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SUBDUCTION TECTONICS AND EXHUMATION OF HIGH-PRESSURE METAMORPHIC ROCKS IN THE MEDITERRANEAN OROGENS

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ABSTRACT. Using the peri-Mediterranean blueschist belts as a case study we discuss the mechanisms of syn-orogenic exhumation of high-pressure (HP) and ultra-high-pressure (UHP) rocks. The Mediterranean examples, within an overall convergent zone, show variations in the rates of convergence, rates of slab retreat, available space, and various stages of maturation of accretionary complexes. After a compilation of the deformation history and the kinematic boundary conditions and their evolution through time, we discuss P-T-t paths. Most of the structures found in the field today relate to the exhumation stage and often to quite superficial events related to syn-orogenic detachments. However a significant vertical motion occurred from the depth of eclogites (or UHP eclogites) to the depth of the blueschist or greenschist facies along cold P-T paths. Two types of mountain belts are described: those where a single thermal gradient was recorded throughout its' history, like the Franco-Italian Alps, which suggest a steady-state evolution; and those where the thermal gradient has changed through time (and space), such as the Aegean region, which suggest a non-steady-state evolution. Slab retreat within a subduction complex does not lead to the exhumation of UHP rocks because the open subduction channel allows for fast circulation and detachment of sediments from the subducting basement. Subducted sediments lubricate the subduction channel and the basement is thus not involved in the return flow. Early exhumation is fast and the thermal regime in the subduction channel is partly controlled by the velocity of convergence and the velocity of slab retreat as well as by the nature of the subducted material. Final exhumation occurs within the accretionary complex at a much slower rate below extensional detachments. The removal of the overburden is achieved primarily by extension in the upper part of the accretionary complex. Extensional faults and shear zones are rooted in the brittle-ductile transition of the accretionary complex. Deeper "extensional" shear zones result from shearing along the base of the upper plate. We discuss a model with several levels of circulation of subducted material and compare it with available thermo-mechanical models.

INTRODUCTION

The discovery of ultra-high-pressure metamorphic minerals such as coesite (Chopin, 1984; Smith, 1984; Caby, 1994; Wain, 1997; Ernst and Liou, 2000; Katayama and others, 2000; Liou, and others, 2000) has forced us to reconsider our views on mountain belts and continental subduction. The goal of this paper is to review the first order constraints on exhumation coming from observations in Mediterranean blueschist belts and to discuss a model based on the various natural settings offered by the Mediterranean region. The Mediterranean region allows comparison between various types of subduction zones, some with fast slab retreat and others in steady-state (Le Pichon, 1982; Malinverno and Ryan, 1986; Morley, 1993). Arcuate orogenic belts, partly disrupted by Neogene backarc extension, border the Mediterranean basins. Fragments of these belts drifted apart hundreds of kilometers (Alvarez and others, 1974) during the formation of backarc basins.

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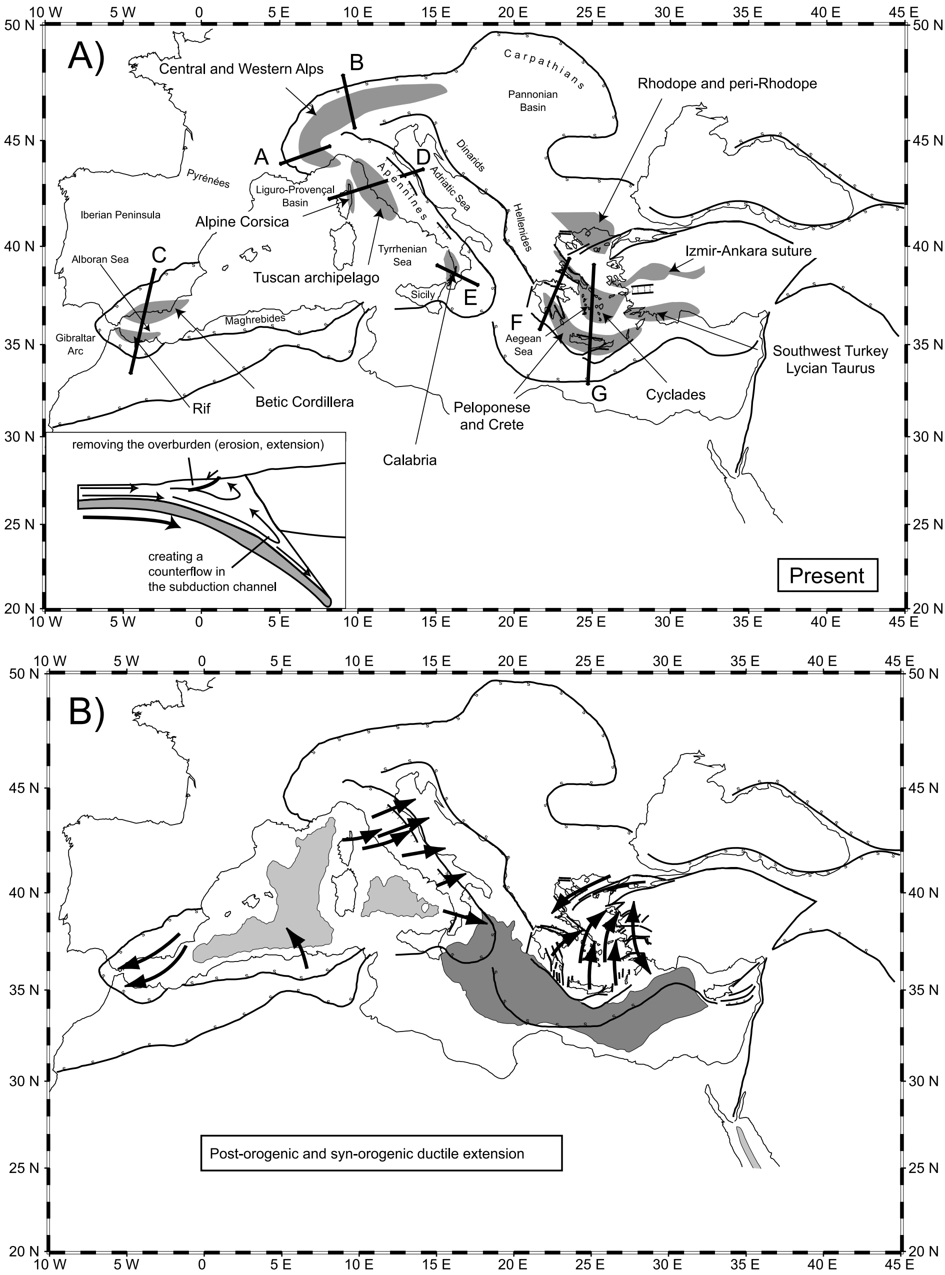


Fig. 1. (A) Location of the main blueschist belts in the Mediterranean region. Insert: Schematic sketch of a subduction complex with the two basic questions to be answered, overburden removal and counterflow in the subduction channel. (B) Directions and sense of shear of syn- and post orogenic extension related to backarc basin formation. Data from (Frizon de Lamotte and others, 1991, 1995; Gautier and Brun, 1994a; Jolivet and others, 1994b, 1996, 1998a; Vissers and others, 1995).

Understanding the mechanisms of exhumation of high-pressure metamorphic rocks requires answering two basic questions (fig. 1): (1) how is it possible to create counter flow in the subduction channel to get rocks moving upward, and (2) is the overburden removed either by erosion or extension? To answer these questions, we try to investigate which parameters control the temperature gradient in the subduction complex, or when, where, and under what mechanisms blueschist rocks return to the surface, and how the overburden is removed. The formation of blueschists and eclogites seems to occur most often during subduction of an oceanic plate or during the first stages of collision of two continental plates. This process has been recognized in recent and old chains, even in some proven Precambrian examples (Liou and others, 1990). The major reason for working with recent orogens such as the Mediterranean is that the kinematic boundary conditions are known with some precision. Although various possible plate configurations are still under discussion (Dercourt and others, 1986, 1993; Dewey and others, 1989; Stampfli, 2000), kinematic reconstructions have proven compatible with the present-day 3D structure of the mantle deduced from seismic tomography (de Jonge and others, 1993; Wortel and Spakman, 2000).

The internal zones of Alpine belts have large outcrops of metamorphic rocks that followed a variety of exhumation paths. Blueschists and eclogites were exhumed along cold P-T gradients that allowed for the preservation of HP-LT parageneses, whereas high temperature gneiss exhumed along warmer gradients often have only relics of HP-LT parageneses. A simple hypothesis would consider well-preserved blueschist and eclogites as characteristics of a cold syn-orogenic exhumation mode (by syn-orogenic we mean within the subduction complex whatever its geometry and internal dynamics, as opposed to post-orogenic which relates to collapse tectonics). HT rocks would be exhumed after subduction had ceased or after the subduction front had migrated to a different position allowing for thermal relaxation. Although this view is valid in some cases, the reality is in general more complex because the thermal regime of a subduction complex depends upon various factors, such as the velocity of convergence, the heat flow from the mantle, the density structure of the crust, the amount of heat generated inside the accretionary wedge, fluid behavior and the scale of fluid circulation.

Published exhumation models consider (1) forces mainly due to geometry and the kinematics of forced advection in the subduction channel (corner flow) (Shreve and Cloos, 1986; Cloos and Shreve, 1988; Polino and others, 1990; Allemand and Lardeaux, 1997), (2) the increase of buoyancy forces during burial (Chemenda and others, 1995), (3) the dynamic of the orogenic wedge itself (Platt, 1986; Thompson and others, 1997) and (4) coupled thermal and mechanical evolutions (Burov and others, 2001). The exhumation of dense pods of eclogite can be envisaged inside a less dense and low viscosity matrix made of metasediments or serpentinite (Ernst, 1971; Maekawa and others, 1995; Guillot and others, 2000, 2001; Gerya and others, 2002). The removal of the overburden is either due to erosion (Brandon and others, 1998; Ring and others, 1999) or to extension (Platt, 1986, 1993) or both (Johnson and others, 1997). Most authors consider metamorphic pressures as good indications of the maximum burial; other authors suggest that a significant part of the total pressure could be due to tectonic overpressures (Mancktelow, 1995; Petrini and Podladchikov, 2000).

Most models of accretionary complexes use simple kinematics of burial along the subduction plane and vertical exhumation in the internal part of the wedge without mechanical considerations. These simple models seem to explain quite well the exhumation P-T paths of blueschists in the case of subduction of oceanic crust or warmer paths such as those of the Lepontine dome in the case of continental accretion (Shi and Wang, 1987; Bousquet and others, 1997; Goffé and others, 2003). Giunchi and Ricard (1999) in addition modeled orogenic wedges by considering the thermomechanical evolution and the thermodynamics of phase changes, but this model does not consider the formation and exhumation of ultra-high-pressure rocks, being limited to the crust.

Some estimates of erosion rates in the Cascades or Taiwan seem to indicate that erosion can be the main agent of exhumation (Davis and others, 1983; Brandon and others, 1998). The case of the Aegean where unmetamorphosed units still rest on top of the accretionary complex shows that erosion in some cases plays only a minor role and that extension is the major mechanism for removing the overburden. However the mechanics of this extension is not well understood. Wilett (1999) showed that extension parallel to the direction of convergence is feasible in some particular configurations of a viscous and weak accretionary prism and that the mechanisms depend strongly upon the chosen rheology for the model. Post-orogenic extension is equally important. It can be due to a real post-orogenic collapse after convergence has ceased, to convective removal during the formation of the mountain belt, or to slab retreat (Malinverno and Ryan, 1986; Dewey, 1988; Platt, 1993). Extensional detachments above high-pressure metamorphic complexes can form either during the contractional stage (Platt, 1986), hereafter called syn-orogenic detachments, or during post-orogenic stages (Wernicke, 1981; Lister and others, 1984), hereafter called post-orogenic detachments. In the Aegean Sea post-orogenic detachments may earlier have been syn-orogenic detachments (Parra and others, 2001; Trotet and others, 2001a).

Chemenda's model (Chemenda and others, 1995, 1996) is based on an analogue experiment where the lithospheric material is simulated by mixture of paraffin wax with a plastic rheology. Buoyancy forces and erosion represent the main engine for the extrusion of a rigid slice of crustal material. A piece of continental crust is carried along the subduction plane until the low density crust cannot resist the stresses created by Archimede forces and it is extruded upward between a new frontal thrust and a normal fault above, the overburden being eroded. Buoyancy forces are also the main agent for exhumation but the deformation is more distributed in the model of Burov and others (2001), which uses a numerical approach with coupled thermal and mechanical evolution.

Mancktelow (1995) has proposed a model of subduction and exhumation of sediments with a fixed geometry of the subduction channel between a rigid subducting plate and a rigid upper plate. The reduction of the channel width produces significant tectonic overpressures that can be as large as the lithostatic pressure. Such a model suggests that metamorphic pressure does not result from burial. Yet as shown by Burov and others (2001), for more realistic rheological parameters the subduction channel adapts its width throughout the convergence and no significant overpressures can build up because the strength of the rocks in the channel walls is too low to preserve a constant shape. In some particular configurations such as lithospheric buckling, significant tectonic overpressure might exist (Petrini and Podladchikov, 2000), but this case is not considered here. Furthermore, Schreyer (1995) argues against tectonic overpressures noticing that the pyrope-coesite paragenesis of the Dora Maira massif developed at the expense of a phyllosilicate-rich matrix that could not stand significant overpressures. These observations are in good agreement with the earlier observation by Etheridge (1983) that the formation of synmetamorphic tensile fractures imposes very low differential stresses in metamorphic rocks.

LARGE-SCALE STRUCTURE OF THE MEDITERRANEAN ALPINE BELTS, DEFORMATION HISTORY

The Mediterranean alpine belts extend from the Betic Cordillera and the Rif in the west to the Lycian Taurus in the east. The Rif and the Betic cordilleras are part of the Gibraltar arc that surrounds the Alboran Sea. The Betic Cordillera and the Rif formed by collision between the southern margin of the Iberian Peninsula and the northern margin of Africa or other continental blocks such as the hypothetical

"Alkapeca" landmass (Bouillin and others, 1986). The Alboran Sea formed by collapse of the Betic-Rif orogen. The Pyrénées formed by contraction between the Iberian block and the main body of Eurasia but they do not contain any HP-LT metamorphic rocks. The Alps-Carpathians belt corresponds to the underthrusting of the European crust below the Apulian block. The internal parts of the Carpathians have collapsed to form the Pannonian basin. The Apennines is partly a southern extension of the Alps in its internal zones (Tuscany) as well as a more recent accretionary complex formed at the expense of the Apulian plate. The internal zones have collapsed and extension has given birth to the Tyrrhenian Sea. Remnants of the Alps-Apennines can be found in Corsica and the Tuscan Archipelago. The southern Apennines connects with the Maghrebides through Calabria and Sicily. The internal parts of this southern belt have also been affected by post-orogenic extension. The Dinarides, Hellenides and Taurides correspond to the closure of the Tethyan Ocean and the subsequent underthrusting of Apulia below Eurasia. The eastern part of the belt has collapsed to form the Aegean Sea.

Almost all HP metamorphic complexes in the Mediterranean region are sandwiched between a basal thrust system and one or several extensional shear zones above (figs. 1 and 2). J.P. Platt (1986) first noticed this fact in the Western Alps, the Betic Cordillera, and the California Coast Range. A similar situation is found in most HP belts over the world (Maruyama and others, 1996). Although pressures recorded by these rocks were strongly underestimated in Platt's paper, the main picture still holds although his interpretation of the Coast Range has been challenged based on the analysis of kinematic data (Ring and Brandon, 1994).

Abnormal stratigraphic superpositions have long been interpreted as nappe structures that remain the major characteristic of orogenic wedges (Bertrand, 1884, 1887; Argand, 1916). However, after the work of Wernicke (1981) and Platt (1986), many studies using large-scale mapping and small-scale deformation analysis were devoted to the characterization of extensional structures. One quite *a posteriori* disappointing observation is that most of the deformation seen today in the internal zones is related to exhumation and often to extension (or, at least, thinning of the nappe pile), and that little is left from the prograde deformation stage other than the abnormal, old above young, superpositions. In the following we first review the criteria used here to define the nature of the tectonic contact, then review the arguments in favor of extension and finally discuss structure remaining from the earlier shortening stage. Figure 2 shows the main extensional structures (syn- and post-orogenic) in the peri-Mediterranean mountain belts discussed in the following section.

First it is necessary to define syn-orogenic extension. Although the syn-orogenic detachments have all the characteristics of normal faults (Platt, 1986), they work during the contractional episode and do not lead to crustal thinning. They only accommodate the exhumation of deep units and/or prevent too much crustal thickening, or result from gravitational instabilities in the upper part of the prism. It is however convenient in the field to consider them as extensional structures. Furthermore some regional examples show a continuum from syn-orogenic extension near the front of subduction to post-orogenic extension in the backarc domain [Aegean or Northern Tyrrhenian for instance (Jolivet and others, 1998a; Jolivet and Patriat, 1999)].

The Alps

In the Alps (figs. 2 and 3) HP metamorphism occurred before and during collision (Hunziker and others, 1989). The basal contact of HP metamorphic units (Penninic front) brings the HP Briançonnais (or Penninic) nappes, on top of the more external and little metamorphosed Dauphinois (or Helvetic) domain (Roure and

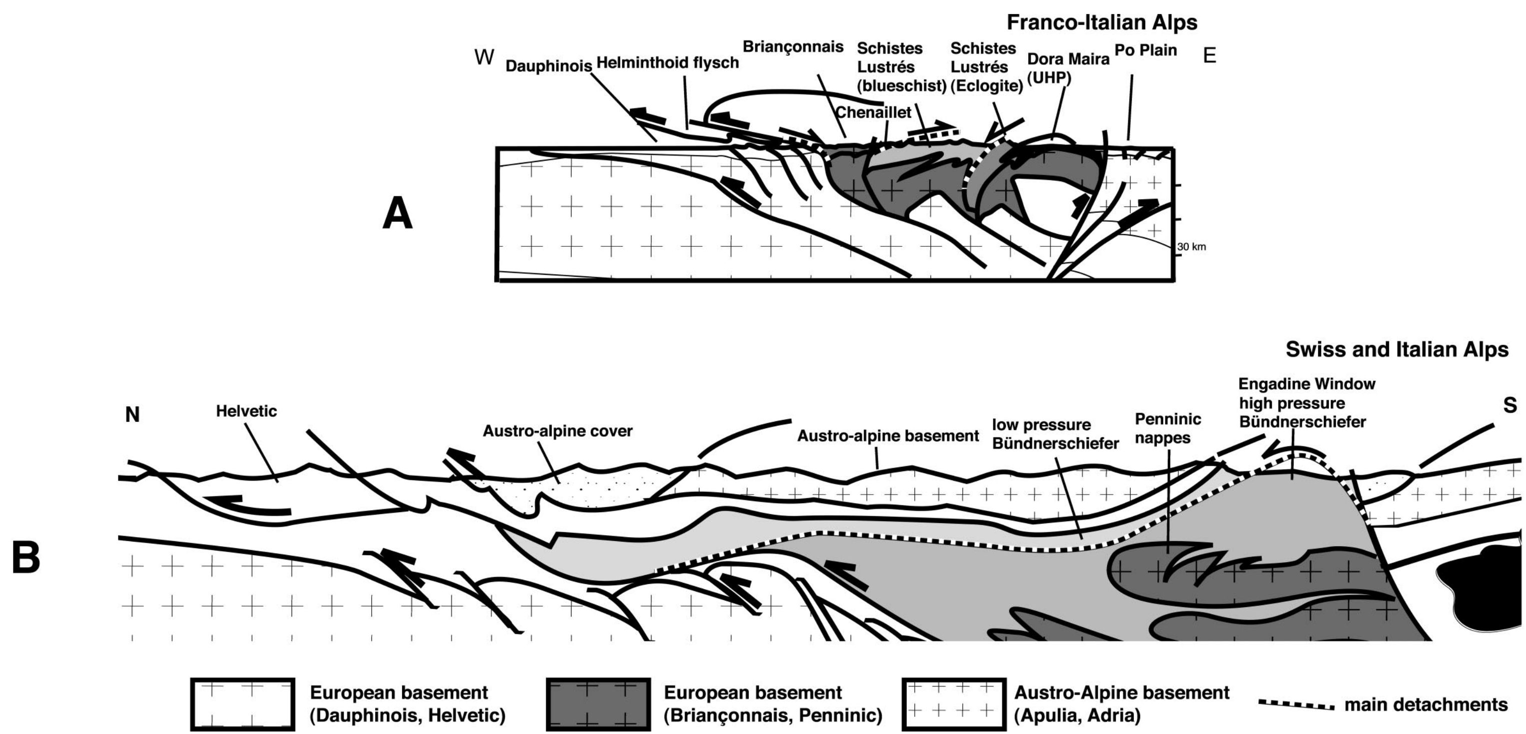
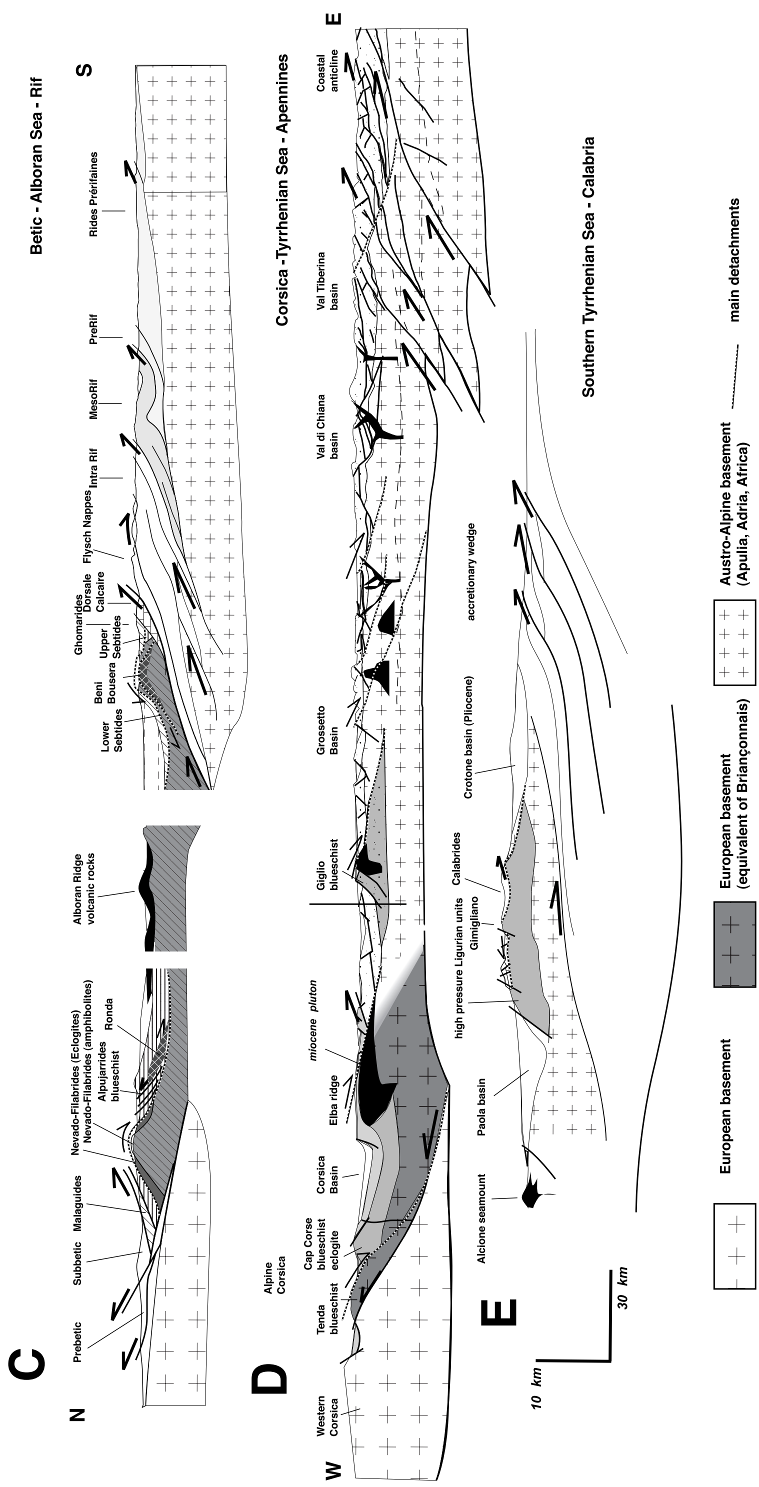


Fig. 2. Crustal-scale cross-sections of the transects discussed in the text adapted from (Trümpy, 1972; Ring and others, 1990; Vissers and others, 1995; Jolivet and others, 1998a; Boncio and others, 2000; Schwartz, 2000; Chalouan and others, 2001). In this figure and all the following figures detachments are shown as dotted lines whereas solid lines represent thrusts.



