ELSEVIER

Earth-Science Reviews 45 (1999) 167–208

EARTH-SCIENCE REVIEWS

Orogens and slabs vs. their direction of subduction

C. Doglioni [a,*], P. Harabaglia [b], S. Merlini [c], F. Mongelli [d], A. Peccerillo [e], C. Piromallo [f]

[a] Dipartimento di Scienze della Terra, Università La Sapienza, 00185 Rome, Italy
[b] Centro di Geodinamica, Università della Basilicata, 85100 Potenza, Italy
[c] AGIP-ENI, 20097 San Donato Milanese, Italy
[d] Dipartimento di Geologia e Geofisica, Università di Bari, 70125 Bari, Italy
[e] Dipartimento di Scienze della Terra, Università di Perugia, 06100 Perugia, Italy
[f] Istituto Nazionale di Geofisica, 00143 Rome, Italy

Received 28 October 1997; accepted 17 December 1998

Abstract

Subduction zones appear primarily controlled by the polarity of their direction, i.e., W-directed or E- to NNE-directed, probably due to the westward drift of the lithosphere relative to the asthenosphere. The decollement planes behave differently in the two end-members. In the W-directed subduction zone, the decollement of the plate to the east is warped and subducted, whereas in the E- to NNE-directed, it is ramping upward at the surface. There are W-directed subduction zones that work also in absence of active convergence like the Carpathians or the Apennines. W-directed subduction zones have shorter life (30–40 Ma) than E- or NE-directed subduction zones (even longer than 100 Ma). The different decollements in the two end-members of subduction should control different PTt paths and, therefore, generate variable metamorphic assemblages in the associated accretionary wedges and orogens. These asymmetries also determine different topographic and structural evolutions that are marked by low topography and a fast 'eastward' migrating structural wave along W-directed subduction zones, whereas the topography and the structure are rapidly growing upward and expanding laterally along the opposite subduction zones. The magmatic pair calc-alkaline and alkaline–tholeiitic volcanic products of the island arc and the back-arc basin characterise the W-directed subduction zones. Magmatic rocks associated with E- or NE-directed subduction zones have higher abundances of incompatible elements, and mainly consist of calc-alkaline–shoshonitic suites, with large volumes of batholithic intrusions and porphyry copper ore deposits. The subduction zones surrounding the Adriatic plate in the central Mediterranean confirm the differences among subduction zones as primarily controlled by the geographic polarity of the main direction of the slab. The western margin of the Adriatic plate contemporaneously overridden and underthrust Europe toward the 'west' to generate, respectively, the Alps and the Apennines, while the eastern margin subducted under the Dinarides–Hellenides. These belts confirm the characters of the end-members of subduction zones as a function of their geographic polarity similarly to the Pacific subduction zones.

Keywords: tectonics; subduction zones; orogens; Mediterranean geodynamics

* Corresponding author. Tel.: +39-06-4991-4549; Fax: +39-06-4454-729; E-mail: doglioni@axrma.uniroma1.it

0012-8252/99/$ - see front matter
PII: S0006-8993(98)01372-9

1. Introduction

Subduction zones occur where the lithosphere descends into the asthenospheric mantle (Fig. 1). The presence of a subduction zone has been firstly recognized by seismicity (Fig. 2). The knowledge of these zones rapidly increased since the papers by Wadati (1935) and Benioff (1949). Since then, subduction zones have been investigated via seismology, seismic tomography, seismic reflection, surface geology and other means such as magmatic petrology and geochemistry, gravity, etc. (e.g., Menard, 1964; Barazangi and Isacks, 1971; Forsyth and Uyeda, 1975; Schubert et al., 1975; Bally, 1983; Giardini and Woodhouse, 1986; Jarrard, 1986; Von Huene, 1986; Fowler, 1990; Dziewonski et al., 1993; Bebout et al., 1996, and references therein). In this paper, we do not intend to discuss all the main peculiarities of subduction zones or to make the state of the art of 'subductology', themes that have been the aim of other classic papers (e.g., Jarrard, 1986; Peacock, 1996). We rather want to focus on the differences between subduction zones as proposed for the Pacific margins (Nelson and Temple, 1972; Dickinson, 1978; Uyeda and Kanamori, 1979), integrating geophysical, geological and volcanological data. The usual explanation for these differences is the variable age of the down-going lithosphere, (e.g., Forsyth and Uyeda, 1975; Jarrard, 1986).

We alternatively propose to test whether the geographic polarity influences the gross feature of the subduction zone and the associated orogen. This could be related to the main flow of plate motion (Fig. 1) and the 'westward' polarization of this flow,

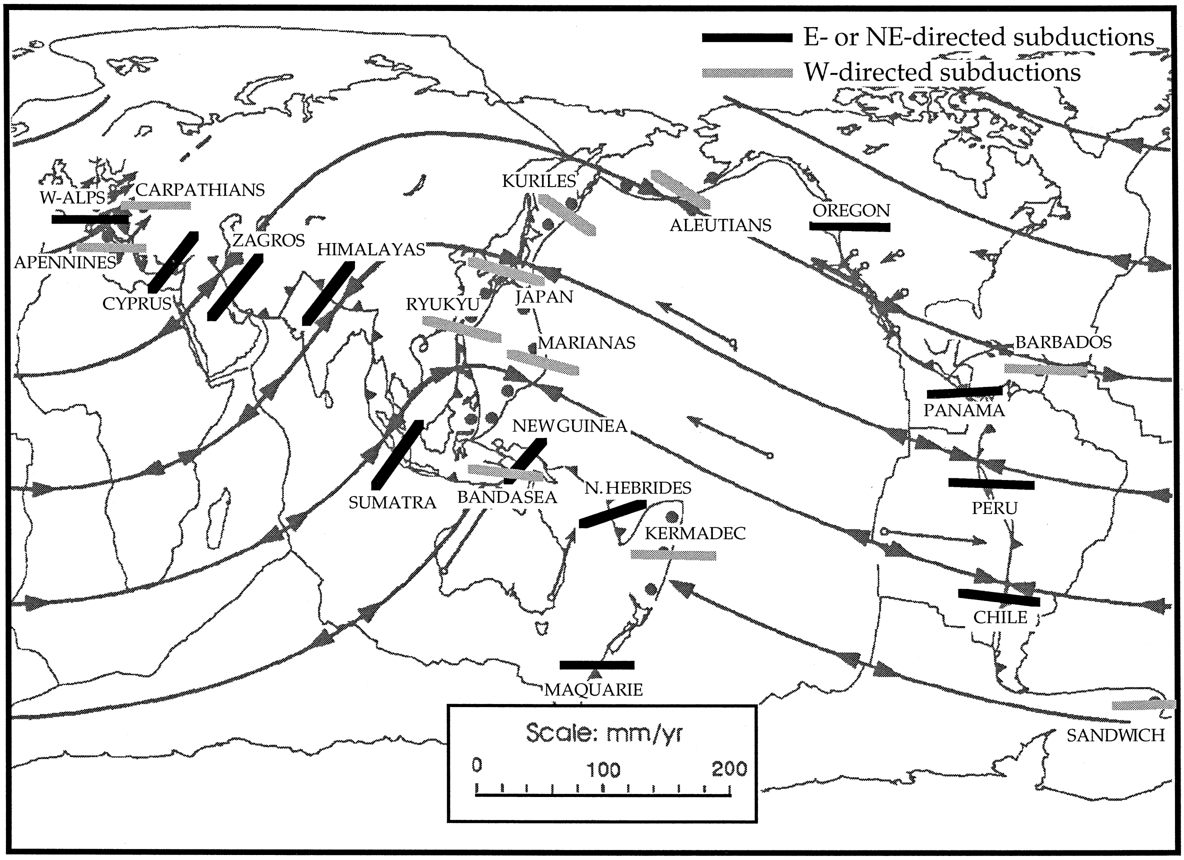


Fig. 1. Location of the subduction zones discussed in the text. The lines represent the mainstream of absolute plate motion (modified after Doglioni, 1993).

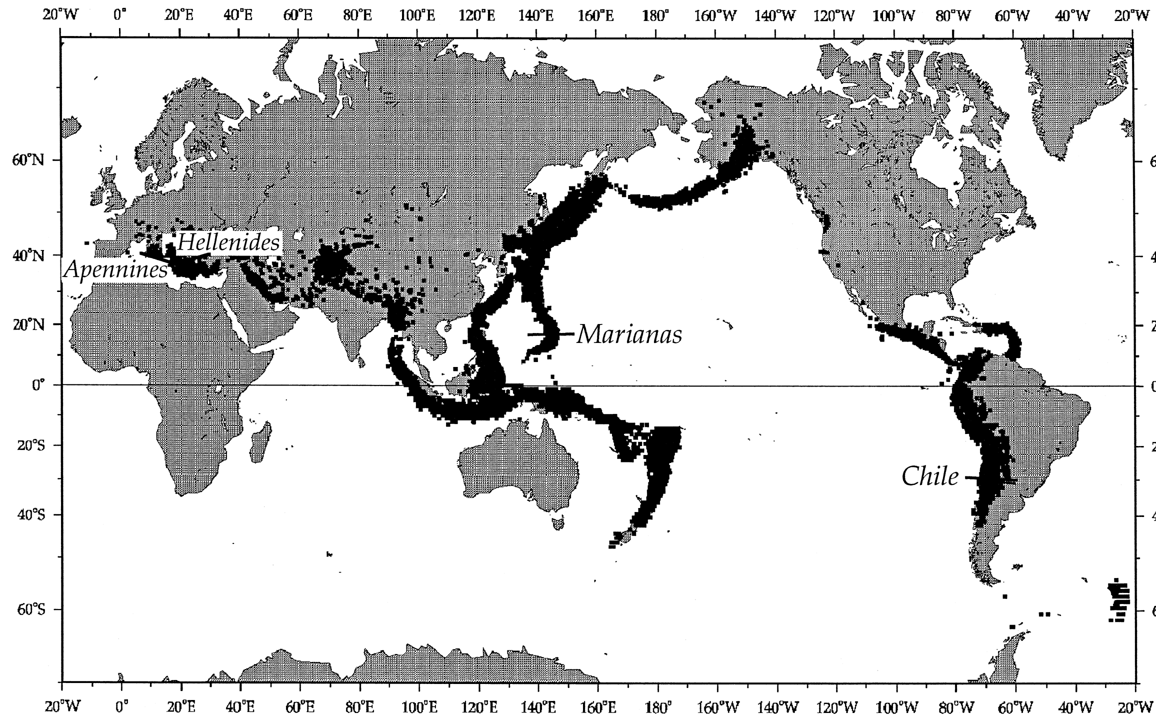


Fig. 2. Map of deep and intermediate earthquakes (> 40 km); they delineate most of the subduction zones. Events are from the NEIC catalogue (1974–1997). Lines refer to sections of Fig. 3.

owing to the westward delay of the lithosphere relative to the underlying mantle. The two classes of subduction zones mentioned in the text have to be seen no strictly W-directed vs. E-directed, but following or opposing the mainstream proposed in Fig. 1.

2. General remarks on subduction zones

The subducting lithosphere may have different composition and thickness. As much as it is thick and old, the oceanic lithosphere is the most easily subducting one. There is a simple, let say trivial observation indicating that among two plates, the denser is the one which subducts. For a recent analysis of the comparative buoyancy values of different types of lithosphere, see Cloos (1993). Therefore, when a denser oceanic lithosphere is located west of a lighter one, the subduction will dip to east, and vice versa. The Pacific margins (e.g., Lallemand, 1995) are clearly showing this control, e.g., the oceanic lithosphere subducting eastward underneath the continental Cordillera. Oceanic lithosphere and thinned continental lithosphere which underwent magmatic underplating are the segments of lithosphere which are observed along subduction zones. The descending oceanic lithosphere has variable thickness (mean values between 30 and 90 km). There are zones where oceanic ridges interact with subduction zones (Lallemand et al., 1992). Continental lithosphere carrying a crust of more than 30 km thickness shows drastic reduction of the subduction activity and convergence rates when it encroaches the trench in all types of subductions.

Clearly, sensu latu subduction zones occur where two plates converge, like in the Andes or in the Himalayas. However, there are subduction zones that developed in geodynamic settings with very low or not convergence in the direction of the slab retreat. Examples are the Caribbean, the Apennines and the

Carpathians. These subduction zones have in common the average W-directed orientation of the slab.

We prefer to use the term direction rather dip of the subduction zone because along an arcuate subduction the dip may vary of 180°, whereas the direction of plate motion is defined by a more precise angle. For example, along the arc of the Apennines the subduction dips to the southwest (northern Apennines), or to the west–south–west (southern Apennines), or to the northwest (Calabria), or even to the north (Sicily and Maghrebides).

In the Pacific, W-directed subduction zones are steep (up to 90°) and deep (down to 670 km) with respect to those directed to the east which are on average shallower and less inclined (Isacks and Barazangi, 1977; Lundgren and Giardini, 1994). This asymmetry (Fig. 3) was defined in terms of Mariana type and Chilean type subduction zones (Dickinson, 1978; Uyeda, 1981). This comparative subductology of Uyeda (1981) has been challenged by many authors (e.g., Jarrard, 1986; Scholz and Campos, 1995). These differences have been mainly interpreted in terms of different age of the subjecting lithosphere, i.e., older, colder and denser in the W-Pacific subductions where the slabs are steeper; other attempts were related to the dimension, composition and velocity of plates, mantle convection, etc. (e.g., Hager and O'Connell, 1978). Alternatively, Nelson and Temple (1972), Uyeda and Kanamori (1979) and Ricard et al. (1991) interpreted this asymmetry in terms of a relative eastward mantle flow or the westward drift of the lithosphere. Doglioni (1990, 1992) also described the geological signatures associated with this asymmetry and proposed an undulate

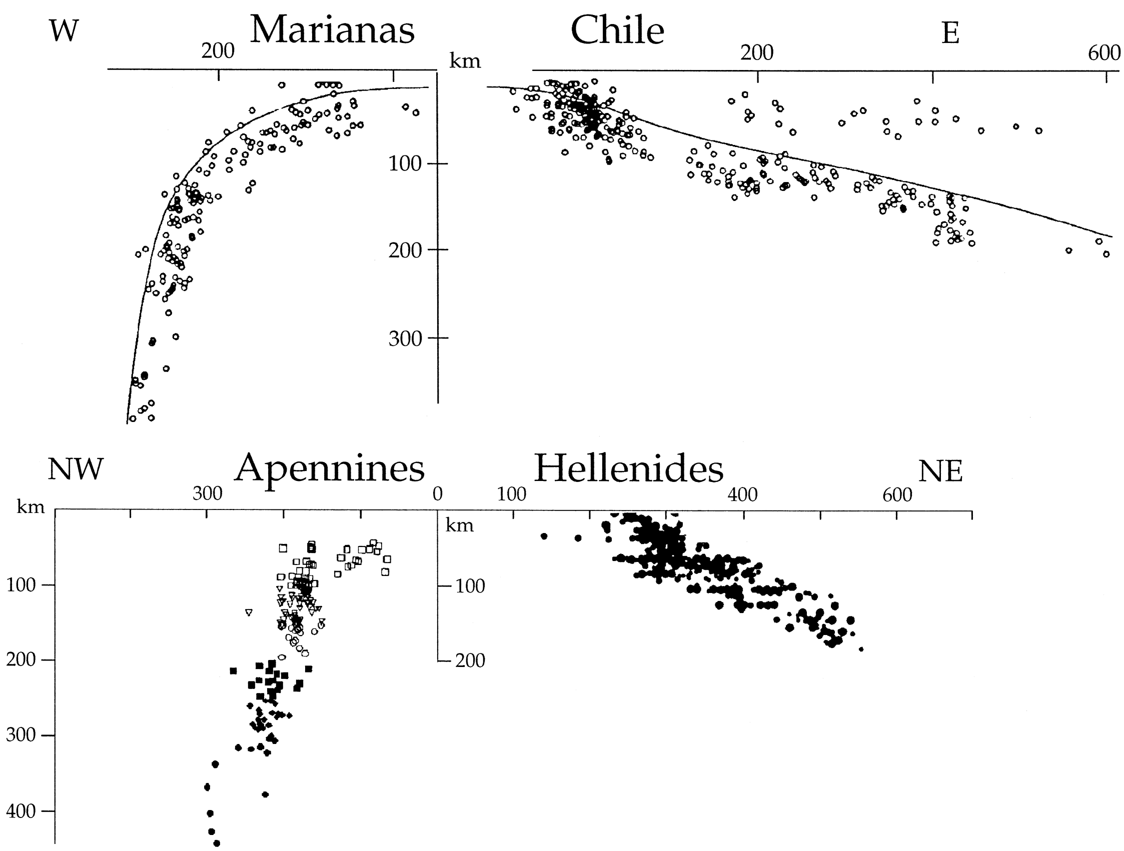


Fig. 3. Ipocenters of the Marianas and Chile subduction zones in the Pacific (after Isacks and Barazangi, 1977), compared with the seismicity of the Apennines (Selvaggi and Chiarabba, 1995) and Hellenides (Papazachos and Comninakis, 1977) opposed subduction zones. The Pacific asymmetry is present also in the central Mediterranean subduction zones where the Ionian oceanic lithosphere is subducting contemporaneously both underneath the Apennines and the Hellenides. Location of the sections in Fig. 2.

flow of plate motion on the Earth's surface, along which the asymmetry of the western and eastern Pacific subduction zones persist as a function of the geographic polarity in terms of geophysical and geological differences along the other subduction zones of the world, even in complicated areas such as the Mediterranean or Indonesia (Fig. 1).

Most of these differences have been so far explained in terms of slab pull and age of the subducting oceanic lithosphere, or rates of convergence between plates (e.g., Royden and Burchfiel, 1989). Waschbusch and Beaumont (1996) proposed that the two end-members of orogens develop when the convergence rate is either faster (double vergence orogen, no back-arc spreading, e.g., the Alps) or slower (single vergence and back-arc spreading, e.g., the Apennines) with respect to the subduction hinge roll-back.

