**Zeitschrift:** Eclogae Geologicae Helvetiae

**Herausgeber:** Schweizerische Geologische Gesellschaft

**Band:** 89 (1996)

**Heft:** 1

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**DOI:** https://doi.org/10.5169/seals-167894

Eclogae geol. Helv. 89/1: 43–80 (1996)

0012-9402/96/010043-38 $1.50 + 0.20/0
Birkhäuser Verlag, Basel

# Did the Western Alps develop through an Oman-type stage? The geotectonic setting of high-pressure metamorphism in two contrasting Tethyan transects

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Dedicated to the memory of Carlo Sturani

*Key words:* HP-LT metamorphism, tectonics, subduction, obduction, collision, ophiolites, Western Alps, Oman, Tethyan belts

ABSTRACT

In the Western Alps, the Penninic metamorphism was considered as the result of a Cretaceous obduction similar to the Oman Mountains one. The analysis of the geodynamic evolution of both these belts leads us to abandon this hypothesis.

At the onset of convergence between Africa and Eurasia, the Oman-Makran transect involved a broad ocean, partly old, partly very young (< 5 Ma) and with a thick crustal section. Two subduction zones *sensu lato* accommodated convergence, one being of the andean type, the other one being intra-oceanic and shallow-dipping (detachement type) and resulting in the obduction of the young oceanic lithosphere onto the Arabic margin. High-pressure, low-temperature metamorphism was effective in both zones during the Late Cretaceous.

In the Western Alps, sedimentation on the internal crystalline massifs (ICM, i.e. European distal margin) and the adjacent oceanic lithosphere lasted at least until the Late Cretaceous, most likely the Paleocene and locally the Eocene (age of the allochthonous flyschs of Ligurian origin). Eclogitic metamorphism in the Sesia zone (Austro-Alpine, Apulian margin) is Mid-Cretaceous in age. The age of eclogitic metamorphism in the ophiolites and ICM was long considered as Eoalpine (Cretaceous) on the basis of K/Ar and Ar/Ar isotopic data; it must be Paleocene to Late Eocene respectively according to recent isotopic data (Sm/Nd and "SHRIMP") and in agreement with the stratigraphic record.

In contrast to the Tethys ocean between Oman and Makran, the Ligurian ocean at the onset of convergence was narrow (400 km?), old (50 Ma), and had a thin crust or no crust at all through tectonic denudation of the mantle. The Alpine ophiolites, devoid of high-temperature metamorphic sole, are minor syn-collisional slivers. The metamorphism in the continental Penninic crust is not obduction-related but collision-related. The P-T-t paths in Oman and the Penninic Alps show similarities, they are both concave toward the pressure axis and both show rapid initial cooling. However, the final cooling is slower in Oman (< 10 °C/Ma) than in the ICM (15–20 °C/Ma). In conclusion, the idea of considering the Oman obduction as a transient but necessary stage of the evolution of collision belts must be abandoned. Such an obduction stage never occurred during the evolution of the Western Alps.

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RESUME

Dans les Alpes occidentales, le métamorphisme pennique a été considéré comme le résultat d'une obduction crétacée semblable à celle de l'Oman. L'analyse de l'évolution de ces deux chaînes nous amène à abandonner cette idée.

Le transect Oman-Makran présentait au début de la convergence Afrique-Eurasie un océan large, en partie ancien, en partie très jeune (< 5 Ma) et à croûte épaisse. Deux zones de subduction *sensu lato* ont accommodé la convergence, l'une de type andin, l'autre intra-océanique et à faible pendage (détachement), évoluant en obduction. Le métamorphisme HP-BT est présent dans les deux zones au Crétacé supérieur.

Dans les Alpes occidentales, la sédimentation s'est poursuivie sur les massifs cristallins internes (ICM, marge européenne distale) et sur la lithosphère océanique adjacente jusqu'au Crétacé supérieur au moins, et même très probablement jusqu'au Paléocène et par place jusqu'à l'Eocène (âge des Flyschs allochtones d'origine liguro-piémontaise). L'âge du métamorphisme éclogitique de Sesia (Austro-Alpin, marge apulienne) est Crétacé moyen. L'âge du métamorphisme éclogitique dans les ophiolites et les ICM, considéré comme éoalpin d'après de nombreux résultats K/Ar et Ar/Ar, doit être resitué au Paléocène (ophiolites)-Eocène (ICM), d'après les données isotopiques récentes (notamment Sm/Nd et «SHRIMP») et en accord avec la stratigraphie.

Contrairement à l'océan téthysien entre Oman et Makran, l'océan ligure était, au début de la convergence, étroit (400 km?), ancien (50 Ma) et à croûte mince ou absente (dénudation tectonique du manteau). Les ophiolites alpines, dépourvues de sole métamorphique HT, sont des écailles mineures syncollisionelles. Le métamorphisme de la croûte continentale pennique n'est pas un métamorphisme d'obduction, mais un métamorphisme de collision. Les chemins P-T-t omanais et penniques montrent des formes semblables, concaves vers l'axe des pressions, et le refroidissement est d'abord rapide dans les deux cas. Par contre, la fin de l'exhumation est plus lente en Oman (< 10 °C/Ma) que dans les ICM (15–20 °C/Ma). En conclusion, il faut renoncer à considérer le transect Oman-Makran comme l'image d'un stade transitoire toujours présent dans l'histoire des chaînes de collision. Un tel stade est en particulier absent de l'histoire des Alpes occidentales.

## 1. Introduction

The Oman Mountains are famous for their giant ophiolitic nappe thrust onto the Arabian margin during Late Cretaceous (Allemann & Peters 1972, Glennie et al. 1974). This obduction event (Coleman 1971) followed some sort of intra-oceanic subduction *sensu lato* (intra-oceanic detachement) and occurred without continental collision. On the Oman transect, Africa is still separated from Eurasia by a relic of Tethyan lithosphere (Gulf of Oman) which is being consumed by the Makran subduction (Fig. 1). The rest of the peri-Arabian ophiolitic belts (Zagros, Taurus) developed through an Oman-type stage before being crushed by the Arabia-Eurasia collision during Tertiary times (Ricou 1971). This nourished the idea that obduction could constitute a transient, initial stage in the evolution of many collision belts (Coleman 1971, Dewey 1976). Considering the Alpine collision belt itself, Mattauer & Proust (1976) suggested that high-pressure–low-temperature (HP-LT) metamorphism would have been controlled in the European margin by its transient subduction below oceanic material, during the blocking stage of an intra-oceanic subduction. Such a hypothesis was greatly strengthened by the discovery of HP-LT metamorphism in the Arabian material below the Oman ophiolite (Boudier & Michard 1981, Lippard 1983, Michard 1983, Le Métour et al. 1986, Lippard et al. 1986, Goffé et al. 1988). Thereby it was proved that HP-LT metamorphism can affect continental material in relation with an obduction event and before any collision does occur. Accordingly, Debelmas et al. (1983) explained the HP metamorphism of the European margin by the "obduction of a huge ophiolitic slab", 80 Ma ago. Mattauer et al. (1987), although without reference to the Oman case, suggested that eclogitic metamorphism in the internal

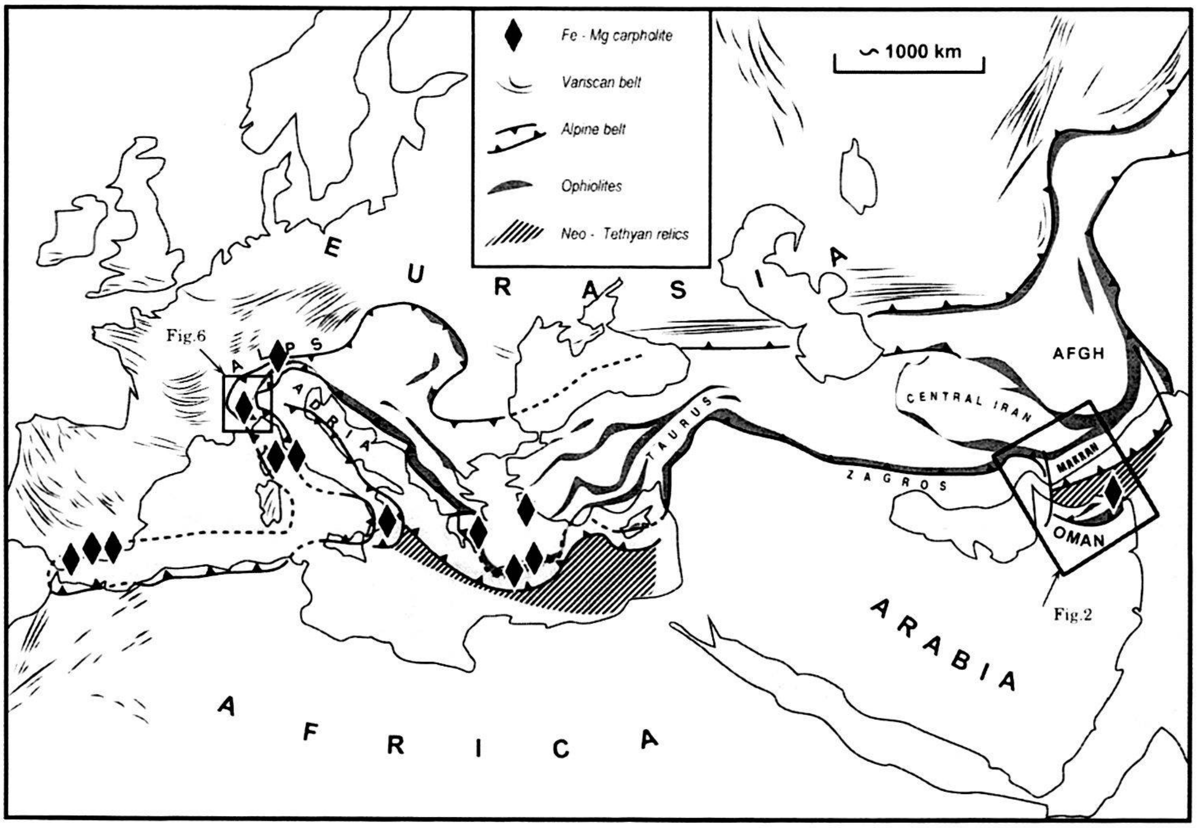



Fig. 1. Location of the Western Alps and Oman-Makran transects in the Tethyan (Alpine) mountain belt. Ophiolite stripes after Knipper et al. (1986). The Fe-Mg carpholite occurrences (mainly after Goffé 1984, Goffé & Chopin 1986, Goffé et al. 1988) visualize the continuity of blueschist facies metamorphism along most of the belt, included Oman.