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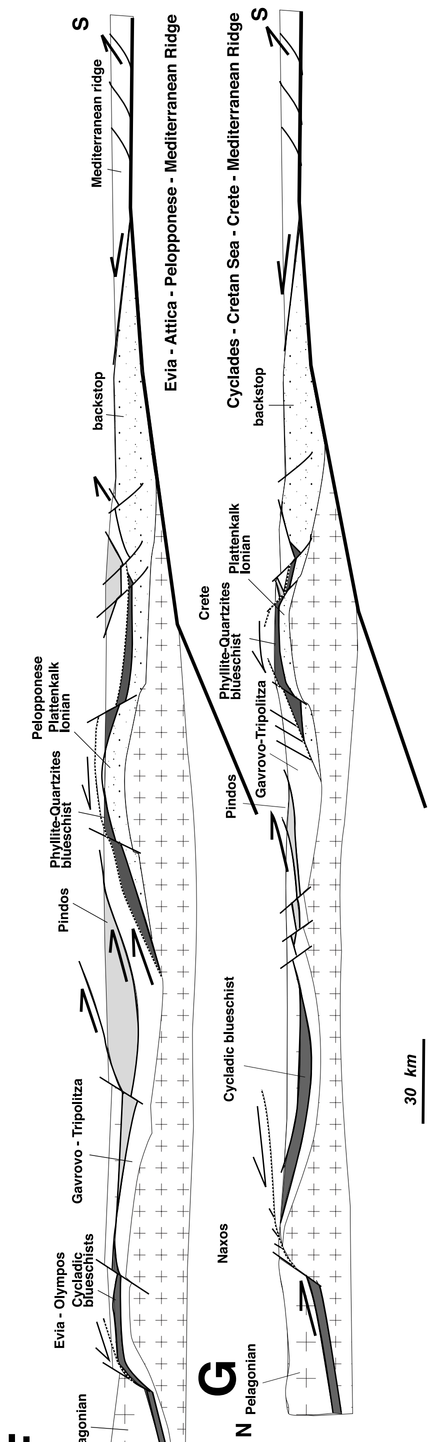

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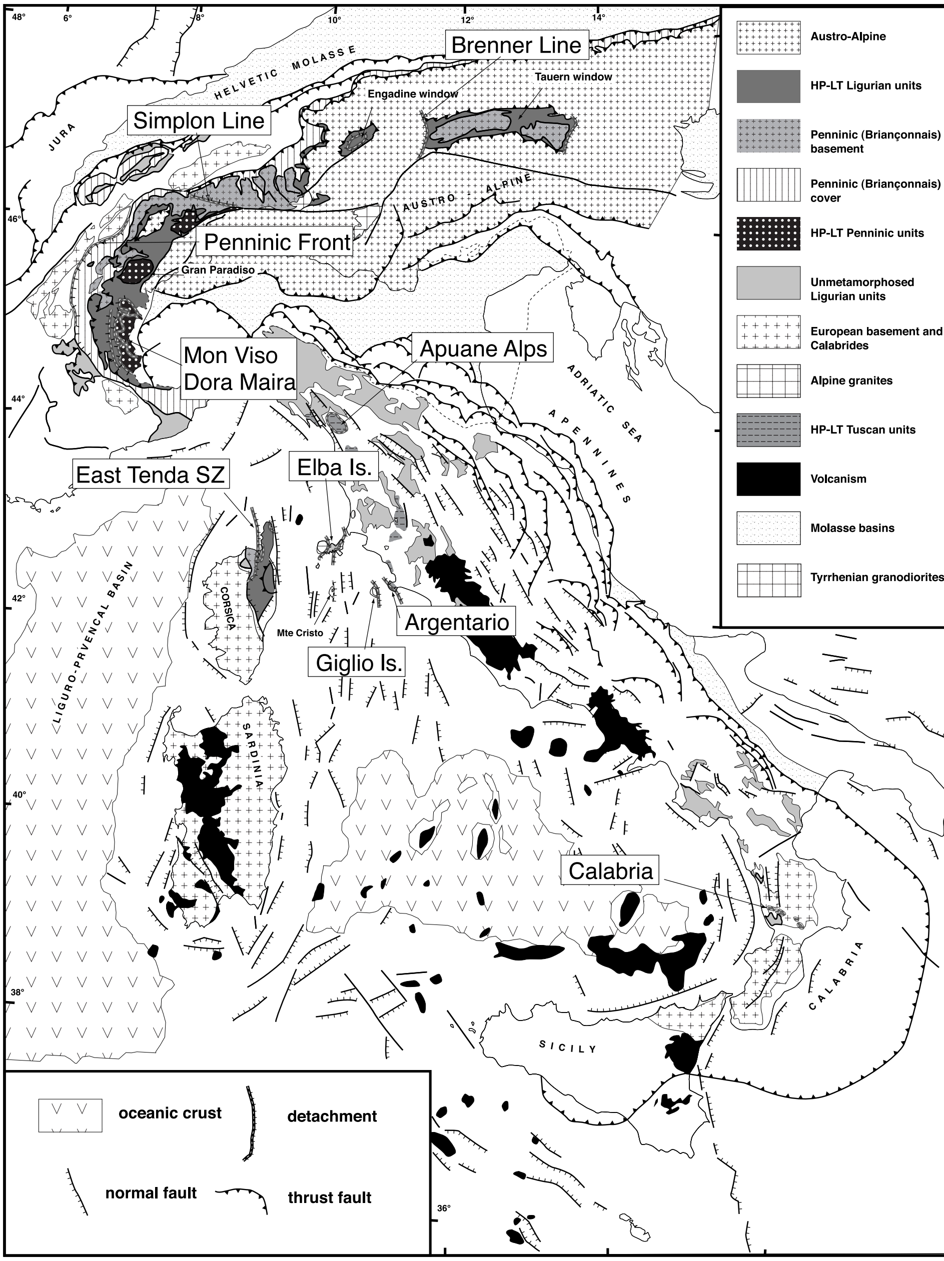

Fig. 3. Tectonic map of the Alpine-Tyrrhenian region adapted from (Bigi and others, 1989; Jolivet and others, 1998a).

others 1996; Pfiffner and others, 1997). This thrust has been reactivated as an extensional shear zone in many places (Sue and Tricart, 1999; Sue and others, 1999). On top of the Briançonnais units are the HP Piemontese and Ligurian units (for example the oceanic domain and the transition with the continental margin). The contact between the two domains is also a HP thrust.

Each of the large domains corresponds to paleogeographic zones of the Ligurian ocean and its margins: the Dauphinois (Helvetic) is the European platform, the Briançonnais (Penninic) the upper part of the margin, the Piemontese is the lower part of the margin and the Ligurian domain is the oceanic crust. Each of these domains contains variably metamorphosed units. Although the metamorphic grade remains in general low in the Dauphinois domain its internal part close to the contact with the internal zones has recorded P-T conditions around 300°C and 2-3 Kbar (Jullien and Goffé, 1993). The Briançonnais has weakly metamorphosed units near the front as well as blueschists and eclogites in the internal zones (Chopin, 1984; Goffé and Velde, 1984; Goffé and Chopin, 1986; Lardeaux and Spalla, 1991; Chopin and Schertl, 2000), and the Ligurian and Piemontese domains also have a considerable variation in metamorphic grade, ranging from the low pressure Chenaillet ophiolitic unit (Mevel and others, 1978) to the UHP Zermatt unit (Reinecke, 1991; Agard and others, 2001).

The uppermost contact of the metamorphic domain is the base of the Austro-Alpine nappes in Switzerland and Austria. The lower part of the Austro-Alpine nappes contains high-pressure parageneses (Monte Emilius) (Dal Piaz and others, 1983), demonstrating that part of the Austro-Alpine was deeply buried during convergence. The results of recent investigations suggest that at least parts of the high-pressure parageneses were formed during the Late Cretaceous (Miller and Thöni, 1995, 1997; Sölva and others, 2001). Some authors favor the possibility of two successive subduction episodes, in the Late Cretaceous and Eocene, in the Alps (Froitzheim and others, 1996; Michard and others, 1996). Most Austro-Alpine eclogites seem to belong to the Cretaceous stage except for some outliers such as Monte Emilius (Cortiana and others, 1999). The base of the Austro-Alpine nappes has been reworked as an extensional shear zone that contributed to the exhumation of the HP rocks (Selverstone, 1988; Merle and others, 1989). The Combin fault is a thrust fault that reactivated an earlier detachment fault of Late Cretaceous - Early Eocene age (Ballèvre and Merle, 1993). The most significant of these extensional shear zones separates the eclogitic Schistes Lustrés from the blueschist Schistes Lustrés (Ballèvre and others, 1990; Philippot, 1990). The sense of shear is top-to-the-west and the greenschist facies deformation corresponds to thinning of the Schistes Lustrés nappe during the exhumation of the Dora Maira massif 35 to 40 Myr ago (Duchêne and others, 1997; Schwartz, 2000; Agard and others, 2001, 2002). An earlier stage (49-43 Ma) that had a large component of E-W coaxial stretching and a component of top-to-the-east shear is associated with the first stages of retrogression of the HP minerals. This contact can be interpreted as a top-to-the-east detachment responsible for the first exhumation of the blueschists below the non-metamorphosed Chenaillet ophiolite (Agard and others, 2001). The Sesia Zone and the Mont Rose area also contain evidence for extensional shear zones that participated in the exhumation of high-pressure rocks to some extent (Reddy and others, 1996, 1999; Wheeler and others, 2001).

A possible evolution of the Franco-Italian Alps in cross-section is presented in figure 4 (Michard and others, 1996; Agard and others, 2001). 50 Myrs ago deep subduction occurred along the European continental margin. The most external parts of the Briançonnais were still at shallow depth while the internal zones were within the subduction channel. The Dora Maira massif was rising above the internal Briançonnais, within the eclogite-facies Schistes Lustrés (Michard and others, 1996). In the upper parts, the Schistes Lustrés were exhumed in the accretionary complex below east-dipping extensional shear zones, which separated them from unmetamorphosed Ligurian units (Chenaillet). Upwelling of the Adriatic mantle started in the middle and late Eocene and the orogen became bivergent with a closed subduction channel. Backthrusts formed as well as west-dipping extensional shear zones that unroofed the internal domains while the thrust front propagated westward. The recent stage of

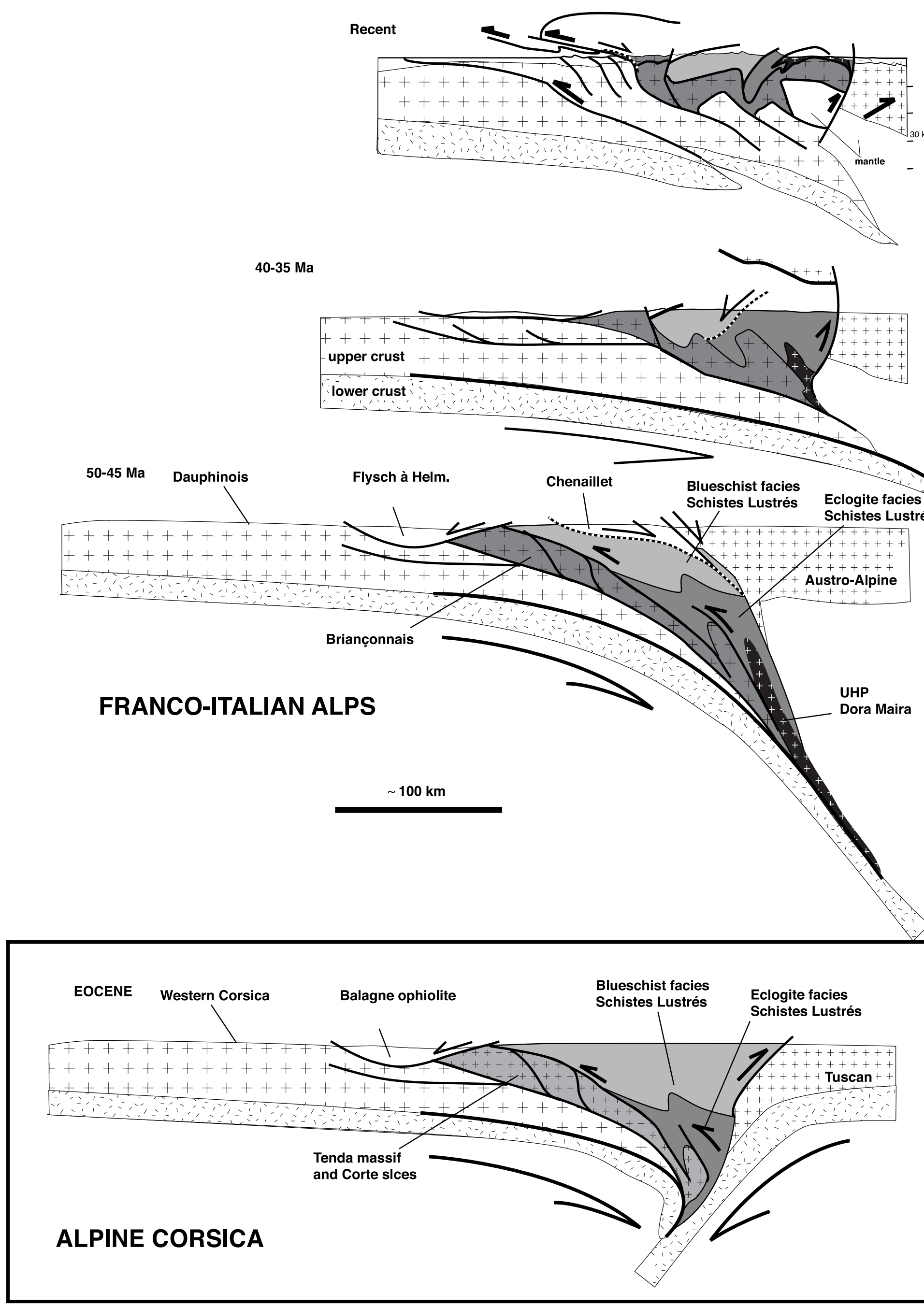

Fig. 4. Schematic evolution of the Western Alps in three stages from Eocene to recent and the geometry of the Corsica-Apennine transect in the Eocene (Agard and others, 2001, 2002). The first stage illustrates the maximum burial for the Dora Maira massif and the overburden removal by a top-to-the-east detachment in the accretionary complex below the unmetamorphosed Chenaillet ophiolitic nappe. The accretionary complex is built mostly at the expense of the oceanic domain and the thinned part of the margin. In the second stage thrusting has migrated inside the European crust and indentation by the Adriatic mantle (Schwartz, 2000) exhumes the internal domain below west-dipping detachments. The main one is localized between the eclogitic Schistes Lustrés and the Blueschists Schistes Lustrés. The last stage is the present situation where the continental margin of Europe is deeply engaged in the accretionary complex and the shortening affects the whole crust.

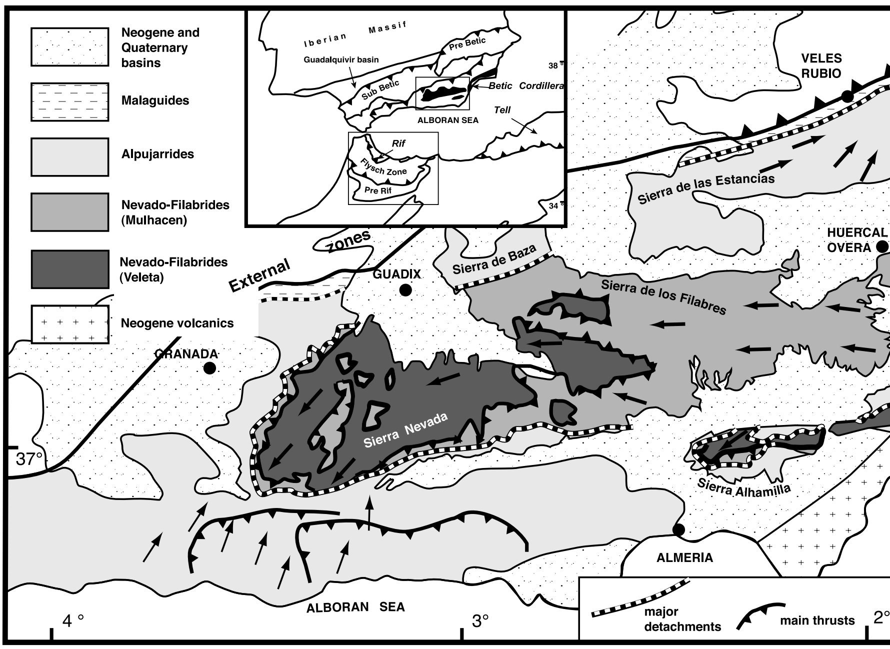

Fig. 5. Tectonic map of the Betic Cordillera around the Sierra Nevada core complex (Jabaloy and others, 1993; Vissers and others, 1995). Thick arrows represent post-orogenic sense of shear in the Nevado-Filabrides and thin arrows the syn-orogenic shears in the Alpujarrides.

deformation followed a similar scenario with the Dauphinois domain involved in the thrusting and the Penninic front reactivated as an extensional shear zone.

