However, there are also a number of substantial differences in geometry and kinematic evolution of the two transects (see Schmid *et al.* [1996] for an extensive discussion of the kinematic evolution along the eastern transect).

1. The Adriatic Moho descends northward and toward its contact with the lower (European) crust in the eastern transect, while this same Adriatic Moho rises toward the surface when approaching the contact zone with the European lithosphere in the western transect (in this case, toward the WNW, see Figures 5 and 6). This contrast finds its expression also in the surface geology. In the eastern transect the southern Alps form an impressive south vergent foreland fold and thrust belt riding above the Adriatic lower crust [Schönborn, 1992; Schumacher *et al.*, 1997], while this same Adriatic lower crust is exposed in the Ivrea zone [Schmid, 1993], situated at the ESE end of the western transect. The presence of lower crustal material in the Ivrea zone is the surface expression of the so-called Ivrea



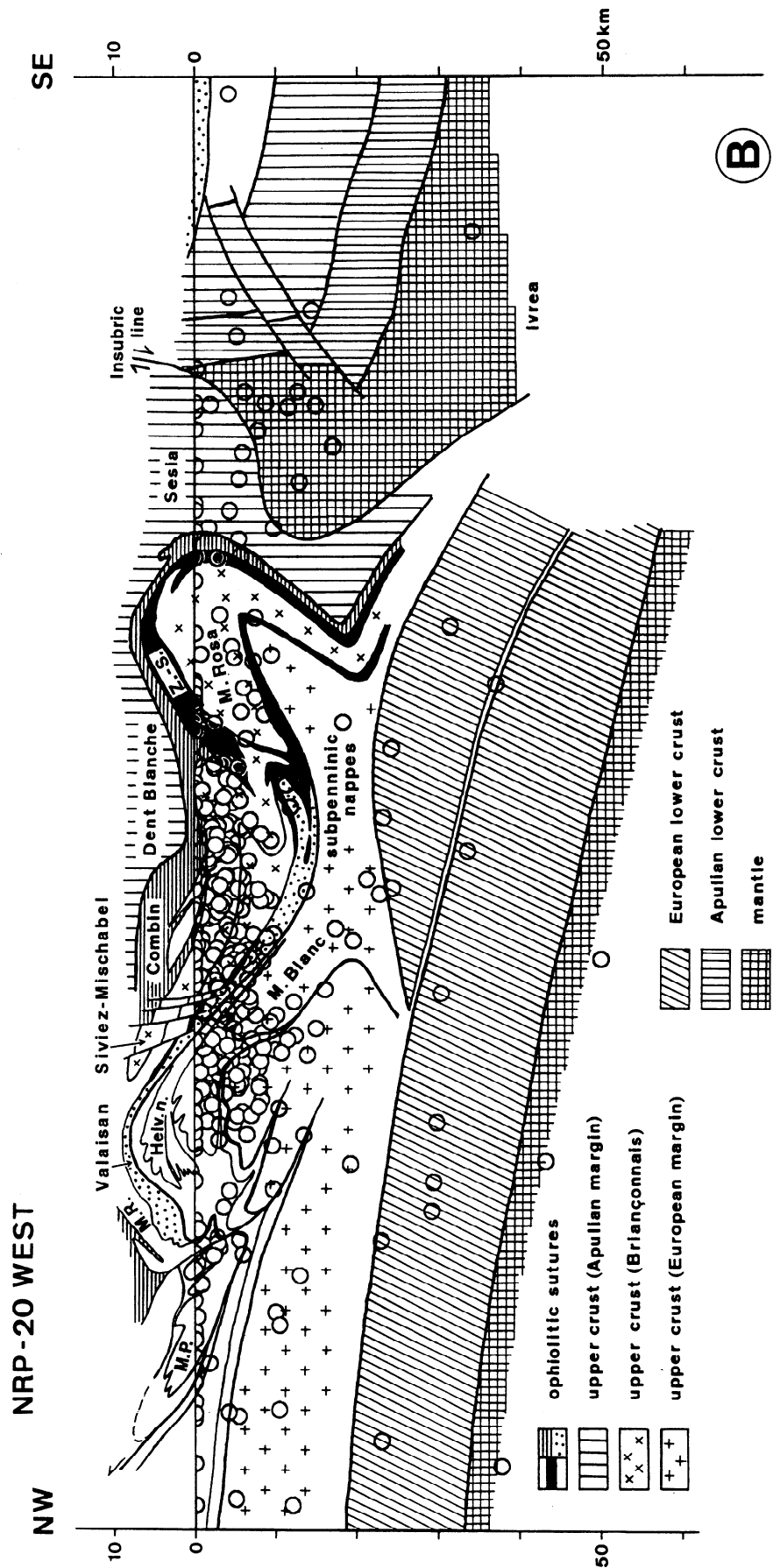


Figure 4. (Continued)