crystalline units of the Western Alps was a result of subduction of the European continental margin below the oceanic lithosphere of the Piemont-Ligurian ocean during Late Cretaceous time. According to Mattauer et al. (1987), even the Eoalpine metamorphism of the Sesia-Lanzo zone would have occurred in this obduction-type setting. This was at odds with the hypothesis of Dal Piaz et al. (1972), more widely accepted (e.g. Frisch 1979, Homewood et al. 1980, Laubscher & Bernoulli 1982, Fry & Barnicoat 1987), which considers an eastern subduction zone at the Ligurian ocean-Apulian block boundary as the only geotectonic site for HP-LT metamorphism prior to the Tertiary (Mesoalpine) collision.

In an attempt to resolve these contradictions, and taking advantage of the comparison with the Oman-Makran transect, Avigad et al. (1993) suggested a two-subduction model for the Eoalpine evolution of Western Alps, with i) an east-dipping, intra-oceanic subduction formed within the Ligurian (Neo-Tethyan) ocean, then encroaching on the European paleomargin, and ii) another east-dipping subduction along the Ligurian ocean/Apulian block (Adria) boundary, carrying at depth slivers of the Apulian paleomargin. This model was designed to explain the eclogitic metamorphism of the Upper Penninic (Dora-Maira) and Austro-Alpine (Sesia) continental nappes, supposed to be of Eoalpine age in both nappe complexes. The aim of the present paper is to revisit the

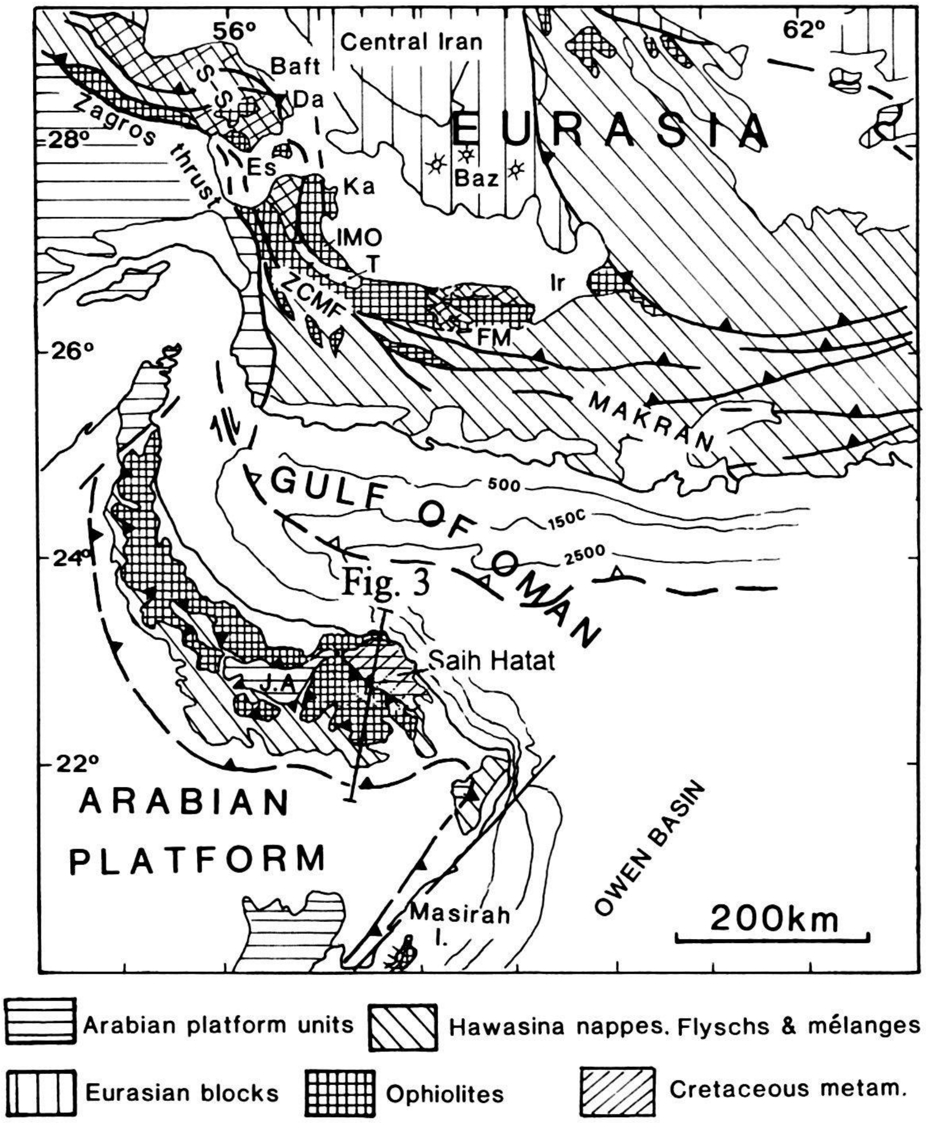



Fig. 2. Tectonic map of Oman and southeastern Iran, after Coleman (1981) and Glennie et al. (1990), with location of Cretaceous metamorphic rocks in Oman after Michard et al. (1994), and in Makran after Delaloye & Desmons (1980) and Okay (1989). Baz: Bazman volcanoes; Da: Dargaz; Es: Esfandagheh-Dowlatabad; FM: Fanuj-Maskutan area; Ir: Iranshar; IMO: Inner Makran ophiolite; J.A.: Jabal Akhdar; Ka: Kahnu ophiolites; S-S: Sanandaj-Sirjan zone; T: Tiab nappe; ZCMF: Zone of Coloured Mélanges and Flyschs.

comparison between the Oman and Western Alps transects and to discuss some critical points of the tectonic model by Avigad et al. (1993). Firstly, where was the equivalent of the Gulf of Oman in the Late Cretaceous Alpine realm, i.e. where is the suture in the present-day Western Alps cross-section? Reconstructions at odds with that retained by Avigad et al. (1993) have been recently developed by Mattauer et al. (1994) and Stampfli & Marchant (1995). Secondly, is the timing of the alleged paired subductions correctly constrained? An increasing amount of geochronological data tends to support a Mesoalpine (instead of Eoalpine) age for the eclogitic metamorphism in the European paleomargin (Tilton et al. 1991, Gebauer et al. 1992, Gebauer 1995). Does the Oman-Alps

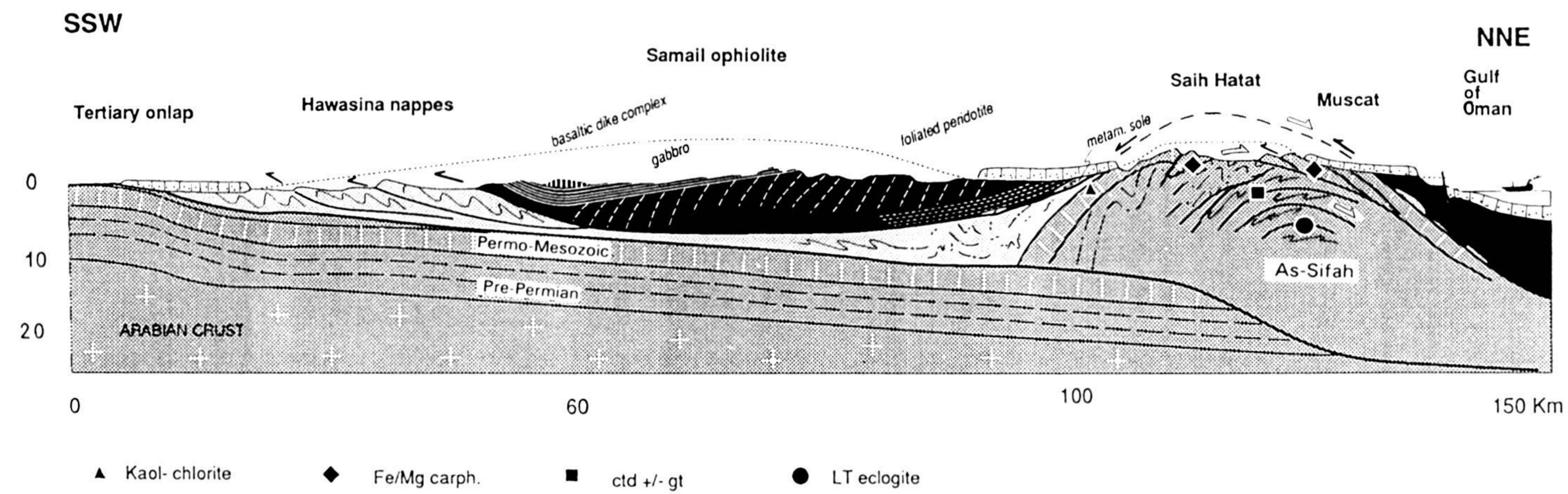



Fig. 3. Schematic cross-section of the Oman Mountains (Fig. 2 for location), slightly modified after Goffé et al. (1988) and Nicolas (1989, p. 152). Bold arrows: early thrusts; empty arrows: late orogenic displacements (Michard et al. 1994). The steeply dipping foliation in the peridotite is related to mantle flow perturbations along transform faults, prior to the detachment of the ophiolite. Foliation in the metamorphic sole parallels that of the juxtaposed mylonitic periodotites. The structure below the 8 km-depth is highly speculative.