The Adriatic plate in the Mediterranean is overriding Europe to form the Alps and it underthrusts Europe to generate the Apennines; therefore, the same lithosphere determines different thrust belts as a function of the polarity of the subduction. Moreover, the Adriatic continental lithosphere and the Ionian possibly oceanic lithosphere are subducting both under the Apennines (steep W-directed subduction) and under the Dinarides–Hellenides (shallow NE-directed subduction), as shown by tomography in Fig. 4. The two related thrust belts follow the east and west Pacific rules, in spite of the same variations in age and thickness of the subducting lithosphere. In the Pacific itself, the W-directed subductions are the fastest in the world and the slab is steep, while the Andean subduction is active since the Mesozoic and the slab is shallow. Those examples are clearly in contradiction both with the age of the subduction,

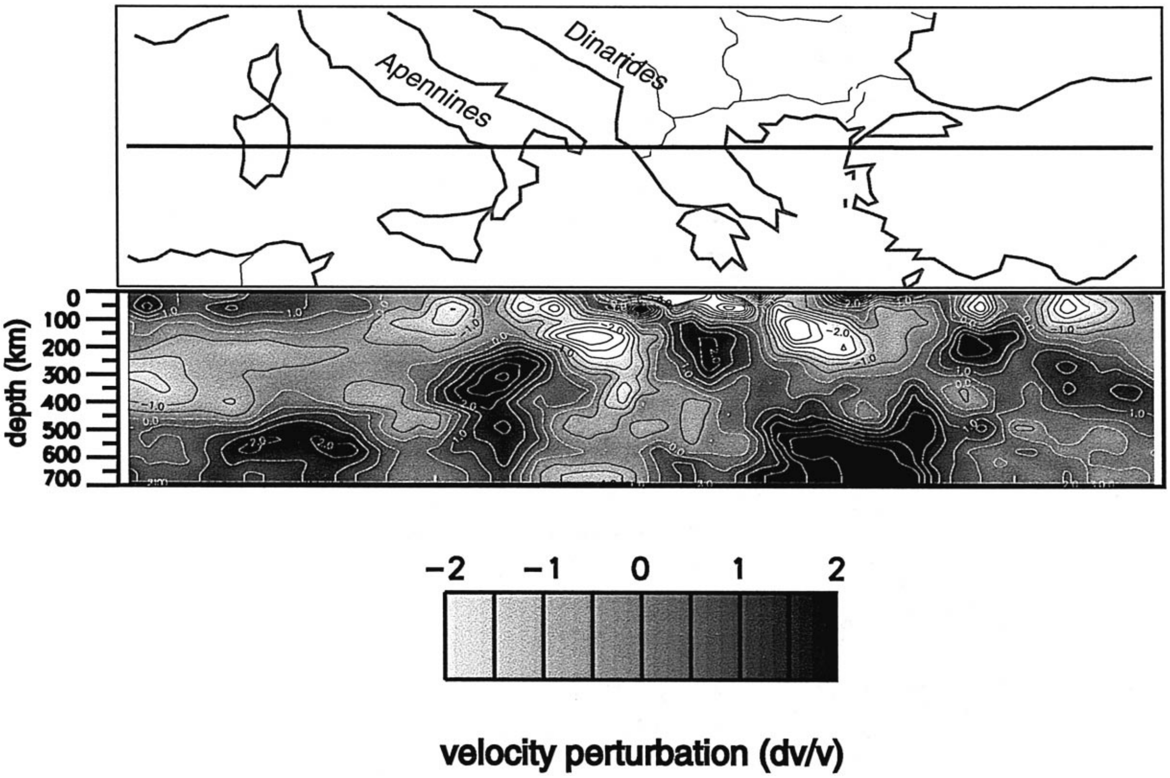


Fig. 4. Tomographic image of the Mediterranean upper mantle in a roughly E–W cross-section. Top: location of the profile (solid line) on the map for orientation. Bottom: P-wave velocity structure in percent deviation from reference velocity model SP6. Darker (lighter) shading denote positive (negative) anomalies, and contour levels are 0.5% apart. The depth axis is given in kilometers without vertical exaggeration. Regardless the continuity of the slabs which might be function of the reference velocity model, we interpret in this section the asymmetry between the steep high velocity body beneath the Apennines and the less inclined high velocity body along the Dinarides subduction zones.

and the age and thickness of the subducting lithosphere as the basic constraints of the tectonic style.

The present Apennines subduction shows focal mechanisms with down-dip compression parallel to the slab (Frepoli et al., 1996). This contrasts with a negative buoyancy of the supposed slab pull effect which is invoked also for the detachment of the slab into the mantle: the slab pull should generate extension along the down-dip direction of the slab and this is not observed for the Apenninic subduction zone which is the recent evolution of the western Mediterranean slab that retreated from west to east during the Neogene and Quaternary. Delamination processes (Channell and Mareschal, 1989) with subduction of the lithospheric mantle and in some cases of the lower crust have also been invoked.

A very popular concept in the literature is the slab detachment, or slab breakoff, a process that would imply mantle substitution to the falling subduction body (e.g., Platt and Vissers, 1989; Davies and Von Blanckenburg, 1995). However, in the case of the Alboran sea, where this model has been widely applied, the extension that should have resulted from this supposed detachment of the slab is oblique and cross-cut the Betic orogen. This indicates an independent origin of the extension with respect to the Betic subduction zone (Doglioni et al., 1997).

3. Subduction zones and the westward drift of the lithosphere

The westward drift of the lithosphere relative to the underlying mantle could explain the asymmetries among the subduction zones (Fig. 5). The 'westward' net rotation of the lithosphere has been detected in the hot spot reference frame (Le Pichon, 1968; Gripp and Gordon, 1990; O'Connell et al., 1991; Ricard et al., 1991; Cadek and Ricard, 1992). In particular, O'Connell et al. (1991) perform a spherical harmonic expansion of tectonic plate motions. They observe that the toroidal term of degree one, representing the net rotation of the lithosphere, follows the same pattern of the spectrum whenever the hot-spot reference frame is considered. Since any other choice of reference would have resulted in a value of the degree one term discordant with the rest of the spectrum, they argue that it probably has significance.

This is in agreement with the finding of Ricard et al. (1991). According to them, summing the vectors

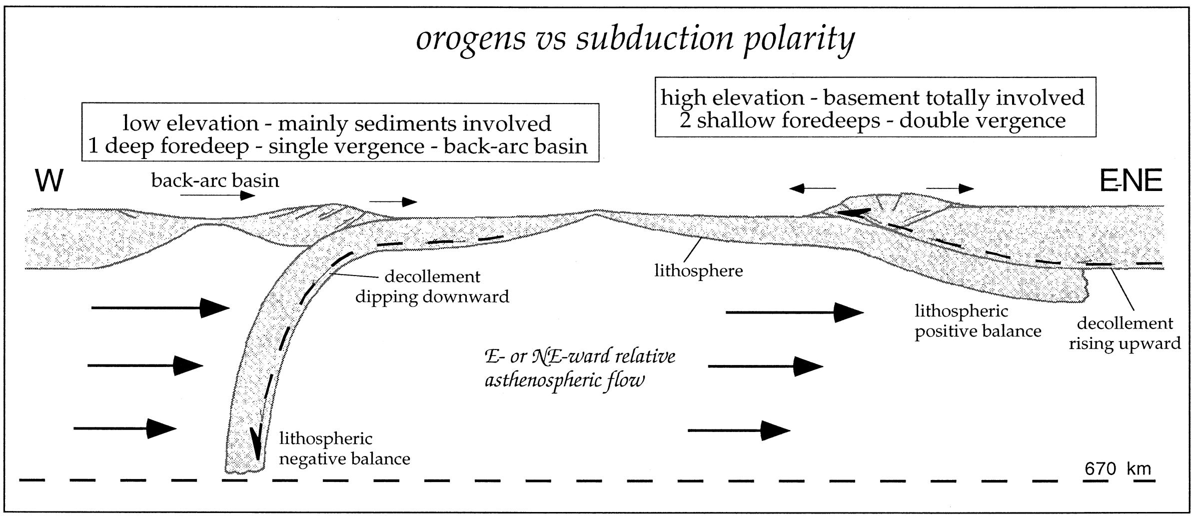


Fig. 5. W-directed subduction zones are steeper and deeper with respect to the E–NE- or NNE-directed subduction zones. Note that the decollement plane of the eastern plate is warped and subducted in case of W-directed plane, whereas it ramps toward the surface in the E–NE-directed subduction, enabling the uplift of deep seated rocks: this asymmetry may be explained by the 'westward' drift of the lithosphere relative to the mantle and controls the strong differences in morphology, structure and lithology of the related thrust belts.

of plate motions in the hot spot reference frame a westward component of the lithospheric motion of a few centimeters per year remains. This is mainly due to the high speed of the Pacific plate toward the 'west' which is not entirely compensated by the other plates which are moving in the opposite or different directions. Ricard et al. (1991) show that lateral heterogeneities in the asthenospheric viscosity are a possible mechanism to justify this phenomenon, although its real causes are not yet fully understood.

It is important to observe that the term 'westward' net rotation refers in fact to an 'undulated' motion with a pole at about 84°E and 56°S (Ricard et al., 1991). Therefore, its larger effects should be observed in the Mediterranean, in Japan, in the Marianas and in Chile. However, plate velocities relative to the mantle are variable probably due to lateral heterogeneities and viscosity variations both at the base of the lithosphere and in the asthenosphere.

This rotation might imply a relative 'eastward' directed mantle flow (Bostrom, 1971; Nelson and Temple, 1972; Uyeda and Kanamori, 1979; Uyeda, 1981). The consequence of the 'westward' delay of the plates with respect to the underlying mantle, confirm that plates may have a general sense of motion, a sort of mainstream (Fig. 1), and that they are not moving randomly. If we accept this postulate, plates are moving along this trend at different velocities toward the 'west' relative to the asthenospheric mantle. Rather than exactly west it would be better to say moving generally 'westward' (SW, WWN, etc.) along flow lines, which undulate and are not E–W parallel (Fig. 1). This phenomenon can account for the different dip of subduction zones (steep W-directed and shallow E–NE-directed) and the peculiar geological signatures of the opposite subduction zones.

The roll-back of the slab along W-directed subduction zones implies a substitution of the lithosphere by the mantle. Since the retreat of the subducting lithosphere is eastward directed, an equivalent amount of mantle should move eastward to replace the lithospheric loss. We might argue that vertical motions of the mantle could compensate this loss without invoking the lateral eastward migration of the mantle, but the steep attitude of the W-directed subduction zones and the strong increase of the viscosity value at the 670 km discontinuity between upper and lower mantle (Hager, 1990) make the mantle adjacent to the seismically detectable slab an isolated system, with inhibited communication among eastern and western sectors of the slab, and the upper and lower mantle underneath. This supports the notion that as the slab vertically retreats eastward, we should kinematically expect a contemporaneous migration of the mantle toward the east. This does not exclude that the eastward moving mantle is the cause for the eastward slab retreat and not a consequence of it. An actively eastward pushing mantle agrees with the westward drift of the lithosphere relative to the asthenosphere detected in the hot-spot reference frame (Ricard et al., 1991), and it supports an eastward oriented push at depth on the slab in order to generate the arcuate shape of the subduction zone like an obstacle in a river.

Subduction angles change with depth. According to Marotta and Mongelli (1998), the main factors in determining them are the slab pull, the pressure due to the subduction induced mantle flow and the pressure exerted by the asthenospheric motion relative to the lithosphere. The authors consider the slab as a thin elastic plate of finite length embedded at its initial point of immersion in the asthenospheric mantle. W-directed slabs subduct with larger angles because they do it with a counterflow trend with respect to the lithosphere–asthenosphere relative motion. Conversely, E-directed subductions yield smaller angles because of the uplift due to flow-ward motion. The W-directed subduction zones have a negative balance of lithospheric accretion. In other words the subducted lithosphere could be anchored by the relatively eastward moving mantle and eventually annihilated. In contrast, E- or NE-directed subduction zones provide a thickening of the hangingwall lithosphere from underneath, producing a positive balance of lithospheric growth.

It is noteworthy that seismicity along subduction zones is decreasing toward the polar regions; this would confirm a general polarization of plate motion that agrees with a westward drift of the lithosphere relative to the mantle which as stated above, does not coincide with the geographic poles.

There are orogens and subduction zones which do not follow the flow proposed in Fig. 1, i.e., the E–W trending southern Carribbean belt or the Pyrenees (e.g., Teixell, 1998). Those are orogens related to second order rotations of plates. However, those features have in general geometries similar to the orogens associated with E-directed subduction zones.

4. Seismicity along subduction zones

The shallow seismicity indicates that seismic coupling at convergent plate boundaries can vary enormously from one subduction to another (Scholz and Campos, 1995). The depth of the decollement planes could also control the seismic coupling. In the accretionary wedge, the depths of the decollements are function of the thickness and rheology of the involved rocks. No clear distinction can be performed between W-directed and E- or NE-directed subductions, although it seems that boundaries where at least one of the plates is continental normally host larger earthquakes than those where there is only oceanic lithosphere. According to Lay and Kanamori (1981), the two end-members are the Marianas subduction where large earthquakes are absent, and the Chilean subduction that is characterised by some of the largest events ever observed (including the largest event ever recorded, the Great Chile Earthquake, 1960, $M_w = 9.5$). Ruff and Kanamori (1980) show in fact that there is a relation between the presence of back-arcs and low seismic coupling.

Plate boundaries deform in a complex way. Punctual estimates of the strain field at the boundary, such as those obtained from focal mechanisms, might be highly misleading. Moreover, the latter can provide strong kinematics indicators even in areas where earthquakes activity is not particularly relevant. To reduce this bias, it is, therefore, necessary to reconstruct the whole strain field integrating geophysical observations with geological ones. In the Apennines arc, for instance, the observed compression–extension wave is generated by the 'eastward' roll-back of the subducting Adriatic–Ionian–African lithosphere. The convergence between Africa and Europe can be estimated within a few millimeters per year, that is one order of magnitude lower than the velocity of the Apenninic subduction roll-back toward the east. The Apennines arc generated and generates compression and extension all around the arc, being the tectonic pair independent from what Africa and Europe are doing one relative to the other. The N–S compression in Sicily is located in the southern arm of the Apenninic arc and do not necessarily entirely reflects the relative motion of Africa and Europe. Support to this model comes also from VLBI and GPS data at sites well within the plates and, therefore, not affected by plate deformation. It follows that seismicity around arcs of W-directed subduction zones, which are simply controlled by the roll-back of the slab, may not give reliable information on plate motions.

Regarding deeper events, their rupturing processes are still poorly understood. Petrological and rheological transformations are expected along the subducting body (Wortel, 1982; Wortel and Vlaar, 1988; Gaherty and Hager, 1994). A number of possible explanations are given by Green and Burnley (1989), Meade and Jeanloz (1991), and Kirby et al. (1991) all based or triggered by various phase transitions. A comprehensive review is given by Frohlich (1989). Details on the rupture process can also be found in the papers of Houston and Williams (1991), Vidale and Houston (1993), and Houston and Vidale (1994). It is noteworthy that some deep earthquake seem to be located outside of known slabs. According to Lundgren and Giardini (1994), they might take place on some deflected portion of slab. Green (1993) and Green and Zhou (1996) demonstrated that polycrystals undergoing the ilmenite $\leftrightarrow$ perovskite transformation show transformation-induced faulting occurring only in the exothermic direction of the reaction. They suggested the structural and thermodynamic analogies between that transformation and those expected in the silicates of the earth's mantle. They argued strongly that these results are relevant to transformations occurring in subducting lithosphere. If so, it follows that faulting by this mechanism cannot occur in the lower mantle, providing a natural explanation for the termination of earthquakes just before 700 km depth.

5. Lithospheric properties and geometry of subduction zones

In particular, we observe that W-directed slabs have dip angles ranging from 40° of the fastest western Pacific slabs to verticality of the blocked Carpathians slab (Oncescu, 1984). The 11 W-directed subduction zones we examine (Fig. 1) should in fact be divided in two sub-groups. The first comprises the Pacific slabs and is characterised by active convergence (from 4 to 10 cm/year) as well as slab retreat, the second comprises the two Atlantic and two European slabs where the dominant mechanisms is slab retreat.

According to Isacks and Molnar (1971), the distribution of down-dip extension and compression within the slab can be related to the maximum depth of the slab and ultimately to the slab pull. It is noteworthy, however, that Frepoli et al. (1996) question the role of the slab pull for the Tyrrhenian slab on the base of the focal mechanism distribution; its rate of descent is small and this probably means that it reaches a state of thermal equilibrium at a shallower depth. This in turn has a great role in determining the density difference between the slab and the surrounding mantle and might be the reason because the slabs apparently do not sink, or it sinks slower. The Atlantic and European slabs are not so well known as the Tyrrhenian. It is, however, possible on the basis of the few available data that their asset might be similar. Moreover, their shallow depth might be only apparent; the deeper portion might simply be no longer distinguishable from the surrounding upper mantle.

It has long been known (e.g., Wadati, 1935; Benioff, 1949) that deep events delineate slab geometry and that intermediate and deep seismicity is confined within subducting plates (Fig. 2). In fact earthquakes hypocenters delineate what is termed Wadati–Benioff zone and their mere presence is assumed as a sure proof of the existence of subducting lithosphere (Fig. 3).

However, a problem in defining slab geometry in general is caused by discrepancies between tomographic imaging and earthquake locations; see, for example, the Aegean subduction by Wortel and Spakman (1992) where the subduction angle resulting from tomographic imaging is about 20° steeper and deeper (670 km) than that resulting from earthquakes (200 km). The latter are defined within few kilometers in depth. Conversely tomographic imaging is smeared out and can be severely biased by an inappropriate starting model.

In particular in the Apennines subduction, the several hundreds kilometers deep seismicity is concentrated in the southeastern Tyrrhenian sea (Peterschmitt, 1956; Caputo et al., 1970, 1972; Selvaggi and Chiarabba, 1995). However, few scattered earthquakes have been recorded also below the northern Apennines (Amato et al., 1993). The seismicity appears located along the northwestward paleogeographic prolongation of the Ionian Mesozoic lithosphere (there are authors which interpret the Ionian sea as composed of oceanic lithosphere, e.g., de Voogd et al., 1992; but there are other authors which consider it as continental in character, e.g., Cloetingh et al., 1979; Farrugia and Panza, 1981; Suhadolc and Panza, 1989; Cernobori et al., 1996).