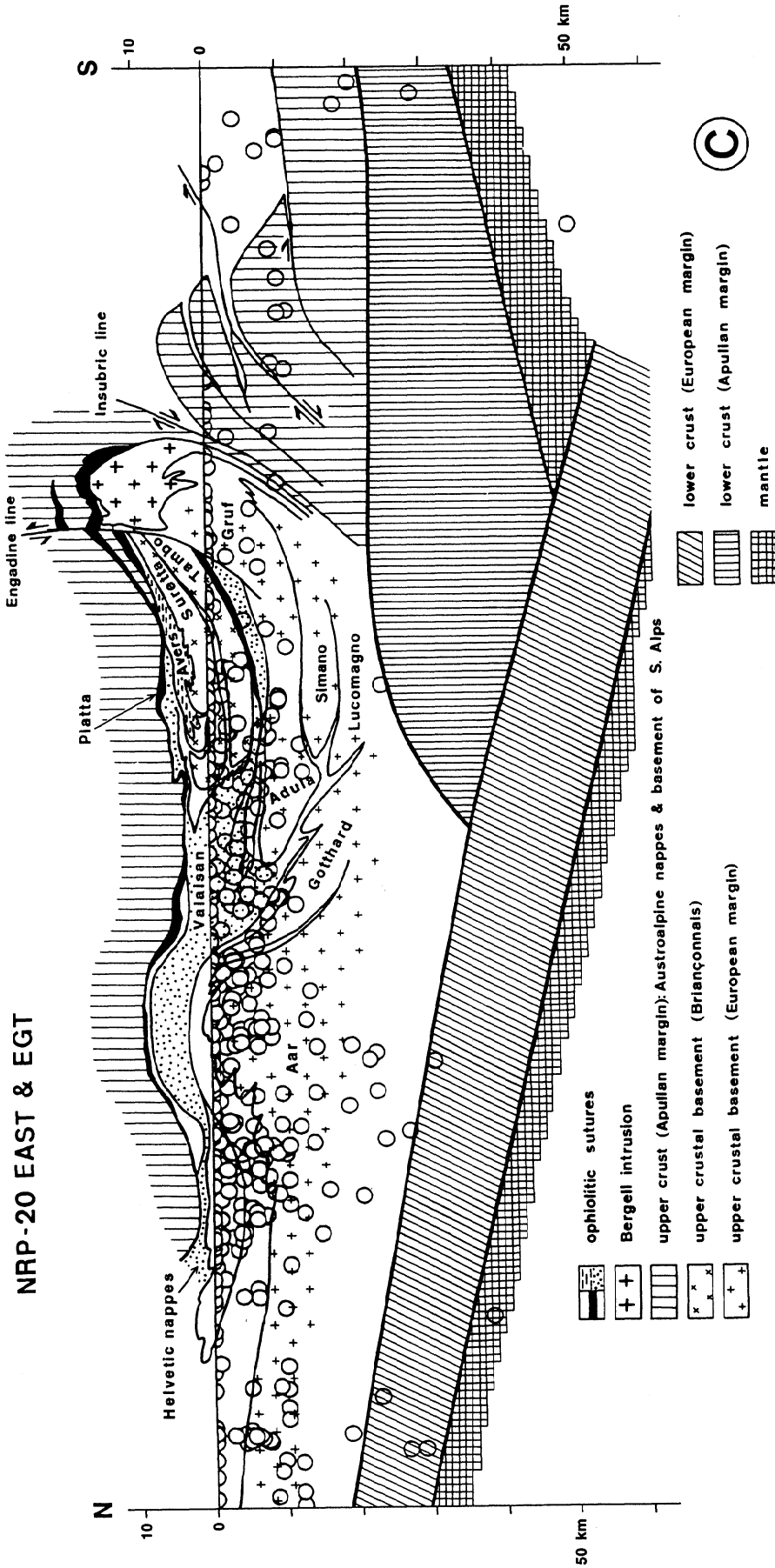


Figure 4. (continued)

body, a complex of high-velocity, high-density, and high magnetic susceptibility rocks of lower crustal and probably largely upper mantle ("Ivrea mantle" in Plate 1) origin [Kissling, 1984; Solarino *et al.*, 1997]. The Ivrea zone and Ivrea geophysical body wedge out eastward and do not extend into the area covered by the eastern transect. The subvertical to steeply inclined western margin of the deep-reaching Ivrea body (Plate 2) marks the western limit of the Adriatic microplate, which acted as a west moving indenter during Alpine orogeny [Laubscher, 1991], owing to the presence of high-strength Ivrea mantle material at shallow depth.

2. A number of geophysical and geological arguments extensively discussed by Holliger and Kissling [1992], Pfiffner *et al.* [1997], Schmid *et al.* [1996], and Valasek and Mueller [1997], and not repeated here, suggest the presence of a slice of Adriatic lower crust situated above European lower crust and below European upper crust at its northern tip (Figure 4c). According to the kinematic reconstruction of Schmid *et al.* [1996], this slice of Adriatic lower crust was wedged into the European lithosphere during the Miocene, opening a gap between lower and upper crust. According to the previously discussed interpretation of the western transect, a somewhat similar process of wedging is inferred for the ECORS-CROP transect (Fig. 4a). In this latter case, however, the lower crustal wedge is interpreted to be derived from the European lithosphere. Hence the two wedges of lower crust seen in the eastern and western transects, respectively, cannot be laterally connected according to these interpretations. The observation discussed above, which implies that the Adriatic lower crust descends below the Penninic nappe stack in the eastern profile while it rises to the surface in the western transect (see rising Adriatic Moho in vicinity of the Ivrea body in Plate 1), independently supports the conclusion that the two wedges are not laterally connected.

3. The eastern transect exhibits substantial back thrusting and backfolding of all the Penninic nappes, including the Valais suture zone, in an area immediately north of the Insubric line, associated with exhumation of the amphibolite-grade Lepontine dome [Schmid *et al.*, 1989] and the deep-seated Bergell intrusion [Rosenberg *et al.*, 1995; Berger *et al.*, 1996]. In the western transect, however, back thrusting does not affect the Valais suture zone and appears to be restricted to the units above this suture (within the Briançonnais upper crust, Figure 4a). The fact that Barrovian-type amphibolite-grade rocks have not been exhumed to the surface in the western transect, where the Insubric line only exhibits minor vertical offset during its activity [Schmid *et al.*, 1989], supports our interpretation given in Figures 4a and 4c and lends further support for postulating major differences regarding the deep structure of the two transects.

4. The orogenic lid of the Austroalpine nappes system (part of the Apulian passive margin), under which Penninic and Helvetic nappes were accreted in the eastern transect, is absent in the western transect west of the Insubric line, except perhaps for the Dent Blanche klippe. The Dent Blanche klippe, however, was interpreted to be part of an extensional allochthon, separated from the distal Apulian margin already during Jurassic rifting [Froitzheim *et al.*, 1996] and partly incorporated within the accretionary wedge formed by the ophiolitic Piemonte-Liguria units (eclogitic micaschists of the Emilius, Torre Ponton, and Glacier Raffrey units and the Sesia

zone in Plate 1). In the western transect all remnants of the Apulian passive margin remained within the Adriatic microplate and hence ESE of the Insubric line.

After this comparison between the eastern and western transects we now discuss a third transect (NRP-20 West profile referred to as "central" transect; see Figures 1, 4b and 5) in order to decide whether the major structural features of this transect fit more into the characteristics of one or the other of the transects discussed above. Note that both the quantity and quality of geophysical and, in particular, of seismic data on deep structure in this third transect are much reduced when compared to the eastern and western transects. Direct inspection of some unmigrated key seismic features imaging the deep structure below and in front of the Penninic frontal thrust in two NRP-20 seismic lines on either side of the central transect (lines W5 and W1-W2 in Figures 3a and 3b, for positions, see Figure 5) [Steck *et al.*, 1997] demonstrates remarkable similarities with the western profile (Figure 3c) [Damotte *et al.*, 1990]. In all three seismic profiles depicted in Figures 3a and 3b, a foreland-dipping major reflector is visible in a depth interval of 7-10 s two-way travel time (TWT) (point W in Figure 3). The reflector near point W abuts against top European lower crust of the descending European lithosphere (points M and C in Figure 3) underneath the hinterland-dipping Penninic front (point P in Figure 3). Hence the characteristic seismic structure of the crust found in the western transect seems to extend northwestward into and beyond the central transect (i.e. profile W1-W2). As in the case of the ECORS-CROP profile, we regard the reflector at point W in Figures 3a and 3b to be caused by the frontal tip of a slice of European lower crust, which therefore extends beyond the central transect but not into the eastern transect. Strong similarities between the western and central profiles are, in fact, expected since the central transect also runs through the Ivrea body at its SE margin. Hence it is not surprising to find close similarities between the western and central transects in respect to almost all the major features which were found to differ between the eastern and western transects.

This close similarity led us to also reinterpret the deep structure below the central traverse as proposed by Escher *et al.* [1997] in the light of our interpretation of the western traverse (compare Figures 4a and 4b). The near-surface geology given in Figure 4b, however, follows the interpretation of Escher *et al.*. The interpretation of the deep structure along the central profile depicted in Figure 4b is based on (1) the Moho depths as given by Waldhauser *et al.* [1998], only modified in the vicinity of the Ivrea body, (2) hand migration of the key reflectors depicted in Figures 3a and 3b and (3) the contour map depicted in Figure 5, which will be discussed in section 5. The interpretation of the deep structure differs from that given by Escher *et al.* in that a large part of the space between the European Moho and the base of the Valais suture zone is taken up by duplicated European lower crust rather than by complexly refolded subpenninic nappes.