comparison still hold if we accept this new datation of the Penninic metamorphism? We answer hereafter negatively and we abandon our earlier proposal (Avigad et al. 1993). A review of the stratigraphic data that constrain the palaeogeographic and tectono-metamorphic evolution of both studied transects will be presented. A model for the tectonic evolution of the Western Alps will be inferred, substituting that of Avigad et al. (1993) and Michard et al. (1993). In the following, we will use the Odin & Odin (1990) time scale.

## 2. Evolution of the Oman-Makran transect

The Gulf of Oman is a deep oceanic embayment inserted between the Arabian and Iranian (Lut) continental blocks (Fig. 2). Widely connected eastward to the Jurassic (?)-Cretaceous Owen basin, the Gulf of Oman is arguably underlain by a Cretaceous oceanic lithosphere covered with a ca. 4 km-thick sedimentary pile (Jacobs & Quittmayer 1979, White 1982, Mountain & Prell 1990). Both the southwestern (Oman) and northeastern (Makran) margins of the gulf show wide ophiolitic outcrops. The stratigraphy of the ophiolites and that of the associated sedimentary units, summarized hereafter, supports their origin from Neo-Tethys, a relic of which would correspond to the Gulf of Oman (Coleman 1981, Dercourt et al. 1993).

### *2.1 Stratigraphic constraints*

#### 2.1.1 Oman Mountains

The Oman ophiolite sequence (Samail nappe, Fig. 3, 4F) documents an Early Cretaceous (Aptian-Albian and possibly older) to early Late Cretaceous (Cenomanian) oceanic accretion (MORB gabbros and lower basalts V1 in Fig. 4F) followed by a differentiated volcanism, dated from Turonian to early Campanian (V2, V3) (Ernewein et al. 1988, Béchennec et al. 1992). The amphibolites of the metamorphic sole, which record the early, intra-oceanic detachment of the future ophiolite, gave K/Ar ages in the range 100–93 ± 4 Ma (late Albian-Turonian), while the lower, greenschist facies slivers yielded Coniacian to Campanian ages (Montigny et al. 1988, Boudier et al. 1988). The youngest, transitional to calc-alkaline volcanism has been considered as the signature of a subduction setting and the Oman ophiolite interpreted as a back-arc lithosphere detached above an intra-oceanic subduction zone (Lippard et al. 1986). Alternatively, a model of detachement at a ridge axis has been postulated (Ceuleneer 1986, Boudier et al. 1988, Michard et al. 1991), the ophiolite having formed by a typical ridge accretion process before being thrust southwestward over the adjoining oceanic crust and capped by a seamount volcanism (V2, V3). In any case, a northeast-dipping shear surface, coeval with a restricted arc-type volcanism in the upper plate, i.e. an intra-oceanic subduction *sensu lato*, operated from Cenomanian-Turonian to Campanian time below the future Samail ophiolite (Fig. 5A). The lower plate of this intra-oceanic, weakly-dipping subduction zone consisted of Permian-Triassic oceanic (Neo-Tethyan) lithosphere, bounding to the northeast the Arabian, Permian-Triassic passive margin. This is documented by the Permian-Triassic alkaline to MORB basalts and pelagic sediments in the Hawasina nappes (Bernoulli et al. 1990b, Béchennec et al. 1990, 1992). These nappes occur beneath and in front of the Sa-

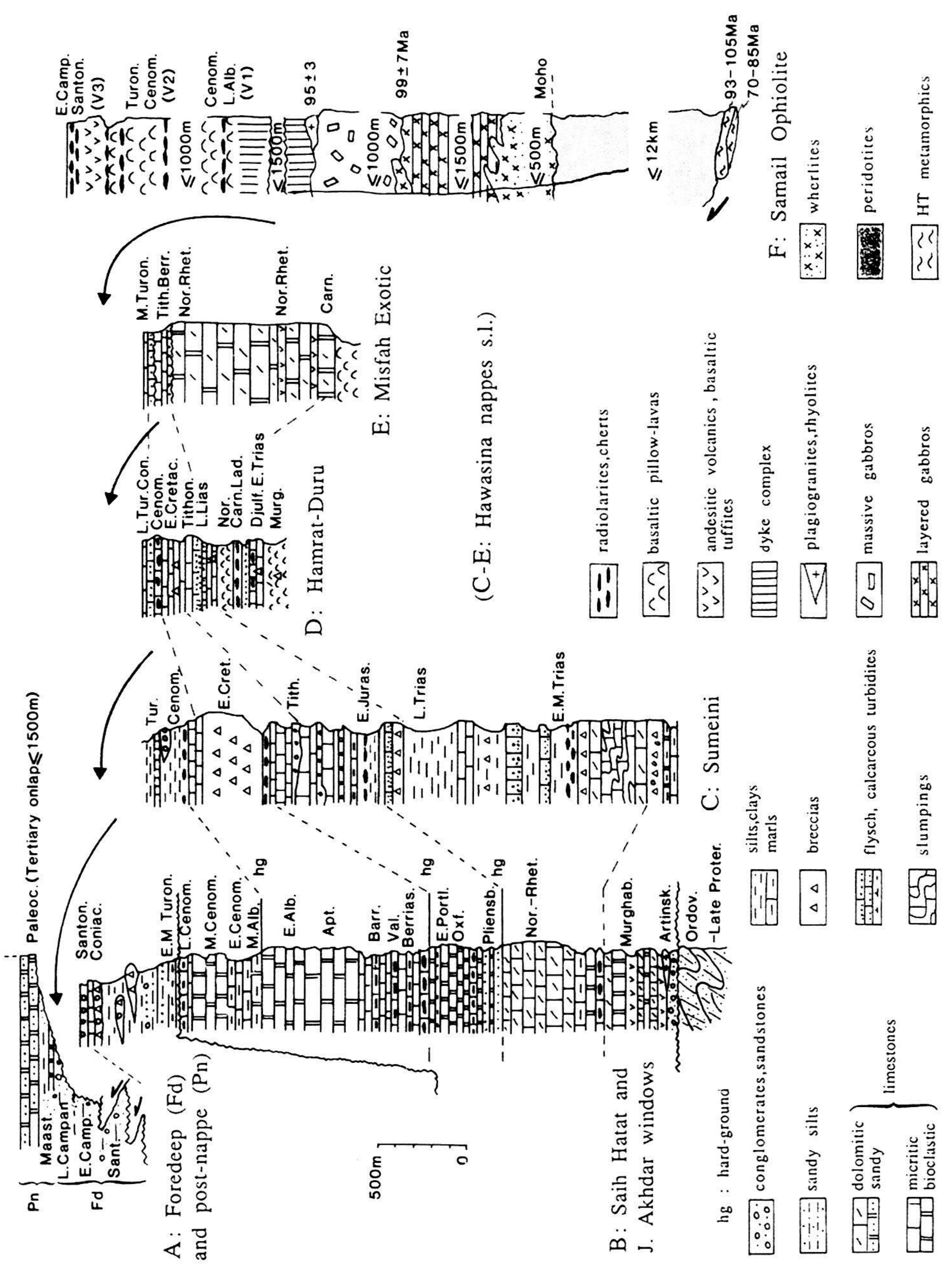



Fig. 4. Stratigraphy of the main tectonic units of the Oman Mountains. References as follows; A: Warburton et al. (1990), Nolan et al. (1990), Béchennec et al. (1992). B: Rabu et al. (1990), Béchennec et al. (1992), Pillevuit (1993), C: Watts (1990), Pillevuit (1993). D: Béchennec et al. (1990, 1992). E: Béchennec et al. (1990, 1992). F: Boudier et al. (1988), Ernewein et al. (1988), Montigny et al. (1988), Béchennec et al. (1992).