The shortening in the Apennines accretionary wedge is maximum in northern Calabria and southern Apennines and decreases along the opposite arms of the Apenninic arc. Therefore, the shortening which is visible in the accretionary prism where the deep slab seismicity is lacking or attenuated suggests that subduction has occurred all along the Apenninic arc. This is questioned by Marson et al. (1995) and Du et al. (1998) who propose the absence of the deep slab underneath the southern Apennines based on gravity interpretation.

The Mesozoic facies piled up in the belt indicated that they were lying on thinned continental crust. The missing crust in the Apennines and the shortening in the belt support the subduction of those volumes of continental lithosphere.

The continental crust has lower temperature of the brittle–ductile transition (300–400°C) than the oceanic crust (500–650°C). The paucity of deep seismicity along the southern and northern Apennines could be attributed to the more ductile rheology of the quartz–feldspar rich Adriatic continental lithosphere with respect to the olivine–pyroxene rich Ionian sea subducting underneath Calabria with a more brittle behavior and generating a more elevated seismicity.

Recent tomographic investigations of the Apennines slab show a much more continuous cold body underneath the Apennines than so far imaged (Amato et al., 1993; Piromallo and Morelli, 1997). The differences in the subducting lithosphere, i.e., continental below the central–northern Apennines and oceanic below Calabria are supported also by the magmatism that shows clearly different sources (Peccerillo, 1985; Serri et al., 1993). The decrease in velocity of the arc migration toward the northern Apennines and Sicily should correspond to a decrease of the stress acting on the slab plane as well. Therefore, the mechanical differences between the Adriatic and Sicilian continental lithospheres and the Ionian possibly oceanic lithosphere, and the lower stress moving away from the Calabrian arc can account for the decrease in slab seismicity along the Apenninic arc.

In Fig. 4, a roughly E to W vertical cross-section through a 3-D P-wave velocity model of the mantle below the Calabrian arc and the Dinarides is presented, down to 700 km depth. The 3-D velocity model shown here is derived by tomographic inversion of P travel times of a large number of selected regional and teleseismic earthquakes with shallow focus reported in the International Seismological Center bulletins (Piromallo and Morelli, 1997). The model is parameterized by a grid of nodes with about 50 km spacing, both horizontally and vertically. Two evident high (positive) velocity anomalies are detected: an E-directed shallow dipping body and a W-directed steeper anomaly, that can be interpreted as images of the Dinaric and Apennines subducted slabs, respectively. The scale of the anomaly contouring is in percentages of the ambient (reference) mantle velocity given by the 1-D reference model SP6 (Morelli and Dziewonski, 1993). Though the images must be regarded as blurred mapping of actual velocity structures, due to the combined effect of data errors, inhomogeneous sampling by seismic rays, model parametrization, linear approximation and inversion algorithm, the resolution of the model in the central part of the cross-section (eastern Tyrrhenian to western Turkey) is fair and the general shape of the subducted lithosphere can be resolved. The images which best show slab thickness and dip, are usually obtained by vertical cross-sections perpendicular to its strike. The section of Fig. 4 is probably slightly oblique to the Apennines and Dinarides slabs. Nonetheless, the figure summarizes the difference in the dip between the two subductions, being shallower and less steep the Dinaric slab.

The velocity anomaly distribution along the strike of the Apenninic chain could be influenced by the presence of lateral variations in thickness and nature of the lithosphere which underwent subduction. In particular, the lower velocities beneath the Southern Apennines detected by various tomographic studies in the shallow layers of the models (Amato et al., 1993; Spakman et al., 1993; Piromallo and Morelli, 1997) and interpreted as evidence for the detachment of the slab (Spakman et al., 1993) could be explained by a difference in the nature of the subducting lithosphere, this being of continental origin underneath the southern Apennines and oceanic or thin continental underneath the Calabrian arc.

Steep roots underneath the central Alps down to about 230 km have been proposed by Panza and Mueller (1979), Mueller and Panza (1986) and Du et al. (1998). The central–eastern Alps have been interpreted as the right-lateral transpressive segment of the Alpine orogen (Laubscher, 1971, 1983). Their steep roots may be interpreted as the dextral lateral ramp of the subduction zone which is in any case shallower than the Apennine ones.

6. Thermal state of W- vs. E–NE-directed subduction zones

The W-directed subduction zones present among the lowest and the highest heat-flow values of the earth, respectively, in the foredeep or trench and in the back-arc basin. The low values in the foredeep (up to 30 mW/m^2) are due to the deflection of the isotherms with the subduction and to the sediments filling the foredeep, whenever they occur. The accretionary wedge is generally formed of superficial thrust sheets mainly made of sedimentary cover and, therefore, a low thermal gradient is expected. A rapid increase of heat-flow is observed from the foredeep toward the back-arc basin (e.g., in the southern Apennines, up to 80 mW/m^2, Doglioni et al., 1996). Back-arc basins are sites of very high surface heat flow, up to 150 mW/m^2, which is the consequence of the thinning of the lithosphere. This has generally been considered as due to pure-shear extension and, therefore, characterized by a dome-shaped symmetrical structure (McKenzie, 1978). Wernicke (1985) and Lister et al. (1991) proposed also an alternative asymmetrical thinning of the lithosphere. A more careful examination of the distribution of the heat-flow in a back-arc basin reveals two important features: (1) the maximum value is not located in the central sector of the basin, but it is laterally displaced eastward, toward the subduction hinge; (2) the heat flow distribution follows that of the sub-basins and boudins with a series of highs and lows. This may be interpreted by supposing that the extension of the lithosphere and the upwelling of the asthenosphere is a pulsating phenomenon which occurs following the roll-back of the subducting slab. Probably each sub-basin may open by pure-shear or simple-shear extension. An example is represented by the back-arc basin related to the Apenninic subduction, which is in fact the entire western Mediterranean, where clearly heat-flow data and thermal modeling show that the maximum heat-flow values are encountered in the sub-basins: 120 mW/m^2 in the eastern Alboran (Polyak et al., 1996), 90–100 mW/m^2 in the Valencia trough (Foucher et al., 1992; Fernandez et al., 1995), and more than 200 mW/m^2 in the Tyrrhenian sea (Rehault et al., 1990; Mongelli et al., 1991).

E-directed subduction zones conversely present, on average, high heat flow values. Taking the central Alps as an example, we observe a heat flow of about 90 mW/m^2 in correspondence of the Molasse basin and the Helvetic Nappe, that decreases to about 70 mW/m^2 between the Penninic Nappe and the Insubric Line, and finally lowers to about 40 mW/m^2 in the Po Plain. This apparently contrasts with a thicker crust in the case of E-subductions that should yield average lower heat flow values than W-subductions. However, it can be explained as due to a crustal doubling and a corresponding doubling of the radiogenic layers. For a more detailed discussion of the thermal state of the Alpine subduction, see Cermak and Bodri (1996) and references therein.

Shallow dipping subduction zones should in principle be cooler than the steeper slabs, due to their longer permanence in the cooler upper parts of the upper mantle. This could also account for their stronger seismicity (e.g., Andes).

7. Main geological characters of subduction zones

Fold and thrust belts or accretionary wedges are the typical expression of subduction zones. They are the area where material of both the footwall and hangingwall plates is juxtaposed and shortened. The transfer from the footwall to the hangingwall plate is usually termed 'accretion', whereas the possibly temporary transfer from the hangingwall to the footwall plate is defined as 'erosion' (Von Huene and Lallemand, 1990; Von Huene and Scholl, 1991; Von Huene et al., 1994). Material eroded and transported down in subduction may eventually be re-exhumed by more advanced and deeper thrusts, this cycle providing a prograde and later retrograde metamorphic path.

Analogic models of accretionary wedges (e.g., Malavieille et al., 1992) show the importance of the distance between thrust ramps, depth of the decollement planes and the viscosity contrasts in the involved rocks. Accretionary wedges form on top of regional monoclines which dip toward the hinterland of the orogens. The dip may range between very low angles (1–3°) to higher values (7–15°). These steeper angles are more typical of W-directed subduction zones.

Accretionary wedges and orogens in general are zones of earth's degassing due to mantle processes, but they are also areas where there is a general expulsion of squeezed fluids contained in the sediments and rocks in general (e.g., Peacock, 1987, 1990; Le Pichon et al., 1992; Hyndman et al., 1993; Martin et al., 1995; Cochrane et al., 1996).

The orogens have variable width which is mainly function of the age of the subduction, depth of the decollement planes, and polarity of the subduction. We have the largest orogens such as the Andes and Himalayas which are active since hundreds of million of years. They have decollement planes that reach the base of the lithosphere and the subduction is E- or NE-directed. On the other hand, along the Marianas subduction zone, the accretionary wedge is poorly developed and where it exists is very narrow. That subduction is mainly Neogene–Quaternary in age, the accretionary wedge on top of the Pacific plate has very shallow decollement planes, mainly at the base of the sedimentary cover, and the subduction is W-directed.

Geologically, W-directed subductions are less known because they are mainly below the sea-level. W-directed subduction zones have arcuate shape and average length of 2000–3000 km, with the convexity verging toward the east. It was proposed that any subduction on a sphere generates an arc (e.g., Fowler, 1990). However, this geometric observation does not explain why E- or NE-directed subduction zones do not show such a shape (e.g., Cordillera, Himalayas).

The W-directed subduction zones are associated with low structural elevation (1–5 km of erosion), mainly single 'eastward' vergence (or southeast, or northeast) of the accretionary prism, mainly sedimentary cover involved, one foredeep with high subsidence rates, and a back-arc basin to the west.

The E–NE-directed subduction zones are instead characterized by orogens with high structural elevation (20–50 km of erosion), double vergence, basement rocks deeply involved by deformation, and two foredeeps with low subsidence rates.

W-directed subductions occur both in case of the highest convergence rates among plates (e.g., W-Pacific examples) and no-convergence (e.g., W-Atlantic examples). There are also W-directed subductions where the slab retreats eastward without E–W convergence along margins of plates independently traveling toward the NE (e.g., the Banda arc at the northwestern Australian margin, Veevers et al., 1978; Charlton, 1991).

Thrust belts associated with W-directed subductions have a shallow new Moho beneath the thrust belt with respect to the deeper and pre-subduction Moho of the foreland. Moreover, the accretionary wedges are composed of rocks scraped off the top of the subducting plate and basement rocks from pre-existing structural highs or inherited in the hangingwall from earlier orogenic belts. In some sections of such belts, the crust may be missing due to subduction. Associated with a W-directed subduction there always occurs a back-arc basin, with fast eastward progradation (50 mm/year, e.g., the Tyrrhenian sea).

On the other hand, thrust belts associated with E- or NE-directed subductions exhibit thickened crust along the orogenic belt (up to 60–70 km), large outcrops of crystalline basement, a double vergence (the frontal thrust belt synthetic to the subduction and the back-thrust belt antithetic to the subduction).

The main differences between the two thrust belts end-members may be interpreted as resulting from the polarity induced by the different behavior of the decollement planes associated with subductions opposing or following the mantle flow. In the W-directed subduction zones, the decollement at the base of the lithosphere is subducted and only shallow upper layers of the lithosphere are accreted. Moreover, the subducted lithosphere is replaced by new asthenospheric material in the back-arc. In E–NE-directed subduction zones, the decollement at the base of the lithosphere is ramping upward toward the surface and allows the uplift of deep seated rocks (Doglioni, 1992).

There is a vaste literature on the metamorphic rocks associated with orogens and subduction zones in general (see Ringwood, 1976; Peacock, 1996, with selected references therein for an updated list). According to the distinction that we are supporting in this article, the different behavior of the decollement planes in the opposite subduction zones should generate different metamorphic evolutions in the two end-members. The different kinematic systems should generate variables mechanisms of uplift or burial determining different PTt paths as a function of the subduction polarity and the location of the starting rock within the evolving system (Fig. 6). It will result that HP/LT metamorphism is intrinsic more with the frontal thrust belt of E- or NE-directed subduction zones, whereas the HT/LP metamorphism is more typical of the hangingwall of W-directed subduction zones where the asthenosphere replaces the slab at shallow levels (Fig. 6). In the hangingwall of the Apennines subduction (western Italian peninsula), a shallow asthenosphere has been documented (Calcagnile and Panza, 1981; Della Vedova et al., 1991; Doglioni, 1991; Mele et al., 1997). In a following chapter, we will see that W-directed subduction zones appear to generate to the east of pre-existing E-directed subductions. Relics of the Cordillera or Alpine type orogens are scattered respectively in the hangingwall and back-arc of the Barbados and Apennines subduction zones. They recorded the metamorphic history of both accretionary wedges, first the E-subduction related and later the W-subduction related thermal and pressure evolution. Therefore, it could be useful to discriminate the PTt history of orogens as a function of the subduction polarity: Within polyphased orogens the PTt history can help to unravel the kinematics of the opposed subduction zones. However, the W-directed subduction zones are relatively less studied, due to their general location below the sea-level. The Japanese and Apennines W-directed subduction zones are outcropping cases where to study the metamorphic history of this type of subduction. Nevertheless,

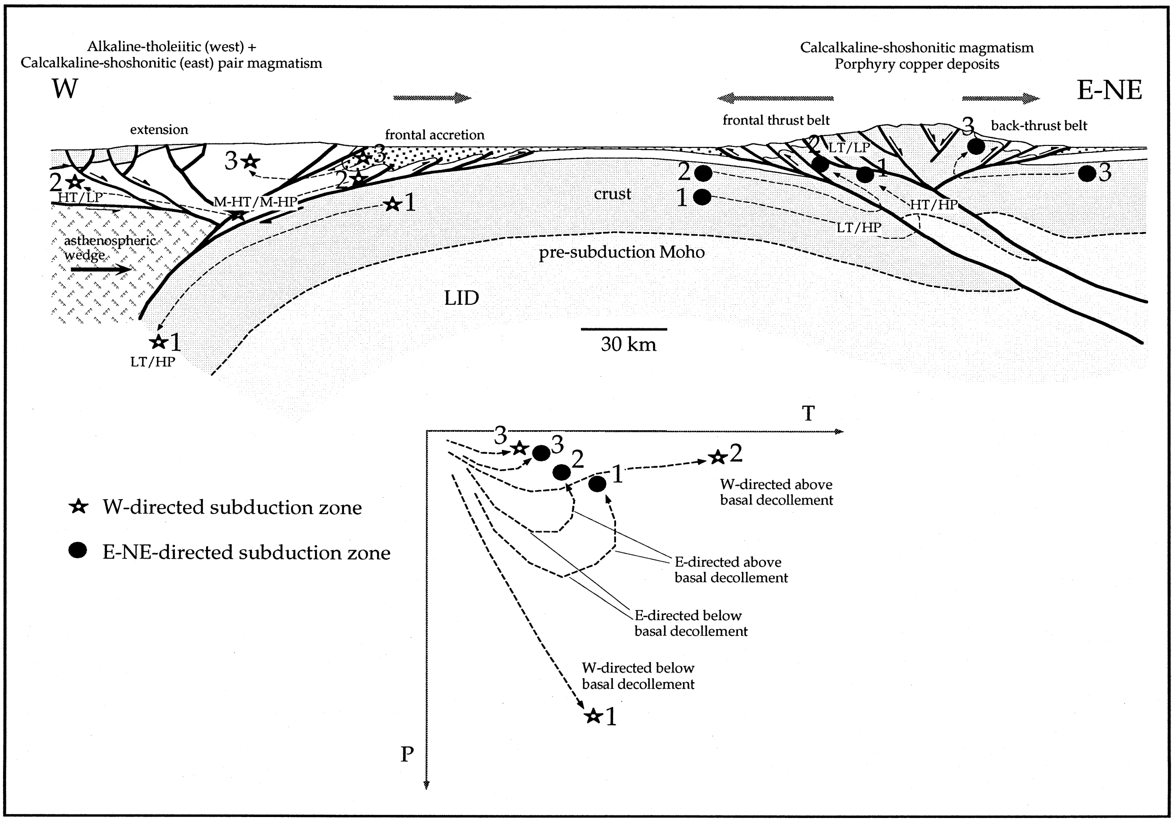


Fig. 6. Main structural differences among orogens due to W-directed and E–NE-directed subduction zones. The paths of three possible markers in the two systems illustrate different end-members of possible PTt metamorphic evolutions.

there are not univoque interpretations on the origin of the metamorphism in those belts: There are models proposing that the metamorphic rocks cropping out in Japan or in the Apennines were formed and uplifted during 'andean and alpine' E-directed subduction (Sillitoe, 1977; Cadet and Charvet, 1983; Doglioni et al., 1998), but there are also papers proposing that these rocks were generated by the normal evolution of the present W-directed subduction zone, without any flip of the direction of subduction (Jolivet et al., 1998; Rossetti et al., 1998).

There are evidences that the two opposite subductions may coexist (e.g., the Central America subduction zones, Nicaragua vs. Barbados). This should generate interference patterns both in terms of structures and PTt paths. This interplay probably occurred also in the Mediterranean where the W-directed Apennines subduction developed while the E-directed Alpine subduction was still active. This would explain the occurrence of HP/LT Early Miocene occurrences of Alpine affinity within the internal Apennines. However, there are authors that consider the HP/LT occurrences within the Apennines (Theye et al., 1997) as intrinsic of the Apennines subduction itself and not as indicators of the Alpine subduction history (Rossetti et al., 1998).

The interference between the oblique eastward migrating Apennines back-arc extension and the westerly migrating Betics compression (Doglioni et al., 1997) should have generated a variegate set of inversion structures and PTt paths in the western Mediterranean (Betic cordillera and Alboran sea).