The Betic Cordillera and the Rif

The Betic-Rif orogen (figs. 2, 5, and 6) is a bivergent pile of nappes thrust toward the European basement in the north and toward the African basement in the south (Lonergan and White, 1997). Subduction is still active below the Gibralter arc (Gutscher and others, 2002). The center of the belt collapsed in the Miocene and is now occupied by the Alboran Sea (Platt and Vissers, 1989; Comas and others, 1992). High-pressure and low-temperature parageneses in the Betic-Rif orogen are found in the Alpujarrides (or Sebtides in the Rif) units (Goffé and others, 1989; Bouybaouene and others, 1995; Azañon and Goffé, 1997; Vidal and others, 1999). The Alpujarrides-Sebtides also contain the Ronda and Beni Bousera peridotite bodies and associated kinzigites (Kornprobst and Vielzeuf, 1984; Kornprobst and others, 1990). The Alpujarrides are sandwiched between the non-metamorphosed Malaguides-Ghomarides units above and the Nevado-Filabrides below. The uppermost unit of the Nevado-Filabrides (Mulhacen) contains eclogites whereas the lower one (Veleta) contains only greenschist and amphibolite parageneses with no traces of an earlier HP stage (Gomez-Pugnaire, 1987; de Jong, ms, 1991). The HP-LT units (Alpujarrides and Mulhacen) are thus sandwiched between two major contacts with pressure gaps (fig. 5). The upper contact is a brittle extensional or a low-temperature ductile structure (Lonergan and

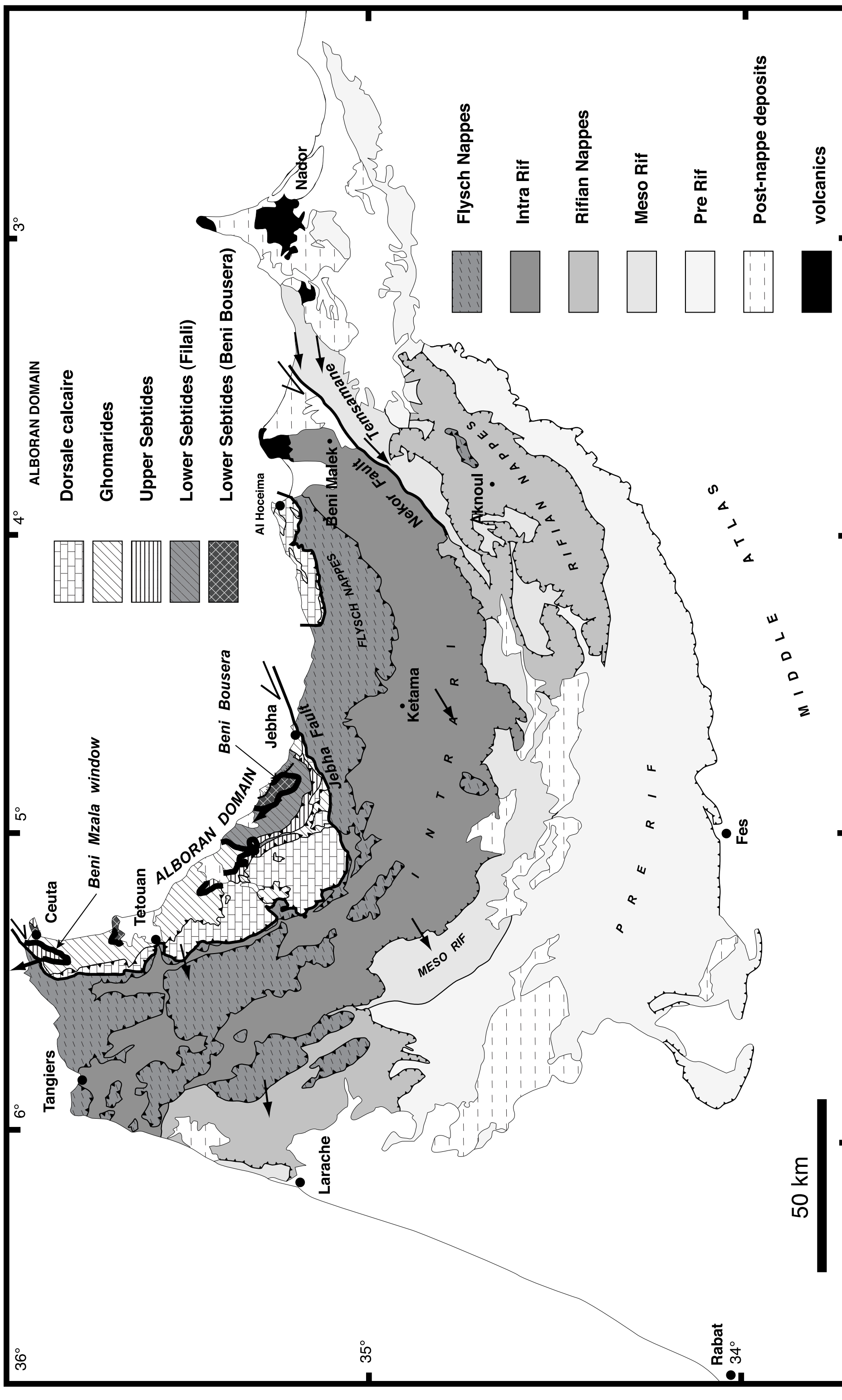

Fig. 6. Tectonic map of the Rif (Wildi, 1983; Chalouan and others, 2001).

Platt, 1995) that puts LP units on HP units, whereas the lower contact (between Mulhacen and Veleta) puts HP units on top of lower-P units and is thus a thrust. Post-HP thrusts are present also within the Alpujarrides, but the detailed geometry is still under discussion (Balanya and others, 1997; Martinez-Martinez and Azañon, 1997; Platt, 1998; Azañon and Crespo-Blanc, 2000). There is a significant pressure gap represented by the contact between the Alpujarrides and the Nevado-Filabrides. The contact is interpreted as the major post-orogenic extensional shear zone (Betic movement zone) (Platt and Vissers, 1989; Jabaloy and others, 1993; Vissers and others, 1995). Considerable post-HP thinning is recognized to have occurred in the whole nappe pile (Martinez-Martinez and Azañon, 1997; Azañon and Crespo-Blanc, 2000) both before and during the rifting of the Alboran Sea from the Late Oligocene (27 Ma) (Platt and Whitehouse, 1999) to the Early Miocene. North-south or northeast-southwest stretching lineations are associated with top-to-the-North sense of shear and thrusting and exhumation in the Alpujarrides (Crespo Blanc and others, 1994). Extension in the internal zones proceeded while the thrust front propagated toward the external zones (Lonergan and White, 1997). A strong thermal event is associated with this late extension as indicated by the clustering of radiometric ages around 19 to 22 Ma obtained from various methods such as Ar/Ar on micas or U/Pb on zircons (Monié and others, 1991; Platt and Whitehouse, 1999; Zeck, 1999). Platt (1998) has challenged the existence of pre-extension thrusts preserved in the Alpujarrides (Balanya and others, 1997). Our own observations south of the Sierra Nevada metamorphic core complex confirm the existence of post high-pressure thrusts that bring Fe-Mg-carpholite-bearing units and HT units above weakly metamorphosed units. The age of these thrusts is not clearly ascertained because the HP event is not well dated. Their relation to the 20 Myr old cooling event is not clear either because it is not certain that this event has been recorded in the low-temperature units of the Alpujarrides. Obviously, the last displacement along the contacts is related to extension (Balanya and others, 1997; Platt, 1998).

Martinez-Martinez and Azañon (1997) and Azañon and Crespo-Blanc (2000) have gathered the available data and proposed the following timing: (1) nappe stacking and high-pressure metamorphism from 37-40 Ma to 23 Ma, contemporaneous with thinning of the nappe stack and exhumation of high-pressure metamorphic rocks, (2) two extensional events in the Late Burdigalian – Langhian (18-15 Ma, N-S extension) and the Serravalian [15-10 Ma, E-W extension along the Betic Movement Zone (Jabaloy and others, 1993; Vissers and others, 1995)]. Fission-track-data, from the Sierra Nevada complex, show that east-west extension lasted until the Tortonian (8-9 Ma) (Johnson and others, 1997).

The retrograde P-T paths of most Alpujarrides units do not show any excursion toward high temperature (Azañon and Goffé, 1997; Azañon and Crespo-Blanc, 2000). This observation suggests that the units were exhumed into the upper crust before the high temperature event occurred, prior to the opening of the Alboran Sea. This event is thus syn-orogenic exhumation before the Miocene (Azañon and Crespo-Blanc, 2000). The age of the peak of HP metamorphism is not well constrained but is generally considered to be 40 to 30 Ma (Monié and others, 1991; Sanchez-Rodriguez and others, 1996).

The geological structure and history of the Rif (fig. 6) has close similarities with the Betics (Chalouan and others, 2001). Tectonic units attributed to the Alboran domain (internal Rif) were thrust over the flysch nappes and parautochtonous units scraped off the African margin. The Ghomarides (= Malaguides) do not show any sign of significant alpine metamorphism, whereas the underlying Sebtides (= Alpujarrides) have a variety of metamorphic grades. HP-LT parageneses have been recently found in the northern Sebtides in the Beni Mzala window (Bouybaouene and others, 1995) with an upward decrease in pressure with strong pressure gaps represented across the contacts implying post-HP thinning. The southern upper Sebtides do not show clear HP-LT parageneses (El Maz and Guiraud, 2001) The contact between the upper Sebtides and the Filali micaschists is thus originally a thrust. Gneiss that crop out below the Filali micaschists was derived from the kinzigite and high-temperature eclogites associated with the underlying Beni Bousera and Ceuta allochtonous peridotite body (Kornprobst and others, 1990; Chalouan and others, 2001). The basal contact of the Filali micaschists thus corresponds to downward pressure increase with a considerable pressure gap and is likely a detachment.

Completely different deformation fields are recorded in the internal Rif and underlying units (Frizon de Lamotte and others, 1991). The internal Rif is characterized by N120°E stretching lineations except in the lower part of the Beni Bousera unit where the trend is N60°E. NE directions are found in the Temsamane regions where ductile deformations are associated with thrusting in the most internal part of the parautochtonous units (Frizon de Lamotte and others, 1991). Once the counterclockwise rotation of the internal Rif has been removed the N120°E lineations become roughly N-S (Saddiqi, 1995; Saddiqi and others, 1995), that direction is not significantly different from the early lineations in the Betics. The lineations were thus passively transported above the basal thrust of the internal Rif. The major direction of thrusting is NE-SW across the Rif and the internal zones of the Betics.

The Tyrrhenian Sea, Corsica, the Apennines and Calabria

High-pressure and low-temperature metamorphic rocks crop out on both margins of the Tyrrhenian Sea in Alpine Corsica (Caron and others, 1981; Fournier and others, 1991), in the Tuscan archipelago, in Tuscany (Theye and others, 1997; Jolivet and others, 1998a; Rossetti and others, 1999, 2001a) and in Calabria (Rossetti and others, 2001b) (figs. 2 and 3).

In Alpine Corsica HP/LT rocks belong to the Schistes Lustrés nappe with an early history similar to that of the Alps (Mattauer and others, 1981; Caron, 1994) and a late history controlled mainly by post-orogenic extension and the opening of the Liguro-Provençal and Tyrrhenian basins (Jolivet and others, 1991, 1998a; Daniel and others, 1996). The most « external » HP rocks in Corsica belong to the Tenda massif [Pmax < 9 kbar (Lahondère, 1996) or up to 1.1 kbar (Tribuzio and Giacomini, 2002)], a piece of the European continental basement thrust toward the west onto western Corsica (Jourdan, ms, 1988). The oceanic domain (Schistes Lustrés nappe and Balagne nappe) rest on top of the Tenda massif. The uppermost low-pressure Balagne oceanic unit that can be compared with the Chenaillet massif in the Alps (Durand Delga, 1984) was emplaced at the front of the orogen in the foreland basin during the Middle and Late Eocene (Egal, 1992). The high-pressure metamorphism dates back to the Late Cretaceous, Eocene and Early Oligocene (Lahondère and Guerrot, 1997; Brunet and others, 2000). The basal contact of the Balagne nappe (lato sensu) in the internal zone is a detachment with unmetamorphosed units resting directly on top of blueschists and eclogites (Jolivet and others, 1990). A post-orogenic extensional event reactivated the Tenda-Schistes Lustrés contact 32 Ma ago and continued until at least the Middle Miocene (Daniel and others, 1996; Jolivet and others, 1998a; Brunet and others, 2000). This post-orogenic stage is responsible for the exhumation of greenschist parageneses while the syn-orogenic stage exhumed eclogites and blueschists.

High-pressure rocks are found farther east within the Tyrrhenian Sea on the island of Gorgona, but with a slightly younger age for the pressure peak (Rossetti and others, 2001a). Farther east HP metamorphism affected both the Ligurian oceanic units and the Tuscan continental units that belong to the Adria plate (Theye and others, 1997; Jolivet and others, 1998a). Radiometric ages are younger to the east and suggest that a HP-LT accretionary complex was under construction in Tuscany while post-orogenic extension and backarc rifting was active in the west (Brunet and others, 2000). HP units in the Tuscan archipelago are topped with unmetamorphosed units and syn-exhumation deformation is mostly extensional. The Apuane Alps farther north also show a syn-orogenic detachment at the top above the syn-exhumation thrusts (Carmignani and Kligfield, 1990; Jolivet and others, 1998a). Extension migrated toward the east from the late Oligocene-Early Miocene in Corsica to the Late Miocene in the Tuscan archipelago, the Pliocene in Tuscany and Quaternary in the Apennines while the thrust front also migrated eastward (Jolivet and others, 1998a).

In contrast to the Alps, this orogen has been bivergent at least from the Late Oligocene or even earlier (Carmignani and Kligfield, 1990) (fig. 4). The Apulian domain here was never thrust on top of the Ligurian domain. The main subduction was west-dipping and corresponds to the underthrusting of the Apulian lithosphere below the accretionary complex whereas the alpine subduction can be considered a backthrust behind the main convergence front. This geometry, of a bivergent wedge, lasted until the Oligocene (~ 35-30 Ma). Before the first backarc extension started, the internal zones were reworked by extension, and the thrust front migrated eastward.

A high-pressure province also exists in Calabria (figs. 2 and 3) (Rossetti and others, 2001b). HP Ligurian units are found beneath a flat extensional contact underlying the Calabrides units of continental affinities. Originally formed as a thrust, it was reworked as a ductile-brittle extensional shear zone contemporaneous with the isothermal exhumation of the HP units between 30 Ma and the Neogene sedimentation (middle-upper Miocene). The pressure peak is dated at 35 Ma and the exhumation took place between 30 Ma and the sedimentation of post-orogenic basins in the middle-upper Miocene. The upper units contain Variscan metamorphic correlated with the continental basement of Sardinia. HP-LT parageneses of Eocene age are found in the lower units of the Calabrides (Rossetti and others, 2001b). A thrust contact carried the HP units above unmetamorphosed Ligurian units. Lower temperature HP parageneses dated at 35 Ma characterize the Ligurian unit observed in the lower tectonic complex. Extensional detachment have been described also in the Calabrian nappe complex (Platt and Compagnoni, 1990).

The Aegean Sea and the Hellenides

High-pressure metamorphic rocks are distributed throughout the whole Aegean region (figs. 2 and 7) from the Rhodope massif in the north (Liati and Seidel, 1996) to Crete and the Peloponese in the south (Seidel and others, 1982; Theye and Seidel, 1991), and from Mt. Olympos in the west (Schermer, 1990) to the Izmir-Ankara suture (Okay and Tüysüz, 1999), the Menderes massif and the Lycian nappes of Turkey in the east (Bozkurt and Oberhänsli, 2001; Oberhänsli and others, 2001; Rimmelé and others, 2003). The scattering is due to a southward migration of the thrust front and subduction through time, and to the post-orogenic Aegean extension (from 30 Ma to the Present) (Jolivet and Faccenna, 2000) (fig. 2). Ultrahigh-pressure relics (microdiamonds, polygonal quartz, former majoritic garnets) have recently been suggested in the Rhodope massif (Mposkos and Kostopoulos, 2001). SHRIMP U/Pb zircon ages (Liati and Gebauer, 1999) that range between 73 Ma (for east Rhodope) and 42 Ma (for west Rhodope) are interpreted as peak pressure ages. For the west Rhodope zircon ages suggest a fast cycle of burial and exhumation with peak pressure at 42 Ma and a peak temperature during exhumation at 40 Ma. There is in fact no firm evidence that the growth of zircon is controlled by pressure. It is thus not certain that the ages correspond to the peak of pressure. A recent synthesis of available radiochronological data (Moriceau, ms, 2000) shows an eclogitic stage at 50 Ma or earlier and progressive exhumation until ~30-35 Ma. Similar ages can be found in the equivalent of the Rhodope massif cropping out on the island of Thasos (Wawrzenitz and Krohe, 1998).

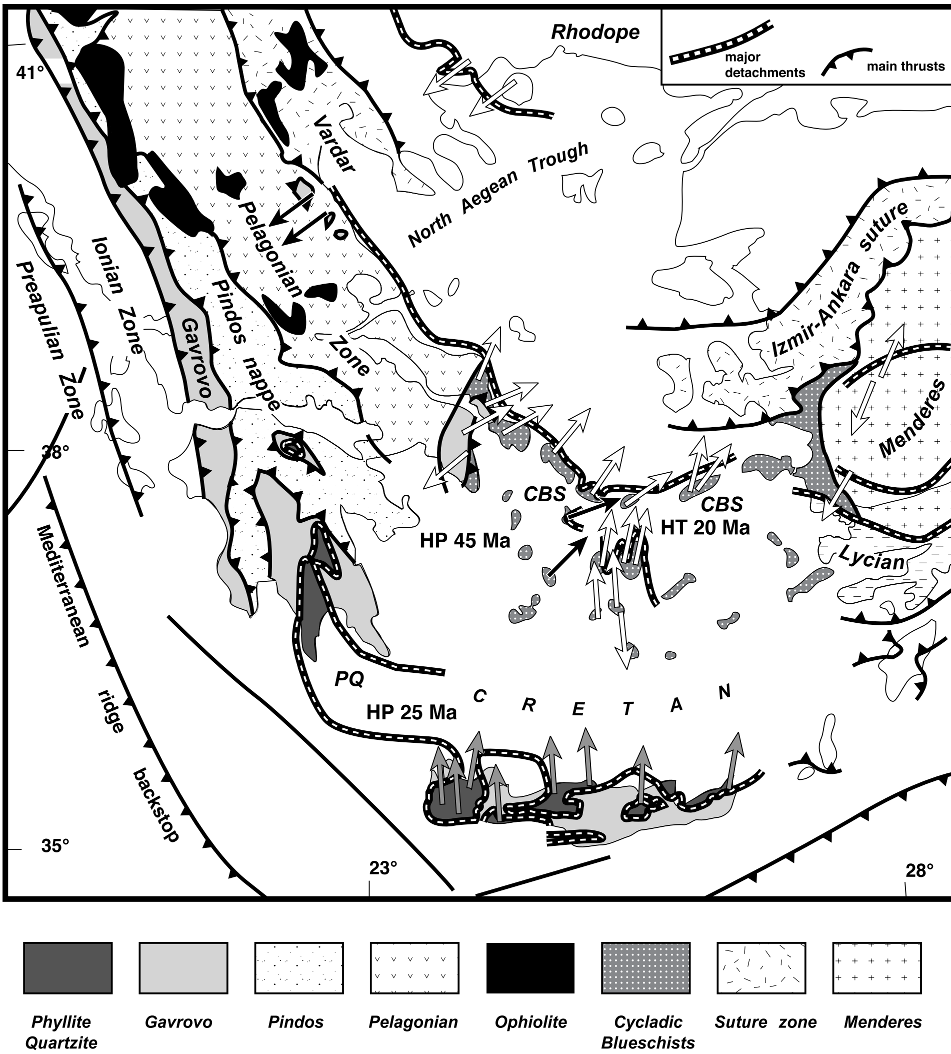

Fig. 7. Tectonic map of the Aegean region and kinematic data (Buick and Holland, 1989; Faure and others, 1991; Lee and Lister, 1992; Gautier and others, 1993; Gautier and Brun, 1994a; Jolivet and others, 1994b, 1996; Foster and Lister, 1999). CBS: Cycladic Blueschists, PQ: Phyllite-Quartzite Nappe, HP: high pressure, HT: high temperature.

The Cycladic blueschists crop out from Mt. Olympos to the Cyclades islands. They have yielded scattered ages, ranging from Late Cretaceous for some eclogites (70 Ma in Syros) (Bröcker and Enders, 1999) to Eocene for most blueschists (Wijbrans and McDougall, 1986, 1988). Although a part of the P-T history is without doubt Late Cretaceous, the major part of it cannot be older than Eocene as attested by the occurrence of nummulites at the top of the Gavrovo unit in the core of the Olympos window where the thrust contact is associated with blueschist metamorphism (Schermer and others, 1990; Schermer, 1993). The blueschists are contained within a stack of tectonic units sandwiched between the underlying Gavrovo-Tripolitza carbonate unit and the overlying Pelagonian units (Bonneau and Kienast, 1982). A gneissic basement also crops out in the core of extensional complexes in the Cyclades (Naxos-Paros-Ios) (Lister, and others 1984; Buick and Holland, 1989). The basal contact of the Cycladic blueschist unit is a large syn-HP thrust that was active in the Paleocene and Eocene. The top contact was originally a thrust that was reactivated as a detachment during the formation of the accretionary complex (Eocene) as well as later during the backarc extension (Late Oligocene and Miocene). The Eocene exhumation has preserved blueschist and eclogite parageneses in the upper units of the Cycladic Blueschists while post-orogenic later events exhumed mostly highly retrogressed eclogites, amphibolites or migmatites which are contained in the lower units (Jolivet and Patriat, 1999; Trotet and others, 2001a).

A younger HP nappe is found farther south in Crete and Peloponese, the Phyllite-Quartzite nappe (Bonneau, 1984). A detachment brings the unmetamorphosed Mesozoic limestones of the Gavrovo nappe directly down on top of the HP metapelites (Fassoulas and others, 1994; Jolivet and others, 1994a, 1996). The lower thrust puts the Triassic-Liassic metapelites on top of the Ionian Plattenkalk and the Oligocene flysch. Both the Phyllite-Quartzite and the Plattenkalk contain HP-LT parageneses in Crete, while in Peloponese the Plattenkalk did not reach blueschist facies (Theye and Seidel, 1991; Trotet, 2000). The pressure peak is dated at 25 Ma in Crete (Jolivet and others, 1996), in good agreement with the stratigraphic constraints on the age of thrusting. The P-T paths in the Phyllite-Quartzite are colder near the contact than farther down in the section where severe retrogression is observed associated to top-to-the-north sense of shear. This exhumation occurred in the early and middle Miocene while thrusting was active deeper in the accretionary complex. In the Peloponese the P-T paths have similar shapes but they are hotter by 100° to 120°C (Trotet, 2000). Despite this difference the upper detachment is present with the same top-to-the-NE sense of shear (once the Peloponese has been rotated back to its pre-rotation position) (Kissel and Laj, 1988).