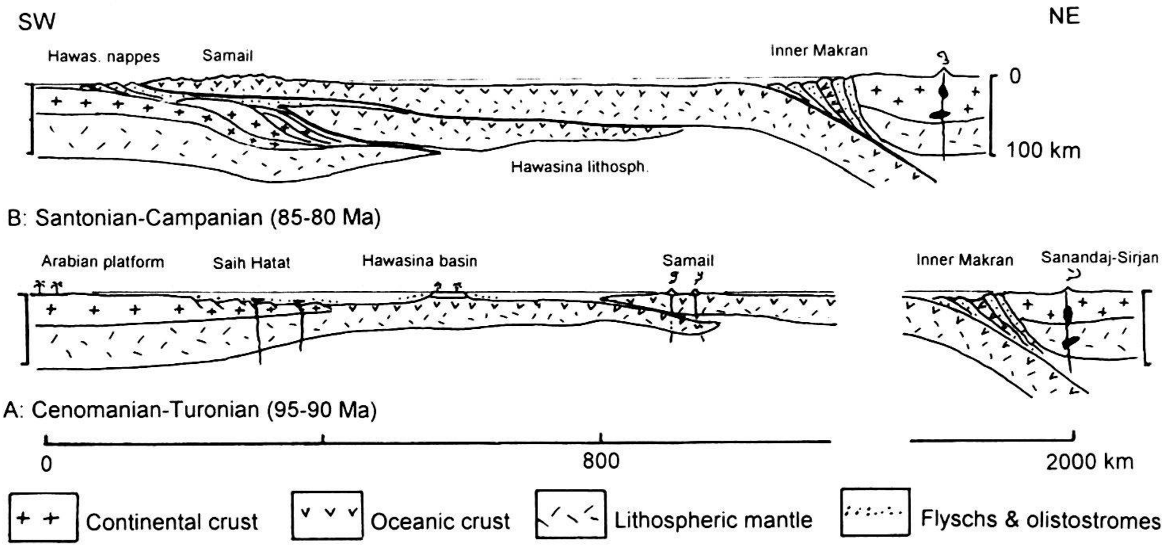



Fig. 5. Restored cross-sections of the Oman-Makran transect during the Early Turonian (A), and the Santonian-Campanian (B). Arabian margin and Hawasina after Bernoulli et al. (1990b), Béchennec et al. (1990), Pillevuit (1993). Samail detachement after Boudier et al. (1988). Makran accretionary prism according to Kazmin et al. (1986) and Glennie et al. (1990). Position of Samail with respect to Arabia after Perrin et al. (1994); position of Makran with respect to Arabia after Dercourt et al. (1993). Duplication of the oceanic lithosphere thrust onto the Arabian margin after Nicolas (1989, p. 88) and Michard et al. (1994).

mail nappe and were progressively bulldozed by the oceanic upper plate during its southwesterly thrusting. According to Stampfli et al. (1991) and Pillevuit (1993), the Hawasina basin opened during the Early Permian (Artinskian) and the Arabian paleomargin would have been of the flexural type (upper plate in a simple shear rifting process). The Hawasina basin *sensu lato* included deeps and atoll-type highs (Fig. 4D, E). Slope sequences (Fig. 4C; Watts 1990) were deposited on attenuated continental crust close to the Arabian margin, which was characterized by shallow-water carbonates (Fig. 4B; Rabu et al. 1990). The Hawasina basin (except its innermost part) was to retain its main morphological features until the Coniacian. Encroachment of the Samail wedge on the innermost Hawasina lithosphere was probably responsible for the break in carbonate sedimentation on the platform margin after late Cenomanian. The Samail nappe reached the margin during Santonian-Campanian time, with development of a foredeep as a response to loading (Fig. 4A; Warburton 1990). Post-nappe, shallow-water sediments were deposited onto the Samail nappe from Maastrichtian onward (Nolan et al. 1990). The compressional evolution of the former intra-oceanic subduction, converted into obduction during Santonian–early Campanian time, was over by that time.

### 2.1.2 Makran

On the other bank of the Gulf of Oman, ophiolites crop out along two juxtaposed stripes (Knipper et al. 1986, Glennie et al. 1990). The outer (southeastern) stripe corresponds to a zone of coloured mélange and flyschs (ZCMF). The rock assemblage of the coloured mélange has many similarities with the Hawasina-Samail nappes (faunas, lithologies), but it has passed through higher metamorphism and is imbricated with Maastrichtian and early Tertiary flyschs. To the south, the ZCMF is bounded by post-Oligocene flyschs deformed in the accretionary prism of the active Makran subduction (Jacobs & Quittmeyer 1979, White 1982). The ZCMF prolongates northwestward into the Zagros crush zone-Eastern Taurus suture zone, where ophiolites are similar to those of Oman (Ricou 1971, Knipper et al. 1986). The Inner Makran ophiolites are thrust over the ZCMF, either directly or through a carbonate nappe (Tiab zone), or (western part of the belt) through metamorphic basement units belonging to the southern tip of the Sanandaj-Sirjan zone (Glennie et al. 1990). The rocks of the upper sub-zone of Inner Makran ophiolites, or Kahnu ophiolites, are less metamorphosed and covered by tintinnid and radiolaria-bearing Tithonian to Berriasian calcareous mudstones (Glennie et al. 1990). The Inner Makran ophiolites bound the southern part of the Lut (Central Iran) continental block and trend north into the Cretaceous Baft-Nain ophiolite belt. The latter belt is pinched between the Sanandaj-Sirjan and Central Iran continental blocks (Knipper et al. 1986). Early and Late Cretaceous dacitic lava flows occur in the Sanandaj-Sirjan zone (Dargaz area). Late Cretaceous-Tertiary arc-type volcanism occurs along the inner border of the same zone and along the southeastern border of Central Iran (Bazman area). These extrusives may have been associated with the subduction of Neo-Tethys under its northeastern flank since Early Cretaceous time. The Lut block would have provided the source for much of the flysch deposition south of the trench (McCall & Kidd 1982).

### *2.2 Geotectonic setting of HP-LT metamorphism*

The main elements of the plate-tectonic history in the Oman-Makran sector can be summarized as follows (Kazmin et al. 1986, Glennie et al. 1990, Dercourt et al. 1993). During the Permian and Triassic, Central Iran-Sanandaj-Sirjan separated from Gondwana to create a Neo-Tethys branch. Then Arabia drifted relatively southwestward, away from the Neo-Tethys spreading axis. In contrast, the northern continental block came closer to the ridge due to subduction of the Neo-Tethys northern flank (Fig. 5A). Splitting of the northern continental block occurred at about the same time (Early Cretaceous) as the oceanic ridge jumped southwestward, into a position close to the Arabian paleomargin. Easterly, an intra-oceanic detachement, or shallow-dipping subduction was initiated close to the ridge axis in the Albian-Cenomanian (Boudier et al. 1988). Obduction of the Hawasina-Samail nappes (Oman) and Coloured Mélange (Zagros) ceased in the late Campanian (Warburton et al. 1990, Nolan et al. 1990). Therefore, during the Late Cretaceous, the Arabia–Central Iranian blocks convergence was accommodated by two subduction zones *sensu lato*: a typical Benioff-type subduction to the east and a shallow-dipping, intraoceanic, later on continental subduction to the west (obduction). The width of the Permian-Cretaceous oceanic lithosphere to be consumed was between 2000 km (Le Pichon et al. 1988) and 3000 km (Dercourt et al. 1993). Paleomagnetic studies evidenced large rotations of the Lut block (Westphal et al. 1986) and Samail nappe (Thomas 1991, Perrin et al. 1994) during plate convergence.

HP-LT metamorphism developed in both subduction zones during Late Cretaceous. The Inner Makran ophiolites include a lower sub-zone of metamorphic peridotite-gabbro masses thrust over the Sanandaj-Sirjan zone prior to deposition of Eocene sandstones.

Amphibole schists and micaschists from these units were dated (K/Ar, micas) at 98 to 125 ± 4 Ma in the Esfandageh-Dowlatabad area (Delaloye & Desmons 1980). In the Fanuj-Maskutan and Iranshar area, at the southern margin of the Central Iran block, blueschist-facies metabasalts and quartzitic schists are found in coloured mélanges. The HP-LT rocks gave K/Ar mineral ages between 81 ± 2 and 101 ± 9 Ma, which fall in the same age range as the amphibolite and greenschist-facies rocks from the same area or from the Sanandaj-Sirjan zone further to the NW (Delaloye & Desmons 1980). Higher-grade rocks are not exhumed in the Makran area.