4. Topology of Deep Crustal Features in the Transition Zone Between Western and Central Alps

The contour maps of Figures 5 and 6 depict the 3-D shape of some important interfaces characterizing the deep structure

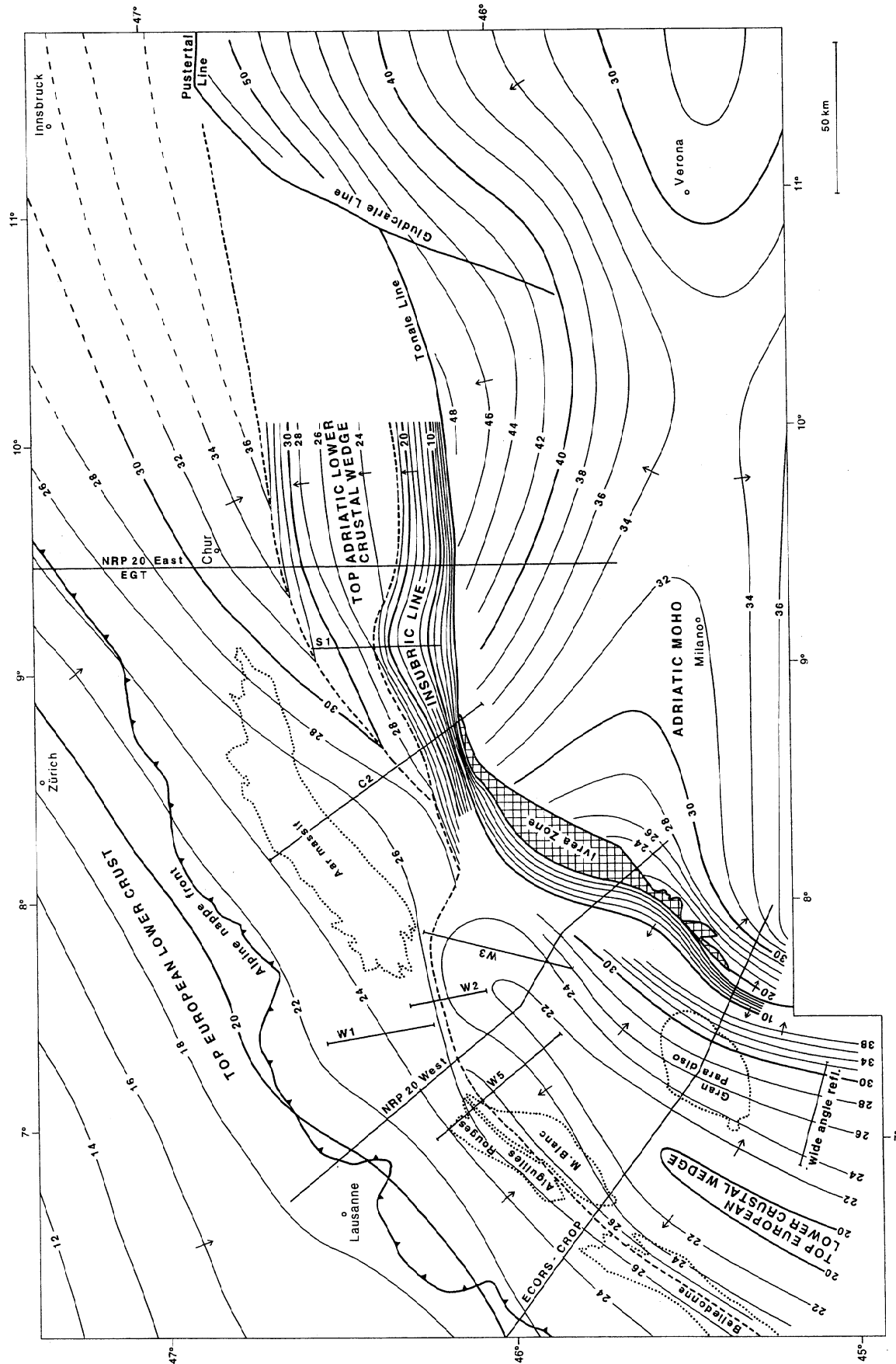


Figure 5. Contour map of deep crustal structure in the western Alpine region (depth in kilometers). See text for discussion.

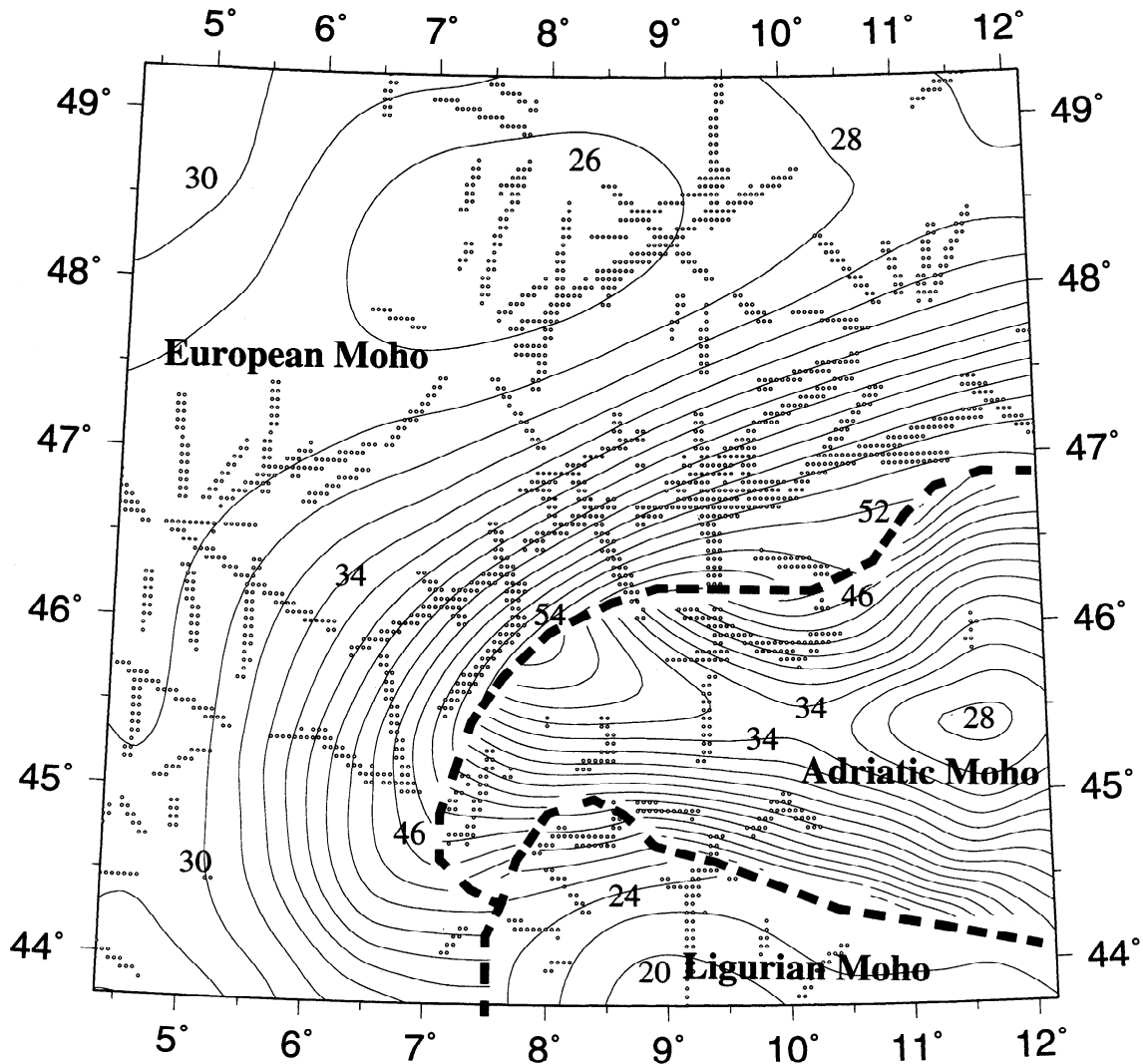


Figure 6. Map of crust-mantle boundary with contour intervals of 2 km [Waldhauser *et al.*, 1998]. Dashed line denotes boundaries between the European, Adriatic, and Ligurian Mohos. Shaded areas denote regions sampled by controlled source seismic profiling.

in the hinge zone between the N-S striking western Alps and E-W striking eastern Alps. These contours are drawn according to the principle of deriving the smoothest and laterally most continuous interfaces possible [Kissling, 1993; Waldhauser *et al.*, 1998], and they are based on the following data.