Foredeeps are also very different depending upon whether they are associated with thrust belts generated by W-directed or E–NE-directed subduction zones. Along W-directed subduction zones, the subsidence rate of the foredeep may be up to 1–1.6 mm/year (e.g., Apennines and Carpathians), while along E- or NE-directed subductions they are of the order of 0.1–0.3 mm/year (e.g., Alps and Dinarides, Doglioni, 1994). This observation enables us to interpret the slow filling of foredeeps of the first type with huge flysch deposits (Apennines) or very limited deposition (Marianas trench), and the much faster filling of foredeeps of the second type which are quickly filled by the classic flysch–molasse sequences and then by-passed. In the first case the origin of the foredeep appears to be controlled by the eastward roll-back of the subduction hinge resulting from the eastward mantle push. In the second case, the westward or south–westward roll-back of the subduction hinge is probably controlled by the lithospheric load and by the downward component of the advancing upper plate, contrasting with the upward component of the eastward–northeastward mantle flow. The cross-sectional area of thrust belts associated with W-directed subduction is smaller in comparison with the area of the foredeep. The area of thrust belts associated with E–NE-directed subduction zone is instead always greater than the area of the foredeep (Doglioni, 1994). All these observations appear to maintain their general validity both in the case of subduction of oceanic lithosphere and of thin continental lithosphere.

In the foredeep of W-directed subduction zones, the subsidence rate due to roll-back of the subduction hinge is so high that there are cases where the frontal folds of the accretionary wedge uplift slower than the regional subsidence (Doglioni and Prosser, 1997). That can be called as negative fold total uplift (Fig. 7), to be distinguished from the typical case in which folds growth faster than the regional subsidence in Alpine type orogens (positive fold total uplift). The negative fold total uplift is characterised by envelope to the folds crest dipping toward the hinterland. There are evidences of negative fold total uplift even in trenches or foredeeps unfilled or poorly filled by sediments indicating that the subsidence is not simply due to the load of the clastic deposits filling the basin (e.g., Timor, along the southern arm

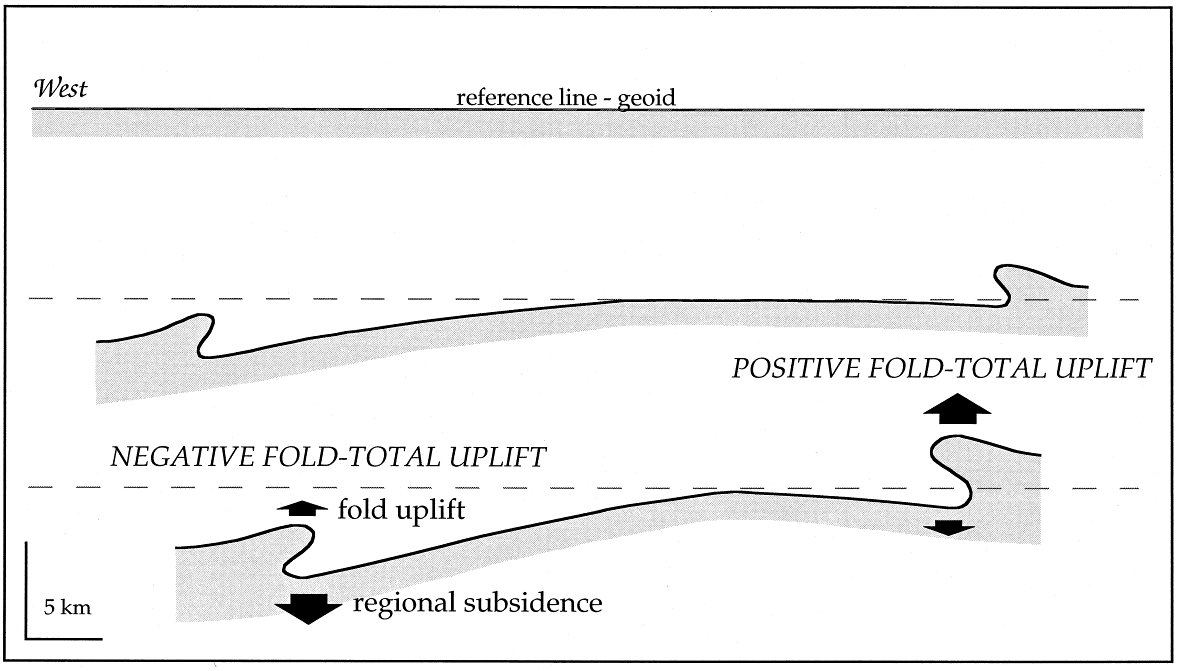


Fig. 7. The fold-total uplift may be defined as the single fold uplift rate and the regional subsidence. This value can be either positive or negative. This last case may occur along the front of W-directed subduction zones where the subsidence rates may be faster than the fold uplift rate.

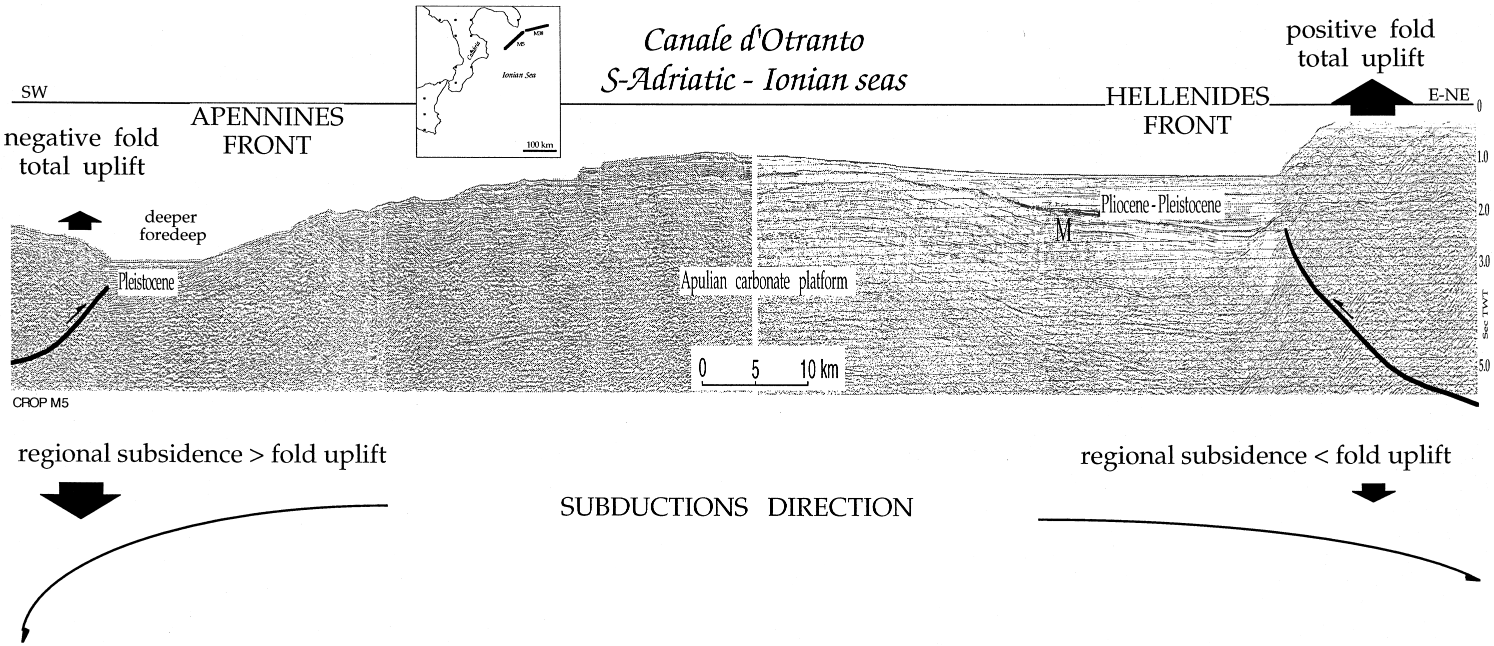


Fig. 8. Comparison between the Apennines and Dinarides–Hellenides front in the southern Adriatic–Ionian seas. Note the deepest foredeep along the Apennines subduction and the higher structural elevation of the Hellenides front. M, Messinian. The Apenninic front has negative fold-total uplift, whereas in the Hellenides it is positive.

of the Banda arc W-directed subduction zone). Fig. 8 is an example of the comparison of the two end-members; in the front of the Apennines accretionary wedge, the fold growths slowly than the regional subsidence, whereas at the front of the Dinarides the fold growths faster than the regional subsidence. Therefore, the fold tends to outcrops from the sea.

Close observations on these differences between orogens have been pointed out also by Royden and Burchfiel (1989) and Royden (1993). They associated these differences to the efficiency of the slab pull, regardless of the subduction geographic polarity. However, the westward drift of the lithosphere relative to the asthenosphere can contribute to produce these differences. In fact the polarity imposed by the westward delay of the lithosphere should generate different kinematics of the decollement planes along opposed subduction zones (Doglioni, 1990, 1992). These determine that along a W-directed subduction zone much part of the down-going plate is lost and only superficial layers are involved in the accretionary wedge (e.g., Apennines). The decollement at the base of the down-going plate to the east is subducted. In fact the volume of the orogen in the hangingwall of W-directed subduction zones is always minor. In a section of the Apennines in northern Calabria it has been calculated in about 2.900 km^2 vs. about 62.000 km^2 of subduction. Only a little amount of the crust is involved in the orogen and much part of it is subducted. We observe thin-skinned compressive tectonics at the frontal accretionary wedge mainly involving only the sedimentary cover, whereas the belt is characterized by thick-skinned extensional tectonics cross-cutting the entire crust. On the contrary, along opposed E–NE-directed subduction zones, the thrusts entirely involve the upper and lower plates, thickening in the orogen both the crust and the lithospheric mantle (e.g., Alps, Polino et al., 1990; Roure et al., 1990, 1996). That is why we observe large slices of basement in the Cordillera or similar orogens. Therefore, the orogens are characterized by thick-skinned compressive tectonics throughout the double verging orogen, apart the frontal parts which frequently exhibit thin-skinned tectonics (e.g., Jura at the front of the Alps, Rocky Mountains) which are, however, always rooted in basement ramps along more internal zones. Extensional tectonics in those orogens are confined to the upper layers, while compression is acting at depth (e.g., Dewey, 1988).

The kinematics of W-directed subduction zones generate a loss of lithosphere which is eastward retreating. This movement has to be compensated by the asthenosphere which goes to replace the rolling-back slab. A shallow asthenosphere in the hanging-wall of W-directed subduction zones is commonly observed and can explain the high heat flow of these areas (Doglioni et al., 1996).

8. Topography and gravimetry of subduction zones

An analysis of topographic profiles associated with satellite free air gravimetric anomalies profiles along subduction zones (Harabaglia and Doglioni, 1998) shows that average low topography (-1250 m) and pronounced gravimetric anomalies characterise W-directed subduction zones. The average depth of the trenches is -5000 m (Fig. 9). On the contrary, average elevated topography (1200 m) and smoother gravimetric waves are peculiar to E- or NE-directed subduction zones. The average depth of the trenches is about -3000 m for the E- or NE-directed subduction zones. The W-directed subduction zones are characterized by a narrow arc-chain in the hangingwall of the subduction (about 200–300 km in section), that rises to 2000–3000 m over the mean plate height. The subducting plate is typically only 1000–2000 m lower than the overriding one. If the subducting plate is oceanic, there is always a pronounced trench, whereas for continental plates, the trench may be filled (e.g., Carpathians) or partly filled (Banda) by sediments. The associated volcanic arc is typically well defined.

W-directed subductions show strong negative free-air gravimetric anomalies with an asymmetric shape (150–200 mgal) along the trench (Fig. 9), and a prominent positive signature (over 100 mgal) corresponding to the arc-chain, and similar gravimetric values on both plates immediately off the trench–arc system.

Conversely E–NE-directed subductions are characterized by less pronounced negative gravimetric anomalies (about 100 mgal or less) with a strong asymmetric shape relative to the trench (Fig. 9), less

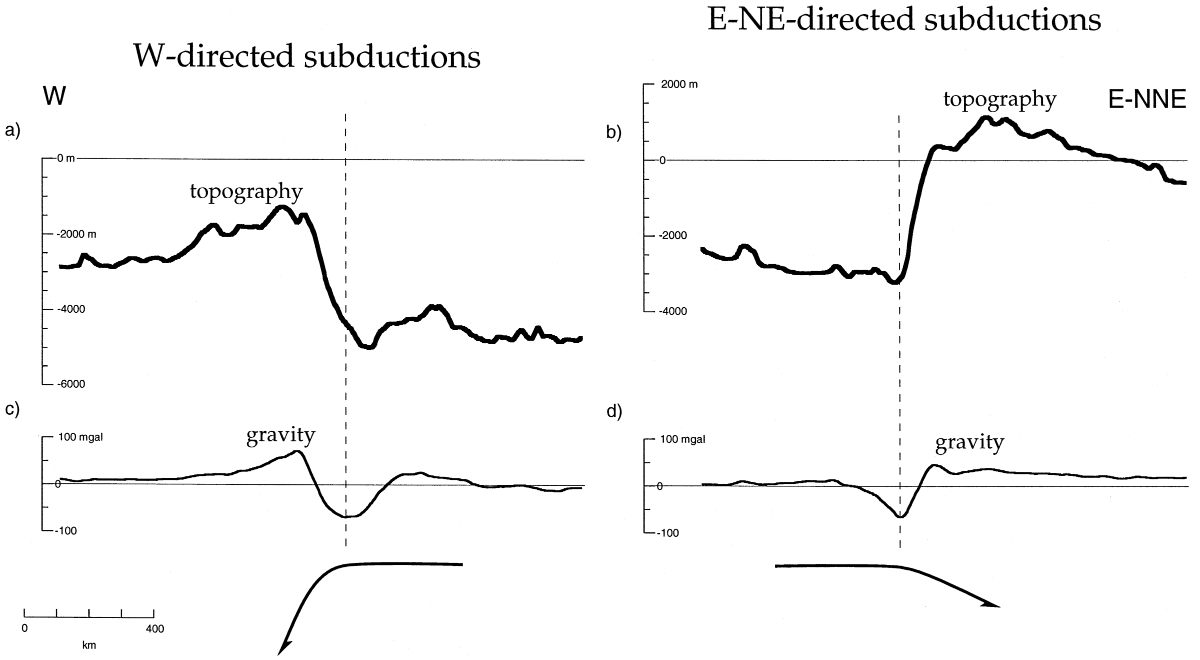


Fig. 9. Average topographic profiles (a), (b), and free-air gravimetric profiles (c), (d) across the main subduction zones of the earth. These confirm the presence of two classes of subduction zones of which differences are related to the geographic polarity of the subduction zone (after Harabaglia and Doglioni, 1998).

pronounced positive anomalies (less than 100 mgal) corresponding to the orogen, and higher gravimetric values (20–40 mgal) on the overriding plate immediately off the back-thrust belt with respect to those observed on the subducting plate immediately before the plate bulge. In case of subducting continental plates, there seems to be a large compensation.

Some of the subduction zones do not present all the peculiarities of the W- or E-class. However, none of them differs by more than a couple of characteristics from the standard of its assigned class. Therefore, topography profiles across subduction zones show two distinct signatures.

The average gravity profiles across the two end-members of subduction zones (W- and E–NE-directed) are very different both in terms of amplitude, shape and relationship with topography. In the Apennines (W-directed subduction zone), for example, the minimum gravimetric values are located more toward the foredeep rather than exactly below the mountain range (Mongelli et al., 1975), while they are more coupled with the topography in the Alps (Nicolas et al., 1990) where a lithospheric root has been detected (Mueller and Panza, 1986).

All these differences are particularly evident along the Pacific margins, but they persist also along the other subduction zones of the world in the Atlantic, Mediterranean, Himalayas and Indonesia regions, and they are primarily function of the geographic direction of the subduction, regardless of the involved lithosphere. The Italian geodynamics is shaped by two opposite subduction zones, i.e., the western Alps and the Apennines, which fall respectively into an E-directed subduction zone and a W-directed subduction zone. The two belts show the characters of the global signatures in terms of topography and gravimetry as a function of the subduction polarity, having the same Adriatic plate involved, i.e., overriding Europe in the Alps, underthrusting Europe in the Apennines (Fig. 10). Obviously, this conclusions may not be so straightforward if we were to look at different data sets. For example, Marson et al. (1995) obtain rather different results by forward modelling of bouguer anomalies. In particular, they note that

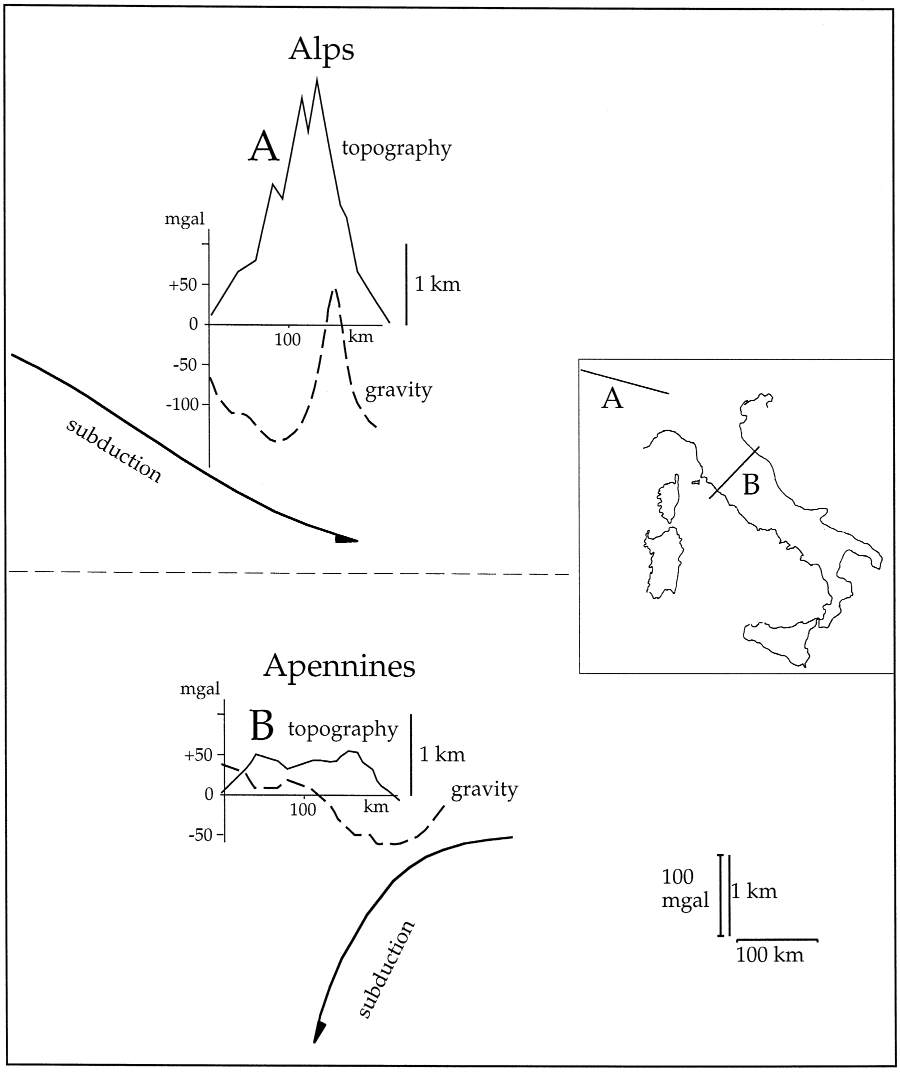


Fig. 10. Application of the topographic and gravimetric differences between orogens to the Italian thrust belts. The Alps show elevated topography and gravimetric minimum relatively coupled with the belt (A, after Buness, 1991). The Apennines rather show lower topography and a gravity minimum displaced toward the foredeep of the belt (B, Mongelli et al., 1975).

the slab in the Southern Apennines should be much shallower than along the Calabrian arc and Tuscany. This is not in contradiction with Harabaglia and Doglioni (1998) since the latter limited themselves to observe that a W-directed subduction would be characterized by a steeper slab than an E-directed one and this would reflect on the general trend of the satellite free air anomaly. Also Marson et al. (1995) assume that the slab penetrates at a high angle; they only disagree on how deep the penetration is in the various portions of the subduction. Therefore, topography and gravimetry appear to confirm the existence of two separate classes of subduction zones regardless the age and nature of the subducting slab.