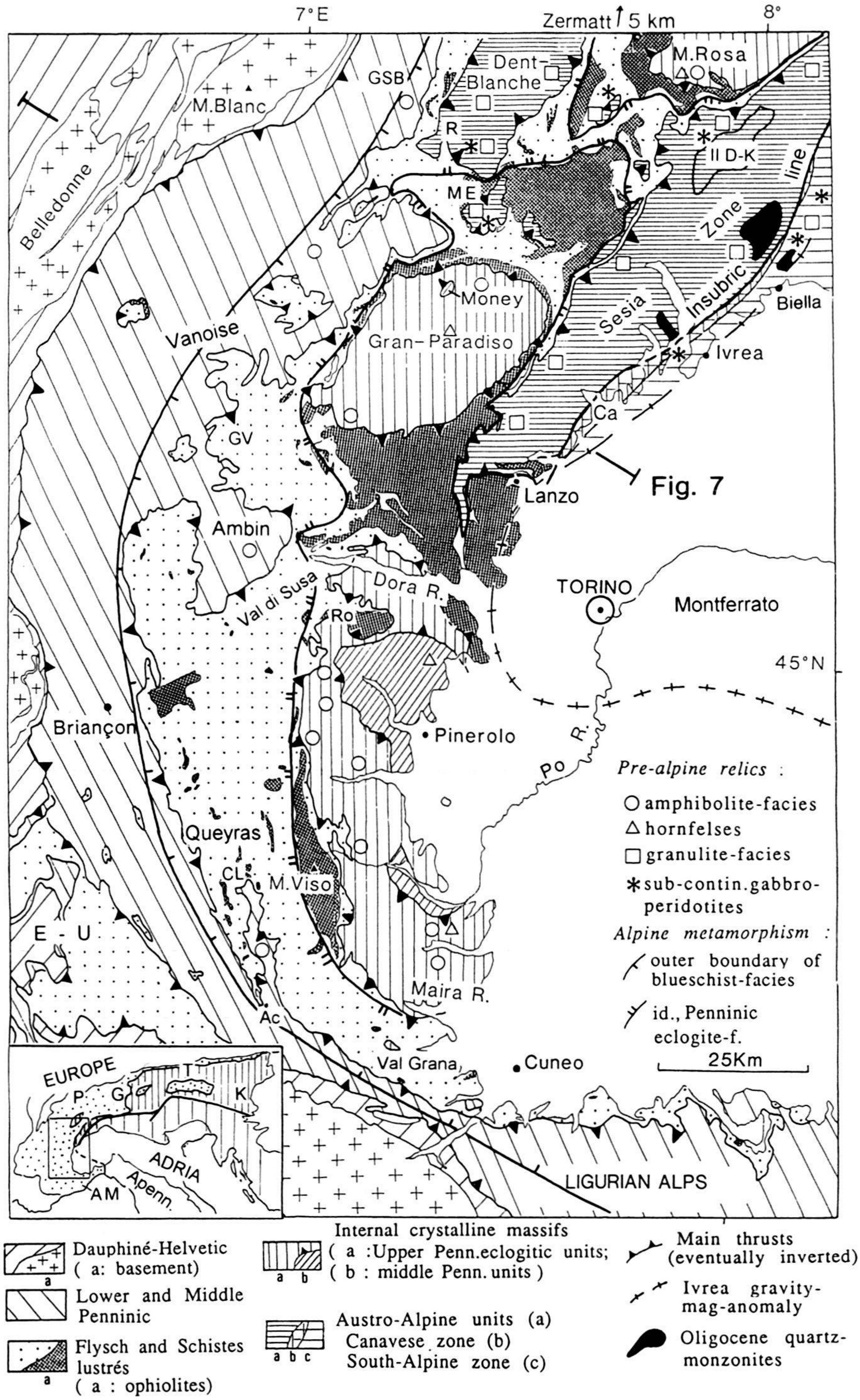
On the southern bank of the Tethys relic (Gulf of Oman), HP-LT metamorphic rocks crop out in the Saih Hatat window, in a tectonic position below the Samail ophiolite and the associated Hawasina nappes (Lippard et al. 1986, Le Métour et al. 1986, 1990; Goffé et al. 1988, Béchennec et al. 1992). The HP rocks represent metamorphic equivalents of Proterozoic-Paleozoic basement and Permo-Mesozoic cover rocks from the Arabian paleomargin. They consist of i) low-grade, carpholite- or lawsonite-crossite-bearing metasediments or metavolcanics in the uppermost units; ii) high-grade, eclogite-facies micaschists and metabasites in the lowermost units and iii) intermediate grade schists and metabasites in the intervening units (Fig. 3; Goffé et al. 1988, Figs. 2, 3). P-T conditions range from about 10 kbar–300 °C (Muscat nappes) to 23 kbar–500 °C (As-Sifah unit) (Goffé et al. 1988, Wendt et al. 1993, El Shazly 1994, Searle et al. 1994). HP metamorphic rocks of the Saih Hatat reveal complex K/Ar and Ar/Ar results. Typical K/Ar ages on phengite range from 80 Ma to 131 Ma (Montigny et al. 1988). Ar/Ar step heating of white micas yielded rather intricate, convex-upward age spectra (Montigny et al. 1988, El-Shazly & Lanphere 1992), with weighted mean "plateau" ages of 106–111 Ma. However, the Saih Hatat upper units contain HP-metamorphic Coniacian-Santonian rocks, while post-orogenic late Campanian-Maastrichtian sediments are seen on the overlying Hawasina-Samail nappes. Then HP-recrystallization of the Saih Hatat upper units occurred during Santonian–early Campanian times (from 85 ± 2 to 75 ± 2 Ma). This is also likely to be the age of HP-recrystallization of the deeper units of the same metamorphic window (Lippard et al. 1986, Le Métour et al. 1986, Goffé et al. 1988), while the occurrence of older K/Ar and Ar/Ar ages can be explained by the presence of extraneous Ar (Montigny et al. 1988, Michard et al. 1991). The HP-metamorphism of the whole Saih Hatat window can be linked to the obduction process (Lippard et al. 1986, Goffé et al. 1988, Michard et al. 1991). More precisely, it can be envisaged that metamorphism in the Arabian margin was triggered by its subduction under a thickened oceanic lithosphere, the leading edge of which corresponds to the Samail nappe (Fig. 5B; Nicolas 1989, Michard et al. 1994).

## 3. Evolution of the Western Alps transect

### *3.1 Position of the Alpine suture*

Since Argand (1911, 1916) and Staub (1924) (e.g. Schaer 1991, Trümpy 1991), most authors consider that the Western Alps suture is represented by the ophiolite-bearing Schistes lustrés, which are found usually in a position east of and/or overlying the Penninic continental nappes (the Middle Penninic "Zone houillère"–Briançonnais–Grand Saint Bernard nappes and the Upper Penninic Monte Rosa–Gran Paradiso–Dora-Maira nappes, also referred to as Internal crystalline massifs, ICM), but west of and underlying the Austro-Alpine nappes, including the Sesia-Lanzo zone and the Dent-Blanche nappe system (Fig. 6, 7). However, Mattauer et al. (1987, 1994) argued that the Sesia-Lanzo zone (shortly Sesia) is an European massif comparable with the ICM and that the Alpine suture runs along the Canavese zone, as first suggested by Aubouin et al. (1977). This interpretation has been once more refuted by Avigad et al. (1994). Indeed, the contact between the Sesia schists and the Lanzo meta-ophiolites (Lagabrielle et al. 1989) was severely deformed, such that the local occurrence of serpentinized ultramafics of the Lanzo massif on the southern tip of the Sesia zone is inconclusive (Spalla et al. 1983). In contrast, the thrust contact of Sesia *over* the ophiolitic zone is well exposed along the western boundary of the Sesia zone (Caby & Comes 1975, Bigi et al. 1990). The ECORS-CROP seismic reflexion profile did not unravel the geometry of the Sesia zone at depth, probably due to the steepness of the structures (Bayer et al. 1989, Tardy et al. 1990). However, a northern prolongation of the Lanzo massif under Sesia is consistent with gravity modelling of the Ivrea positive anomaly (Bayer et al. 1989). Ophiolitic rocks, which are widespread west of the Sesia zone, are not present in the Canavese zone. The latter zone forms a steeply-dipping belt of crystalline schists, Permian granites and volcanics and Mesozoic cover of the Southern Alps. This zone appears to be an element of the Insubric Line, bounded to the west by mylonites of the Sesia zone and to the east by mylonites of the Ivrea zone (Zingg et al. 1976, Laubscher & Bernoulli 1977, Schmid et al. 1987). The Insubric line is definitely an intra-continental transpressive fault-zone along which mylonitization recorded backthrusting followed by dextral motion around 20–30 Ma ago (Heitzmann 1987, Schmid et al. 1987). The rare serpentinized lherzolites observed there can be regarded as infra-continental mantle rocks. According to Pognante (1989), the blueschists, orthogneisses, granulites and serpentinized lherzolites of the Rocca di Canavese thrust sheets belong to the innermost Sesia zone and derive from pre-Alpine crustal sections and associated mantle rocks. Sturani (1973) pointed out that the low-grade, Mesozoic sediments of the Canavese zone are "inequivocabilmente deposta su di un substrato sialico". Sesia itself shares common lithological characteristics with both the Ivrea zone and Dent-Blanche nappe, containing relics of pre-Alpine granulite-facies and mantle-derived rocks, while the Penninic basement nappes contain relics of pre-Alpine amphibolite facies and hornfelses (Fig. 6; and Lemoine et al. 1987; Ballèvre & Merle 1993, with ref. therein). This striking contrast could result from asymmetrical stretching of the Variscan crust during the Neo-Tethyan opening (see next section).