The depth of the European Moho and the Adriatic Moho (Figure 6) is constrained by a new approach to infer 3-D model topography of seismic subsurfaces using 2-D derived controlled source seismic reflector data [Kissling *et al.*, 1997; Waldhauser *et al.*, 1998]. Kissling *et al.* [1997] and Waldhauser *et al.* [1998] concluded that the European and Adriatic Moho are offset in an area roughly coinciding with the Insubric line (Figure 6). In Figure 5 the contours of the Adriatic Moho have only been slightly modified by us in a seismically poorly constrained area near the eastern margin of the Ivrea body according to geological criteria and in the light of recent results from local earthquake seismic tomography [Solarino *et al.*, 1997]. The Insubric fault plane is contoured according to

field data, migrated seismic reflection lines S1 and C1 [Valasek, 1992], and 3-D gravity profiles obtained from 3-D modeling of the Ivrea body [Kissling, 1980; Schmid *et al.*, 1996]. The top interface of the Adriatic lower crustal wedge traversed by the eastern transect intersects the downward projection of the Insubric fault plane and is controlled by seismic reflection lines S1 and C1 [Holliger and Kissling, 1992; Valasek, 1992]. The top interface of the allochthonous European lower crust, visible in the western and central transects, is contoured on the basis of our reinterpretation of the ECORS-CROP profile and of the hand-migrated reflectors depicted in lines W1, W2, W3, and W5 (Figure 3 and the work of Steck *et al.* [1997]) in the vicinity of the central transect. The top interface of the autochthonous European lower crust is largely contoured after Valasek [1992] and Valasek and Mueller [1997]. Their contours were modified in the vicinity of the intersection with the top of the allochthonous wedge of European lower crust.

The following important features emerge from inspection of Figures 5 and 6.

1. The Adriatic Moho, sandwiched between the European Moho to the north and west and the Ligurian Moho to the south (Figure 6), is updomed below the Po plain and uptilted (Figure 5) in the vicinity of the Ivrea zone [see also *Giese and Buness*, 1992; *Ye et al.*, 1995]. There is therefore an important topological change in the Adriatic Moho surface in the vicinity of the point where the Insubric line changes from an E-W strike (Tonale line) to a NE-SW strike (Canavese line). Westward, the northerly dip toward the central Alps gradually changes into an easterly to southeasterly dip away from the central and internal western Alps. Given the resolution of the data, however, these contours may in reality change orientation in a more discontinuous manner.

2. The western termination of the Adriatic lower crustal wedge under the central Alps, underthrusting the Insubric fault plane according to our interpretation (Figure 4c), roughly coincides with the eastern termination of the European lower crustal wedge under the western Alps (Figures 4a and 4b). It also coincides with the area of bending of the Insubric line (Figure 5).

3. The line of intersection between "autochthonous" and "allochthonous" European crust, i.e., the NNW tip of the European lower crustal wedge of the western Alps (Figure 5) follows an important shear zone visible at the Earth's surface: the Simplon shear zone (Figure 1) [Steck, 1990], which merges with the western termination of the EW running segment of the Insubric line (the Tonale line).

These features suggest that the central Alps - western Alps transition is a rather abrupt one, occurring at the western termination of the Tonale line, a site which closely coincides with the eastern termination of the Simplon shear zone. Hence both these shear zones, discussed in section 5, seem to play a key role for understanding the formation of the arc of the western Alps.

5. Dextral Strike-Slip Movements Along the Tonale and Simplon Shear Zones

The term "Tonale line" denotes the E-W striking segment of the Periadriatic or Insubric line (Figure 1), separating the southern and central Alps, which is offset by the sinistral Giudicarie line in the east and continues into the NE-SW striking Canavese line in the west [Schmid *et al.*, 1989]. The Tonale fault zone (including the northeasternmost segment of the Canavese line) represents a 1 km wide greenschist facies mylonite belt, which accommodated a combination of back thrusting and dextral strike-slip movements [Schmid *et al.*, 1989]. The maximum age of dextral movements along the Tonale line is given by radiometric age data from the eastern margin of the Bergell pluton (32-30 Ma [von Blanckenburg, 1992]), a deep-seated intrusion, whose subsequent exhumation is closely related to a combination of back thrusting and dextral strike-slip movements along the Tonale greenschist mylonites [Berger *et al.*, 1996; Schmid *et al.*, 1989]. A minimum age is provided by the offset of the Tonale line by the younger Giudicarie line. According to Schönborn [1992] and Schmid *et al.* [1996], this offset occurred at around 19-16 Myr ago, contemporaneous with post-early Miocene south

directed shortening in the Milan belt of the southern Alps, kinematically linked to underthrusting of the Adriatic lower crustal wedge underneath the Insubric fault plane and the central Alps (Figure 4c).

Two lines of evidence, however, indicate earlier dextral movements within an EW corridor situated immediately north of the future Tonale line. Firstly, ascent and final emplacement of the Bergell pluton in the time interval 35-30 Ma were shown to have taken place in a dextrally transpressive regime [Rosenberg *et al.*, 1995; Berger *et al.*, 1996; Davidson *et al.*, 1996] within a steep zone (the Southern Steep Belt or root zone of the Penninic and Austroalpine nappes). Secondly, dextral transpression in this Southern Steep Belt can be directly shown to mostly predate Oligocene magmatic activity farther to the west (between Domodossola and Locarno, i.e., in a transition zone between Tonale and Canavese lines). Oligocene dikes, dated between 29 and 26 Ma [Romer *et al.*, 1996], cut this amphibolite-grade dextral shear zone, which according to Steck and Hunziker [1994] was active in the 35-30 Ma time interval. We use the term "Tonale shear zone" in order to denote the totality of dextral movements occurring after 35 Myr ago and north of the Tonale fault zone.