The uppermost unit of the nappe pile in the Cyclades and Crete is a dismembered nappe (Vari unit in Syros, Asteroussia unit in Crete) (Bonneau, 1984), which does not show HP recrystallization of Alpine age. It contains ophiolite and high temperature gneiss, which have yielded ^{39}Ar-^{40}Ar ages as old as 70 Ma (Bonneau, 1972, 1984; Maluski and others, 1987; Patzak and others, 1994), and locally contains Late Jurassic HP-LT metamorphic rocks (Crete). These observations allow correlation with the Pelagonian units that rest upon the Cycladic blueschists in the north (Bonneau, 1972). The existence of remnants of this nappe shows that the upper part of the accretionary complex has not been removed everywhere and that erosion was thus not the principal exhumation mechanism.

Post-orogenic extension has thinned the crust and reworked earlier syn-orogenic extensional contacts from the late Oligocene until the Pliocene (Jolivet and Faccenna, 2000). It is now active in western Turkey and around the Gulf of Corinth (Armijo and others, 1996; Davies and others, 1997; Kahle and others, 2000). Ductile N-S to NE-SW on N-dipping detachments was found in several islands of the archipelago (Lister and others, 1984; Buick and Holland, 1989; Faure and others, 1991; Lee and Lister, 1992; Gautier and others, 1993; Gautier and Brun, 1994a; Jolivet and others, 1994b; Foster and Lister, 1999) coeval with the intrusion of granodiorites in the Miocene. Finite extension was already very significant in the Early Miocene as shown by the presence of local Early Miocene marine basins. There is no large-scale observation supporting the idea that the crustal thickness has not changed since the Early Miocene over the entire Aegean region (Avigad and others, 2001) but it had been seriously reduced in some areas.

In most islands the earlier Eocene HP mineralogy is strongly overgrown by later greenschist facies parageneses but several examples show unretrogressed eclogites and blueschists which preserved Eocene Ar/Ar radiometric ages suggesting that the rocks were exhumed early (Wijbrans and McDougall, 1988). The examples of Syros and Sifnos suggest that north- or northeast-dipping extensional shear zones also controlled exhumation (Avigad and others, 1992; Trotet and others, 2001a). The rocks on these two islands show superposition of eclogites over blueschist and of blueschists over greenschists suggesting post HP thrusting (Lister and Raouzaios, 1996). However, a careful study of the P-T evolution of these three units show that all of them went through HP-LT conditions and that the deeper units were more intensely reworked because they followed warmer P-T paths during exhumation (Trotet and others, 2001a). Radiometric ages are also younger toward the base, Eocene in the eclogites and Miocene in the greenschists (Maluski and others, 1987). A continuum of shear is thus seen from the Eocene to the Miocene exhumation with top-to-the-north or -NE shear during decompression along a detachment cutting deep in the accretionary complex during the Eocene and accommodating post-orogenic extension in the Oligo-Miocene (Jolivet and Patriat, 1999).

The sequence of events shown on figure 8 is compatible in terms of kinematics with the reconstructions discussed later in this paper, and is partly based upon the works of Trotet (Trotet, 2000; Trotet and others, 2001a), Jolivet and Patriat (Jolivet and others, 1996; Jolivet and Patriat, 1999), and Moriceau (Moriceau, ms, 2000). Soon after the final closure of the Vardar Ocean around 50 Ma the continental margin of Apulia was subducted below the active margin of Eurasia. A bivergent orogen was developing from the northern Rhodope to the Apulian carbonate platform. This stage was characterized by deep (>200 km) circulation of crustal material as could be attested by the presence of UHP parageneses (Mposkos and Kostopoulos, 2001) if this discovery is confirmed. The Cycladic blueschists formed within the subduction complex via the progressive accretion of Apulian platform (Gavrovo) and Pelagic (Pindos) sediments. 35 Myr ago some of the Cycladic blueschist had already been exhumed close to the surface below detachments. Post-orogenic extension started in the Rhodope massif. The upper nappe (Pelagonian and the obducted ophiolite) devoid of HP metamorphism was affected by syn-orogenic extension and partly dismembered. After 30 Ma a fast southward slab retreat started. The subduction complex integrated the last units of the Apulian block (Phyllite-Quartzite nappe and PlattenKalk of Crete and Peloponese) and the oceanic domain located north of the African margin began to subduct. During the progressive accretion in the south final exhumation of the Cycladic blueschists occurred in the backarc domain in a warmer environment below north-dipping detachments. The Phyllite-Quartzite nappe has been exhumed below north-dipping detachments in the late Oligocene – early Miocene. At present the subduction front has migrated until it touches the northern margin of Africa and a thick accretionary complex has built up in the eastern Mediterranean basin (Mediterranean ridge) (Lallemant and others, 1994; Chaumillon and others, 1996; Le Pichon and others, 2002).

The Aegean case is an example of a migrating subduction zone where tectonic units are first incorporated in the accretionary wedge at the subduction front and finally exhumed in the backarc region. Some of these units were exhumed early in the migration process, that is within the subduction complex (Eocene for the Cyclades, Miocene for Crete and Peloponese), others wait until they are brought up to the surface by post-orogenic extension (Miocene in the Cyclades). Post-orogenic exhumation follows a warmer P-T path as shown by the central Aegean core complexes (Naxos-Paros).

What is left of contractional ductile deformation?

The intensity of syn- or post-orogenic extension is often strong enough to have erased most small-scale structures related to the construction of the accretionary

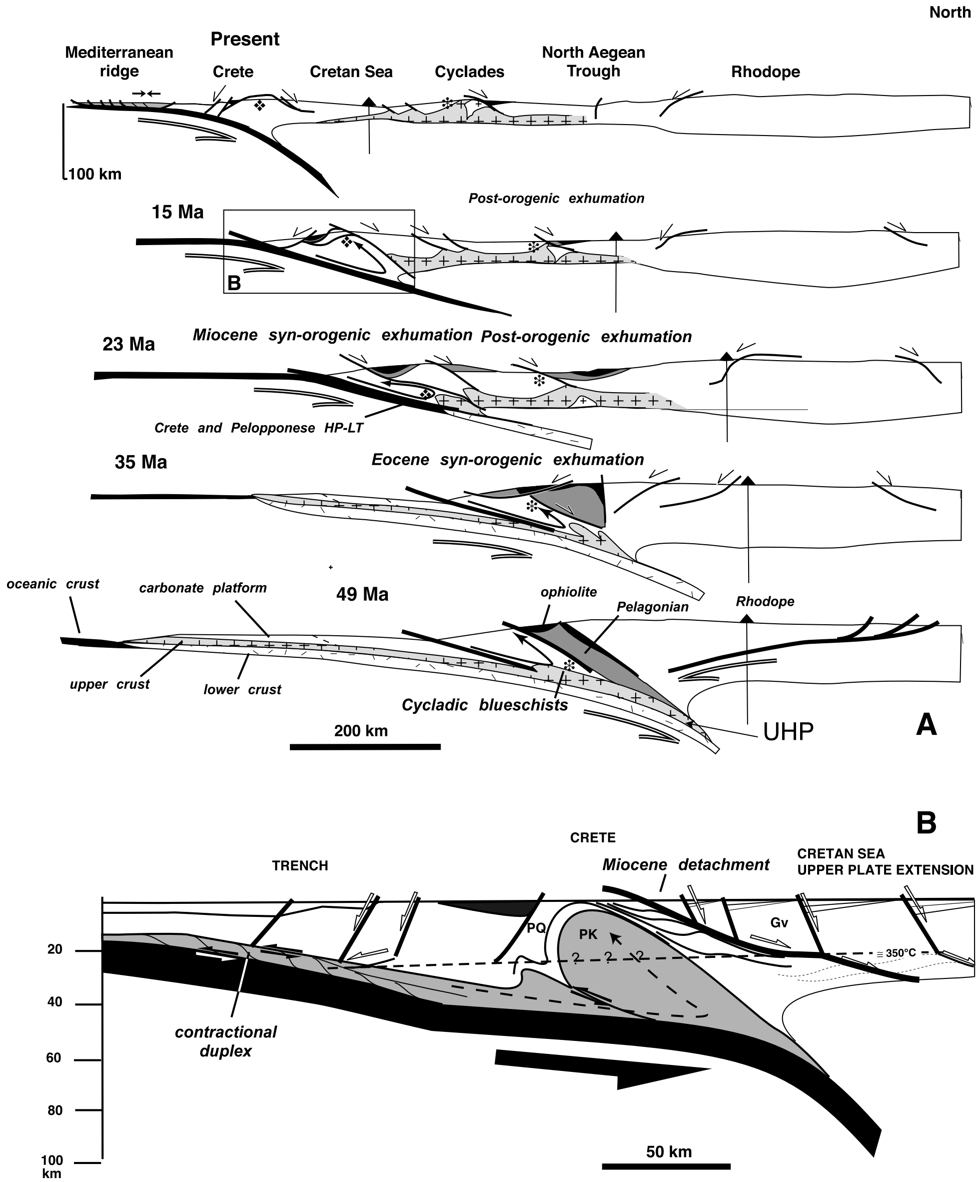

Fig. 8. Schematic evolution of the Aegean region in cross-section.

wedge. The abnormal superpositions are preserved but the ductile deformation observed along these major contacts often dates from the more recent extension. Determining the kinematics of thrusting is thus difficult in the internal metamorphic domains. However, in some cases, contractional structures are preserved in the nappe stack above the detachment. In the Betic cordillera, the internal fabric of the Alpujarrides units was acquired during a top-to-the-northeast shear before the Alboran Sea extension. This early deformation is preserved above the main post-orogenic extensional contact below which all early deformation has been erased. In Calabria, MP-LT units are characterized by top-to-north-east sense of shear related to the growth of glaucophane-bearing assemblages (Rossetti and others, 2001b). These units are located in the hanging-wall of the main detachment and this early HP event pre-dates not only the extension but also the first HP episode of the footwall rocks. A similar case can be found in the Norwegian Caledonides where contractional deformation is preserved mainly above the main late-orogenic detachment (Andersen and Jamtveit, 1990; Andersen and others, 1994). The Rif shows a preserved contractional ductile deformation in the Temsamane where WSW-ENE stretching lineations and top-to-the-WSW kinematic indicators seem to correlate with nappe emplacement in the deep parts of the external Rif (Frizon de Lamotte and others, 1991).

In the Aegean, the major thrusts of the Pelagonian zone above the Cycladic blueschists and of the Cycladic blueschists above the Gavrovo platform crops out in the Olympos, Ossa and Almyropotamos tectonic windows (Godfriaux, 1965; Bonneau and Kienast, 1982; Godfriaux and Ricou, 1991; Schermer, 1993). The top-to-the-southwest syn-thrusting HP deformation is preserved in the Olympos and Ossa windows showing that in some cases syn-orogenic extension does not affect the whole pile and that some deep units reach the surface without major reworking. Inside the Aegean Sea however most shear zones have been reworked either by syn-orogenic or by post-orogenic extension. Avigad and Garfunkel (1989) have correlated the deepest unit of Tinos with the Almyropotamos window. However, the deformation along the contact is entirely extensional and dates back to the late Oligocene - early Miocene post-orogenic episode (Gautier and Brun, 1994b; Jolivet and Patriat, 1999). Much the same could be said for the original contact between the Cycladic blueschists and the continental basement with its platform sedimentary cover in Naxos. Almost nothing has been preserved of the original thrusting fabric. The case of the Peloponese preserves syn-HP top-to-the-SW sense of shear along the main contact between the high-pressure Phyllite-Quartzite nappe and the underlying lower pressure Ionian marbles (Trotet, 2000).

Extensional Shear Zones

The first large-scale evidence for extensional structures is an abnormal superposition through a tectonic contact of young over old terranes with a significant gap, or of low-pressure onto high-pressure metamorphic rocks. Alternatives to normal faulting exist to explain such abnormal superpositions. If the structure is tilted before the formation of the fault, even a reverse fault can cut down section and give the impression of attenuation of the column (see a synthesis in Ring and others, 1999). However when the contact has a regional extent such as in Crete or in the Engadine window, the most likely explanation is a normal fault or a ductile extensional shear zone. It is however always necessary to assess the nature of the contact using the P-T-t evolution of metamorphic rocks on both sides of the contact. A significant pressure gap associated with a significant difference in radiometric ages (old above young) on a regional scale is indicative of a large-scale detachment. A regional variation of the shape of P-T paths is also obtained in some examples, especially in the Mediterranean region, but also in the Himalayas or Oman (Hodges and others, 1993; Jolivet and others, 1998b). The examples of Crete and the Engadine window show a distinct difference between cold exhumation paths immediately below the contact and paths with an isothermal decompression deeper in the structure. This regionalization of the P-T evolution can be interpreted either as a cooling effect of the cold upper plate gliding above the warm lower plate (Hodges and others, 1993; Jolivet and others, 1998b) or as an effect of fluid convection within the shear zone (Morrison and Anderson, 1998).

Shear zones that operate mostly at the greenschist facies or at more brittle conditions often accommodate post-orogenic extension. This was true during the Miocene in the Aegean Sea (Jolivet and Patriat, 1999), and is still true in the Gulf of

Corinth region (Rigo and others, 1996). The Tyrrhenian Sea shows the same geometry (Jolivet and others, 1998a). Syn-orogenic extension often follows the same rule. Some extensional shear zones such as those seen in the Schistes Lustrés or Bündnerschiefer in the Alps started to function in the stability field of Fe-Mg-carpholite around 350° to 400°C in high pressure conditions and then evolved toward low-pressure and temperature conditions during exhumation (Bousquet and others, 1998). In Crete the main detachment between the PQ nappe and the underlying Gavrovo-Tripolitza nappe also shows the same evolution (Jolivet and others, 1996). Other examples outside the Mediterranean region such as the Saih Hatat window in Oman show a similar structure (Jolivet and others, 1998b). The depth involved is variable but the temperature is often around 350° to 400°C that is close to the brittle-ductile transition in the crust. Below this transition the deformation is more distributed in the whole exhumed nappe pile. Above the transition zone extension is exclusively localized along shallowly dipping detachments or steeply dipping normal faults.

However various models of exhumation involve deep extensional shear zones, which cut the entire lithosphere (Andersen and Jamtveit, 1990; Dewey and others, 1993; Chemenda and others, 1995). In the Cyclades the continuum of extensional deformation seen in the islands of Sifnos and Syros from the depth of the transition eclogite-blueschist to the surface might be one example (Trotet and others, 2001a) of such deep shear zones. In this case the sense of shear is toward the internal zones of the belt.

Shear direction along the extensional shear zones is either perpendicular to the belt (Engadine window, Mon Viso) (Ballèvre and others, 1990; Bousquet and others, 1998) or parallel to the belt (Simplon shear zone, Tauern window) (Selverstone, 1988; Merle and others, 1989). Sense of shear for perpendicular extension is either toward the external zones [Schistes Lustrés, stage 2, or Engadine window (Ballèvre and others, 1990; Bousquet and others, 1998)] or toward the internal zones [Schistes Lustrés stage 1, Aegean (Jolivet and Patriat, 1999; Agard and others, 2001)].

CONCLUSION

From the above review of the structure of the main Mediterranean HP belts we draw the following conclusions:

1. All HP units have been highly reworked by extensional deformation either during the formation of the accretionary complex or during later crustal collapse. Most kinematic indicators, with a few exceptions, are related to exhumation below large-scale detachments rather than shortening and burial.
2. Exhumation is accommodated by two processes: underthrusting of more external units and overburden removal by extension (with a minor contribution of erosion). Large overthrusts are evident from abnormal superpositions but the related contractional small-scale deformation has often been erased during subsequent exhumation.
3. Main thrusts and contemporaneous detachments migrate outward at a rate which varies significantly, slow in the case of constrained mountain belts such as the Alps, fast above retreating subduction zones. In the second case, post-orogenic exhumation and extension follow crustal thickening in the backarc region.
4. Detachments often root in the brittle-ductile transition zone.

KINEMATIC BOUNDARY CONDITIONS

Kinematic boundary conditions are important to understand the dynamics of subduction complexes. Fast subduction is expected to result in a colder thermal regime. In the same manner, slow subduction, characteristics of the first stages of subduction (Faccenna and others, 1999), is expected to result in a lower P/T ratio. It

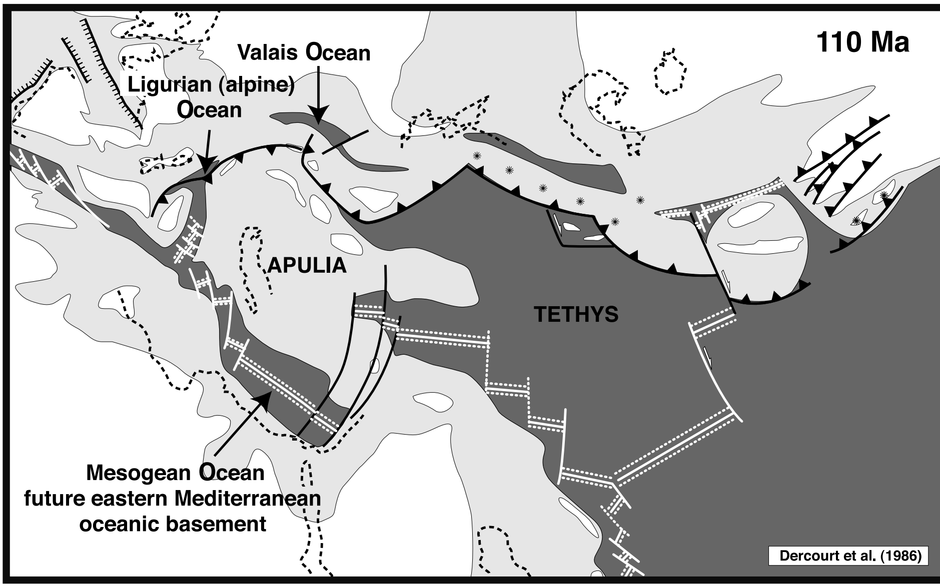

Fig. 9. The Apulian plate 110 Myrs ago (Dercourt and others, 1986).

must be noted however that the fast emplacement of a young and hot ophiolite may also result in a low P/T ratio (Hacker and others, 1996). The motion of the trench with respect to the mantle is also important. A retreating slab can influence the behavior of the subduction conduit (Beaumont and others, 1999; Ellis and others, 1999). HP-LT metamorphic rocks in the Mediterranean region formed in the deep parts of accretionary complexes during the subduction of either the Cretaceous Mesogean Ocean now represented by the Ionian and Levantine basins (Hellenides, Taurides, Apennines, Maghrebides), or the Ligurian ocean (Alps, Alpine Corsica). Figure 9 (Dercourt and others, 1986) illustrates the position of the Apulian Plate (sometimes named "Adria") between Africa and Eurasia. Different recontructions are possible, they differ by the age of separation between Apulia and Africa, thus by the age of the Mesogean Ocean (Triassic or Cretaceous) (Stampfli and others, 1998; Stampfli, 2000; Stampfli and Borel, 2002) or by its size. Because the northern margin of the African continent had an irregular shape, collision was diachronous. Some parts of the Mesogean Ocean are still to be subducted while the collision started as early as the Late Cretaceous along the northern margin of the Adria (or Apulian) plate. In the western Mediterranean and Aegean Sea post-orogenic extension has thinned the crust and dispersed the blueschists exhumed during the syn-orogenic stage. Some parts of the accretionary wedges formed during slab rollback and important outward migration of the thrust front while others remained quite steady. We present a set of reconstructions from the Late Cretaceous to the Present with Eurasia fixed (figs. 10 and 11), which are significantly modified and simplified from earlier works (Dercourt and others, 1986). The Africa-Eurasia kinematics were taken from Dewey and others (1989) and the Apulian plate was attached to Africa as suggested by paleomagnetic data throughout the Cenozoic (Van der Voo, 1993). The kinematics of the Anatolian block for the last 5 Ma is from Le Pichon and others (1995). Before 5 Ma Anatolia was attached either to Africa (before 23 Ma) or to Arabia (after 23 Ma). The kinematics of Arabia with respect to Africa is from Jestin and Huchon (1992). The successive positions of trenches were estimated

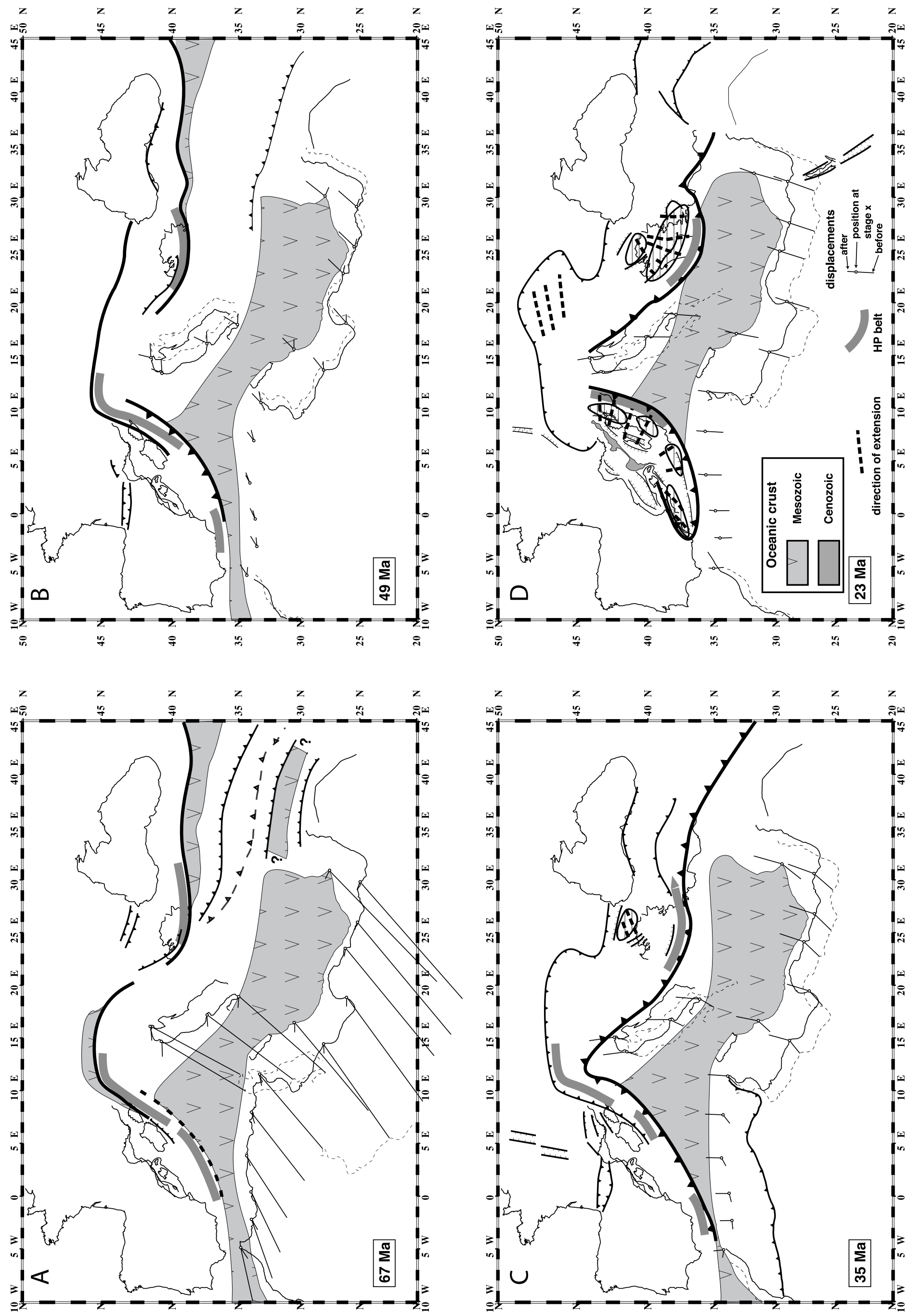

Fig. 10. Kinematic reconstructions (see text for references).