Hunziker et al. (1989) and Polino et al. (1990) proposed that the Briançonnais and Upper Penninic basement nappes (ICM) would have been part of the Austro-Alpine paleomargin, the European paleomargin being restricted to the Dauphinois-Helvetic domain. Such a proposal conflicts with many stratigraphical arguments (next section), and neither accounts for the position of the ophiolitiferous Schistes lustrés between the Penninic and Austro-Alpine units, nor with the contrasting pre-Alpine lithology of these continental units. Less extreme solutions were also proposed, leaving the Middle Penninic nappes attached to the European paleomargin, but not the ICM. In one of these solutions, the ICM would derive from a continental block isolated within the oceanic realm between Europe and Apulia, while the Antrona and Monte Viso ophiolites (below Monte Rosa and west of Dora-Maira, respectively) would originate from the oceanic arm west of the ICM block (Homewood et al. 1980, Platt 1986, Philippot 1990). Pfeifer et al. (1989) indeed pointed to minor differences between the Antrona and Zermatt-Saas


7°E
Zermatt 5 km
8°
M.Blanc
GSB
Dent-Blanche
M.Rosa
R
II D-K
ME
Belledonne
Vanoise
Money
Gran-Paradiso
Sesia
Zone
Insubric
line
Biella
Ivrea
Ca
GV
Fig. 7
Lanzo
Ambin
Val di Susa
Dora R.
Ro
TORINO
Montferrato
45°N
Briançon
Pinerolo
Po R.
Pre-alpine relics :
amphibolite-facies
hornfelses
granulite-facies
sub-contin.gabbro-peridotites
Alpine metamorphism :
outer boundary of blueschist-facies
id., Penninic eclogite-f.
25Km
Queyras
CL
M.Viso
E - U
Maira R.
Ac
Val Grana
Cuneo
EUROPE
P
G
T
K
ADRIA
AM
Apenn.
LIGURIAN ALPS
Dauphiné-Helvetic
( a: basement)
Lower and Middle Penninic
Flysch and Schistes lustrés
( a : ophiolites)
Internal crystalline massifs
( a :Upper Penn.eclogitic units;
( b : middle Penn. units )
Austro-Alpine units (a)
Canavese zone (b)
South-Alpine zone (c)
Main thrusts
(eventually inverted)
Ivrea gravity-mag-anomaly
Oligocene quartz-monzonites

(main Ligurian) ophiolites, but also emphasized that both ophiolite belts could represent parts of the same oceanic basin, axially separated by transform faults.

The stratigraphic and metamorphic homologies between Zermatt and Monte Viso ophiolites (e.g. Deville et al. 1992), together with their common position above the ICM, do not support the hypothesis of two distinct Ligurian arms. Stampfli & Marchant (in press) proposed an alternative solution with a palinspastic position of the ICM south of the Ligurian rift. In this recent reconstruction, at variance with the broadly "Argandian" reconstructions of Stampfli & Marthaler (1990) and Stampfli (1993), the ICM are considered as extensional allochthons, i.e. pieces of the foot of the upper-plate margin (European upper-crust) left back on the Apulian lower-plate margin, northwest of the Dent-Blanche and Sesia zones, during the Neo-Tethyan (Ligurian) opening. However, this proposal raises questions regarding the present-day geometry, and we prefer to assume that the present position of the Zermatt and Piemont-Ligurian ophiolites between the Dent-Blanche and Sesia nappes *above* and the ICM *below* does not result from some sort of mélange tectonics or later backthrusting but reflects the original palaeogeographic arrangement of these units. Overthrusting of the Gran-Paradiso and Dora-Maira nappes upon the Briançonnais-type Money and Pinerolo windows (Compagnoni et al. 1974, Borghi et al. 1985, Henry et al. 1993), without interposition of ophiolitic remnants, also pleads for a palinspastic position of the Piemont-Ligurian ocean east of all the Penninic continental units. In the following we accept that the Neo-Tethyan suture in the Western Alps runs between the Briançonnais-ICM (below) and the Sesia-Dent-Blanche Austro-Alpine nappes (above). The Canavese zone was either juxtaposed to Sesia along the same transect (e.g. Lemoine et al. 1987, Stampfli & Marthaler 1990), or located about 200 km southwest of Sesia until the Mid-Cretaceous (hypothesis of an Early Cretaceous sinistral motion along a proto-Insubric strike-slip fault: Stampfli & Marchant in press).

## *3.2 The stratigraphic record*

### 3.2.1 Ligurian oceanic realm

In the frame of the current structural model of the Western Alps just discussed, the rifting, drifting and early convergence evolution of this Tethyan transect can be deduced from a wide range of stratigraphical data. The oceanic, or Ligurian (= Piemont-Ligurian)

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Fig. 6. Tectonic and metamorphic sketch map of the Western Alps, showing the distribution of pre-Alpine relics in the Penninic and Austroalpine nappes. Structural contours after Bigi et al. (1990). Ivrea gravimetric-magnetic anomalies after Bozzo et al. (1992). *Insert:* Location in the Alpine belt (dotted: Penninic domain and External Alps; ruled: Austro-Alpine and South-Alpine domains). References for the nature of pre-Alpine relics as follows. Middle Penninic nappes: Caby (1974), Lefèvre & Michard (1976), Thélin et al. (1993). Upper Penninic nappes: Borghi et al. (1985), Sandrone et al. (1988), Henry (1990), Cadoppi (1990), Biino & Compagnoni (1991, 1992), Ballèvre & Merle (1993), Bouffette et al. (1993). Austro-Alpine nappes and Ivrea zone: Caby et al. (1978), Brodie et al. (1987), Ballèvre & Merle (1993). Ac: Acceglio; AM: Alpi Marittime; CL: Col du Longet; II D-K: Seconda zona dioritico-kinzigitica; E-U: Embrunais-Ubaye; G: Graubünden; GSB: Grand Saint Bernard; GV: Pointe du Grand Vallon; K: Koralpe-Saualpe; ME: Mont Emilius; P: Prealps; R: Roisan; Ro: Rocciavre; T: Tauern.

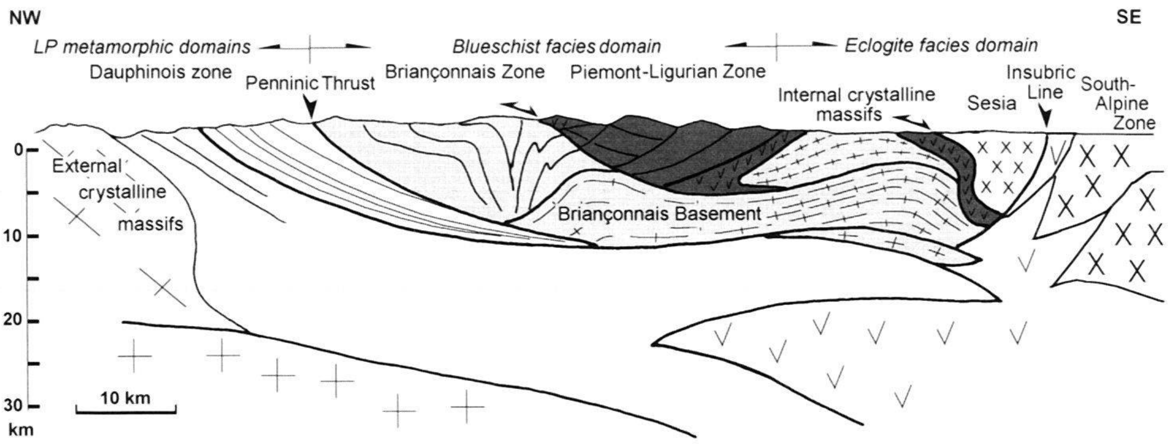
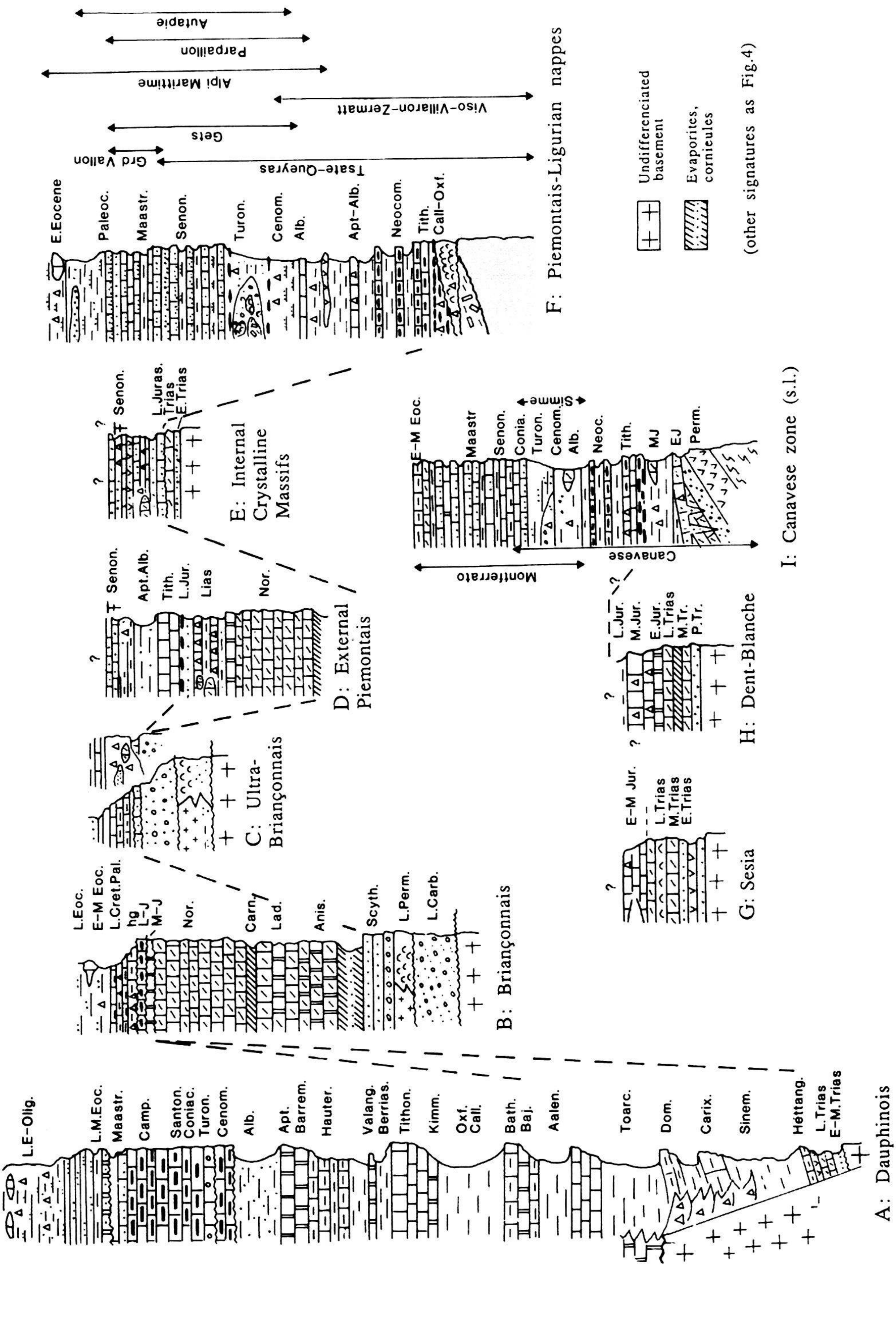