The term "Simplon ductile shear zone" (Figure 1) is used to denote the totality of dextral movements which were initially distributed within an approximately 10 km thick deformation zone, which started to be active at around 35 Ma [Steck, 1984, 1990; Steck and Hunziker, 1994]. These movements subsequently were localized along the "Rhône-Simplon line" (Figure 1), a fault zone formed under retrograde (ductile to cataclastic) conditions. Mancktelow [1992] showed that the Simplon line in the strict sense (i.e., the NW-SE striking part of the Rhône-Simplon line) represents a normal fault which accommodates orogen-parallel extension that is contemporaneous with NW-SE shortening and crustal thickening. Thermal modeling [Grasemann and Mancktelow, 1993] suggests that normal fault displacement rates were particularly high between 18 and 15 Ma. Strain localization and normal fault activity along the Simplon line in the strict sense is the result of orogen-parallel extension related to the exhumation of the western part of the Lepontine metamorphic dome. In agreement with Steck and Hunziker [1994], however, we regard the Simplon shear zone as a first-order shear zone which was continuously active since 35 Ma. After the cessation of strike-slip movements along the Tonale line (19-16 Ma), ongoing dextral shear along the Rhône-Simplon line was kinematically linked to orogen-parallel extension in the central Alps and, as will be discussed in section 6, sinistral rotation of the Adriatic microplate. Fault plane solutions based on earthquake data [Maurer *et al.*, 1997] indicate that the Rhône-Simplon shear zone continues to be active up to the present day.

The area of transition between the Tonale and Simplon shear zones is a very broad one. Part of the ductile dextral movements along the Tonale mylonite belt is also taken up along the northernmost Canavese line. The amount of ductile back thrusting and dextral strike-slip movements along the Canavese line, however, continuously diminishes southwestward [Schmid *et al.*, 1989, Figure 7], dextral movement along the Tonale and Canavese lines being gradually transferred onto the Simplon shear zone.

We conclude that the Tonale and Simplon shear zones accommodated substantial dextral strike-slip movements

between the Adriatic microplate in the south and the Penninic realm of the central Alps situated north and east of the Simplon shear zone since 35 Myr ago. This is interpreted to imply that since that time, the western Alps situated south and west of the Simplon shear zone are kinematically part of the westward moving Adriatic microplate. Strain within the Simplon shear zone is distributed over a very wide area (up to 45 km in map view, Figure 1).

Unfortunately, an assessment of the amount of dextral displacement is difficult and can only be a very qualitative one. Regarding the Simplon shear zone, *Steck and Hunziker* [1994] provide an estimate of the order of 80 km, based on integrating strain magnitude over the width of the shear zone. Regarding the Tonale line, *Heitzmann* [1987] assumed that the Oligo-Miocene depocenter containing eroded Bergell boulders was originally situated south of the present-day outcrop area of the Bergell pluton, and he thereby estimated an offset of 60 km across the Tonale line. This is a minimum value, however, that does not take into account earlier pre-Bergell dextral displacements. *Laubscher* [1991] estimated a total of 150 km westward movement of the Adriatic microplate, 100 km being taken up along the Rhone-Simplon line and 50 km west of the Ivrea zone and along the Canavese line. This estimate is based on correlative features north and south of the Insubric line. *Lacassin* [1989] estimated a dextral offset of the order of 110 km on the Tonale shear zone, based on a large-scale kinematic analysis. In section 6 we will use an estimate of 100 km dextral strike slip across the Tonale and Simplon shear zones, being aware that this estimate is a very crude one.

6. Kinematic Evolution and Formation of the Arc of the Western Alps

The post 35 Ma dextral strike-slip movements along the Tonale and Simplon shear zones, together with the present-day topology of the deep crustal features (Figures 5 and 6) discussed in sections 4 and 5, suggest that the west directed component of thrusting in the western Alps was kinematically linked with the west directed component of movement of the Adriatic microplate, as has been proposed by previous authors [*Laubscher*, 1991; *Lacassin*, 1989; *Steck*, 1990]. A direct estimate of the amount of shortening along the western profile by profile balancing is difficult in view of the complex structure of the internal domains. The duplication in thickness of the European lower crust (Plate 1), however, indicates a minimum amount of 80 km shortening and the retrodeformation given in Figure 7 implies a total of 124 km shortening, an estimate which is based on the following simplified assumptions (see Figure 7).

1. Post-35 Ma shortening in the area along the western transect is assumed to be due to displacements parallel to this transect, i.e., towards azimuth 305°.

2. The displacement vector leading to shortening in the western transect is taken to be identical with the displacement of the Adriatic microplate (undeformed part of the basement below the Po plain) relative to stable Europe (AE in Figure 7). The amount of displacement (124 km) is obtained by assuming that the west directed component (100 km) of displacement between the Adriatic microplate and the central Alps (AC in Figure 7) represents the pure strike-slip component of

dextrally transpressive convergence. The 71 km N-S shortening in the central Alps north of the Insubric line (CE in Figure 7) is a corollary of these simplifying assumptions.

Assumption 1 is justified by the trend of the stretching lineations, which is dominantly WNW-ESE regarding post-high pressure (greenschist and subgreenschist facies) deformation in the area covered by the western profile [e.g., *Choukroune et al.*, 1986]. Additionally, the strike of the major tectonic features related to post-35 Ma shortening in the area of the western transect (i.e., external massifs, Penninic front, Ivrea zone, and northern part of the Ivrea geophysical body) is perpendicular to the transect. Assumption 2 is supported by the fact that movements along the Tonale shear zone are dextrally transpressive rather than pure strike slip [*Steck*, 1990; *Schmid et al.*, 1989; *Berger et al.*, 1996], and its validity may be checked by comparing the length of vector CE (71 km, see Figure 7) against the amount of NS shortening north of the Insubric line, as independently estimated on the basis of a kinematic reconstruction along the eastern transect [*Schmid et al.*, 1996]. *Schmid et al.* [1996] estimated a total of 119 km N-S shortening after 32 Ma, 56 km being due to shortening within the southern Alps. This leaves 63 km shortening north of the Tonale shear zone, a figure which fits quite well with the 71 km shortening resulting from the assumptions made above.

The retrodeformation along the western transect proposed in Figure 7 respects the depth of all major tectonic units, as given by *Hurford et al.* [1991] on the basis of cooling ages. Furthermore, it attempts to partition the estimated total of 124 km post-35 Ma shortening into the three episodes discussed in section 2. The 30 km post-12 Ma shortening during episode 3 corresponds to the shortening in the Jura mountains. Shortening associated with back thrusting of the Briançonnais basement over the Gran Paradiso unit between 35 and 30 Ma is estimated to be associated with westward thrusting of the Gran Paradiso unit during episode 1 (30 km and along a thrust marked by reflector I, depicted in Plate 1). This leaves about 60 km shortening, mainly taken up by early Miocene movements along the Penninic frontal thrust during episode 2.

The kinematic reconstruction of Figure 7 indicates two interesting features: (1) Detachment of the "alloclithonous" wedge of European lower crust appears to be a young feature, associated with the late stages of WNW directed motion of the Adriatic microplate. (2) Pre-35 Ma nappe stacking within the Penninic units, associated with the closure of the Valais and Piemonte-Liguria oceans, occurred in a relative position significantly farther to the east with respect to the external massifs.