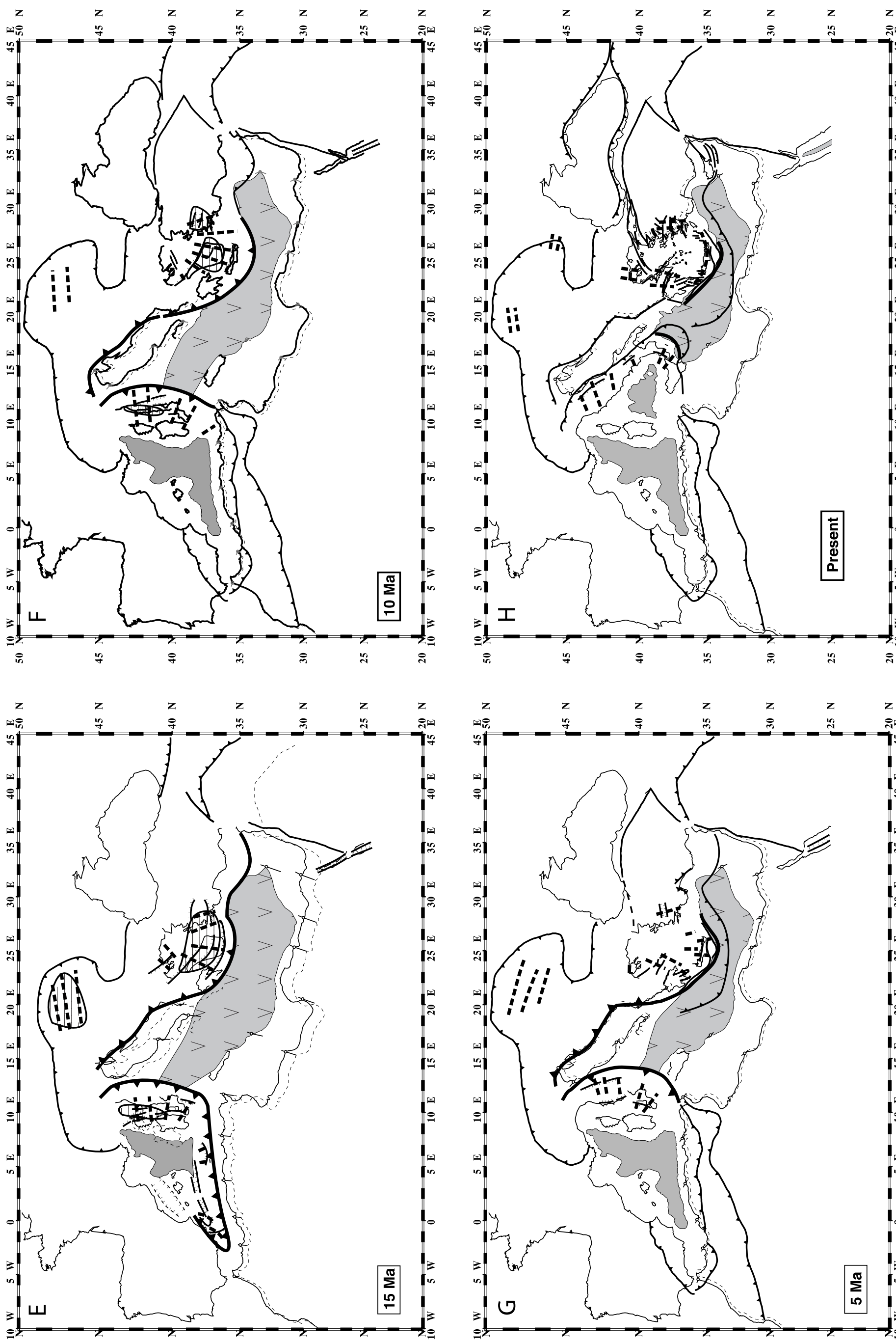

Fig. 10. (continued)

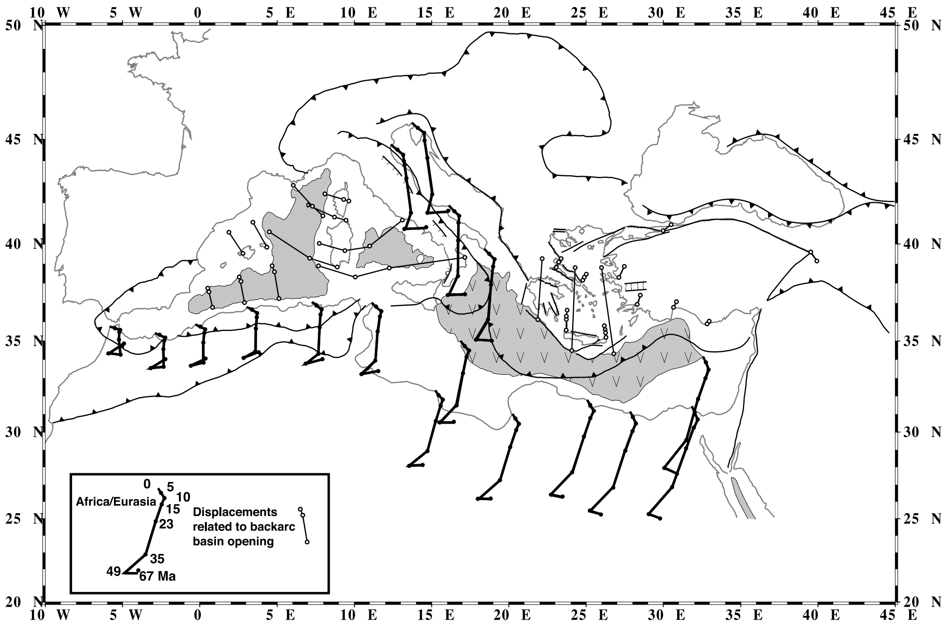

Fig. 11. Displacement vectors for the reconstructions shown in figure 11 (kinematic data from Dewey and others, 1989).

from various sources: volcanic arc migration (Fytikas and others, 1984; Serri and others, 1993; Lonergan and White, 1997; Jolivet and others, 1998a), the length of subducted slabs from tomography (Spakman, 1990; de Jonge and others, 1993; Spakman and others, 1993; Piromallo and Morelli, 1999; Wortel and Spakman, 2000), and the timing of backarc basin opening (synthesis in Jolivet and Faccenna, 2000). The backward motion of subduction zones is constrained by the migration of volcanic arcs and HP metamorphic rocks. Figure 12 shows the case of the Aegean Sea. A fast migration starts at 30 Ma after a long period of steadiness of the magmatic domain in the Rhodope massif (Fytikas and others, 1984; Moriceau, ms, 2000). A similar scheme can be described for the northern or southern Tyrrhenian Sea (fig. 13) (Jolivet and others, 1998a). The case of the Alboran Sea and the Betic-Rif orogen is different as no significant migration of the volcanism or a limited one is observed. The reconstructions show the position of the main trenches, of the main HP metamorphic belts, post-orogenic metamorphic core complexes, and the direction of extension in backarc regions (Frizon de Lamotte and others, 1991; Jabaloy and others, 1993; Gautier and Brun, 1994a; Jolivet and others, 1994b, 1996; Vissers and others, 1995; Gautier and others, 1999). Although HP metamorphic rocks older than 70 Ma do exist in the Mediterranean region, their ages are often suspect so we start our description of the geodynamic setting in the Maastrichtian.

In these reconstructions we consider information provided by seismic tomography. For instance the presence of several HP belts in the Aegean from the Rhodope to Crete must be explained with a single subduction zone throughout the Cenozoic. Tomographic images show a single low temperature anomaly from the surface down to the depth of at least 1000 kilometers with an average northward dip of ~45 ° (Van der Voo and others, 1999; Wortel and Spakman, 2000). The length of this slab suggests that the same subduction zone has been active through most of the Cenozoic and probably before.

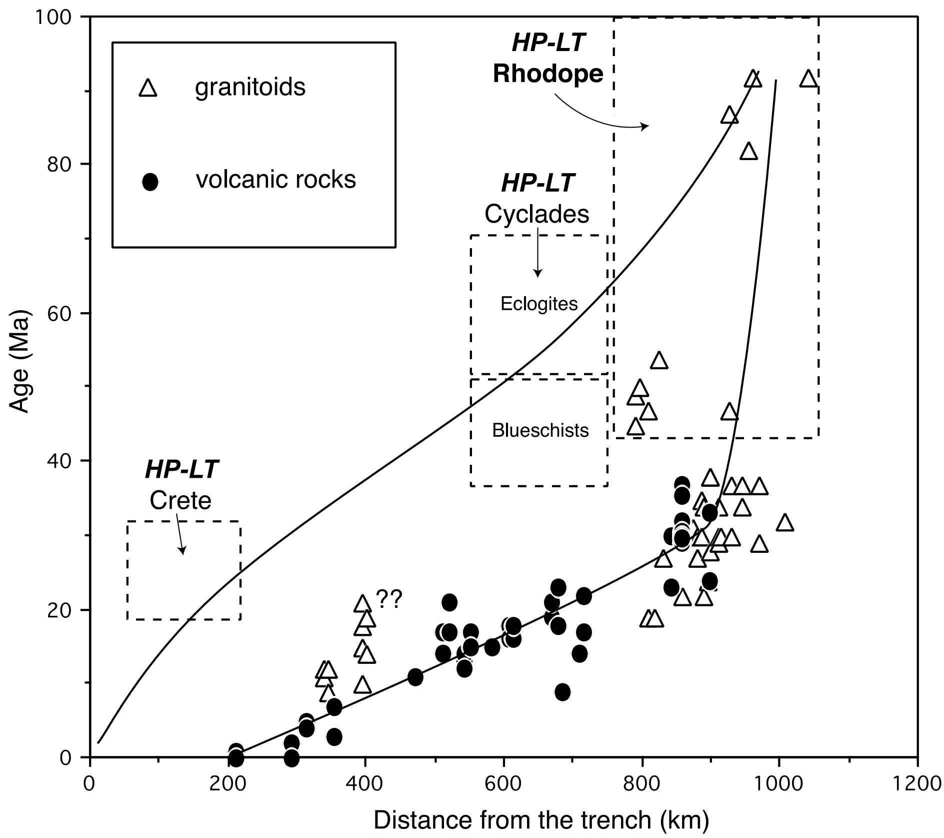

Fig. 12. Age/distance graph showing the migration of magmatic and metamorphic events along a N-S transect of the Aegean region from the Rhodope to Crete (see references in text). Dotted boxes represent the ages of HP-LT events along the transect. The two lines show the main trends of migration of magmatism and metamorphism.

Late Cretaceous - Early Eocene, 67 to 49 Ma (fig. 10A and 10B): the obduction of the Tethys ophiolites on the northeastern margin of Africa (which now belong to Arabia) is contemporaneous with the collision in the Alps with the closure of the Ligurian ocean. Latest Cretaceous ophiolites are found from Turkey to Oman (Ricou and others, 1986). In the eastern Mediterranean they root in the Izmir-Ankara suture and in the Vardar suture in northern Greece (Ricou and others, 1998; Okay and Tüysüz, 1999). Paleocene nappes of blueschists and eclogites also root in the same suture in Turkey. During the same period HP metamorphic rocks form in the Alps in the Ligurian accretionary complex from the central Alps to Corsica, Calabria and probably the Maghrebides. This episode follows a period of fast displacement of the African plate. After 67 Ma the velocity is slow until 35 Ma (fig. 10C). During this time interval the subduction progressively migrates westward and a slow convergence is recorded below Calabria, the Maghrebides and the future Alboran domain.

Late Eocene - early Oligocene, 35 to 30 Ma (fig. 10C): The last remains of the ocean north of the future Arabian plate are consumed and the hard collision between Africa

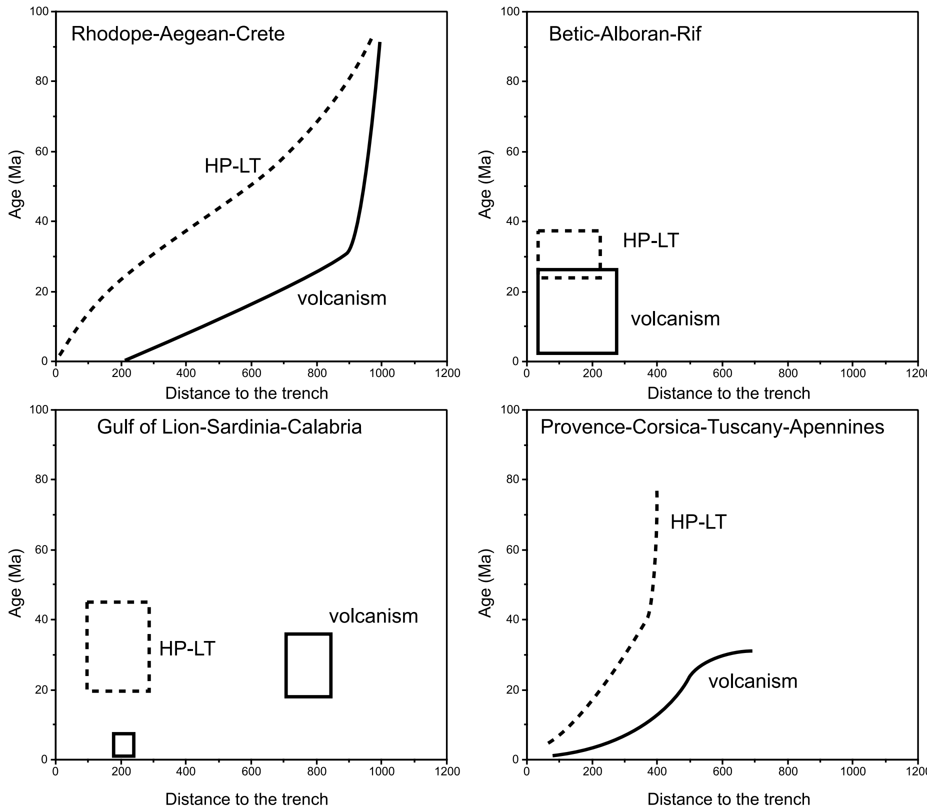

Fig. 13. Age of HP metamorphic and volcanic vents versus distance to the trench for the various Mediterranean regions.

and Eurasia starts. A strong reduction of the absolute northward motion of Africa ensues and the tectonic regime changes completely after 30 Ma [as illustrated by the 23 Ma (fig. 10D) stage] (Jolivet and Faccenna, 2000). The last HP metamorphic rocks are exhumed in the Alps and a transition from flysch to molasse is observed when the frontal thrust starts to propagate westward in the European crust (Tricart, 1984). In the Hellenides thrusts propagate southward inside the Apulian continental crust and the last Cycladic blueschists are exhumed in the accretionary complex. Extension is already active in the Rhodope massif associated with magmatism.

Late Oligocene - middle Miocene, 30 to 10 Ma (fig. 10D to 10F): the reduction of the absolute northward velocity of Africa induces a sharp change in the tectonic regime. The oceanic domain still not subducted is locked between two collision zones and sinks in the mantle with a fast retrograde motion of the subduction hinge (roll back), leading to the opening of backarc basins (Jolivet and Faccenna, 2000). Distributed extension ensues and this period is characterized by the formation of extensional metamorphic complexes in the Aegean Sea, Tyrrhenian Sea, Alboran Sea and Pannonian basin. The direction of extension is radial, always perpendicular to trenches. Blueschists still form and are exhumed in the frontal accretionary complexes (Crete and Peloponese, Tuscany). Backarc extension leads to the emplacement of oceanic crust in the Liguro-Provençal basin (Burrus, 1984; Gueguen and others, 1998), and to distributed crustal thinning elsewhere where the crust has already been significantly thickened (Faccenna and others, 1997). Post-orogenic extension in the Tyrrhenian and Aegean Seas completes the exhumation of HP metamorphics with an excursion of P-T paths toward higher temperatures.

Middle-Miocene – Present, 10 Ma to Present (fig. 10F to 10H): Subduction zones are still migrating toward the external zones but the free face is being progressively reduced as the oceanic lithosphere still available for subduction progressively disappears. At present it concerns only the southernmost part of the Apennines chain (Calabrian arc) and the Hellenic trench. A change in the direction of extension is recorded during the late Pliocene or Quaternary along the Hellenic trench from N-S to E-W suggesting incipient collision with the African margin (Armijo and others, 1992). The narrow oceanic part of the Tyrrhenian Sea opens with very high rates. Extension is active in the whole Apennines and convergence has almost stopped there (Frepoli and Amato, 1997). At 5 Ma or even 2 Ma, fast westward motion of the Anatolian block starts giving a faster convergence along the Hellenic trench (Le Pichon and others, 1995; Armijo and others, 1999).

From these reconstructions we draw the following remarks:

— An important change in the kinematic boundary conditions of the subduction complexes occurred 30 Myr ago. Before this, subduction zones were steadier, after this slab retreat was the rule. It cannot be proven that no slab retreat was active before but we can say that it was much slower. The young HP-LT metamorphic rocks of Tuscany and Crete were thus exhumed in the context of active slab retreat as opposed to the older Cycladic blueschists. Because the available space was small in the Alboran region, slab retreat obviously had a less important impact and, if any, it had to be much slower than in the Tyrrhenian or Aegean subduction zones.

— Highly variable subduction velocities can be derived from the reconstructions. This is due to the increase of the Africa-Eurasia convergence rate eastwards and to the rate of slab retreat after 30 Ma.

P-T-t constraints

A compilation of maximum P-T values and exhumation P-T paths shows a high variability (fig. 14) (for references see figure caption). Most peak metamorphic conditions fall in the lawsonite blueschist, epidote blueschist and greenschist facies. Maximum P-T conditions are quite variable although most points plot below 600°C. Higher temperature corresponds either to UHP parageneses (Dora Maira) or to lower crustal material such as in the Beni Bousera and Ronda units.