Fig. 7. Schematic cross-section of the Western Alps (Fig. 6 for location), based on geology and on ECORS seismic profiling (Roure et al. 1989, Tardy et al. 1990), after Henry et al. (1995), modified. Large V pattern: Apulian lithospheric mantle. Small v pattern: ophiolites.

units comprise a relatively thin ophiolitic basement overlain by a thick cover sequence (Fig. 8F). In contrast to the Oman ophiolite characterized by highly depleted harzburgites, serpentinized lherzolites, part of them weakly depleted (Pognante et al. 1986), prevail in the Ligurian mantle section. Moreover, the sheeted-dike complex is lacking. The pillow-lavas do not form a true basaltic layer, but are associated with ophicarbonate rocks, ophiolitic breccias and radiolarites overlying a previously deformed oceanic basement (serpentinized peridotites, partly foliated and recrystallized gabbros; e.g. Lemoine et al. 1986, Fig. 21). In the Ligurian domain, part of the oceanic spreading probably resulted from tectonic uncovering of the mantle (Decandia & Elter 1972, Lemoine et al. 1987, Saby et al. 1988, Trommsdorff et al. 1993). The formerly sub-continental peridotites were exhumed by shallow-dipping extensional shear-zone systems which penetrated a significant part of the upper mantle (Vissers et al. 1991). The magmatic sequence (gabbros, basalts) shows N-MORB affinities, and probably formed in small oceanic basins compartimented by normal and strike-slip faults and underlain by highly depleted peridotites (Lemoine et al. 1986, Pognante et al. 1986). Part of the gabbros correspond to layered intrusions emplaced at the crust-mantle boundary beneath the continental margins and sheared under granulitic to amphibolitic conditions during further mantle uncovering, as evidenced by Trommsdorff et al. (1993) in the eastern Central Alps. Gab-

Fig. 8. Stratigraphy of the main tectonic units of the Western Alps. References as follows. A: Lemoine et al. (1986). B: Lemoine et al. (1986), Tricart et al. (1988), Barféty et al. (1992). C: Lefèvre & Michard (1976), Jaillard (1988), Allimann (1989). D: Lemoine et al. (1984), Vanossi (1990). E: Marthaler et al. (1986), Deville (1989), Deville et al. (1992). F: Lemoine et al. (1986), Deville et al. (1992), Homewood (1983), Marini (1988). G: Venturini et al. (1994). H: Ayrton et al. (1982), Ballèvre et al. (1986). I: Sturani (1973, 1975), Homewood (1983), Bernoulli et al. (1990a).


A: Dauphinois
B: Briançonnais
C: Ultra-Briançonnais
D: External Piemontais
E: Internal Crystalline Massifs
F: Piemontais-Ligurian nappes
G: Sesia
H: Dent-Blanche
I: Canavese zone (s.l.)
Autapie
Parpaillon
Alpi Marittime
Viso-Villaron-Zermatt
Gets
Grd Vallon
Taste-Queyras
Montferrato
Canavese
Simme
Undifferenciated basement
Evaporites, cornieules
(other signatures as Fig.4)

bros and associated ferrodiorites and plagiogranites were dated isotopically from 193–163 Ma (Liassic–early Bathonian) by Fontignie et al. (1982). The radiolarian cherts intercalated with the ophiolitic breccias and pillow basalts, which yielded the first known fossils from the Schistes lustrés (Parona 1891), allowed De Wever & Caby (1981) and De Wever et al. (1987) to establish Callovian and Oxfordian ages. The coeval detrital formations (serpentinite sands, ophiolitic breccias) are a consequence of the mobility of the oceanic basement during sea-floor spreading (Lagabrielle et al. 1984, Deville et al. 1992). Some of them contain granitoid boulders which suggest the proximity of an attenuated continental crust (Saby et al. 1988). The radiolaritic sedimentation is followed by a widespread deposit of pelagic limestones (Tithonian-Berriasian), overlain by shales with siliceous limestone beds (Palombini shales, Neocomian).

The pelagic beds grade upward into Mn-bearing black shales, phyllitic marbles and calcschists containing ophiolitic olistoliths and olistostromes, banded detrital metabasites and minor phyllitic quartzites. This detrital deep-water formation is comparable to the Middle Cretaceous "basal complex" of the Gets nappe in the Swiss Pre-Alps and to the Embrunais-Ubaye and Alpi Marittime Helminthoid Flysch nappes, split off the Ligurian basin during Late Cretaceous time (Sturani 1975, Marini 1988, Homewood 1983). The detrital deposits show mixed provenances from oceanic and paleomargin areas. They indicate the presence of scarps of ophiolitic material (from which basic-ultrabasic olistoliths of all sizes, sometimes kilometres long, have been shed) in a narrowing Ligurian ocean (Lagabrielle 1987, Deville et al. 1992). The stratigraphic sequence of some of the Ligurian nappes (Zermatt–Saas Fee, Monte Viso) seems to terminate at about this time (Albian-Cenomanian), possibly in relation to a subduction process. In contrast, the sequence of some other Piemont-Ligurian units continues upward in flysch facies with Late Cretaceous *Globotruncanidae* fossils (e.g. Queyras blueschist units, Lemoine et al. 1984, Deville et al. 1992; Tsaté nappe, Marthaler 1981, 1984; Deville et al. 1992). A metamorphic Maastrichtian flysch has even been recognized on top of the "schistes lustrés" west of the Gran Paradiso massif (Grand Vallon unit, Deville 1986, Deville et al. 1992). Senonian and Maastrichtian-Paleocene flyschs are well preserved in the Helminthoid Flysch nappes. The provenance of their detrital material is mostly from WSW and SE (External Piemontais, inner Briançonnais, Corso-Sardinian basement) in the originally external units (Parpaillon nappe in Embrunais, Kerckhove 1969; San Remo and Alassio nappes in the Alpi marittime, Marini 1988, Vanossi 1990). In the originally internal units such as the Autapie nappe of Embrunais (Kerckhove 1969), the Simme nappe in the Swiss Pre-Alps (Sturani 1973, Homewood 1983) and the Cassio-Caio North-Apenninic nappes (Sturani 1973, Vanossi 1990), the detrital material was provided simultaneously by ophiolitic sources and by an innermost continental ridge with Insubrian (Canavese) affinities. The uppermost, olistostromic levels of the Alassio nappe are dated from Early Eocene and record the early emplacement of the more internal flysch and "schistes lustrés" units (Marini 1988).

### 3.2.2 European paleomargin

The stratigraphic record of oceanic rifting, spreading and closure can also be recognized in the paleomargin units. As illustrated early by Sturani (1973, Fig. 2) and Laubscher & Bernoulli (1977), the European paleomargin mainly includes the Dauphinois-Helvetic and Briançonnais realm, partly separated by the Sub-Briançonnais-Valaisan rift and bounded easterly by the external Piemont and Dora-Maira units, while the Apulian (Adriatic) paleomargin corresponds to the Sesia and Dent-Blanche nappes and the Canavese zone, passing eastward to the South-Alpine domain. In the proximal part of the European paleomargin, i.e. the Dauphinois-Helvetic domain, local volcanic flows (alkaline basalts of the Pelvoux area) testify that extension took place there as early as Late Triassic (Fig. 8A). In the same domain, several normal faults were active during Liassic-Middle Jurassic time and have been preserved between tilted blocks and half-grabens (Lemoine et al. 1986). The Middle-Late Triassic carbonate platform which extended to the Briançonnais-Piemont domain and to Adria was broken during Early-Middle Jurassic time. The Briançonnais zone shows numerous Liassic-Middle Jurassic faults usually sealed by Upper Jurassic-Berriasian pelagic limestones (Lemoine et al. 1986, Michard & Henry 1988, Tricart et al. 1988). The Briançonnais emerged during late Liassic-Bajocian time (Fig. 8B), probably due to its thermal uplift as a part of the upper plate of a lithospheric scale, low-angle normal fault responsible for the opening of the Ligurian ocean (Lemoine et al. 1987, Stampfli & Marthaler 1990). Uplift of the northwestern shoulder of the Tethyan rift (Ultra-Briançonnais = Acceglio Zone, Lefèvre & Michard 1976) caused its deep erosion, allowing the break-up unconformity to be located on top of Lower Triassic or even pre-Triassic rocks (Fig. 8C). Huge slope breccias accumulated during Early-Middle Jurassic time in subsiding basins of the External Piemontais domain (Fig. 8D), probably close to the uplifting Ulta-Briançonnais shoulder (e.g. Val Grana, Michard 1967; Brèche nappe, Lemoine et al. 1986, Fig. 19; Arnasco-Castelbianco nappe, Vanossi 1990). However, an origin from the Apulian side of the Ligurian ocean can also be advocated for the Brèche unit (Stampfli & Marthaler 1990).