Figure 8 represents an attempt to retrodeform the present-day position of some key tectonic features into their positions at 35 and 50 Myr ago, respectively. Regarding the first step of retrodeformation (post-35 Ma), an additional complication had to be built in. As mentioned above, retrodeformation along the eastern transect implies 56 km N-S shortening within the southern Alps south of the Tonale line, not taken into account by the vector triangle of Figure 7. As proposed by *Schönborn* [1992] and *Schmid et al.* [1996], this shortening is interpreted to be associated with the emplacement of the wedge of Adriatic lower crust underneath the Insubric line (see "top Adriatic lower crustal wedge" in Figure 5). Since this deformation is of middle Miocene age [*Schönborn*, 1992], it is

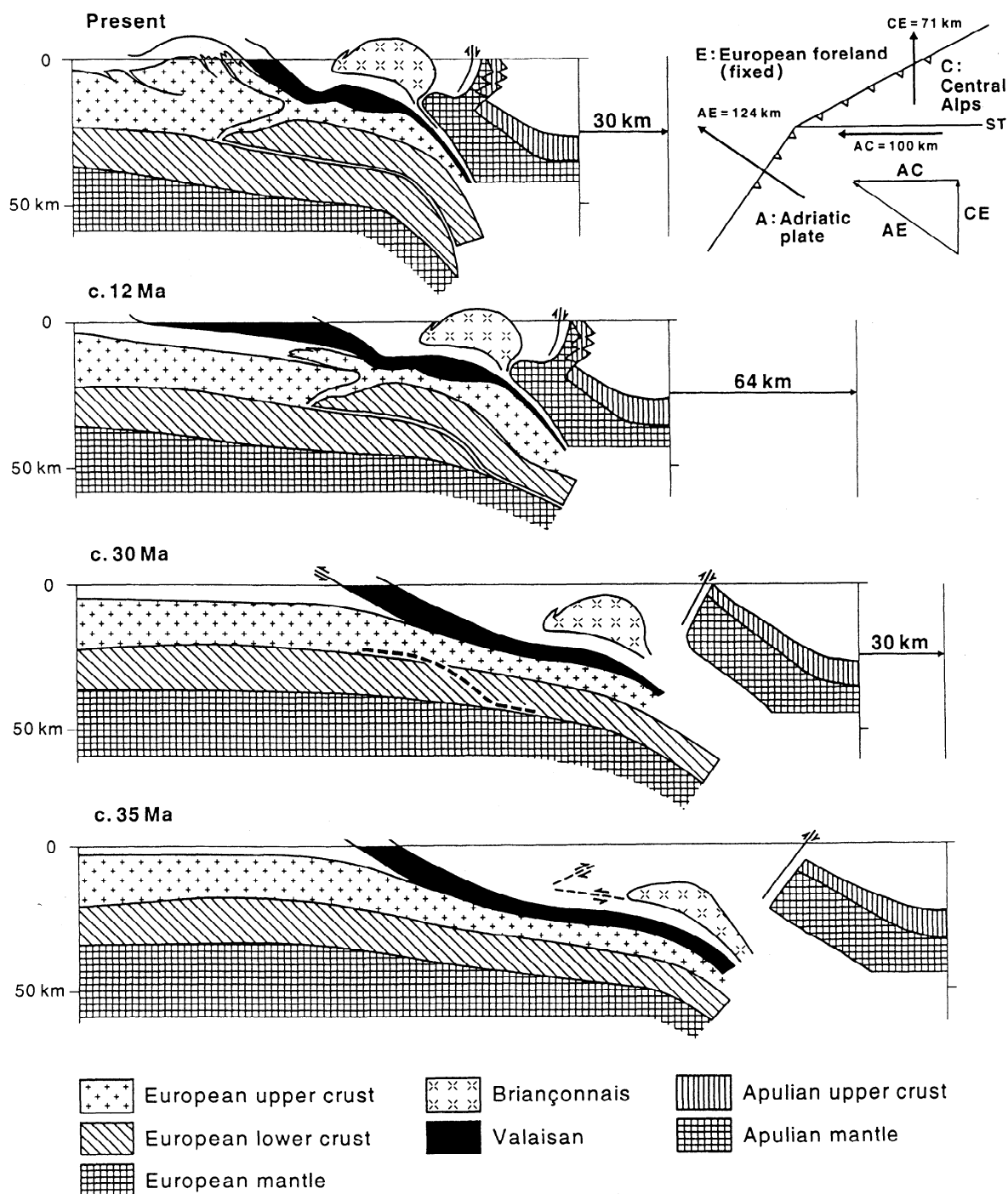


Figure 7. Kinematic reconstruction of postcollisional (post-35 Ma) shortening along the western transect in three episodes. The vector triangle (top right) illustrates how the assumed total amount of postcollisional shortening (124 km), corresponding to vector AE (movement of the Adriatic microplate in respect to Europe), was obtained from resolving 100 km dextral strike slip between the Adriatic microplate and the central Alps (vector AC) onto the western transect. Vector CE results in 71 km N-S shortening within the central Alps, parallel to the eastern transect. See text for discussion.

kinematically unrelated to earlier (early Miocene) shortening within the southern Alps at the eastern termination of the western transect (Plate 1).

Several independent lines of evidence suggest a counterclockwise rotation of the Adriatic microplate after the

early Miocene. Schönborn [1992] observed that the amount of shortening in the southern Alps south of the Tonale line decreases from east to west, which implies a 15° counterclockwise rotation of the Adriatic microplate situated south of this fold and thrust belt and relative to stable Europe.