9. Back-arc basins

Back-arc basins are typical features forming in the hangingwall of subduction zones. They may show very low Q values (Barazangi and Isacks, 1971), which indicate low viscosity in the mantle wedge at the top of the subduction hinge (Mele et al., 1997). They are notably well developed in the western Pacific (Karig and Sharman, 1975; Zonenshain and Savostin, 1981; Brooks et al., 1984; Tamaki and Honza, 1985, 1991; Taylor, 1993; Honza, 1995; Hawkins, 1995; Taylor and Natland, 1995), and they are particularly associated with W-directed subduction zones (Uyeda and Kanamori, 1979; Doglioni, 1991). They are characterized by semicircular or triangular shape, the highest subsidence rates for extensional environments (up to 600 m/Ma), and they have mainly ages ranging from early Tertiary to Recent (Doglioni et al., in press). The extension in the back-arc propagates eastward like the roll-back of the subduction zone. The stretching is discontinuous, producing basins and swells that recall the boudinage (Ricard and Froidevaux, 1986; Gueguen et al., 1997). An example is the western Mediterranean which is composed by sub-basins which are younger from west to east, and they developed in the hangingwall of the W-directed and eastward retreating Apennines subduction (Fig. 11). The basins are often floored of oceanic crust which is also rejuvenating toward the east (e.g., Parece Vela basin, Caribbean sea and western Mediterranean).

The western margin of the back-arc basin has a convexity opposed to that of the subduction arc, i.e., toward the west. In other words, the shape of the back-arc is in some way specular to the one of the main arc. The reentrance of the western margin of the back-arc basin is particularly visible on the Asian margins of the Japan sea and the South China sea, and the Cordillera margin of the Caribbean sea. The westernmost initiation of the back-arc extension probably corresponds to the latitude of the location of the first onset of the W-directed subduction to the east and in the end to the largest amount of subducted slab and the widest back-arc extension.

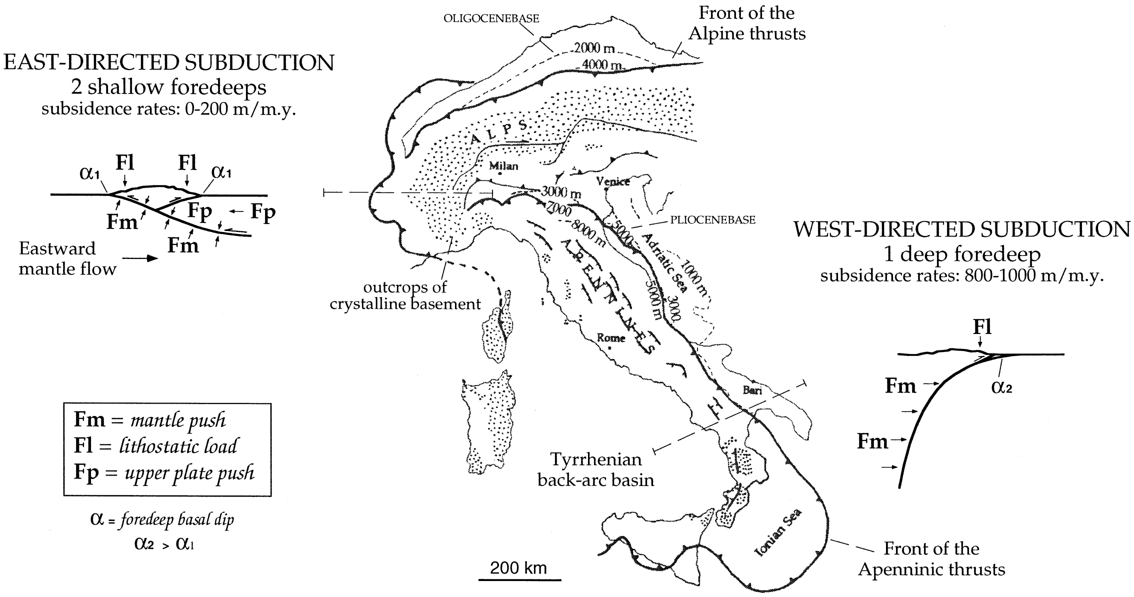


Fig. 11. Alps and Apennines are thrust belts with very different characters: Alps have a shallow foredeep, broad outcrops of metamorphic rocks, and high structural and morphologic relief. Apennines have deep foredeep, few outcrops of basement rocks (partly inherited from earlier Alpine phase), low structural and morphologic elevation, and Tyrrhenian back-arc basin. These differences mimic asymmetries of Pacific east-dipping Chilean and west-dipping Marianas subductions. For east-dipping subduction foredeep origin may be interpreted as controlled by lithostatic load and downward component of upper plate push, minus upward component of the 'eastward' mantle flow. For the west-dipping subduction where subsidence rate is much higher, foredeep origin may rather be interpreted as due to horizontal mantle push and the lithostatic load (after Doglioni, 1994).

Back-arc basins have maximum width variable between 800 and 1500 km. The maximum and minimum widths of the back-arc basin correlate to the west at the maximum and minimum amount of subduction present to the east. For instance, the largest opening of the Tyrrhenian sea corresponds to the deepest part of the Apennines slab. The eastward migration of the back-arc extension is accommodated by asymmetric rifting, with low-angle normal faults dipping to the east. These faults have been recognized in the western Mediterranean back-arc setting, e.g., in the western margin of the Provençal basin (Benedicto et al., 1996), in the northern Apennines (Barchi et al., 1997) and in other western Pacific back-arc basins. E-dipping normal faults are usually spaced (10–50 km) and sometimes they isolate large boudins of thicker continental crust generating a lithospheric boudinage both in the Pacific, Atlantic and Mediterranean back-arc basins (Gueguen et al., 1997).

Back-arc spreading associated to the W-directed subductions may develop both within the former orogen or even far into the foreland of the frontal thrust-belt of the earlier E-directed subduction zone, probably as a function of the width of the orogen (e.g., Japan sea, Provençal basin). Part or the entire orogen of the former 'E'-directed subduction zone is stretched and boudinated in the hangingwall of the W-directed subduction, in the back-arc region. HT-LP metamorphism associated to asthenospheric wedging in the back-arc underneath the former orogen commonly overprints HP-LT metamorphic assemblages.

The western Mediterranean back-arc basins associated to the W-directed Apenninic subduction provide a complete set of these variations even along strike in the opening of the back-arc basins oblique to the former Alpine orogen. The back-arc spreading associated to these subductions may jump from one segment of lithosphere to another, generating large scale boudinage of the lithosphere. The subducted lithosphere is replaced by the uprising and laterally compensating asthenosphere in the back-arc basin. There are several interpretations of back-arc basins associated also to E–NE-directed subduction zones, such as the Aegean sea, the Basin and Range province, the Andaman sea. Although those basins are in the hangingwall of the subduction, they present geodynamic settings which deeply differ from the back-arc basins of W-directed subduction zones. They usually have thicker crust, there are two or even three plates involved in the system and they show very different magmatic and tectonic evolution. Moreover, the extension more rarely arrives to oceanization and it is not always coeval with the subduction. For a review of those extensional environments, see Doglioni (1995).

The asymmetrical opening of a back-arc basin maybe generated by a series of factors, the most relevant being the slab induced mantle flow, combined with the differential drag between the eastward intruding asthenosphere and the overlying crust, and, to a lesser extent, thermo-mechanical factors that might cause a differential drag. The mantle drag interferes with the induced flow. The former can be viewed either as a resistive force or as a driving force in plate dynamics depending on its magnitude and direction with respect to plate motion. Ricard et al. (1991) have shown that on average the lithosphere has a delay of about 2 cm/year with respect to the mantle. This means that in case of a 'W'-directed subduction it would add its effect to that of the induced flow in the arc-region while in case of an 'E'-directed subduction it would counteract the induced flow effect. A rough estimate of the mantle drag due to this delay is:

$$\tau_d = \eta \frac{\Delta V}{\Delta h}$$

where η is the mantle viscosity, Δh is the upper mantle thickness, and ΔV is the velocity variation within it. Assuming a viscosity $\eta = 4 \times 10^{19}$ Pa s, a mantle thickness of $h = 550$ km, and a velocity that varies from 2 cm/year at the base of the lithosphere to 0 at the bottom of the upper mantle, we obtain a mantle drag estimate of about 46 kPa away from the slab region. However, the value proposed by Ricard et al. (1991) is an average estimate and it might be larger in certain regions. Moreover an estimate of the drag in the arc-region is not straightforward; the different geometry of the system might play an important role and increase its effect as it probably does the fact that the slab contemporaneously retreats.

According to Turcotte and Schubert (1982), a sinking slab produces a flow with horizontal velocity component:

$$u = -B - D\tan^{-1}\frac{z}{x} + (Cx + Dz)\left(\frac{-x}{x^2 + z^2}\right)$$

where B, C, and D are constants that depend on boundary conditions, x is the distance from the subduction hinge and z is the depth. Assuming that the mantle behaves as a Newtonian fluid with constant viscosity η, the shear stress is:

$$\tau = \eta\frac{\mathrm{d}u}{\mathrm{d}z}.$$

At the base of the lithosphere ($z = 0$), the derivative of the horizontal component of velocity with depth reduces to:

$$\frac{\mathrm{d}u}{\mathrm{d}z} = -2\frac{D}{x}$$

A general expression for B, C, and D is not straightforward; they depend on the subduction angle and the hinge migration velocity. According to Harabaglia (1998) given U, the speed at which the subducting plate moves toward the trench, $D = 1.9U$ for a subduction that penetrates with a 30° angle and no trench migration. D triplicates in case of a subduction where there is no active convergence, that is the hinge migration velocity is $-U$, and it decreases abruptly in the opposite case. D also depends on the subduction angle and becomes larger for smaller angles; this would contrast with the fact the most developed back-arc basins are associated to the steep 'W'-directed verging subductions. However, this dependence is less relevant then that due to the hinge migration. Moreover, when a subduction initiates it is always characterised by a low angle $< 35°$. Even when a slab is fully developed, a steep inclination is reached progressively due to the elastic bending of the slab (Marotta and Mongelli, 1998). Consequently, the average slab angle remains low in the shallower portion and the validity of this model should be intact. In particular assuming $\eta = 4 \times 10^{19}$ Pa s and $U = 1$ cm/year at 10 km from the hinge we would have a stress of 4.7 MPa; in case of retreating hinge this value would increase at 14 MPa Since most 'W'-directed subductions are characterised by hinges with retreating velocities of 4–10 cm/year this would mean stresses of about 50–150 MPa. These values should be large enough to cause the breaking of a weakened lithosphere and the opening of a back-arc basin.

We must also consider what should be second order effects due to thermo-mechanical variations in the lithosphere and the mantle in the arc-corner. In case of 'W' subductions, we expect the presence of a mantle wedge; this probably expands adiabatically (Doglioni et al., 1996) increasing the mantle viscosity (may be by an order of magnitude) and ultimately the coupling between mantle and lithosphere in proximity of the subduction hinge.

In conclusion, there is a relevant stress at the base of the lithosphere in the arc-corner of W-directed subductions since these are characterized by an eastward migration of the subduction hinge (Doglioni et al., 1996); this can eventually generate faults that propagate from the base of the lithosphere whenever the latter is already weakened by a previous subduction.

10. Structural evolution of W- vs. E–NE-directed subduction zones

Comparing the structural evolution of W- vs. E- or NE-directed subduction zones, we note a much larger amount of erosion along the last ones. We know that several tens of kilometer of embricated rocks have been eroded in the Alps and Himalayas (Argand, 1924). High rates of uplift in the Betics (5–10 km/Ma) have been reported by Zeck et al. (1992). An uplift of 1.5 cm/year has been documented in Taiwan by Angelier et al. (1998). This did not occur along W-directed subduction zones that never reached high structural elevation. The Apennines are a good example where the lithostatic load has never been very high (Corrado, 1995). W-directed subduction zones mainly develop horizontally moving as a wave toward the 'east' at rates of a few centimeter per year. The uplift in the belt may be in the order of a millimeter per year, but it works only for the short timespan in which the tectonic wave crosses an area; the topography closely matches with the structure. E- or NE-directed subduction zones have rather orogens with persistent and higher uplift

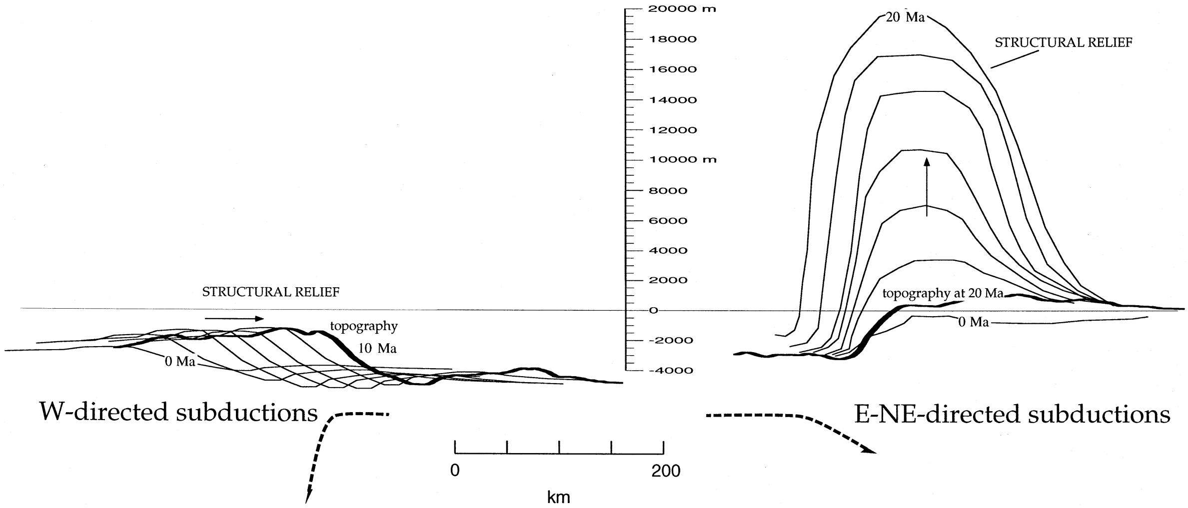


Fig. 12. Different structural evolution along the opposite subduction zones. W-directed subduction zones are characterized by an 'E-ward' migrating structural wave that closely matches topography. The E- or NE-directed subduction zones have a much higher structural elevation. The growth of the waves is constructed assuming conservative values of 2–3 cm/year of eastward migration of W-directed subduction zones and 1 mm/year uplift for the opposite subduction zones. Note how the structural differences among the subduction zones are even more evident than the topographic signatures.

rates; the belts expand laterally in both sides, and they have uplift rates that may be faster than 1–2 cm/year. Fig. 12 simplifies these differences among subduction zones which are even more striking than those marked by the topographic profiles.

11. Coexisting compression and extension along W-directed subduction zones

The active accretionary wedge of W-directed subduction zones usually forms below the sea-floor, it is usually single verging toward the east (or northeast, or southeast). The wedge is followed by an extensional wave cross-cutting the thrusts and folds previously formed in the accretionary wedge (Mazzanti and Trevisan, 1978). The Apennines are a natural laboratory where to study the interplay between compression and extension. The extension has been very well described in the deep seismic profile Crop 03 of the northern Apennines (Barchi et al., 1997) and it shows an asymmetric character with mainly E-dipping undulated normal faults. It has always been a task to interpret the coexistence of this tectonic pair (Lavecchia et al., 1994; Meletti et al., 1995; Frepoli and Amato, 1997). Here, we propose a kinematic interpretation based on the shape and kinematics of the W-directed subduction zones.