It is important to note here that our group obtained a majority of the P-T paths shown in the paper. The same method has been applied (except for the progress of knowledge on the behavior of metapelites during the course of this study) to the same chemical system (mostly metapelites). This procedure ensures that the P-T paths are directly comparable. P-T estimates are based upon the internally consistent thermodynamic database of Berman (Berman, 1988) implemented with new thermodynamic data for high-pressure metapelites (Goffé, ms, 1982; Goffé and Velde, 1984; Goffé and Chopin, 1986; Goffé and others, 1989; Goffé and Oberhänsli, 1992; Vidal and others, 1992; Bouybaouene and others, 1995; Oberhänsli and others, 1995; Jolivet and others, 1996; Azañon and Goffé, 1997; Giorgetti and others, 1997; Theye and others, 1997; Bousquet and others, 1998; Jolivet and others, 1998a; Vidal and others, 1999; Agard and others, 2001; Vidal and others, 2001; Parra and others, 2002). Despite the traditional uncertainties inherent to thermodynamic data in natural environments we explore the consequences of the obtained P-T paths. It must be noted also that when we propose a regional variation of the shape of P-T path inside a single unit or, more often, between units, it does not correspond only to slight differences in the composi-

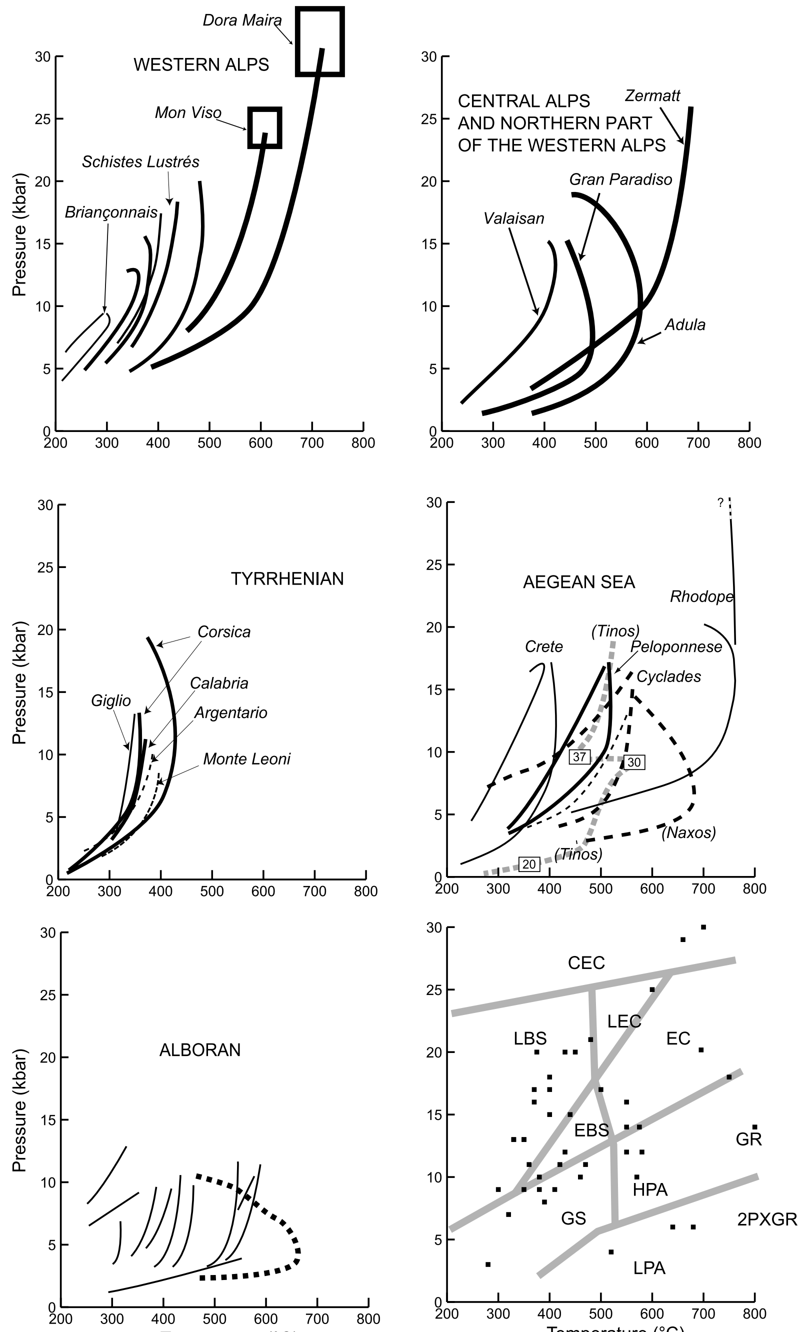

Fig. 14. Compilation of P-T paths in the main HP-LT units from the western Alps (Chopin and others, 1991; Chopin and Schertl, 2000; Agard and others, 2001), central Alps (Lardeaux and Spalla, 1991; Reinecke, 1991; Meyre and Puschnig, 1993; Oberhänsli and others, 1995; Goffé and Bousquet, 1997), Tyrrhenian (Jolivet and others, 1998a), Aegean (Jolivet and others, 1996; Liati and Seidel, 1996; Avigad, 1998; Liati and Gebauer, 1999; Trotet, 2000; Trotet and others, 2001a), and Alboran (Goffé and others, 1989; Bouybaouene and others, 1995; Azañon and Goffé, 1997; Platt and others, 1998; Platt and Whitehouse, 1999; Azañon and Crespo-Blanc, 2000) domains. CEC: Coesite eclogites; EBS: epidote blueschists (Evans, 1990); EC: eclogites; GR: granulites; GS: greenschists; HPA: high-pressure amphibolites; LBS: lawsonite blueschists; LEC: lawsonite eclogite; LPA: low-pressure amphibolite; 2PXGR: two-pyroxene granulites (Bousquet and others, 1997).

TABLE 1

Compilation of some quantitative parameters of Mediterranean subduction systems (see text for references)

	Betics-Rif	Alps	Internal Apennines	Rhodope	Aegean	Crete
Subduction velocity (cm/yr)	0.2	0.5	2	?	1	2.5
Slab retreat	slow	no	fast	slow	slow	fast
P/T (kbar/°)	0.02	0.04	0.026-0.04	0.028	0.029	0.047
Age of peak pressure (Ma)	40 ?	60-45	25	50-100 ?	70-35	25
UHP	-	+	-	+	-	-
Basement	+	+	?	++	+	-
Decompression velocity of eclogites (kbar/Myr)	-	2	-	?	2.5	-
Decompression velocity of blueschists (kbar/Myr)	0.7	0.7	1-2	?	0.3	1.6
Volcanism	+	-	+	+	+	-
Closed/open subduction channel	closed	narrow or closed	open	closed ?	open	open

tion of minerals in a single paragenesis but generally to different parageneses and field occurrence. For instance the mapping of different shapes of P-T paths in the Phyllite-Quartzite Nappe of Crete (Jolivet and others, 1996) is based on the observation of various degrees of preservation of Fe-Mg-carpholite and the presence or absence of chloritoid. The P-T estimates thus corroborate very obvious differences seen in the field. A similar situation exists for the Cyclades where fresh blueschists and highly retrograded blueschists represent very different P-T paths (Trotet and others, 2001a,b). It could be argued that if different parageneses can be observed adjacent to each other in a metamorphic rock then the rock was not fully equilibrated with the P-T conditions, which could make any P-T estimation unrealistic. However detailed investigations of metapelites, which are usually very rich in fluids in the Alpine conditions, suggest that considering local equilibrium between mineralogical phases leads to reasonable P-T estimates (Parra and others, 2001). With this assumption it is then possible to use several parageneses in a single thin-section and derive detailed P-T paths. The absolute P-T values obtained are always associated with significant error bars but the relative position in P-T space of the various estimates are obtained with a single method using an internally consistent database, thus the shape of the P-T path, are probably reliable.

Comparison between the P-T behavior of different Mediterranean mountain belts shows two end-member paths. In the Alps the peak pressures are all aligned on a single steep P-T gradient, whereas in the Aegean peak pressures are aligned on a horizontal line in P-T space. The case of the Alps suggests that the same thermal gradient was maintained during the whole evolution of the belt, implying a sort of steady-state behavior of the subduction complex. This behavior operated throughout time and space in the superficial accretionary complex in the deep parts of the subduction channel for the metasediments, and in the basement units. The Aegean data suggests that the thermal conditions changed during the course of exhumation concomitantly with southward slab retreat. A cold regime was always associated with proximity of the slab and it became even colder when the overburden was removed efficiently by extension, as in Crete. The backarc domain was warmer and the P-T paths followed a loop toward high temperatures. Significantly different peak temperatures evolved in the time of compression before 30 Ma to the time of extension in the upper plate in the high-pressure metamorphic complexes after 30 Ma.

The velocity of subduction is among the important parameters that control the shape of P-T paths. Figure 15A is a diagram of the estimated subduction velocity versus the P/T ratio at maximum burial for several Mediterranean examples. The velocity of subduction was estimated from the reconstructions described above. Subduction velocity here means the velocity of plate sinking into the mantle, not only the velocity of plate convergence. On the reconstruction the contribution of trench retreat to the subduction velocity is taken into account through the opening of backarc basins. For Crete, the Cyclades, Peloponese, Giglio, Argentario and Calabria the maximum P-T conditions are not much different from one part of the nappe stack to the other. For instance in Crete Fe-Mg carpholite is present or recognized as relics in all parts of the Phyllite-Quartzites Nappe and the peak of pressure seems representative of the behavior of the whole nappe. For other examples such as the south-western Alps or the Betic-Rif orogen, which provide several contemporaneous P-T paths for the same belt, we use the P/T gradient on which the various peaks of metamorphism are aligned (fig. 15B, C, and D). There is a considerable difference between the low P/T ratio and slow subduction in the Betics and the high P/T ratio and fast subduction in Crete, and some intermediate points such as the Cyclades or Peloponese suggest that the velocity of subduction is an important factor. This is shown by the P-T estimates as well as the amount and degree of preservation of high-pressure and low-temperature minerals in the rocks. For instance Fe-Mg-carpholite is quite rarely preserved in the Betics while it is still very often present in Crete. The difference between the Betics and Crete is also readily seen from the fact that carpholite appears from alumino-silicates in the Betics at relatively low pressure and from micas in Crete thus at much higher pressure. This is also shown by the fact that HP-LT paragenesis are only found in the uppermost levels of the Betic Alpujarrides units and by a quite strong temperature gradient downward in these units while in Crete or the Peloponese the whole Phyllite-Quartzite Nappe show preserved Fe-Mg-carpholite or its relics. Low temperature was thus preserved only in a narrow band along the subduction contact in the case of the Betics while the accretionary complex was cold in its entirety in the case of the Peloponese or Crete. It thus seems to us that the general regime can be considered colder in Crete than in the Betics and we tentatively relate this difference to the difference in subduction velocity. Some points such as Corsica and the Alps are however quite at odds with this simple tendency as they show high P/T ratio despite a slow subduction. This observation, despite the uncertainties on the estimated subduction velocities, shows that, even with a slow subduction it is possible to obtain and preserve HP-LT parageneses during steady-state subduction and exhumation. This implies that a cold subduction complex can form from the depth of formation of coesite to the surface even where the convergence rate is low. It is important to note here that the Alps-Corsica subduction complex was formed at the expense of the Ligurian ocean and its margins and not the Mesogean Ocean as in the case of the Betics or Hellenids. This paleogeographical difference should also be taken into account.

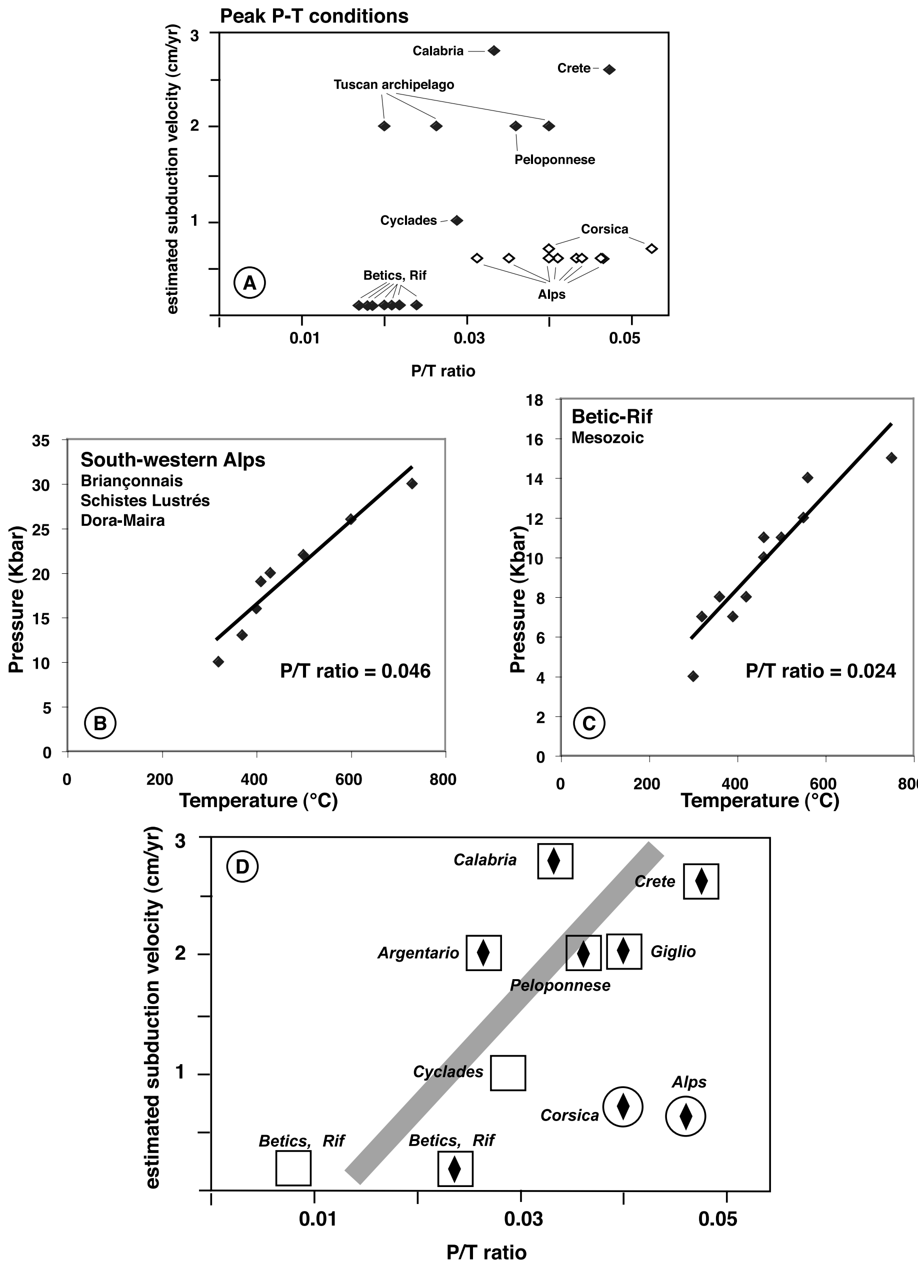

Fig. 15. Relation between estimated subduction velocities and P-T ratio. (A) At peak P-T conditions the estimated subduction velocity is plotted against the P-T ratio. Small black diamands represent units contained in mountain belts issued from the closure of the Mesogean Ocean and small open diamonds those issued from the closure of the Ligurian Ocean (Alpine Ocean). (B) and (C): Peak pressure versus temperature for various units in the Alps and Betic Rif, the slope gives the P-T ratio plotted in figure 15 D. (D) estimated subduction velocity against P/T ration, large black diamonds symbolize Fe-Mg-carpholite. Squares represent belts issued from the closuere of the Mesogean Ocean and circles those issued from the closure of the Ligurian Ocean.

The shape of P-T paths is different from one belt to another. The shape of exhumation P-T paths in the western Alps evolves from a continuous cooling during decompression in the west for units which have not been buried to large depths to almost isothermal decompression with a slight cooling followed by cooling for deep units. Large pressure gaps are seen at the contact between the Mon Viso and the Schistes Lustrés as well as between Mon Viso and the Dora Maira UHP unit (Ballèvre and others, 1990; Schwartz, 2000; Agard and others, 2001). The good preservation of HP-LT parageneses in the western part of the Schistes Lustrés below the poorly metamorphosed Chenaillet ophiolite can be interpreted as the cooling effect of the detachment or as the result of slow exhumation within a cold P/T environment. The slight cooling during the exhumation of the Dora Maira massif from 30 to 8-10 kbar implies that the subduction channel was kept cool even at great depth. That scenario is possible only with a fast circulation of material even though the subduction was slow.

The central Alps show some similarities and differences with the southern transect. The general P/T gradient is about the same, around 20°/kbar. The exhumation path of the UHP Zermatt unit is not different from Dora Maira. The Gran Paradiso and Adula (Lepontine) continental units, however show exhumation with significant heating which is not seen in the southern transect. There are no significant kinematic differences between the rate of convergence in the south and in the north. This difference in the shape of P-T paths is probably not due to a slower exhumation because Burov and others (2001) suggest that a high temperature leads to lower resistance and thus to a faster circulation in the subduction channel. This warmer behavior of the northern part of the Alps has been attributed to the introduction in the prism of thick continental units that produce their own heat by radioactive decay (Bousquet and others, 1997; Goffé and others, 2003). The nature of the material introduced in the accretionary complex is thus crucial to understanding the thermal history.

HP metamorphic rocks found on both margins of the Tyrrhenian Sea show a common P-T evolution with peak P-T conditions significantly colder than in the Alps. The burial and exhumation were mostly achieved before the opening of the Tyrrhenian Sea and, before or during the opening of the Liguro-Provençal basin. The Corsican P-T path is warmer than the Tuscan or Calabrian path and, closer to the Alpine. From west to east, from Giglio to Monte Leoni, the Tuscan HP metamorphic rocks, show a decrease in maximum pressure and a slight decrease in P/T ratio. This suggests a progressive decrease of the thickness of the accretionary complex during the eastward retreat of the trench.

In the Aegean region various kinds of P-T evolution are observed. From north to south and from old to young, the general P/T context gets colder. In the Rhodope massif eclogites are associated with quite high temperatures and the retrograde path goes through the field of migmatites. The Cyclades eclogites (Eocene) contain glaucophane and lawsonite and are much colder than in the Rhodope. Some were exhumed through the blueschist facies and have thus been strongly cooled during exhumation (Syros) (Trotet and others, 2001a) while others, which were exhumed later during the Miocene, were significantly heated and show evidence of partial melting (Naxos) (fig. 14). The island of Tinos shows an intermediate P-T path (Parra and others, 2001): the first part from the depth of the eclogite facies consists in a decompression accompanied with cooling until a pressure about 10 kbar some 37 Myrs ago, then the P-T path shows an isobaric heating until approximately 30 Ma, finally a second stage of exhumation brings the rocks to the surface with a first isothermal decompression or with a slight cooling until 2 kbar followed by the final exhumation along a warmer gradient that is not constrained. The two stages of exhumation, syn-orogenic before the formation of the Aegean Sea (before 30 Ma) and post-orogenic afterward are recorded in the metamorphic rocks of Tinos. In Crete garnet is absent, aragonite is present in the marbles, Fe-Mg-carpholite is well preserved and chloritoid is only locally present showing a much colder evolution. The peak of temperature did not exceed 400°C for pressure as high as 16 to 18 kbar. As in the Cyclades, two types of exhumation P-T paths are recognized (Jolivet and others, 1998b). Immediately below the main detachment a strong cooling is observed while deeper units followed an isothermal decompression. In all cases a reduction of the original thickness and pressure jumps across the main contacts are present. This spatial differentiation between cold and warm exhumation paths can be interpreted in various ways. Footwall cooling can be either due to short distance conduction effects when the cold upper unit glides above the warm lower unit (Hodges and others, 1993; Jolivet and others, 1998b), or to local fluid convection within the detachment or a combination (Morrison and Anderson, 1998) of both. Peloponese blueschists which are contemporaneous with those of Crete also show this difference but with an overall warmer accretionary wedge than Crete (Trotet, 2000). P-T conditions are intermediate between those of Crete and those of the Cyclades. Lateral variations of the thermal regime within a single accretionary complex can be due to various causes: faster removal of the overburden in Crete by a more efficient extension in the center of the Hellenic arc due to slab retreat, faster subduction of the oceanic crust for the same reason, or differences in the material accreted below the Phyllite-Quartzite nappe with more oceanic material or sediments in Crete.

The progressively colder regime observed from the Rhodope to Crete, and from the Late Cretaceous to the early Miocene probably signifies evolution of the subduction regime. The Oligo-Miocene accretionary complex was formed during the retreat of the subduction zone that started some 30 Ma ago. It can be envisaged that the rate of subduction and the efficiency of removal of the overburden by extension led to a colder subduction complex, but other parameters such as the nature of the accreted material are important too. As modeled by Bousquet and others (Bousquet and others, 1997; Goffé and others, 2003), the ratio between the amount of basement with a high radiogenic heat production and the amount of sediments in the accretionary complex is important to determine its thermal regime. In Crete the metamorphic units are indeed less rich in basement lithologies than in the Rhodope and involve instead more calcareous sediments poor in radiogenic elements. The Alps and Corsica (fig. 15) show that even if the convergence is slow a cold complex can be obtained. The Franco-Italian Alps, Corsica and Crete are all characterized by a large amount of weakly radiogenic sediments and little basement. This might partly explain the cold regime in the subduction complex.

The Betic-Rif orogen and the Alboran Sea (figs. 14 and 16) show an evolution strikingly different from the Alps or the Aegean. HP-LT parageneses are found only in the uppermost units of the Alpujarrides metasediments. Deeper units were subjected to intermediate or high temperatures. The P-T peaks are aligned along two gradients, a HP-LT gradient at low T and a warmer one at higher T. The various units of the Alpujarrides were exhumed along parallel isothermal paths with large differences in temperature. These facts indicate a rather warm accretionary complex with only a minor part in the blueschist facies. Those units were exhumed before a sudden thermal event in the early Miocene. In the center of the Alboran domain this event led to an excursion toward high temperature and partial melting during exhumation.