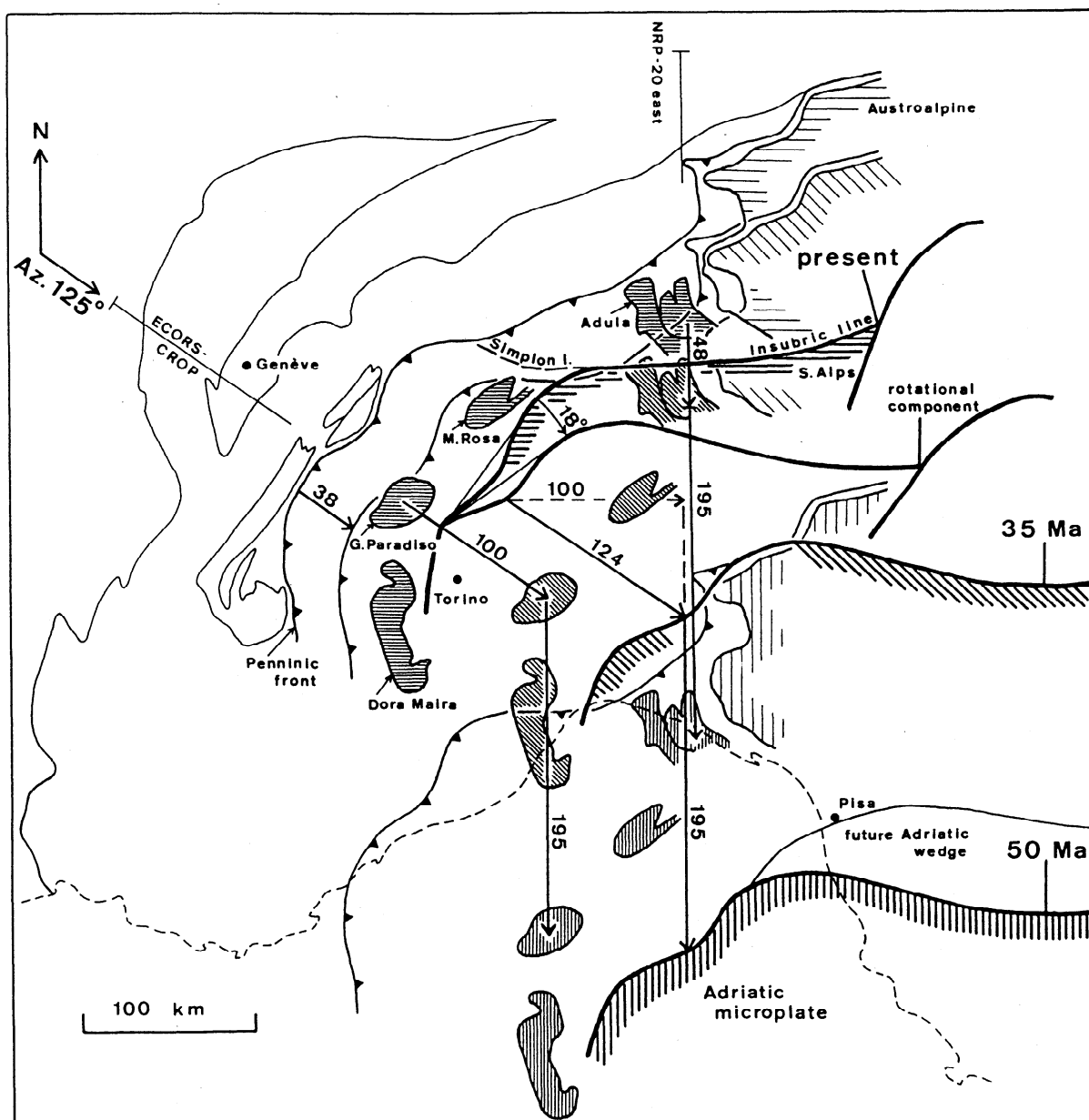


Figure 8. Simple kinematic retrotranslation (length of vectors given in kilometers) of some key tectonic features as they appear in present-day map view, accounting for the estimated amount of convergence between these features and the fixed European foreland (dashed line denotes the Mediterranean coast line for reference). Position at present is shown by horizontal hatching, position 35 Myr ago is shown by cross hatching, and the position 50 Myr ago is shown by vertical hatching. The position of the NW edge of the Adriatic microplate (present-day Insubric line) was obtained by a 124 km retrotranslation toward azimuth 125° (see inset of Figure 7) plus an 18° clockwise retrorotation. The positions of the other tectonic features 35 Myr ago correspond to those given in Figure 7 (ECORS-CROP profile) and by a similar kinematic reconstruction in profile view along the NRP-20 East profile [Schmid *et al.*, 1996], respectively. A common south directed retrotranslation by 195 km between 35 and 50 Ma (corresponding to the N-S convergence between the European foreland and Adriatic microplate for this time interval, as estimated by Schmid *et al.* [1996] along the NRP-20 East profile restores the configuration at the onset of collision in the Alps. See text for further discussion.

Figure 5 shows that the intersection line of the top surface of the Adriatic lower crustal wedge with the top surface of the European lower crust approaches the Tonale line toward the west, suggesting counterclockwise rotation as this wedge underthrusts the Tonale line. *Lowrie* [1986] points out that Istria (part of the undeformed Adriatic microplate) underwent a counterclockwise rotation of about 16° in respect to the southern Alps, on the basis of paleodeclinations recorded in Cretaceous sediments. *Vialon et al.* [1989] postulated a 22° sinistral rotation of the Adriatic microplate, partly on the basis of the curved arrangement of dextral strike slip faults in the external units of the western Alps. *Mancktelow* [1992] suggested that this rotation may be related to the continued opening of the western Mediterranean and Tyrrhenian Seas.

The reconstruction of Figure 8 assumes a late Miocene counterclockwise rotation of the Adriatic microplate around a pole NE of Torino, superimposed on its translation by 124 km toward azimuth 305° . Rotation angle and pole are chosen such as to account for the observed shortening in the southern Alps along the eastern traverse, shortening dropping to zero at the eastern margin of the western profile, an area which lacks middle Miocene shortening. The positions of the other tectonic features depicted in Figure 8 are in agreement with Figure 7 regarding the western transect and a previously published kinematic reconstruction along the eastern profile [*Schmid et al.*, 1996]. This proposed retrodeformation of post-35 Ma rotation and translation has interesting consequences. Firstly, the Dora Maira, Gran Paradiso, and Monte Rosa units align with the Adula nappe along a NNE-SSW oriented corridor. Secondly, diverging displacements between Monte Rosa and Adula units can be made responsible for an orogen-parallel stretch of the order of 60 km. This stretch readily explains the exhumation of the Lepontine dome situated between these two units by inducing normal faulting across the Simplon line.

For restoring the configuration 50 Myr ago, i.e., at the onset of collision in the Alps, a common north directed displacement of 195 km (convergence given by *Schmid et al.* [1996] for the 35-50 Ma time interval) was chosen for the central and western Alps. This last step of the retrodeformation implies substantial sinistral strike-slip movements between the Adriatic microplate and stable Europe in the region of the western Alps, as has been postulated by *Ricou and Siddans* [1986], leading to the disruption of the Briançonnais domain from Corsica-Sardinia [*Stampfli*, 1993]. Presently, the principal extension directions in the western Alps are mostly arranged in an E-W trending radial pattern. As pointed out by *Choukroune et al.* [1986], however, earlier stretching lineations formed under high-pressure conditions and partly reoriented into an E-W orientation by later strain increments do suggest an earlier north to NW directed transport [see also *Philippot*, 1988; *Ballèvre et al.*, 1990; *Platt et al.*, 1989]. In the Valais domain, *Fügenschuh et al.* [1999] document two penetrative phases of deformation associated with north directed transport, overprinted by WNW directed thrusting associated with the Penninic frontal thrust. Hence there is ample evidence that the area of the present-day western Alps formerly represented a NNE-SSW striking corridor of sinistral transpression.

In summary, the present-day arc of the western Alps appears to be prestructured by head-on Tertiary collision in the

central and eastern Alps and oblique collision associated with sinistral transpression in the western Alps. After 35 Myr ago the arcuate shape was accentuated by WNW directed movement and anticlockwise rotation of the Adriatic microplate. Strain partitioning along the Tonale and Simplon shear zones allowed this WNW directed movement to transmit significant amounts of WNW directed transport onto the Penninic units and the European foreland of the western Alps, while shortening continued in a N-S direction in the central Alps. The resulting diverging displacement vectors (Figure 8) led to orogen-parallel extension and exhumation of the Lepontine dome.

7. Discussion and Conclusions

Even in a region with an unprecedented amount of controlled source seismic data, such as in the western Alpine region, the seismic image of the deep crustal structure is incomplete and contains errors of significant size. Hence several different structural interpretations are possible. In earlier interpretations of ECORS-CROP seismic models, a mantle wedge inserted into the European lithosphere and connected eastward with the Ivrea body was proposed [e.g., *Nicolas et al.*, 1990a, b; *Roure et al.*, 1990]. In more recent studies [e.g., *Roure et al.*, 1996a] and in this work, thickening of the European lower crust beneath the western Alps is favored ("European lower crustal wedge" of Figure 5).

Our approach, however, primarily differs from previous studies in the handling of the (still sparse) seismic data: (1) We choose not to project any seismic information on deep crustal structure, while in previous models, projection of seismic structural data along the axis of the mountain range was more the rule than the exception [e.g., *Mueller et al.*, 1980; *Thouvenot et al.*, 1990]. (2) We seek the simplest model that fits all seismic information within its a priori estimated error bounds [*Kissling*, 1993; *Waldhauser et al.*, 1998]. We are fully aware that the simplest model does not necessarily denote the true structure, but faced with intrinsic ambiguity, we believe this to be the most appropriate approach to derive the main features of deep crustal structure.