The boudinage and the extension in the hangingwall of a W-directed subduction may be attributed to the differential drag generated by the slower horizontal component of the mantle flow induced by subduction near the mantle wedge, and by thermal constraints (Fig. 13). Back-arc extension develops as soon as the subduction starts, suggesting prevalent mechanical controlling factors. The shear stress at the base of the hangingwall lithosphere due to the induced flow has an exponential decrease moving westward, away from the subduction hinge and this could explain why E-dipping normal faults in the hangingwall of the subduction nucleate on top of the mantle wedge at the subduction hinge. Moreover, the eastward roll-back of the subduction hinge implies a volume deficit in the hangingwall lithosphere which is following and trying to compensate the hole. This is what can be considered as the trench suction (Forsyth and Uyeda, 1975) and it may be responsible for the main 'eastward' dipping normal faults in the Apennines.

On the other hand, the shortening in the accretionary wedge can be explained as related to the shear between the down-going and retreating litho-

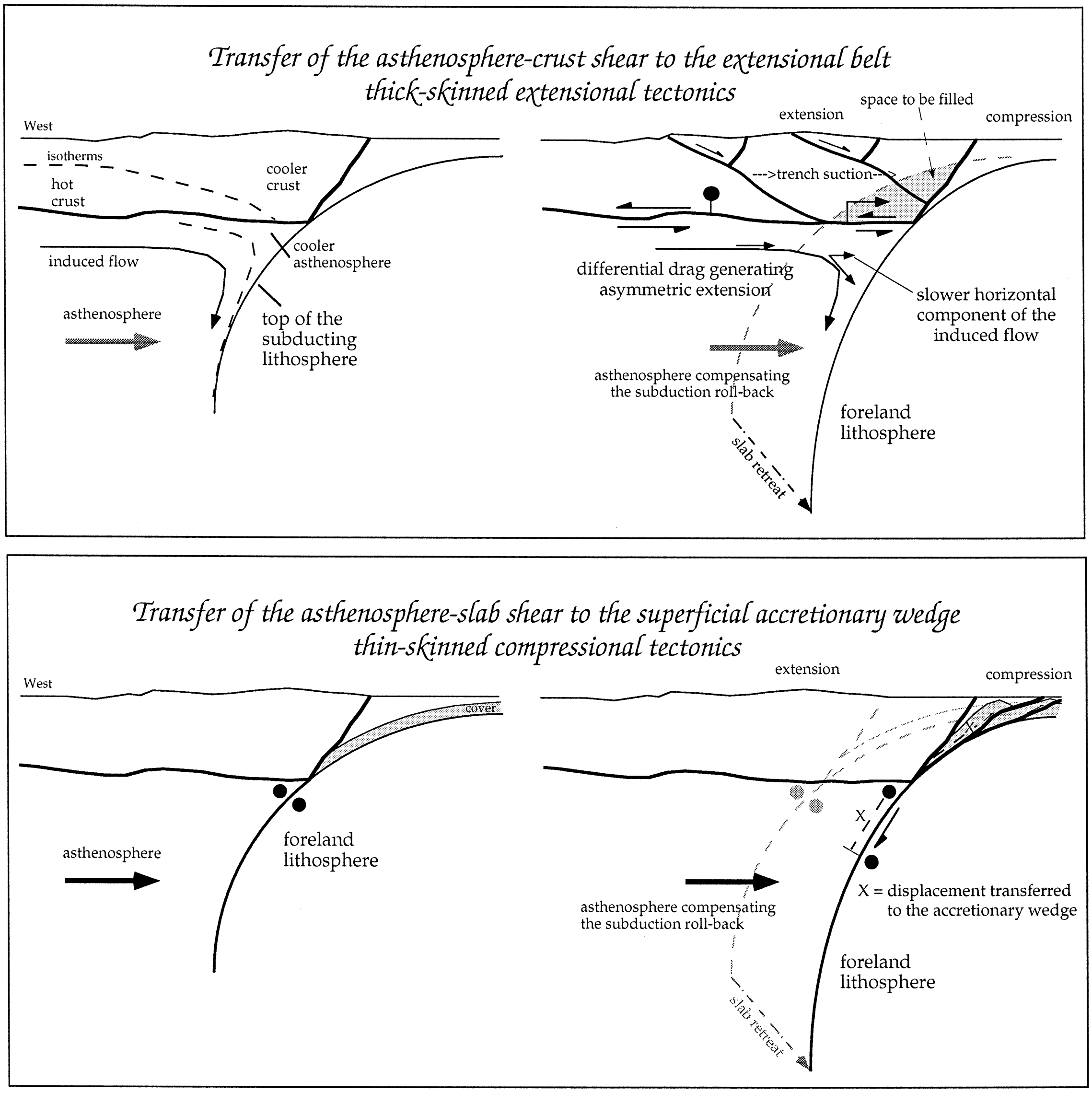


Fig. 13. Kinematic model for the association compression–extension of W-directed subduction zones as the Apennines. The boudinage and the extension in the hangingwall of a W-directed subduction may be attributed to the differential drag between the eastward intruding asthenosphere and the overlying crust. The differential drag may be controlled by the slower horizontal component of the mantle flow induced by subduction near the mantle wedge, and by thermal constraints. Moreover, the retreat of the subduction hinge generates the trench suction. The shortening in the accretionary wedge can be explained as related to the shear between the down-going and retreating lithosphere and the eastward compensating mantle. The displacement is transferred upward and peels-out the cover from the foreland lithosphere.

sphere and the eastward compensating mantle, the displacement being transferred upward and peeling-out the cover from the foreland lithosphere (Fig. 13). These kinematic model for the eastward migrating compression–extension pair is triggered by the subduction roll-back and by the simultaneous asthenospheric replacement. Therefore, the compression in the accretionary wedge may not be an indication of what the foreland plate is doing with respect to the plate located to the west of the back-arc basin. Unfortunately, many kinematic models utilize the focal mechanisms of the accretionary wedge of W-directed subduction zones as an expression of plate kinematics, but those data may be misleading because that compression may be simply related to the local roll-back of the subduction hinge, e.g., the southern arm of the Apennines in Sicily, where compression is usually interpreted only in terms of Africa–Europe relative motion.

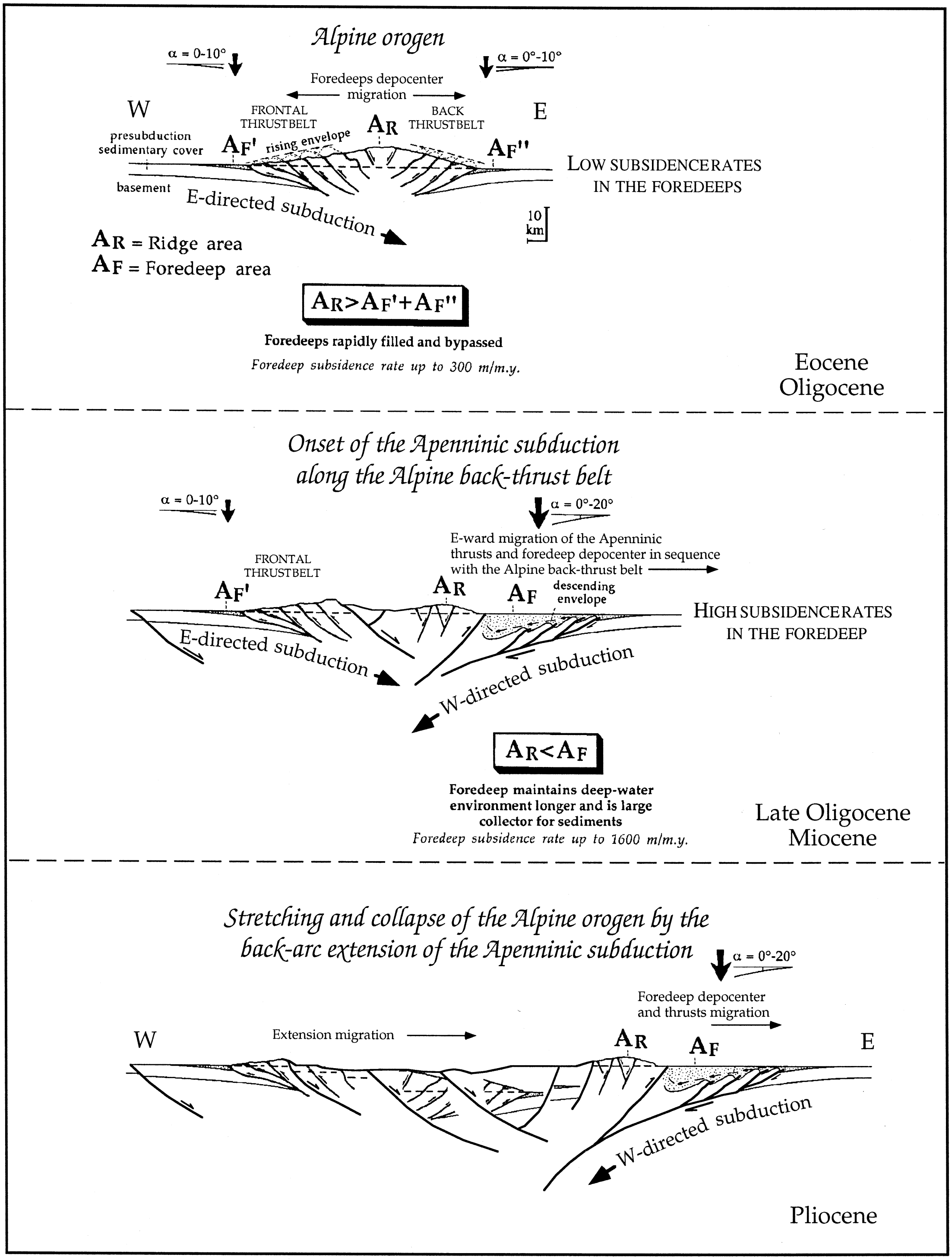
12. Initiation of subduction zones: the Atlantic examples

The initiation of subduction zones has always been a fundamental but complicated mechanical problem (e.g., Cloetingh et al., 1982, 1989). Subduction zones develop either along passive continental margins or at strong lateral variations in thickness and composition of the lithosphere. Considering the Barbados and Sandwich arcs, W-directed subduction zones appear to have nucleated along back-thrust belts of pre-existing E-directed subductions, if oceanic or thinned continental lithosphere were present in the foreland of the 'E'-verging back-thrust belt. The W-directed subductions developed only in Central America and south of Patagonia, where the Northern and Southern American continents narrow. This seems to indicate that the W-directed subductions, which are worldwide Tertiary and Quaternary features, possibly form only in the presence of particular geodynamic constraints, i.e., (1) along the back-thrust belt of earlier E-directed subduction zones, and (2) in the presence of oceanic or thinned continental lithosphere in the foreland of the related back-thrust belt (Doglioni et al., in press). We might apply similar geodynamic parameters for the Apennines–Maghrebides development, i.e., the Apennines W-directed subduction formed along the back-thrust belt of the pre-existing Alps–Betics orogen related to an E-directed subduction zone (Fig. 14). This inversion of the subduction could have occurred only where a branch of oceanic lithosphere was located to the east of the Alps–Betics orogen. Evidence of Mesozoic oceanic crust of the Tethys realm (Bernoulli and Lemoine, 1980; Dercourt et al., 1986; Ziegler, 1988) has been unraveled on the basis of oceanic sediments and slices of ophiolites in the Apennines. Remnants of the former Alpine orogen were passively incorporated into the internal parts of the accretionary wedge related to the younger W-directed subduction. The presence of Alpine Miocene records within the Apennines would support a time of coexistence of the E-directed Alpine subduction and the W-directed Apennines subduction zone, like it was in Central America where the Caribbean subduction initiated while the opposite Andean subduction was and still is active. Remnants of the earlier Andes were later scattered to the east in the Caribbean back-arc domain (e.g., Cuba, Haiti) like the present Alpine relics are spread out in northern Africa and Apennines (e.g., Kabylie, Calabria).

13. Lifetime of W- vs. E–NE-directed subduction zones

Present W-directed subduction zones developed during the Neogene. Some of them started during the Paleogene. Generally speaking, they are younger than 50 Ma (usually less than 30 Ma). The western Pacific back-arc basins are Tertiary in age (Seno, 1985; Honza, 1995), like the Mediterranean analogues (Horvath and Berckhemer, 1982). This does

Fig. 14. Following the Atlantic examples, the Apennines W-directed subduction zone developed along the back-thrust belt of the Alps and then it rapidly retreated eastward. The flip of the subduction from Alpine E-directed to Apenninic W-directed was probably recorded by the sudden increase of the subsidence rates in the Apenninic foredeep filled by the Macigno. In fact W-directed subduction-related foredeeps exhibit subsidence rates about 10 times higher with respect to the foredeeps related to the opposite subduction zones. The three stages show the evolution from the Alpine orogen to its collapse in the back-arc basin related to the W-directed Apenninic subduction. The middle figure, shows the temporary coexistence of the two opposite subductions during the Late Oligocene–Early Miocene (after Doglioni et al., 1998).



Alpine orogen
α = 0-10°
α = 0°-10°
Foredeeps depocenter migration
W
E
FRONTAL THRUSTBELT
BACK THRUSTBELT
AR
AF'
AF''
rising envelope
presubduction sedimentary cover
basement
E-directed subduction
LOW SUBSIDENCE RATES IN THE FOREDEEPS
10 km
AR = Ridge area
AF = Foredeep area
AR>AF'+AF''
Foredeeps rapidly filled and bypassed
Foredeep subsidence rate up to 300 m/m.y.
Eocene Oligocene
Onset of the Apenninic subduction along the Alpine back-thrust belt
α = 0-10°
α = 0°-20°
E-ward migration of the Apenninic thrusts and foredeep depocenter in sequence with the Alpine back-thrust belt
FRONTAL THRUSTBELT
AF'
AR
AF
descending envelope
HIGH SUBSIDENCE RATES IN THE FOREDEEP
E-directed subduction
W-directed subduction
AR<AF
Foredeep maintains deep-water environment longer and is large collector for sediments
Foredeep subsidence rate up to 1600 m/m.y.
Late Oligocene Miocene
Stretching and collapse of the Alpine orogen by the back-arc extension of the Apenninic subduction
α = 0°-20°
Foredeep depocenter and thrusts migration
W
Extension migration
AR
AF
E
W-directed subduction
Pliocene

not imply that W-directed subduction zones did not exist before the Tertiary, but simply that they have a fast and short evolution in the geologic record. On the contrary, E–NE-directed subduction zones may have been active for more than hundred million years. There may be two simple reasons why W-directed subduction zones terminate.

(1) The encroachment of the arc with thick continental lithosphere in the foreland to the east is one case in which the subduction stops due to the high buoyancy values of a thick continental lithosphere which is not subductable. Example may be the Paleozoic Antler orogeny in the western US (Johnson and Pendergast, 1981; Burchfiel and Royden, 1991; Carpenter et al., 1994) and the Neogene Carpathians that reduced their eastward migration during the Pleistocene (Horvath, 1993; Tomek and Hall, 1993; Linzer, 1996). Similar reason has been postulated for the termination of E-directed subduction zones as well, but in this last case we may observe a real collision between upper and lower plate, collision which does not occur along the arc of W-directed subduction zones because the upper plate always remains westward of the back-arc, with an eastward motion slower than the roll-back of the subduction hinge. Slowing of a W-directed subduction zone due to the presence of thick continental lithosphere in the foreland is observed along some segments of the Apennines where the heterogeneity of the foreland controlled the different rates of slab retreat (Doglioni et al., 1994). The forebulge uplift in the foreland is more developed where the lithosphere decreases its subduction rates along segments of thick continental lithosphere.

(2) The second reason for the end of a W-directed subduction zone or the switch to an E-directed subduction zone of the system is implicit in the kinematics of W-directed subduction zones (Doglioni, 1991, 1993). In fact the generation of a back-arc basin introduces two new important lateral discontinuities in the lithosphere at the margins of the basin in the hangingwall of the subduction. Those weak zones are observed to be the main areas of initiation of subduction zones. The presence of those discontinuities can trigger the change in the subduction polarity, e.g., the eastern margin of a back-arc basin with thin or oceanic lithosphere to the west might become the seat for the development of an E-directed subduction zone, like it occurred in the eastern margin of the South China sea. The South China sea probably developed as the back-arc basin of the W-directed Philippine subduction, but since late Miocene or early Pliocene an E-directed subduction zone generated along its eastern margin from Taiwan in the north (Lallemand and Tsien, 1997), to western Borneo in the south. This active 'E'-directed subduction appears to be closing an earlier back-arc basin generated by a W-directed subduction zone.

The self generation of discontinuities in the back-arc basin seems to control the praecox self destruction of the W-directed subduction zones which alternate repetitively as a yo-yo from W- to E-directed subduction polarity, as it occurred in the western Pacific subduction zones during the Phanerozoic. This could explain the young Tertiary age of the W-directed subduction zones.

On the other hand, E–NE-directed subduction zones may remain active for longer periods of time if the converge among plates is maintained and there is subductable lithosphere to the west of the hangingwall plate. These constraints are well documented in the eastern Pacific where the Andean subduction has been active since the Paleozoic.

14. Magmatism and subduction zones

Subduction zones are characterized by extensive magmatic activity. Calc-alkaline rocks are by far the dominating products. These are often associated with a variety of magmatic products, which include island arc low-potassium tholeiites, shoshonites, and Na- and K-alkaline rocks. The calc-alkaline volcanics display variable abundances in potassium and have been divided into normal and high-K calc-alkaline rocks (Peccerillo and Taylor, 1976). In the majority of subduction zones, calc-alkaline andesites represent the most common lithologies (e.g., Gill, 1981; Thorpe et al., 1982), even though mafic or acidic rocks may dominate in some volcanic arcs and sectors of active continental margins.

A scrutiny of the available data on recent to active subduction-related magmatism shows that there are important differences between E- and W-directed subduction zones. These apply to several characteristics including: (1) the complexity of the magmatic setting in terms of number of different rock suites in the two settings; (2) the relative abundances of rock types with variable degrees of evolution (i.e., mafic, intermediate, acidic rocks); (3) the geochemical characteristics of volcanic rocks with the same magmatic affinity and comparable degrees of evolution. Other differences arise from the relative volumes of extruded vs. intruded magmas and from the style of volcanic eruptions.