Figure 16 shows a synthesis of several P-T paths obtained in the Betic and the Rif (Goffé and others, 1989; Bouybaouene and others, 1995; Azañon and Goffé, 1997; Platt and others, 1998; Platt and Whitehouse, 1999; Azañon and Crespo-Blanc, 2000, and this work). We have distinguished between Mesozoic metasediments on one hand and Paleozoic metasediments and basement units on the other. The peaks of pressure

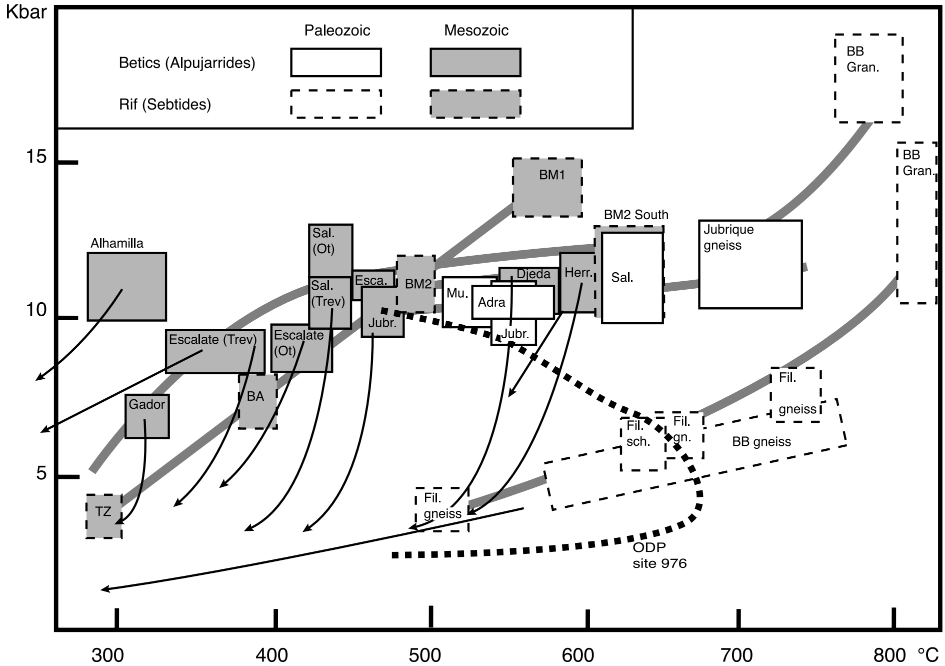

Fig. 16. Detailed P-T paths in the Betic Cordillera and the Rif (Goffé and others, 1989; Bouybaouene and others, 1995; Azañon and Goffé, 1997; Platt and others, 1998; Platt and Whitehouse, 1999; Azañon and Crespo-Blanc, 2000). A distinction is shown between metamorphic rocks derived from Mesozoic sediments on the one hand and those derived from Paleozoic sediments or basement rocks. In Mesozoic metasediments the metamorphic evolution is only alpine and there is no risk of confusion between alpine and possible variscan parageneses. The Rif and the Betics show a very similar evolution for Mesozoic metasediments. Thick lines mark the main P/T gradients. Two main trends are seen: one for all Mesozoic units and the basement units of the Betics, and one for the Rif basement units which seem to have recorded mainly the retrograde low-pressure gradient. See text for explanation. Abbreviations: BA: Boquette-Angeras, BB: Beni Bousera, BM: Beni Mzala, Esca: Escalate, Fil: Filali, Gran: granulite, Herr: Herradura, Jubr: Jubrique, Trev: Trevenque, TZ: Tizgarine.

are aligned on a HP-LT gradient for parts of the Mesozoic metapelites and on a warmer gradient for other Mesozoic units as well as Paleozoic metasediments and basement units. A maximum pressure of ~11 to 13 kbar was reached in the accretionary complex. Higher pressures are recorded in the lower crustal units of the Beni Bousera and Ronda granulite-peridotite units. These values suggest that the accretionary complex was not very thick and that sediments were not buried to great depth as if the subduction channel was closed. The two gradients shown by the pressure peak also show that the deep parts of the accretionary complex were warmer. Two reasons can be invoked: a higher heat production in the basement and Paleozoic metasediments or the injection of warm mantle peridotites and granulites in the accretionary complex. The end of exhumation occurred along a HP-LT gradient. Most rocks following retrograde P-T paths show initial isothermal decompression or a slight cooling before reaching the HT-LP post-orogenic gradient. The P-T paths for rocks of the Alpujarrides units recovered at ODP site 976 instead underwent a strong temperature increase during exhumation, similar to that observed in the central Cyclades (Naxos) for the most recent exhumation.

The association of a given parageneses with an age is an even more difficult exercise (fig. 17). Most geochronometers are thought to be temperature-dependent and the concept of blocking-temperature is widely used. We should thus only have

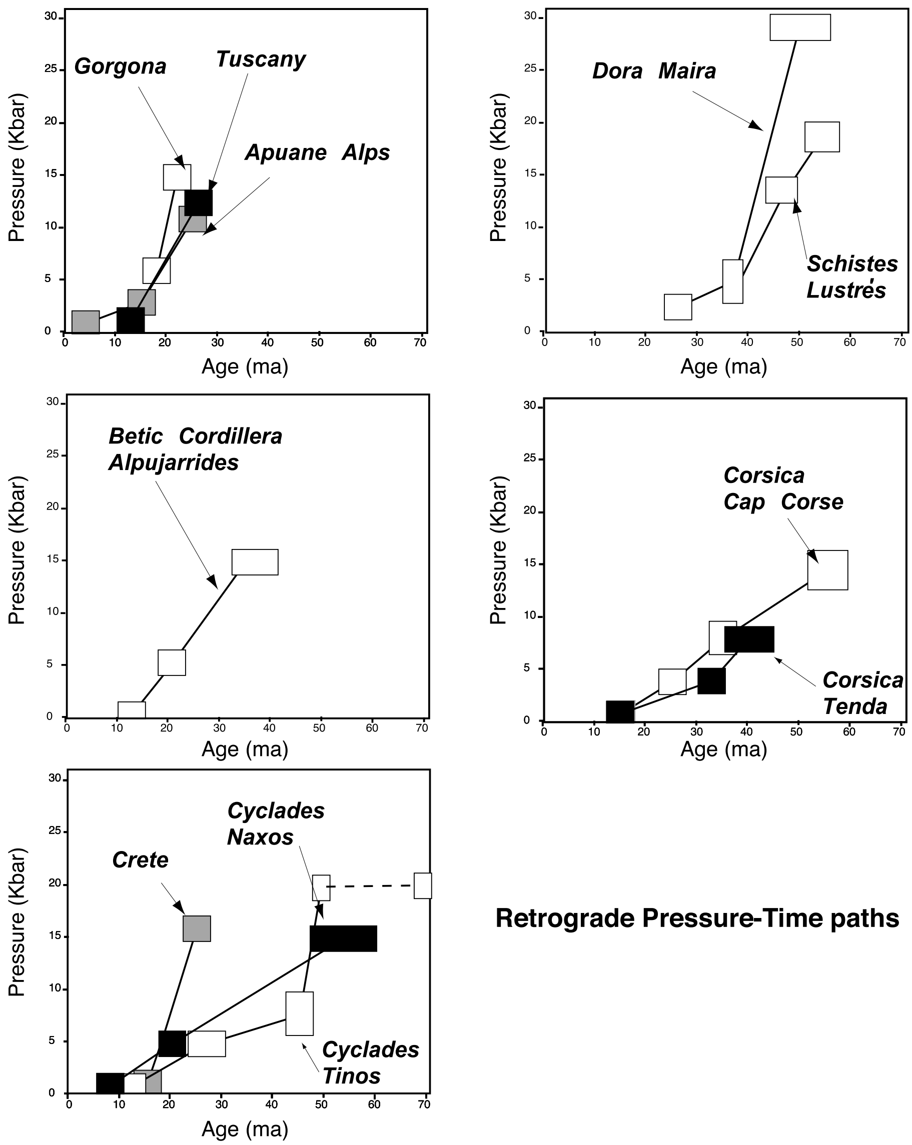

Fig. 17. Pressure/age diagrams (Bröcker and others, 1993; Jolivet and others, 1996; Brunet and others, 1997; Duchêne and others, 1997; Bröcker and Franz, 1998; Platt and others, 1998; Bröcker and Enders, 1999; Platt and Whitehouse, 1999; Azañon and Crespo-Blanc, 2000; Trotet and others 2001a; Agard and others, 2002). Explanation can be found in text.

cooling ages. However in the case of the Mediterranean high-pressure and low-temperature rocks estimated temperatures are often close to the blocking temperature of the Ar/Ar system in micas, these temperatures are only known with large uncertainties. Certainly in the case of the HP-LT rocks of Crete that were equilibrated at very low temperature the Ar/Ar ages of phengites are not cooling ages but rather crystallization ages (Jolivet and others, 1996). In the case of the Alps or Corsica detailed studies of Ar/Ar ages of micas (Brunet and others, 2000; Agard and others, 2001) show that different ages can be obtained from different generations of micas in the same sample or outcrop, thus casting doubt on the concept of blocking temperature, or at least on the value of the blocking temperature. The obtained ages are consistent with the general geological context and the succession in time of deformation stages. We thus use the Ar/Ar ages within the hypothesis that micas can retain crystallization ages rather than mere cooling ages in the HP-LT environments of the Mediterranean. This does not apply of course when we deal with high-temperature parageneses of the Naxos migmatites for instance.

In the Alps the UHP Dora Maira massif shows a two-stage exhumation. The first part of the exhumation path was rapid from the pressure of stability of coesite to 5 kbar. Estimates based on the combination of various dating methods (Lu-Hf, Sm-Nd, Ar-Ar, Rb-Sr) (Duchêne and others, 1997; Amato and others, 1999) suggest exhumation rates which may exceed 1 cm/yr. Similar evidence for a fast exhumation has been described in the Tso Morari eclogite in the Himalaya (de Sigoyer and others, 2000). Similar rates of exhumation are found in the northern Tyrrhenian Sea, Crete or Tinos (Cyclades) for rocks with maximum pressures much lower than in the Dora Maira massif (fig.17). The Schistes Lustrés west of the Mon Viso show a slower exhumation (Agard and others, 2002). The HP event is not well dated in the Betic cordillera because of a thorough thermal resetting around 20 Myrs ago. One amphibole dated at 48 Ma (Ar-Ar) (Monié and others, 1991) suggests an Eocene HP event, which remains to be precisely constrained. The example of Corsica instead shows a quite slow exhumation throughout. Geospeedometry based on the analysis of zoning patterns in garnet in terms of diffusion rates also leads to the general conclusion that exhumation is fast (Perchuk and Philippot, 1997, 2000). In most examples studied so far several hundred °C /Myr, or several cm/yr are calculated. Although the error is difficult to estimate and probably quite large, modelling of diffusion provides an independent observation of fast exhumation rates from the depth of eclogite facies. The main point is that the first stage of exhumation is rapid and that this observation is coherent with the isothermal decompression observed in many examples. The last stage of exhumation, from 5 kbar to the surface, is usually slower (Duchêne and others, 1997; Ring and others, 1999).

DISCUSSION

The geological observations reviewed above suggest that:

— The occurrence of HP and UHP eclogites at the surface implies that rocks buried along the subducting slab were exhumed along the subduction channel, from great depths incompatible with an accretionary wedge.
— The early exhumation of eclogites proceeds at fast rates.
— The surficial part of the exhumation path (30-40 km) occurs within the accretionary complex and is slower.
— Exhumation in the accretionary complex is controlled either by erosion or syn-orogenic extension. In many cases of the Mediterranean region extension seems the most active mechanism because unmetamorphosed units are often preserved as the top structural units.
— Syn-orogenic extension is accommodated by gently dipping detachments with hangingwall motion toward either the internal or external zones. Syn-orogenic extension can be contemporaneous with post-orogenic extension when slab retreat is active and backarc opening occurs (Crete).
— Detachments root in the brittle-ductile transition zone but locally deeper extensional shear zones can be observed (Aegean).
— UHP occurs only in the Alps and the Rhodope, which were highly constrained mountain belts during their formation.

— The thermal regime in the accretionary complex is dependent upon the velocity of convergence and the nature of the subducted material (sediments or basement).
— The velocity of subduction that combines velocities of convergence and retreat further influences the shape of the exhumation P-T path. At least three different behaviors are observed. The Alps show a steady-state evolution with a cold P-T gradient throughout despite a slow subduction velocity, implying a cold subducting crust or that the continental basement was not involved in the accretionary complex during the early stages of accretion. The Aegean shows a non-steady-state evolution with a cooling of P-T gradients with time that can be attributed to an acceleration of subduction and slab retreat. Finally the Betics show a quite warm P-T gradient compatible with slow convergence and a continental subducting crust. Figure 13 illustrates the migration of HP-LT metamorphism (and contractional deformation) and volcanism for the following transects: Rhodope-Aegean-Crete, Betic-Alboran-Rif, and Provence-Corsica-Tuscany-Apennines. Fast migration is observed in the Aegean and Tyrrhenian regions, whereas no significant migration characterizes the Alboran domain. The evolution of P-T conditions in the Aegean region shows a tendency toward a colder regime. The exhumation of the cold HP-LT metamorphic rocks of Crete occurred while the subduction zone was actively retreating after 30 Ma (Jolivet and Faccenna, 2000). A similar conclusion can be drawn from the study of the Corsica-Apennine transect. In the Alboran domain instead a slower migration led to a general low P/T ratio and a clustering of all ages around 2 to 25 Ma. The effect of slab retreat can be three-fold. It leads to a good preservation of the cold HP-LT parageneses and a variety of radiometric ages in the Tyrrhenian and Aegean Seas because the high-temperature event associated with the backarc context does not last long in a given region. It also facilitates the removal of the overburden by extensional processes, thus accelerating exhumation. It may also lead to an open subduction channel where the circulation of particles is easier, enhancing fast burial and exhumation (Beaumont and others, 1999; Ellis and others 1999).

UHP Versus HP, Several Levels of Circulation

While the role played by erosion is difficult to quantify, the end of exhumation in Corsica and the Cyclades was clearly associated with either syn- or post-orogenic extensional tectonics. Extension, probably in association with erosion, appears to be an inescapable factor in removing the overburden, although it appears that erosion played a minor role in removing the overburden in the Aegean. However, true extensional shear zones do not reach the deep portions of accretionary complexes. In any case, it is highly improbable that extensional tectonics reached the depth of the UHP parageneses seen in the Dora Maira massif. A different mechanism must therefore be invoked to explain exhumation from the depth of eclogites to the depth of the blueschist or greenschist facies.

The record of pressures as high as 30 kbar requires that metamorphic rocks have been dragged down along the subduction channel rather than buried below an improbable 100 kilometers thick crust. Present day examples of thick continental crust do not show such large thickness. It is physically improbable that weak crustal material can form bodies thicker than 70 to 80 kilometers because ductile flow in the lower crust should erase such abnormally deep crustal roots (Bird, 1991). A maximum of 70 to 80 kilometers is proposed for the Himalaya (Maggi and others, 2000). It can be argued that some crust is hidden in the mantle because the lower crust has been eclogitized to a large extent and thus has mantle-like seismic velocities (Austrheim, 1987, 1994; Dewey and others, 1993; Le Pichon and others, 1997). However this argument requires that the whole lower crust has been eclogitized to attain such high average seismic velocities. On the other hand, the temperatures suggested by seismic studies at the base of the Himalayan-Tibetan crust are high and partial melting is possible (Romanowicz, 1982; Alsdorf and others, 1998; Alsdorf and Nelson, 1999; Hacker and others, 2000) although the absence of hydrous minerals in xenoliths included in shoshonitic lavas suggest that partial melting is not generalized. The Dora Maira massif does not show evidence of these high temperatures and no evidence of partial melting is observed. In the case of the Alps it is thus likely that the crust never reached such a large thickness. The only remaining explanation, in the absence of tectonic overpressures, is that the rocks followed the subducting lithosphere and were then exhumed within the subduction channel (England and Holland, 1979; Ernst and Liou, 2000). We might thus envisage two levels of circulation: a deep one corresponding to the subduction channel with high rates of exhumation, and a more superficial one, corresponding to the accretionary prism with lower exhumation rates and a different behavior between ductile material at depth and brittle material near the surface.

The analysis of exhumation rates also shows that the first stages are much faster than the last episodes when the tectonic units get close to the surface (see also Duchêne and others, 1997). The last parts of exhumation governed by extension and/or erosion are thus slower. A more efficient mechanism needs to be found for large depths. Extension is active above and within the brittle-ductile transition in general. It can be supposed that at high temperatures the presence of low-resistance material such as sediments or serpentinite leads to low viscosities that favor fast circulation in the subduction channel or in the lower parts of the accretionary wedge. When approaching the brittle-ductile transition, viscosity increases and the exhumation velocity is mostly controlled by the efficiency of overburden removal by extension.

Burov and others (2001) have shown that several levels of circulation can be modeled with an upper level corresponding to the accretionary complex and a lower level to the subduction channel. The velocity of exhumation is much faster in the subduction channel and an "extensional" shear zone is observed at the roof of the subduction channel (fig. 18). Rocks are scrapped off the lower plate by the shearing along the base of the upper plate and the positive buoyancy of the thermally softened material essentially drives exhumation. The displacement of rocks inside the low-viscosity subduction channel induces a component of "extensional" shear along the contact with the upper plate at depth. Two situations have been tested one with a full eclogitization of the lower crust leading to low buoyancy and one with low eclogitization with a higher buoyancy of the subducting lithosphere. In both cases two circulation levels of low viscosity material are observed in the subduction complex. The upper circulation goes down to 40 to 50 kilometers and can be accommodated by natural processes such as an accretionary complex whereas the lower circulation that goes down to 100 kilometers or more is accommodated in the subduction channel. This model cannot be directly compared to the Mediterranean examples because the boundary conditions are probably quite different essentially in terms of temperature distribution. In the model the accretionary complex is too hot compared to what can be deduced from metamorphic parageneses in the Mediterranean case and this explains in part the very low viscosity of the lower crust that is then engaged in the circulation of material in the subduction complex. With a more realistic distribution of temperatures (work in progress) the material would be more viscous and circulate less easily. However the general picture can be used for a comparison with the Mediterranean case.

The rare occurrence of UHP sediments implies, either sediments were difficult to drag down to large depth or they were subducted, but could not make their way back up to the surface. The second solution is difficult to envisage because with a retreating subduction zone such as the Hellenic one the subduction channel should be opened

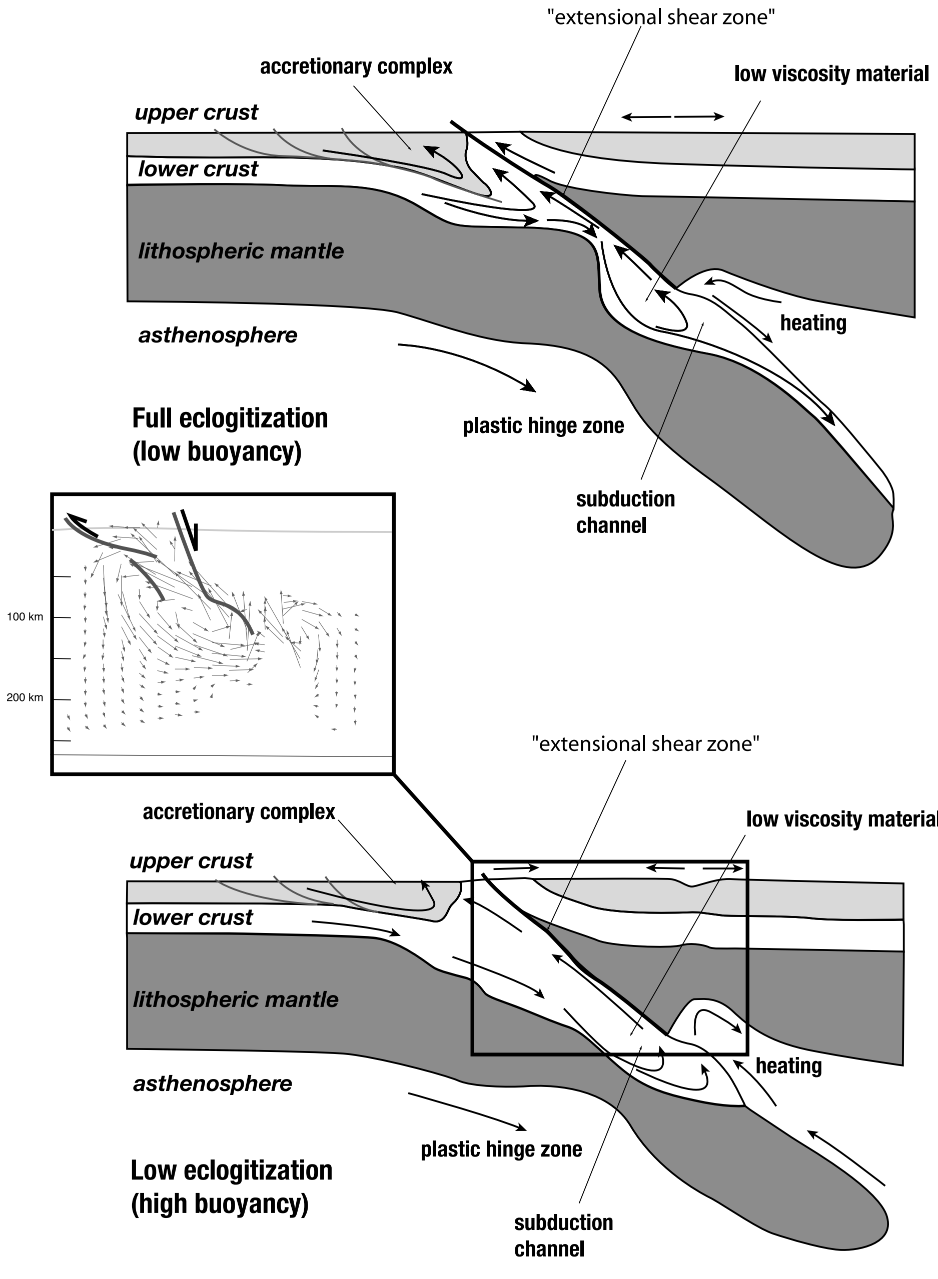

Fig. 18. A numerical model of exhumation after Burov and others (2001). A fully coupled model based on the code Paravoz, itself derived from FLAC (see Burov and others, 2001, and references therein) shows the evolution of a continental subduction zone and the trajectories of particles in the subduction channel. Two end-member situations are illustrated here. Top diagram: the lower crust is fully eclogitized during subduction and the subducting lithosphere thus has a low buoyancy. Bottom diagram: the subducting lower crust is little or not eclogitized thus giving to the downgoing lithosphere a higher buoyancy. In both cases one observes the formation of a subduction channel which allows the circulation of rocks at several levels. The upper circulation can be equated to the accretionary complex (above 40 km) and the lower levels to the subduction channel. The heating of rocks during their subduction lowers their viscosity enough to allow for a fast circulation of materials. The velocity of this circulation is greater at depth where the temperature is higher which matches the observation of fast-then-slow exhumation. An "extensional" shear zone forms at the contact between the subduction channel and the upper plate along which the rocks are exhumed. The insert shows a detail of the particle trajectories in the case of a high buoyancy lithosphere. In all cases there is no build-up of tectonic overpressures beacuse the walls of the subduction channel adapt instantly by ductile deformation.