An important piece of evidence for the original proposal of a mantle wedge along the ECORS-CROP profile was derived from the analysis of the Bouguer gravity field that indicates an excess mass beneath the Penninic realm. The direct vicinity of this high-density volume to the geophysical Ivrea body [*Kaminski and Menzel*, 1968] suggested the interpretation as mantle material and hence as a mantle wedge [*Bayer et al.*, 1989, 1996]. A recent tomographic study [*Solarino et al.*, 1997], however, clearly reveals low-velocity material below and to the west of the Ivrea high-velocity body and thus confirms an early schematic interpretation of refraction seismic data by *Giese* [1968]. As already mentioned above, with respect to resolution of gravity modeling, the gravity effect of the European lower crustal wedge outlined in this study is identical to the effect of the mantle slice proposed by *Bayer et al.* [1989, 1996].

Comparison of the new structural-kinematic model for the western Alps (this study) with the model proposed by *Schmid et al.* [1996] for the central Alps reveals major features common to both transects but also a number of significant differences in deep structure and in kinematic evolution.

Inspection of a third (central) transect (Figure 4b) located between the above two profiles (Figures 4a,c) shows that the western and central transects are similar in most regards, whereas the eastern transect differs. Hence we conclude that the transition from the central to the western Alps occurs in an area immediately west of the Lepontine dome. This area coincides with the western limit of the Adriatic lower crustal wedge and the northern limit of the European lower crustal wedge.

Recently, a database of 30,053 local earthquakes in the Alpine region for the 1980-1995 time period has been compiled by *Solarino et al.* [1997]. Well-located earthquake foci from within a 30 km wide corridor, each running parallel to the three transects (15 km on either side of the transect) have been orthogonally projected onto these transects. Figure 4 compares the location of these events with the large-scale tectonic structures. Comparison between Figures 4a, 4b, and 4c reveals significant differences in the depth extent of the seismogenic regions. As previously noted by *Deichmann and Baer* [1990], the maximum depth of earthquakes is situated near the Moho in the northern and southern forelands along the eastern traverse (Figure 4c). Beneath the Penninic units, i.e., within the Lepontine metamorphic dome, they are restricted to the thickened upper crust. Also note that the Adriatic lower crustal wedge is quiescent.

Coincidence of the lower limit of seismicity with predicted isotherms based on thermal modeling [*Okaya et al.*, 1996; *Bousquet et al.*, 1997] suggests that the 500°C isotherm controls the cataclastic-ductile transition. Quiescence within the Adriatic lower crustal wedge further suggests stress transmission between the European and Adriatic lithosphere in the central Alps to be restricted to upper crustal levels. According to this interpretation, the isotherms are still elevated beneath the central Alps. Thermal modeling [*Okaya et al.*, 1996; *Bousquet et al.*, 1997] suggests that additional heating beneath the central Alps is dominantly provided by two heat sources: (1) frictional heating induced by retroshearing [*Beaumont et al.*, 1994] of nonsubductable upper European crust north of the Insubric line and (2) radiogenic heat production within the accreted upper crustal material (i.e., the Penninic nappes or "upper crustal basement" in Figure 4c). Ascent and emplacement of the Periadriatic intrusions in the context of slab break off [*von Blanckenburg and Davis*, 1995] provide another important heat source that may have affected the southern part of the central Alps.

In contrast, the earthquake distribution along the western and, to a lesser degree, along the central transects exhibits a wide, east dipping corridor of foci affecting the entire transect, including the allochthonous European lower crust (Figures 4a and 4b). Because of station distribution and applied selection criteria (only well-locatable events are considered) by *Solarino et al.* [1997], only few events are visible at middle and lower crustal depths beneath the Penninic realm in Figures 4a and 4b. Studies by *Thouvenot* [1996] and *Eva et al.* [1997], however, document seismic activity down to depths of 40 km beneath the internal region of the western Alpine arc. Thus, coupling and stress transmission between the Adriatic microplate and the European lithosphere occur along a deep-reaching seismogenic zone. This indicates a contrasting (with respect to the eastern transect) present-day thermal regime, primarily caused by the following substantial differences in the

kinematic evolution. Firstly, oblique convergence and collision in the western Alps before 35 Myr ago must have led to a significantly smaller volume of accreted radiogenic upper crustal rocks, as compared to the central Alps, the latter being characterized by head-on convergence and collision. Secondly, diverging displacements of the central and western Alps after 35 Myr ago allowed for orogen-parallel extension in the central Alps (Lepontine dome), associated with updoming of the isotherms. We conclude that the Penninic realms of the central and western Alps differ significantly not only in deep crustal structure but also in the thickness of the seismogenic zone.

According to our kinematic model (Figure 8), late Tertiary crustal shortening in the western Alps postdates collision between 50 and 35 Ma, associated with sinistral transpression [*Ricou and Siddons*, 1986] in the area of the future arc of the western Alps. It may be subdivided into three episodes (Figure 7). From 35 to 30 Ma the Briançonnais basement was back thrust over the Gran Paradiso units, leading to 30 km of crustal shortening (first episode). In the early Miocene, movements concentrated mainly along the Penninic frontal thrust and resulted in about 60 km shortening (second episode). Post-12 Ma shortening within the external massifs is associated with folding in the Jura mountains when the crust was shortened by an additional 30 km (third episode).

The European lower crustal wedge present in the western and central transects is interpreted to have formed during the second episode in response to the 60 km WNW directed movement of the Adriatic micro-plate. This movement and the anticlockwise rotation by about 18° (Figure 8) are inferred to have occurred after the Adriatic microplate had already collided with the European continental margin, and hence it resulted in wedging of lower crustal slices into the buoyant continental crust at midcrustal levels. Wedging of high-density slices into the European crust has been the favorite model to explain both the Bouguer gravity field and Moho topography [e.g., *Kissling*, 1980; *Kahle et al.*, 1980; *Bayer et al.*, 1989, 1996] since a number of years, also for the eastern transect. The shape and extent of this wedging, however, remained speculative until high-resolution seismic data of the NRP-20 project [e.g. *Pfiffner et al.*, 1997] clearly documented such structures beneath the European Geotraverse (EGT) transect (coinciding with our eastern transect, see Figure 4c). Underthrusting of the European crust by Adriatic lower crust and uppermost mantle, as drawn in Figure 4c, implies that the upper crust of the Adriatic microplate has been stripped off. Indeed, upper crustal shortening of about 50 km by back thrusting has been documented for the southern Alps along the eastern (EGT) transect [*Schmid et al.*, 1996] (see also Figure 4c). Along the western edge of the Adriatic microplate (western and central transects, Figures 4a and 4b), however, the situation is entirely different, with the high-density and high-velocity Ivrea body reaching from a depth below the Moho to the surface (see Plate 2 and the work of *Solarino et al.* [1997]). This geometry leads us to propose that the Ivrea body acted as a buttress during the collision of the two plates and that crustal shortening of about 60 km during the second episode was mostly accommodated by wedging involving European lower crust (Figure 7).

We conclude in proposing that the formation of the western Alpine arc was initiated by head-on early Tertiary

collision of the convex European plate boundary zone with the Adriatic microplate in what subsequently became the central and eastern Alps, associated with sinistral transpression in the western Alps. After 35 Myr ago the arcuate shape was accentuated by postcollisional WNW directed movement and anticlockwise rotation of the Adriatic microplate, associated with wedging of lower crustal slices at midcrustal levels both in the western Alps and in the central Alps.

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