The magmatism associated with E-directed subduction zones of Central and South America typically is calc-alkaline to high-K calc-alkaline. It often grades to shoshonitic compositions away from the trench, with significant increase in potassium and other large ions lithophile elements (LILE) (e.g.. Thorpe, 1982). Na-alkaline volcanoes are also present in some areas such as central Mexico (e.g., Robin, 1982; Luhr et al., 1989) and various sectors of the western margin of South America where they occur as scattered centers east of the calc-alkaline volcanic belt. A similar magmatological setting is also observed in the Pacific margin of North America (Cascadia subduction zone).

The magmatic rocks associated with W-directed subduction zones show a larger range of petrologic characters. In several places, such as Japan, Fiji, Aeolian arc, volcanic rocks, although still dominant calc-alkaline, range from island–arc tholeiites and, in some cases, boninites, to shoshonitic, and to Na- and K-alkaline compositions. Boninites and, to a lower degree, arc-tholeiites are very scarce or absent in the E-directed subduction zones. Tertiary boninites s.s. have been only found in Izu-Bonin, Marianas and other western Pacific island arcs (Crawford et al., 1989). On the other hand, shoshonites are generally less abundant in W-directed than in E-directed subduction settings. Na-alkaline rocks are also often associated with W-directed zones in several areas and generally crop out in a back-arc position with respect to arc tholeiites and calc-alkaline volcanics. Examples can be found in various places, such as Fiji (Gill, 1970), Grenada (Arculus, 1978), Aeolian archipelago (Cinque et al., 1988) and in the arc of Japan which represents a classical place where rocks ranging from tholeiitic to Na-alkaline composition crop out at various distances from the trench (Kuno, 1968; Aramaki and Ui, 1982). Undersaturated potassic alkaline rocks occur in Indonesia and in the Aeolian Islands. In the latter, leucite tephrites have been erupted from the same volcanoes as the associated calc-alkaline and shoshonitic products (Keller, 1982; Francalanci et al., 1989; De Astis et al., 1997).

The frequency distribution of various lithologies within the single magmatic associations is also different in the various settings. Along the western margins of North, Central and South America, the dominant rock types are invariably represented by calc-alkaline and high-K calc-alkaline andesites. Basalts generally occur in smaller amounts, whereas acidic dacites and rhyolites are rather abundant and, in some cases such as in Guatemala and Peru, form huge pyroclastic fall and ignimbrite deposits. Along the western Pacific island arcs, mafic calc-alkaline rocks are widespread, whereas dacites and rhyolites are absent or occur in much smaller amounts than in the E-directed subduction zones. In some cases such as in the Tonga–Kermadec islands and South Sandwich, basalts and basaltic andesites represent the only or largely dominant rock types (see Thorpe, 1982).

The geochemical differences between rock types with the same degree of evolution cropping out along the western Pacific island arcs and on the E-directed subduction of Andes, Cascadia and Central America zones are well known. This compositional variability led Jakes and White (1972) to distinguish 'andean-type' and 'island–arc type' calc-alkaline suites. In general terms, rocks erupted on E-directed subduction zones have higher abundances of incompatible elements, especially LILE Rb, Cs, Ba and LREE, than the equivalent rock types erupted in island arcs overlying W-directed subduction zones. Sr also shows higher concentrations in rocks from Andes and Cascadia than at Tonga, New Zealand, Antilles and South Sandwich (Gill, 1981). Other geochemical differences relate to Sr and Nd isotopic signatures that generally show lower ranges of values in W-directed subduction zones, and to REE that are less fractionated in western Pacific island arc rocks than in the andean counterparts. Higher concentrations of Ti, Zr, and P, and lower HREE and Y abundances have been pointed out by Palacio et al. (1983) for andesites from Andes with respect to the same rock types from western Pacific island arcs. However, there are exceptions to these patterns. For instance, the Aeolian arc volcanics have high abundances of LILE and Sr, strongly fractionated REE patterns and a rather large range of Sr isotopic ratios that make these rocks to resemble more closely the Andean than western Pacific island arcs volcanics (e.g., Ellam et al., 1988; Francalanci et al., 1989; Peccerillo and Wu, 1992; De Astis et al., 1997).

An additional significant variation of magmatism in subduction zones applies to the relative volumes of intrusive rocks with respect to extrusive counterparts. It is obvious that the amount of exposed intrusive rocks is heavily dependent on several factors such as the maturity of arcs and the degree of uplift and erosion in various areas, and may not represent a reliable parameter to estimate the role of extrusive vs. intrusive activity. However, a crude consideration of the global distribution of Tertiary and Quaternary volcanic and intrusive rocks along recent to active subduction-related settings point to a stronger role of intrusive over effusive activity in E-directed than in W-directed subduction zones. Although small intrusions are widespread in western Pacific island arcs and the Antilles, the largest concentrations of plutons are observed along the Pacific margins of America where they form an almost continuous belt that runs from Aleutians to Antarctica (e.g., Brown, 1982). On the other hand, the island arcs of the western Pacific are the site of very extensive volcanic activity. For instance, more than 400 Quaternary centers have been identified only in the north western arcs of Japan, Kurili, Izu-Bonin, and Marianas.

The style of eruption of magmas emplaced on W- and E-directed subduction is also worth mentioning. Although calc-alkaline volcanoes are almost invariably characterized by explosive eruptive behavior, those occurring along the Central and Southern America margins are apparently marked by stronger explosive activity, as indicated by the occurrence of large ignimbritic and pyroclastic fall deposits, e.g., in Guatemala and Peru (Thorpe, 1982).

A key problem is that of understanding if the differences of magmatic settings in E- and W-directed subduction zones are a first-order effect of the variable geometry of the undergoing lithospheric slab, or they relate to other parameters such as the nature (i.e., continental vs. oceanic) and thickness of the crust above mantle wedge. There is a consensus that some characteristics of E-directed subduction zones, (e.g., the abundance of rhyolitic vs. intermediate and mafic rocks, the high LILE contents and the variable Sr–Nd isotopic compositions of volcanic rocks, the large amounts of intrusive bodies) depend mainly on processes occurring within the thick continental crust, which is typical of this setting. As suggested by several authors (e.g., Hildreth and Moorbath, 1988; Worner et al., 1988) continental crust acts as a density filter with respect to mantle-derived magmas. Accordingly, mafic liquids are preferentially stored in large magma chambers where they freeze to form intrusive bodies. Evolutionary processes in these reservoirs produce huge amounts of acidic liquids which are eventually erupted to form ignimbritic sheets. Combined fractional crystallization plus crustal assimilation generate strong enrichments in LILE and large ranges of Sr and Nd isotopic signatures in the magmas. The different stress regimes operating in the E- and W-directed subduction zones represent an additional factor that plays a key role in determining the amounts of erupted vs. intruded magmas. Note, however, that stress regimes and thickness of the crust are primary effects of the geometry of subduction zones.

However, there are some petrologic characteristics of subduction-related magmas that are unlikely to be the effects of magma–crust interaction and may be directly related to the geometry of the undergoing slab. For instance, the formation of undersaturated potassic rocks closely associated in space and time with calc-alkaline volcanics, such as in the Aeolian arc in the southern Tyrrhenian sea, may be facilitated by the occurrence of a steep subduction zone. Petrologic and geochemical data strongly suggest that silica-undersaturated ultrapotassic magmas are generated by low degrees of partial melting of anomalous upper mantle at higher pressure than the oversaturated or saturated calc-alkaline and shoshonitic magmas (e.g., Wendlandt and Eggler, 1980). Geochemical data also indicate that arc magma sources have been metasomatized at various degrees by upper crustal material brought into the upper mantle by subducting crust. Shoshonitic and alkaline rocks appear to derive from sources that are more strongly metasomatized than those of arc tholeiites and calc-alkaline magmas. Accordingly, the close association of rocks with variable petrochemical affinity reflects melting at various depths in a vertically zoned upper mantle. The formation of this type of source may be strongly favored by steeply dipping subduction zones.

The occurrence of boninites may also be directly related to the geometry of the subduction zone. It has been suggested that boninitic magmas are formed in the mantle wedge at low pressure and by large degrees of partial melting. These conditions require both the presence of aqueous fluids, which depresses the peridotite solidus, and high temperatures in the uppermost mantle. According to experimental evidence, temperatures between 1100° and 1350°C (e.g., Crawford et al., 1989, and references therein) at pressure around 1 GPa are required to form various types of boninites. It is obvious that these thermal conditions are easily found above steeply dipping subduction zones, in which active convection brings hot material to shallow levels within the upper mantle, with consequent increase of temperatures in the upper part of the mantle wedge. These thermal conditions also favor the formation of arc tholeiites, which also form by large degrees of mantle melting.

In conclusion, W-directed and E-directed subduction zones show significant differences of their magmatological settings. These regard several petrologic, geochemical and field characteristics of magmatic rocks. Most of these differences do not depend directly from the particular geometry of subduction zones but are rather related to the variable tectonic regimes and to the thickness and composition of the overriding crust. Others peculiarities seem to be a consequence of the depth and structure of magma sources which may represent a first-order effect of the dip angle of subduction zones.

Metallogenesis appears also controlled by subduction style (Mitchell and Garson, 1981). Porphyry copper deposits are concentrated in collisional settings and Chilean type subduction zones. The Mariana type subduction is instead characterized by Kuroko or similar volcanogenic sulphide deposits (Nishiwaki and Uyeda, 1983).

15. The western Mediterranean examples

The western Mediterranean basins (Auzende et al., 1973; Biju-Duval et al., 1978; Panza and Mueller, 1979; Horvath and Berckhemer, 1982; Morelli, 1985; Stanley and Wezel, 1985; Robertson and Grasso, 1995) consist of a series of large scale lithospheric boudins which developed in the context of back-arc extension contemporaneous to the Neogene–Pleistocene eastward roll-back of the westerly directed Apennines–Maghrebides subduction zone (Gueguen et al., 1997). The Apennines subduction zone (Caputo et al., 1970, 1972; Scandone, 1980; Rehault et al., 1984, 1985; Malinverno and Ryan, 1986; Royden et al., 1987; Kastens et al., 1988; Patacca and Scandone, 1989; Doglioni, 1991), when considered as a whole with the northern Africa Maghrebides, has a length comparable with the arcs of the western Atlantic and western Pacific subduction zones (1500–3000 km). It is commonly believed that the extension determining its formation developed in a context of relative convergence between Africa and Europe. However, the direction of relative motion is still under debate. Most of the reconstructions (Dewey et al., 1989; Mazzoli and Helman, 1994; Albarello et al., 1995; Campan, 1995) show an amount of shortening of 150 km during the all Tertiary in the western Alboran area, the main difference between these models being the increase of convergence further east ranging from 300 (Dewey et al., 1989) to 250 km (Campan, 1995) in Tunisia. The amount of shortening in the two areas diminishes when computed for the last 20 Ma to 70–80 km in the Alboran and 100–165 km in Tunisia (Dewey et al., 1989; Campan, 1995). It appears that the amount of relative N–S Africa Europe relative motion was in any case five to eight times slower with respect to the eastward migration of the Apenninic arc which migrated eastward about 800 km during the last 23 Ma, i.e., 4–6 mm/year vs. 30–50 mm/year (Gueguen et al., 1997). Recent geodetic data confirm the main directions of Africa and Europe are NE-directed (Smith et al., 1994) as indicated in Fig. 1. Therefore, the Apenninic arc migrated eastward faster than the postulated N–S convergence of Africa relative to Europe. However, new geodetic data on the motion of Africa relative to Sicily indicate that there is active extension NE–SW oriented (Tonti et al., 1998). In other words, Africa is moving away southwestward relative to Europe. These data are considered surprising but they are in fact coherent with the geologic record of the Sicily channel which is in extensional regime since at least the Pliocene. This data could finally clarify this overstated assumption of the northern push of Africa relative to Europe.

In the north–western margin the Adriatic plate is overriding the European plate to form the Alps, whereas along its western margin it subducts westward to form the Apennines. The same lithosphere is undergoing an 'E-directed' subduction (Alps) and a 'W-directed' subduction (Apennines). Elevated topography and gravimetric profile more coupled with the topography characterise the Alps, whereas lower elevation, asymmetric gravimetric profile are typical for the Apennines (Fig. 10). Moreover, the Alps have wider outcrops of crystalline basement than the Apennines (Fig. 11). The Alps have a double vergence, whereas the Apennines single vergence. The Alps have two shallow foredeeps and the Apennines only one and deep. Another paradox is that the Alps have a crust thicker underneath the orogen (even more than 50 km, as usually occurs along similar belts), whereas the Apennines have a much thinner crust (17–35? km, Locardi and Nicolich, 1988; Nicolich, 1989; Scarascia et al., 1994). The two belts best appear to support that the differences between subduction zones are primarily linked to the geographic polarity of the subduction rather than to the differences of the involved lithosphere.

The Alps and the Apennines provide a geodynamic setting in which to compare structural differences between thrust belts and related foredeeps associated with E- vs. W-directed subduction zones (Fig. 11). Laubscher (1988) noted significant differences among the two orogens and defined the Alps as a push-arc and the Apennines as a pull-arc. The Alps and Apennines formed at the margins of the same plates: between the Adriatic plate to the east or southeast, and the European plate to the west or northwest. Note that the two orogens are diachronous and have very different rates of evolution. The Alpine subduction began during Early Cretaceous time and continued until the Pliocene. In contrast, the Apenninic subduction formed during the last 30 Ma. The Alps are more elevated with respect to the Apennines and have a much higher structural relief as also indicated by the outcrops of high grade metamorphic rocks. Erosion eliminated a large part of the uplifted thrust sheets, which would have reached some tens of kilometers of altitude if they had stayed in place. The Apennines, on the other hand, have extensive outcrops of sedimentary cover (Bally et al., 1986), and only a few scattered outcrops of metamorphic rocks, mainly relics of the earlier Alpine phase. In contrast with the Alps, the Apennines did not have a thick pile of a few tens of kilometers of nappes that were eroded. Moreover a back-arc basin, i.e., the Tyrrhenian sea, formed only to the west of the Apennines. Another unexpected difference between the Alps and the Apennines is that paradoxically the Alps have a shallow foredeep with low subsidence rates, in spite of the high topographic relief and consequent higher lithostatic load compared to the Apennines. In the Alpine foredeep (Pfiffner, 1986), the Oligocene base reaches 4000 m (Bigi et al., 1989). Subsidence rates in the Alpine foreland range from 0 to 200 m/Ma. The Southern Alps are the back-thrust belt of the Alpine orogen. The Southalpine foredeep (Massari, 1990) subsided at rates that rarely exceeded 300 m/Ma, determined by dividing the total thickness of the flysch and molasse deposits by duration of the deposition (2–6 km of sediment deposited in 15–40 Ma or more). The average area of the entire Alpine orogen above sea level is about 300 km^2, and the sum of the areas of the foredeeps of the frontal and back thrust belts is estimated as 150 km^2. Thus, the ratio of the area of the orogen to the area of the foredeeps is 2:1 in this case. It is regularly >1, as in thrust belts related to E–NE-directed subduction. It is from the Alps that the terms flysch and molasse were introduced, describing early and later stages of foredeep filling. Flysch and molasse are in general synonymous with deep and shallow clastic facies, respectively. The initial stages of subduction occurred when there was a deep-water environment collector for turbidites coming from the uplifting and eroding wedge. The foredeep was quickly filled, however, because the amount of sediment coming from the orogen was much larger than the area of the foredeep. As a consequence, the foredeep changed to molasse shallow-water conditions, and once the basin was filled, the foredeep was bypassed, and clastic sedimentation occurred in remote areas: for the Alps, in the Rhone and Po deltas, and in the North sea and the Black sea, by means of the Rhine and Danube rivers, respectively. This bypassing occurred because the subsidence rate in the foreland basins around the Alps was insufficient to keep the sediments coming from the orogen. This low subsidence rate did not occur in thrust belts related to W-directed subduction zones, where the subsidence rate was always greater relative to the amount of volume eroded from the internal elevated ridge. The Apennines are characterized by a frontal active accretionary wedge below sea level, whereas the main elevated ridge is the result of uplift and extension. These different tectonic fields are still moving eastward, expanding the Apenninic arc at velocities of 1 to 7 cm/year, rates comparable to those of other arcs related to W-directed subduction zones (e.g., the Banda arc, Veevers et al., 1978). The eastward roll-back of the Adriatic lithosphere accompanies this migration. In the elevated Apenninic ridge, what was previously accreted in the frontal part (mainly Mesozoic cover and deep foredeep deposits) has been uplifted and crosscut by the eastward-propagating extensional wave. Seismic data from the accretionary wedge of the Apennines show that the envelope around the folds crest may dip toward the hinterland (Doglioni and Prosser, 1997). As a result, growth folds are little eroded, and clastic sedimentary rocks of the foredeep onlap the limbs of these folds. In the Alps, on the other hand, the envelope of the fold crest rises toward the hinterland. The front of the thrust belt rises and growth folds are uplifted and deeply eroded during the forward propagation of the orogen, an evolution opposite to that of the folds of a W-directed, subduction-related accretionary wedge. In contrast to the Alps, the Apennines have a very pronounced foredeep; the Pliocene base reaches 8.5 km, indicating subsidence rates of 800–1600 m/Ma. Much of the Apenninic foredeep is located on top of the accretionary wedge, not to its front. Thus, the so-called piggyback basin is often the foredeep for the Apennines. Clastic material in the Apennines is provided not only by their ridge but also by the Alps and Dinarides surrounding the Adriatic plate. The average area of the elevated ridge of the Apennines above sea level, from the water divider eastward, is 40 km^2; the area of the foredeep is 180 km^2, so the ratio is 0.22:1. This ratio is always < 1, as in all W-directed subduction settings.

On the basis of the aforementioned differences, Alps and Apennines may be considered as two different end-members of thrust belts. However, the Apennines developed east of the southward continuation of the present Alps. In fact in the hangingwall of the Apennines subduction there are boudinated remnants of the alpine belt stretched by the extension of the Apennines back arc (Fig. 14). Therefore, within the western internal parts of the Apennines there should be the record of the alpine evolution with alpine rocks of both the frontal thrust belt (e.g., Corsica) and the back-thrust belt (e.g., Cervarola Front?, Doglioni et al., 1998).