(Beaumont and others, 1999) and thus facilitate downward and upward circulations. One can argue that only the UHP alpine metasediments were originally strongly attached to their basement (radiolarian cherts deposited on the Ligurian ocean floor in the Zermatt unit). Other HP sediments instead represent higher levels of the stratigraphy (Permo-Triassic to Late Cretaceous metapelites and marbles with minor metabasites) and might have detached from the basement more easily in the accretionary complex or in the upper parts of the subduction channel. The presence of large amounts of sediments in the subduction channel could explain why no UHP basement has reached the surface except locally in the Alps. Subducted sediments might act as lubricants of the channel and prevent a severe shearing of the basement which then is not included in the return flow. This behavior could explain the difference with "basement-rich" belts such as the Caledonides or the Rhodope.

Density Changes in the Subducting Crust, Delamination

Some Mediterranean chains have little basement cropping out and even less lower crustal material (Le Pichon and others, 1988). When basement units are included in the belt such as in the case of the Alps (Lepontine nappes, Briançonnais basement units) they are usually decoupled from their sedimentary cover. The Ivrea zone is a unique example of lower crustal material exposed in the Alps but its exhumation probably predates the formation of the belt, during the Permian post-Variscan rifting event (Handy and others, 1999). Laubscher (1990) has argued in favor of delamination of the lower crust during the formation of mountain belts and after, during the formation of backarc basins. The relatively thin crust of the Apennines has shortened considerably since the Oligocene. Some parts of the continental crust must thus have been removed, although without more precise data on the pre-contraction crustal thickness it is difficult to calculate the amount of missing lower crust. The present-day thickness of the Adriatic plate (around 30 km) (Morelli, 1998) can give a first-order approximation. Deep seismic sounding experiments across the Italian peninsula show a shortening that can be estimated to be around 30 kilometers at the scale of the whole crust (Barchi and others, 1998a,b) whereas the restoration of extensional processes in the Tyrrhenian Sea gives approximately 120 kilometers of finite extension (Faccenna and others, 2001). These observations suggest that some 90 kilometers of crust might have been removed by delamination during convergence (Contrucci, 1999).

Observations in the Aegean bring some answers to this problem (fig. 19). The present-day mantle structure shows only one 1200 kilometer long slab crossing the upper - lower mantle boundary, although the only real suture is far north of the present-day subduction front (Spakman and others 1988; Wortel and Spakman, 2000). Several deep basins such as the Pindos Ocean were closed during the process of convergence (Bonneau, 1982). Only the sedimentary parts and some upper crustal materials are involved in the accretionary complex. A simple model is to delaminate the lithospheric mantle and part of the lower crust during the formation of the accretionary wedge. Only the upper crust and the sedimentary column of the various paleogeographic domains involved successively are accreted. Given the distance traveled by the front of subduction from the Vardar suture to its present position no delamination would imply the presence of several slabs in the mantle. Their absence pleads in favor of a large delamination.

Delamination is facilitated by density changes in the subducting crust. As shown by Austrheim (1987, 1994), discussed by Le Pichon and others (1992, 1997), and by Dewey and others (1993), lower crustal compositions can reach high densities once eclogitized (see also Goffé and others, 2003). Even though it is difficult (not impossible) to envisage complete eclogitization that would change the lower crust into a "geophysical" mantle as seen through seismic velocities, it is quite probable that some

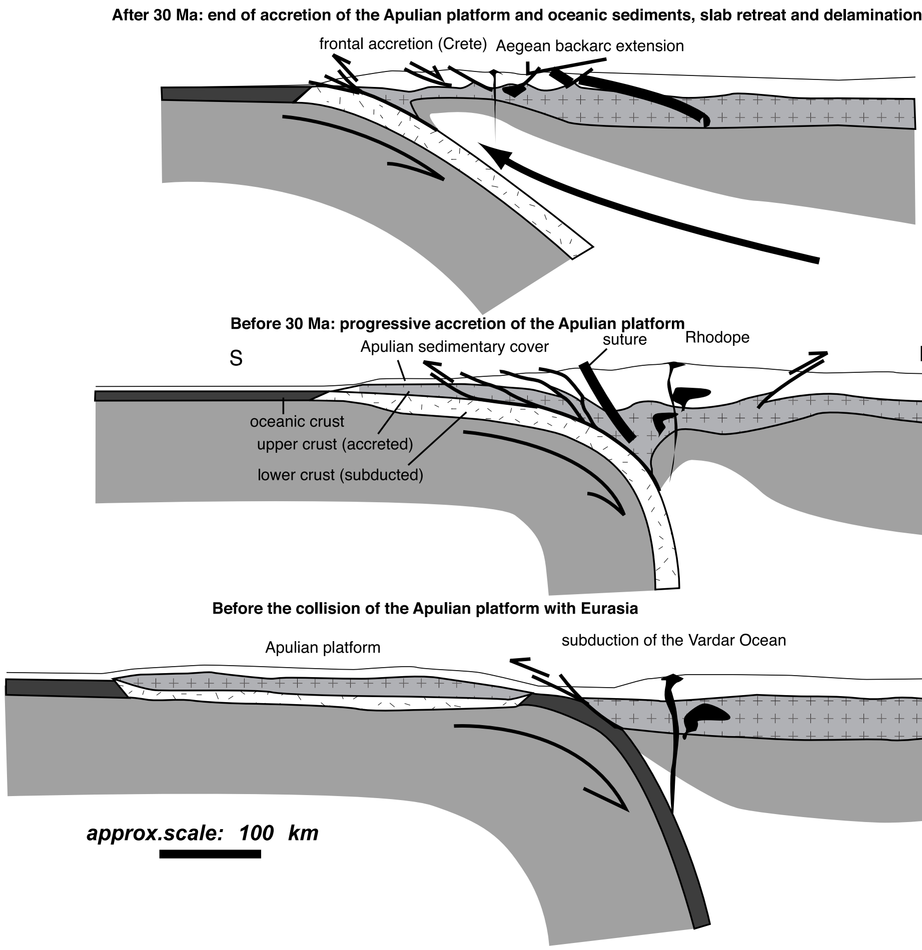

Fig. 19. Delamination in the Aegean region.

parts of the lower crust are dense enough to be subducted and driven definitely into the mantle (Laubscher, 1990; Dewey and others, 1993; Jolivet and others, 1999). It is however not certain that metamorphic recrystallization is complete enough to generate high densities. The availability of fluids seems crucial for recrystallization to occur as shown by the granulite-eclogite transformation in the Bergen arc of the Norwegian Caledonides (Austrheim and others, 1997). Sapin and Hirn (1997) have interpreted seismic data below the Himalayas as indications of the presence of eclogitized lower crustal material. Bousquet and others (1997), based on thermal calculations in a numerical model with imposed kinematics, and taking into account metamorphic recrystallizations rates, suggest a relation between the burial velocity and the formation of eclogites. When the velocity is higher than 4 mm/yr eclogite may form, below this value granulites form instead and partial melting may occur. In the Mediterranean region convergence rates are quite low but mostly above this limit. Although other factors than velocity control the shape of the P-T path, eclogites have formed and probably still form in the Mediterranean subduction zones. There is thus a strong potential for a dense lower crust to delaminate during subduction.

A PROGRESSIVE SUBDUCTION MODEL

We discuss a general conceptual model of burial and exhumation of high-pressure metamorphic rocks, which does not involve any abrupt change in the boundary conditions such as a detachment of the lithospheric root. This model is designed to produce a progressive burial and exhumation of metamorphic rocks.

The model described below (fig. 20) takes into account the major conclusions summarized above. The general framework is a subduction complex, the upper part of which is an accretionary prism (above 30-40 km maximum). The backstop of the prism can dip inward or outward. Assuming a fixed subduction geometry Ellis and others (1999) have shown that the amount of material which enters the subduction channel compared to the amount which does not subduct controls the geometry of the accretionary wedge, whether double vergent or not. The behavior of the subduction channel can be further modified by the retreat of the subduction zone. A limitation of Ellis and others' model is that the imposed condition of fixed upper plate geometry seems incompatible with the low strength of the subducted material (Burov and others, 2001).

Upper parts of the accretionary complex in the internal zones are affected primarily by extension and to a lesser extent by erosion, which remove the overburden. Extensional detachments are restricted to the upper 10 to 30 kilometers of the accretionary complex and they root along brittle-ductile shallow-dipping shear zones near the brittle-ductile transition. They dip either outward or inward, or again along strike. The accretion front progresses toward the foreland for three possible reasons: during the underthrusting of the lower plate the sedimentary and crustal material is progressively accreted above a decollement, body forces tend to enlarge the surface of the thickened region, and in some cases the subducting slab is retreating.

Maximum P-T conditions depend on various factors such as the velocity of convergence, the nature of the accreted material, the efficiency of the overburden removal and the heat flux from the underlying mantle, which can vary strongly if the slab or the orogenic root detach. Most of the lower crust is delaminated and subducted.

In the case of slab retreat the crust will collapse and extension will lead to the formation of a backarc basin. During retreat the internal parts of the accretionary complex will be integrated in the backstop and the extensional shear zones that separate the backstop from the subduction channel will be abandoned and new ones farther outward will accommodate exhumation.

Ultra high pressure and low temperature metamorphic conditions are encountered deep along the subduction channel. Rates of exhumation are high at large depth in low viscosity material and lower near the surface where the viscosity is higher. Several levels of circulation are present, one corresponding to the accretionary complex with slow exhumation and one deeper along the subduction channel with fast exhumation rates. The upper circulation level leads to the exhumation of blueschists, the deep one to the exhumation of eclogites and ultra-high-pressure eclogites. The internal limit of the subduction channel is an "extensional" shear zone that accommodates fast exhumation. The low viscosity of the subducted material allows the circulation within the subduction channel once it is brought to high-temperature conditions. This condition can be met in some belts such as the Norwegian Caledonides where the thermal regime is significantly warmer than in Mediterranean examples and where partial melting during exhumation has been recognized (Labrousse and others, 2001). In the Mediterranean belts partial melting is rarely recognized during the exhumation of blueschists and eclogites. When melting occurs it corresponds to late thermal events such as in the case of Naxos in the center of the Cyclades. One thus has to envisage a different mechanism for the Mediterranean cases. The possible role played by serpen-

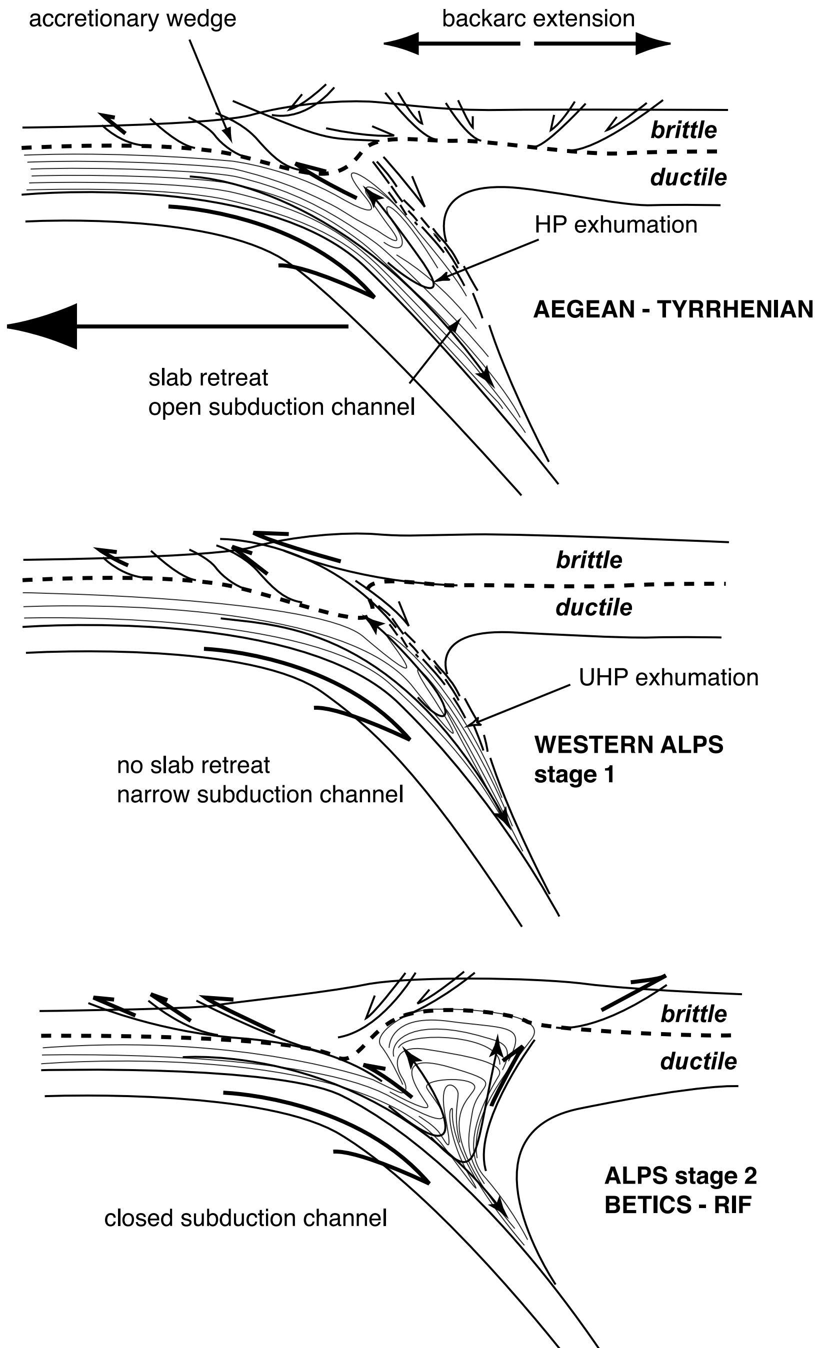

Fig. 20. Three possible situations encountered in the Mediterranean region and the main boundary conditions. These models incorporate some ingredients from the numerical models of Burov and others (2001), of Ellis and others (1999), and Pfiffner and others (2000). The main features are the circulation of rocks inside the accretionary complex above ~40 km and in the subduction channel below that depth. Low viscosity of the heated subducted material allows circulation. The material can be either basement material like in Burov and others (2001) and Labrousse and others (2001), but this would necessitate thermal conditions quite unlikely for the Mediterranean environments, or it can be softer material such as serpentinite (Guillot and others, 2000) or metasediments. A ductile "extensional" shear zone forms along the boundary between the subduction channel and the upper plate which accomodates the upward flow of exhuming material. Extension is observed in the upper part of the accretionary complex (typically above the brittle-ductile transition and within this transition), which removes the overburden with the assistance of erosion. The width of the subduction channel is partly governed by the amount of slab retreat that tends to open it.

tinite has been proposed for the Himalayas (Guillot and others, 2000) and the Alps (Schwartz, 2000). The ubiquitous presence of large metapelitic units suggest low-resistance phases such as white micas may play a similar role in facilitating the circulation of rocks within the subduction channel. They play an important role in localizing strain in the continental crust (Gueydan and others, 2003) and may play the same role in the subduction channel as it is well known that they can resist to very high-pressure conditions (Schreyer, 1995; Chopin and Schertl, 2000).

The importance of the brittle-ductile transition (BDT) as an important rheological boundary has seldom been considered in models of subduction complexes. In our model an entirely different behavior is supposed below and above this transition, which localizes much of the extensional and contractional shear. The accretionary complex possesses a brittle-ductile transition as modeled in Williams and others (1994). The slope corresponds to the brittle part of the wedge and the plateau to the ductile part. In Williams and others' model the transition between the slope and the plateau is characterized by a steeper slope due to an offset between the BDT's of the decollement and of the wedge. Our model is too simple to consider this offset. Fast circulation occurs below the brittle-ductile transition where the material has a low viscosity. In Aegean-type subduction, fast exhumation of HP eclogites occurs in the upper part of the subduction channel below a deep extensional shear zone. Once the exhumed material has entered the brittle-ductile transition and the brittle part of the accretionary wedge, it is exhumed beneath greenschist facies shear zones and brittle detachments that root into the brittle-ductile transition of the backarc region. In the case of the western Alps where UHP parageneses have been exhumed, most of the material detaches from the subducting slab before reaching great depths but some small units keep going down and start to return back to the surface at greater depth. In a first stage the orogen is single vergent (45-50 Ma) and the oceanic material is exhumed below an east-dipping shear zone in the lower parts of the accretionary complex. When the orogen becomes double vergent the exhumation of the metamorphic domain induces the formation of outward-dipping extensional shear zones (Engadine, Dora-Maira, Viso) that root in the brittle-ductile transition zone.

This model has some common characteristics with published models. It contains kinematic ingredients of the model proposed by Platt (1986) for the upper part of the prism. Exhumation is mostly controlled by shallow dipping extensional shear zones, which remove the overburden. The large-scale extensional shear zone that separates the subduction channel from the backstop is similar to the normal fault seen in Chemenda and others' model (1995) during the rigid exhumation of a buoyant rigid crustal slice. It is similar to the thermo mechanical model proposed by Burov and others (2001) with two circulation levels and buoyant exhumation of low viscosity material in the subduction channel, as well as by folding of lithological interfaces in the subduction channel that can be compared to Penninic nappes. In Pfiffner and others (2000) such folding is related to the subduction of heterogeneities. In fact when the shape of the subduction channel is not fixed folding can occur without such heterogeneities simply due to shear along the roof of the subduction channel and buoyant uplift of light material. Burov and others' (2001) model is characterized by several levels of circulation, an upper level where a classical corner flow occurs in the accretionary complex and a lower crustal chamber at a depth of 100 kilometers separated from the upper level by a narrow channel. In the upper accretionary complex the material is excised from the lower descending plate by the shear flow created by the overriding plate and exhumed by the positive buoyancy of the material. In the lower crustal chamber the material goes up because of high buoyancy due to heating and due to the upward shear created by the overriding plate. Between the two systems a narrow subduction channel allows the downward and upward motion of material. The width of this channel can vary during the convergence process. Because the channel width is modified by the stress and resistance variations on either side, no significant overpressures can build up. The velocity of exhumation depends upon the buoyancy of the subducted material and thus upon the degree of eclogitization of the crust. Phase transitions due to continuous P-T changes are not calculated in Burov and others (2001) because the degree of recrystallization during eclogitization is unknown. Only two end-members were considered: a low buoyancy case where all the subducting material is converted to a high density, and a high buoyancy case where the density of the subducted material is kept unchanged. This procedure tentatively models a full eclogitization case versus no eclogitization. Phase changes are taken into account in a more complete way in the kinematic models of Henry and others (1997), but these models do not solve the mechanical part of the problem. Burov and others (2001) predict that low eclogitization leads to faster exhumation rates because buoyancy-driven forces become more important. The upward motion of the exhumed material creates a localized shear zone along the lower boundary of the backstop even at large depth, allowing for the formation of deep "extensional" shear zones. In this model buoyancy forces become predominant at great depth while only forces related to the geometry of the model (corner flow) are active in the upper part. Deep exhumation is thus faster than more superficial exhumation. In this model the removal of the overburden is achieved by erosion only. This is a significant limitation of the numerical models that cannot handle at the same time lithospheric-scale processes and detailed strain localization at the scale of the upper crust.

CONCLUSIONS

A comparative study of Mediterranean mountain belts is used to discuss the mechanisms of exhumation of high-pressure and ultra-high-pressure rocks focusing on syn-orogenic exhumation. The Mediterranean examples, within an overall convergent zone, show a variety of tectonic contexts due to variations in the rates of convergence, rates of slab retreat, available space, frontal or oblique convergence, and various stages of maturation of accretionary complexes that can be used as natural experiments. Two types of behavior are observed, one with steady-state subduction of oceanic and continental units with a constant thermal structure of the subduction complex leading to a single P/T gradient and similar P-T paths throughout the evolution of the belt (western Alps), and one non-steady-state with changes in the geodynamic context such as slab retreat (Aegean) producing different P-T evolutions through time. Most structures were formed during the exhumation stage and are often associated with syn-orogenic detachments. The most important part of exhumation occurs along the subduction plane following cold P-T paths from the depth of eclogites (or UHP eclogites) to the depth of the blueschist or greenschist facies. UHP rocks do not seem to occur in retreating subduction contexts because an easy circulation in an open subduction channel favors an early detachment of sediments from their basement. Subducted sediments also act as lubricants of the subduction channel so that the basement is not affected by a strong shearing and is not involved in the return flow. This early exhumation is rapid and the thermal regime in the subduction channel is partly controlled by kinematic boundary conditions such as the velocity of convergence and the velocity of slab retreat. Final exhumation occurs within the accretionary complex at a much slower rate below extensional detachments. The removal of the overburden is achieved primarily by extension in the upper part of the accretionary complex. Extensional faults and shear zones root in the brittle-ductile transition of the accretionary complex. Some deeper "extensional" shear zones represent the deformation along the roof of the subduction channel. We discuss a model with several levels of circulation of subducted material and compare it with available thermo mechanical models. We conclude that fully coupled thermo-mechanical models with no *a priori* fixed geometry of the subduction channel (for example Burov and others, 2001) best fit geological observations. Further research should concentrate on introducing in such models progressive density and rheological changes due to metamorphic reactions.

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