15.1. Magmatism in the Alps and Apennines

There are striking differences between the magmatism associated with the Alpine orogen and that occurred during the formation of Apennines (Fig. 15).

The Alpine orogenic magmatism is essentially represented by intrusive and hypoabissal rocks, with minor amounts of volcanic products. It was formed during two main stages of activity (Dal Piaz and Venturelli, 1983). A first stage is basically Upper Cretaceous–Lower Eocene in age and consists of andesitic volcanics which were emplaced on the African continental margin. Evidence of this volcanism is only recorded by andesitic s.l. clasts occurring in some flysch and arenaceous sediments (Taveyanne, Schlirenflysch). A second stage is essentially Oligocene in age and mainly consists of granitoid bodies, several dikes and a few, poorly preserved volcanites. This magmatism which is known as the 'periadriatic igneous belt', marks a post-collisional extensional tectonic phase of the Alpine orogen. The plutonic rocks are mainly acid to intermediate in composition with a very few basic products. Intermediate compositions dominate among dikes. Petrochemical affinity is mainly calc-alkaline or high-K calc-alkaline with some shoshonites and few ultrapotassic lamproitic rocks (Beccaluva et al., 1979; Bellieni et al., 1981; Venturelli et al., 1984; Beccaluva et al., 1989; Bellieni et al., 1991). A few dikes of arc tholeiites are also present (Beccaluva et al., 1983). In its eastern sector, the periadriatic belt is represented by granodioritic–tonalitic plutons (Vedrette di Ries, Rensen, Monte Alto, etc., Bellieni et al., 1981, 1991) and by several calc-alkaline to shoshonitic dikes. In the central sector, a few plutons

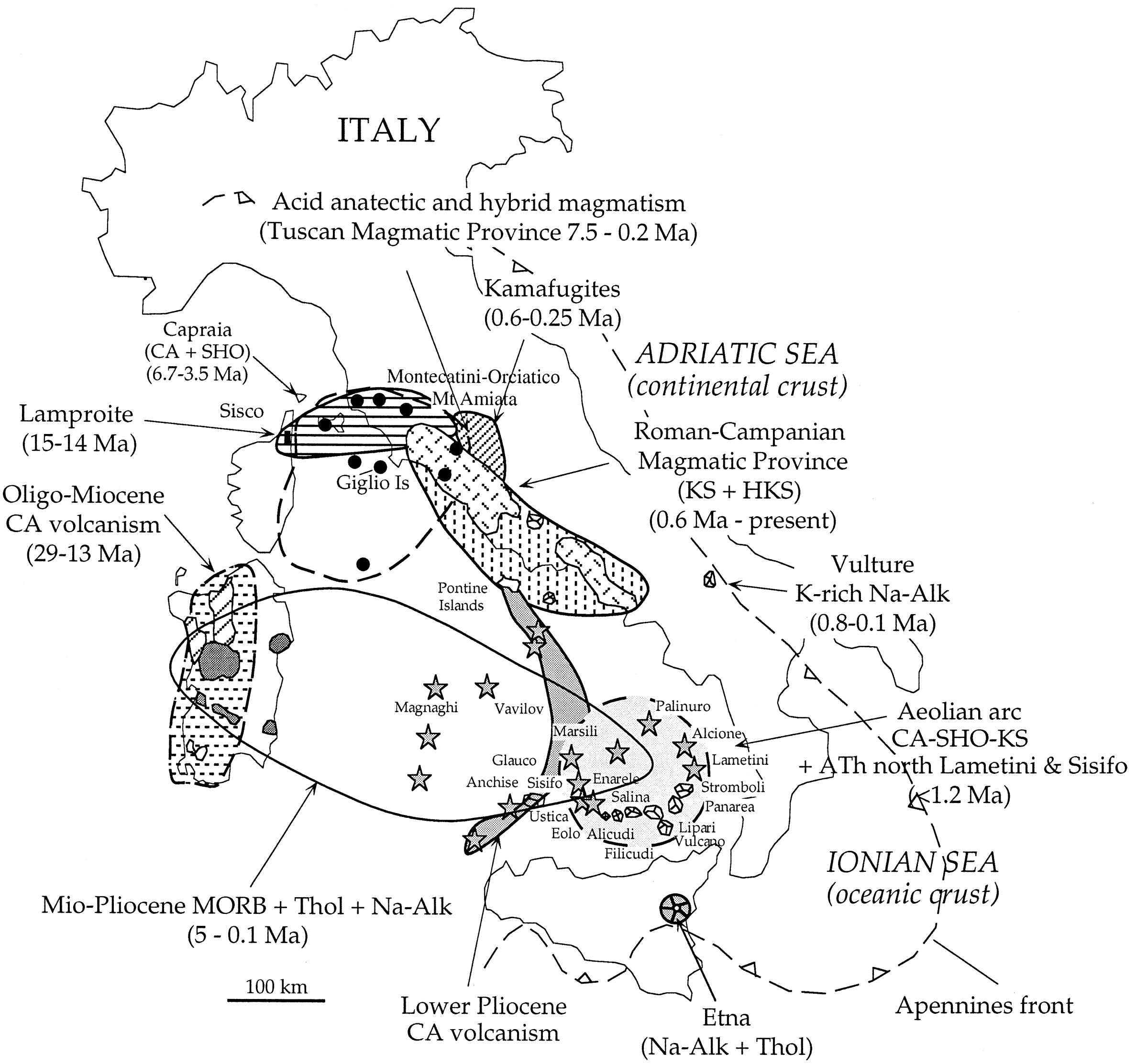


Fig. 15. Schematic distribution of the Tertiary–Quaternary and active volcanism in the Tyrrhenian sea area and Apennines (modified after Locardi, 1991). ATh, arc tholeiitic volcanism; CA, calc-alkaline; SHO, shoshonitic; KS, potassic; HKS, ultrapotassic; MORB, mid-ocean ridge basalts; Thol, tholeiitic; Na-Alk, Na-alkaline; Stars, volcanic seamounts. This complicate setting could be explained with the inherited Alpine orogen in the hangingwall of the Apennines subduction (Peccerillo, 1999) and the different composition of the down-going lithosphere (continental in the Adriatic sea and oceanic in the Ionian sea).

(Bregaglia, Adamello) and a large number of dikes crop out. Compositions range from arc-tholeiitic to calc-alkaline and shoshonitic. Intermediate calc-alkaline and high-K calc-alkaline rocks represent the dominant lithologies (Beccaluva et al., 1983). Finally, the western sector contains few small intrusive bodies, some volcanites and dikes with high-K calc-alkaline and shoshonitic intermediate composition. Ultrapotassic lamproitic dikes are also present in this zone (Venturelli et al., 1984).

Overall, the magmatism associated with the Alpine orogen is represented by dominant plutonic and hypoabissal rocks with variable petrochemical affinity but with clear subduction-related geochemical and petrologic signatures. Volcanic rocks are scarce or absent. Other effusive rocks with transitional to Na-alkaline affinity crop out south of the eastern Alps (Colli Euganei), but their relation with the Alpine orogen is still matter of debate.

The magmatism associated with the Apennines orogen is much more abundant and petrologically complex. Magmatic rocks are spread over a very wide area, from south–eastern France, Sardinia and Corsica, to the Tyrrhenian basin and to central–southern Italy. These rocks are essentially extrusive, whereas intrusive activity is only represented by a few granite–granodiorite bodies (8.5–5 Ma) occurring in southern Tuscany and Tuscan Archipelago (Poli et al., 1989).

Volcanic activity has given MORBs, island arc tholeiites, calc-alkaline, high-K calc-alkaline and shoshonitic products, Na- and K-alkaline rocks. This large spectrum of rock composition makes the circum-Tyrrhenian area one of the most complex magmatic settings in the world. The petrologic and geochemical characteristics of this magmatism have been recently discussed in a number of papers (Peccerillo and Manetti, 1985; Serri, 1990; Conticelli and Peccerillo, 1992) and this review will closely follow these authors.

MORB-type rocks form part of the Tyrrhenian sea-floor and have been recovered by deep sea drilling (see Serri, 1990 and references therein). Their genesis has been related to mantle melting during the opening of the Tyrrhenian back-arc basin. Other rocks with tholeiitic affinity are associated with prevailing Na-alkaline Plio–Quaternary volcanics in Sardinia and represent the lowest exposed products of Mt. Etna.

Plio–Quaternary Na-alkaline rocks occur in Sardinia, Ustica in the southern Tyrrhenian sea, Mt. Etna and in some seamounts arising from the Tyrrhenian sea-floor (e.g., Vavilov, Magnaghi, Locardi, 1991).

Subduction-related volcanism is widespread in the area. The oldest activity occurs in Sardinia (29–13 Ma, Dostal et al., 1982 and references therein) and follows the calc-alkaline volcanism of Provence, in south–eastern France, which dates back to 50–35 Ma (Bellon, 1981). Orogenic rocks in Sardinia range from calc-alkaline basalts to rhyolites, with a dominance of intermediate and mafic rocks (Dostal et al., 1982).

Younger orogenic volcanism progressively shifted eastward up to its present position in the Aeolian arc. This consists of seven main islands and several seamounts (Palinuro, Alcione, Lametini, Enarete, Eolo, Sisifo, Glauco) which define an overall ring structure. Rocks are younger than about 1.2 Ma and mainly consist of intermediate and mafic calc-alkaline to shoshonitic volcanics, with minor leucite tephrites (at Vulcano and Stromboli) and arc tholeiites (at Lametini and Sisifo seamounts). Acidic rocks are concentrated in the last 40 ka. Slightly older calc-alkaline to shoshonitic volcanics form a NE–SW elongated Lower Pliocene volcanic belt (e.g., Anchise seamounts) east of the Aeolian arc (Locardi, 1991). High-K calc-alkaline and shoshonitic volcanics also form the Island of Capraia (6.7–3.5 Ma) in the Tuscan archipelago. The lamproites of Sisco (Corsica) have ages of about 14–15 Ma (Civetta et al., 1978).

Potassic and ultrapotassic magmatism of central and southern Italy (the so-called Roman Comagmatic Province) represents the most striking magmatological feature of the Tyrrhenian area. It spans a time interval between 4 Ma and the present time, with the bulk of potassic volcanics being younger than 0.6 Ma. Potassic and ultrapotassic rocks have variable petrochemical affinity, from Roman type potassic series (KS), Roman type high-potassium series (HKS), lamproites and kamafugites (Peccerillo and Manetti, 1985). KS and HKS rocks represent by far the dominant volcanics, whereas lamproites are restricted to southern Tuscany and Corsica, and kamafugites form a few scattered centers east of the Roman province. KS are basically similar to shoshonites, whereas HKS are undersaturated in silica and show very strong enrichment in potassium and incompatible elements. However, both KS and HKS display low contents of high field strength elements (HFSE) as typically observed in subduction related rocks. Kamafugites show similar incompatible element patterns as HKS, but are more strongly undersaturated in silica and show higher CaO and lower Na and Al than HKS. Finally, lamproites are oversaturated in silica, contain negative anomalies of HFSE and have isotopic signatures that are closer to crustal than to mantle values. However, their high values of Mg# (up to 80), Ni (up to 350 ppm), and Cr (up to 800 ppm) together with the presence of high pressure ultramafic xenoliths strongly suggest a mantle origin. Carbonatitic rocks which have been recently suggested to occur east of the Roman province (Stoppa and Woolley, 1997) may actually represent silicate, kamafugitic or HKS magmas that have undergone strong contamination by the thick carbonate sequences crossed by uprising magmas (Peccerillo, 1998). Overall, the whole ultrapotassic magmatism of central–southern Italy appears to be subduction related and has been hypothesized to derive from an anomalous upper mantle whose composition has been strongly modified by addition of upper crustal material by subduction processes (e.g., Peccerillo, 1985).

In conclusion, a large spectrum of magma types is associated with the Apennines orogen and the opening of the Tyrrhenian sea. Most of these rocks (i.e., island arc tholeiites, calc-alkaline, shoshonitic, potassic and ultrapotassic rocks) are of obvious subduction-related origin. However, also some MORB and Na-alkaline rocks (including Ustica, Etna and Vulture) show evidence for a role of subduction-related geochemical components in their genesis (e.g., Serri, 1990).

16. Conclusions

Differences in subduction styles have been explained as due to variations in convergence velocity, plate thickness and age (e.g., Royden, 1993). However, there are cases where the same plate is subducting with a different style (W-directed or E- or NE-directed) and angle and relative geologic signature depending only on the orientation. One example is the Adriatic microplate (Fig. 3). This plate is sinking toward the west almost vertically beneath the Apenninic arc (Selvaggi and Chiarabba, 1995) whereas on the east it is sinking at a low angle beneath the Dinarides and Hellenides (Papazachos and Comninakis, 1977; Christova and Nikolova, 1993; Piromallo and Morelli, 1997). The same plate determines orogens that fall into the W-class and E–NE-class independently from the nature and age of the down-going lithosphere. For a geodynamic discussion about the E-class nature of the Hellenic–Aegean system, see Doglioni (1995). Topography, gravimetry, structure and all the other geological and geophysical parameters of the two subductions fall respectively into the W-class and E-class. This shows that the nature and age of the down-going lithosphere is not the primary factor in determining the characteristics of the W- and E–NE-classes. Another example is the Kermadec–Macquarie subduction; to the north the Pacific plate subducts westward at a high angle, with low elevation of the hangingwall plate and a deep trench. To the south along the opposite NE-directed New Zealand subduction zone the slab has a low angle, there is high elevation of the hangingwall plate and the trenches are shallower, despite the upper plate is either continental or oceanic. However, in this case, the New Zealand–Macquarie subduction has the Tasmanian sea oceanic lithosphere in the footwall which is younger than the Pacific lithosphere of the Kermadec subduction, but still all the parameters fall into the W-class for Kermadec and E–NE-class for New Zealand subduction zones. On the other hand, along the Sandwich subduction zone the undergoing Atlantic and Antartic oceanic lithospheres show age variations (5–120 ma), but the subduction system maintains the characteristics of the W-class. These examples show that the geographic polarity of the subduction more than any other parameter constrains the different characters observed in the two classes. This poses the question whether subductions can be ascribed only to slab pull or whether they are also influenced by the relative westward drift of the lithosphere with respect to the upper mantle postulated by several authors (Le Pichon, 1968; O'Connell et al., 1991; Ricard et al., 1991). Polarization of the seismic waves in the mantle far away from the subduction zones, e.g., beneath the Nazca plate (Russo and Silver, 1994) and below the Tyrrhenian back-arc basin (Margheriti et al., 1996) could provide evidences for a relative E-ward mantle flow. The mantle polarization would be able to differentiate the opposite behavior of the decollement planes along the W-directed and E- or NE-directed subduction zones and to determine the differences on the orogenic belts (Doglioni, 1992) which may be analyzed also in terms of the ratio between convergence rate and retreat rate of the subduction hinge (Waschbusch and Beaumont, 1996).

One could argue that the two classes of subduction zones are simply sensible to the thickness and composition of the hangingwall and footwall plates. However, we observe that the two classes persist independently from the age and nature of the involved lithospheres, and that they are strictly constrained by the geographic polarity (Harabaglia and Doglioni, 1998).

In summary, W-directed subductions are zones where there is a negative volume balance of lithosphere, in other words, the lithosphere is almost entirely lost in subduction and replaced by the uprising asthenosphere in the back-arc region. Along E–NE-directed subduction zones, the volume balance of the lithosphere is more positive because the hangingwall lithosphere is thickened from the footwall plate which is sliding below and following the shape of the upper plate. This could provide an explanation for their higher structural and morphologic elevation (Fig. 12).

Acknowledgements

The paper benefited from critical reviews by L. Jolivet and another referee. Many thanks to Giuliano Panza for stimulating and improving this paper, and to Albert Bally, Daniel Bernoulli, Enrico Bonatti, Manuel Fernandez, Domenico Giardini, Erwan Gueguen, Andrea Morelli, Giampaolo Pialli, Barbara Romanowicz, Roberto Sabadini, Francesc Sabat and Susanna Zerbini for helpful discussions. The European Community, the Italian CNR and ASI supported this study (HCM Research Networks, 'Geodynamic modelling of western Mediterranean'), grants ERBCHRXCT940607, 96.00279.CT05, 94-RS-61.

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Carlo Doglioni since 1997 is full professor of geology at the University La Sapienza of Rome, Italy. He was formerly at the Universities of Basilicata, Bari and Ferrara. He visited as researcher the Universities of Basel, Oxford and Rice University of Houston. He works mainly in geodynamics and field geology of the Mediterranean area. AAPG distinguished lecturer.

Paolo Harabaglia is researcher of geophysics at the University of Basilicata in Potenza, Italy. He graduated in geology at the University of Trieste and he is Master in geophysics at MIT, Boston, USA.

Saverio Merlini graduated in Geology in 1983 from Milan University (Italy), with a thesis of the Central Alps. He works in San Donato Milanese for ENI/Agip Division since 1984 as regional geologist, studying the structure of Italy by field geology, well data, seismic interpretation and geodynamics. He is Technical Leader of Exploration New Ventures in Italy.

Francesco Mongelli is full professor of geophysics at the University of Bari (Italy). His researches focus on geothermics, geodynamics and on the flexure of the lithosphere. He is member of the editorial board of Geothermics.

Angelo Peccerillo since 1997 is full professor of Petrology at the University of Perugia. He was formerly associate professor of Volcanology at the University of Florence, and full professor of Petrology at the Universities of Messina and Cosenza. His research activity has been dealing with the petrology and geochemistry of igneous rocks, with particular emphasis on arc related magmas and ultrapotassic rocks. He is author or co-author of about 100 scientific publications mostly published on peer-reviewed international journals, and of two books.

Claudia Piromallo, graduated in Physics in 1993 at the University of Bologna (Italy), with a specialization in Geophysics, working on numerical modeling of large scale Earth's dynamics. She is presently working at the Istituto Nazionale di Geofisica (Rome, Italy) where her main research activities relate to Seismology, with particular reference to P-wave seismic tomography of the lithosphere and upper mantle below the Euro-Mediterranean area, using both regional and teleseismic data, and event location methods in heterogeneous media.