The Triassic -Jurassic sequences locally preserved on top of the Upper Penninic continental nappes show internal Briançonnais affinities (Fig. 8E). However, they are followed by Ligurian-type deposits including volcano-clastic levels (prasinites), and detrital-olistostromic, flysch-type calcschists where Senonian *Globotruncanidae* have been identified (Marthaler et al. 1986, Deville 1989, Deville et al. 1992). In contrast, from Berriasian to Late Cretaceous-Paleocene time, the Briançonnais–Ultra-Briançonnais realm generally received thin, fine-grained pelagic deposits (siliceous, marly or calcareous muds, condensed sedimentation interrupted by phosphatized hard-grounds) in a set of shoals and troughs a few kilometres wide (Lemoine et al. 1986, Fig. 20). In the inner Briançonnais and Ultra-Briançonnais zone (or possibly in the adjoining Mont Rose zone), these pelagic deposits comprise detrital intercalations (siliciclastic breccias with Triassic and basement olistoliths: Fig. 8C, right) which record renewals of synsedimentary, likely compressive activity during Late Cretaceous-Paleocene time (Tsanteleina breccias in Vanoise, Ellenberger 1958, Jaillard 1988; Col du Longet, Lefèvre & Michard 1976; Mont-Fort nappe, Allimann 1989, Deville et al. 1992). Finally, the Briançonnais stratigraphic record continues into Late Eocene, with the sedimentation of an olistostromal flysch associated with the westward emplacement of the Ultra-Briançonnais nappes (Barféty et al. 1992). The Eocene flysch basin extended westward onto the Sub-Briançonnais zone, then (diachronically) onto the Dauphinois-Helvetic zone, forming a syntectonic foredeep for the advancing Penninic nappes. The Helvetic flysch terminates by the Early Oligocene in from of wildflysch within which olistostromes are sheared beneath the encroaching Penninic allochthons (Homewood et al. 1980).

### 3.2.3 Adria paleomargin

The southeastern margin of the Tethyan rift was also affected by Liassic-Middle Jurassic normal faulting, but in contrast to the Briançonnais, it subsided rapidly at the beginning of Middle Jurassic times (Sturani 1973, Bernoulli et al. 1990a). This was taken as an argument in support of an asymmetric rifting of the Ligurian ocean, the Austro-Alpine margin corresponding to the lower plate of the detachment normal fault (Lemoine et al. 1987). Indeed, within one of the Austro-Alpine nappes east of the studied transect (Err nappe, eastern Switzerland), part of an upper-crustal, extensional detachment has been observed and dated as Toarcian to Middle Jurassic (Froitzheim & Eberli 1990). This low-angle detachment was linked to syn-sedimentary high-angle faults at the surface and both low- and high-angle faults dipped oceanward in this distal part of the Apulian margin. An older, eastward-dipping detachment system has been active in the proximal part of the margin (Central Austro-Alpine, South-Alpine) during Early Jurassic times (Bernoulli et al. 1990, Froitzheim & Eberli 1990).

The Canavese zone was part of the distal margin, west of the east-dipping, Late Permian to Early Mesozoic Pogallo detachment (Handy 1987). Early to Middle Liassic crinoidal limestones, red shales and breccias unconformably overlie the Triassic dolostones (where they formed the filling of neptunian dikes) and their basement (Fig. 8I). Granitic breccias with Triassic olistoliths were deposited in a pelagic environment (shales with micritic limestones and cherts) during Toarcian to Middle Jurassic times (Sturani 1973, 1975). They are followed by radiolarian cherts with scattered granitic pebbles, which pass upward into calpionellid-bearing limestones with intercalations of graded conglomerates (Tithonian-Berriasian). In the Canavese region itself, the succession ends with a formation identical to the Ligurian Palombini shales (Neocomian). Younger formations from the southern prolongation of the Canavese zone can be observed in the pre-Oligocene basement of Monferrato (Sturani 1973, 1975). Their Albian to Maastrichtian succession (basal complex with continental and ophiolitic olistoliths and conglomerates; Helminthoid flysch) is practically identical to that of some of the Ligurian nappes and can also be recognized in the Simme nappe of the Swiss Pre-Alps (Sturani 1975, Homewood 1983). A calcareous flyschoid formation with Apenninic affinities, Early-Middle Eocene in age, terminates the Monferrato-(Canavese) pre-molassic sequence.

In contrast, the Dent-Blanche and Sesia cover sequences are much more restricted (Fig. 8H, G). The detached cover of the lower series (Arolla) of the Dent-Blanche nappe *sensu stricto* (which is the upper unit of the Dent-Blanche nappe *sensu lato*, Ballèvre et al. 1986), preserved in Mont Dolin, includes Liassic (?) dolomitic breccias overlying unconformably the Triassic dolostones and coarse polymict breccias with elements of crystalline rocks, not older than Middle Jurassic (Sturani 1975, Ayrton et al. 1982). The Roisan zone, to the south of the same Arolla unit, shows distinctive sedimentary sequences, including Permo-Triassic, metamorphic conglomerates, pelites and tuffites, followed by Triassic carbonates and evaporites (Höpfner 1995), as well as Liassic calcschists and breccias, locally followed by phosphatized-manganiferous layered microquartzites with local occurrence of metabasites (Ballèvre et al. 1986). Finally, a sedimentary-volcanic cover sequence has recently been recognized in the Sesia zone, although recrystallized under eclogitic conditions. According to Venturini et al. (1994), the Sesia "monometamorphic cover complex" includes a volcano-sedimentary sequence of Permo-Triassic age, followed by Middle-Upper Triassic dolomitic marbles with intercalated metabasalts, then by conglomeratic calcschists containing blocks of dolomitic marbles, of metabasalts, and of the Sesia Polymetamorphic Basement Complex. By comparison with the breccia formations from the adjoining paleomargin units (Dent-Blanche, Canavese), the Sesia conglomeratic calcschists could be Early to Middle Jurassic.

### *3.3 Age and geotectonic setting of HP-LT metamorphism*

#### 3.3.1 Austro-Alpine HP-LT metamorphism

The development of the Ligurian Neo-Tethys appears to have been linked to the opening of the Central Atlantic through a major, sinistral transform system between Africa and Eurasia (Laubscher & Bernoulli 1977, Lemoine et al. 1986, Le Pichon et al. 1988, Ziegler 1988, Dercourt et al. 1993). The Ligurian ocean began to close during Mid-Cretaceous times, in relation with the SE motion of the southern part of the European plate comprising Iberia, Sardinia, Corsica and the Briançonnais realm (Lemoine et al. 1986, Coward & Dietrich 1989), combined with the progressive change of motion of Africa with respect to Europe from ESE to NE and N (Le Pichon et al. 1988, Dewey et al. 1989), and with the opening of an Eastern Mediterranean basin between Africa and Apulia (Le Pichon et al. 1988, Ziegler 1988, Dercourt et al. 1993, Stampfli & Marchant in press). In the Austro-Alpine system of the Eastern Alps, the Eoalpine compression is documented both by stratigraphical and geochronological data. In the Austrian part of the belt, the fluviatile to marine Gosau beds of Coniacian age transgress already formed nappe structures (Frank 1987). In Graubünden (Swiss Eastern Alps), the onset of flysch sedimentation postdates the Cenomanian-Lower Turonian pelagic "Couches rouges", which are affected by the Eoalpine folding (Froitzheim et al. 1994). Reworked detrital glaucophane was found in upper Turonian flyschs along the northwestern boundary of Northern Calcareous Alps (Winkler & Bernoulli 1986). Eoalpine eclogites have been recognized south of the Penninic Tauern window and in the Koralpe-Saualpe Upper Austro-Alpine units (Neubauer et al. 1995). In the latter area, they were dated at 151–100 Ma (Sm-Nd and Rb-Sr methods, Thöni & Jagoutz 1992). In Graubünden, the radiometric ages indicate a westward migration of the Eoalpine compressive phase from about 100 Ma to less than 90 Ma (Froitzheim et al. 1994).

In the Western Alps there is also widespread petrologic evidence for the conversion of the Austro-Alpine passive margin into a convergent margin during Mid-Cretaceous times. Eoalpine eclogite-facies metamorphism is well developed in the Sesia "micascisti eclogitici" and M. Emilius outliers (Compagnoni et al. 1975, Compagnoni 1977, Ballèvre et al. 1986). Rb/Sr whole-rock and mineral isochrons were used to propose a pressure peak at around 110 Ma and a temperature peak around 85 Ma, followed by a cooling phase down to 300 °C at about 60 Ma (Oberhänsli et al. 1985, Hunziker et al. 1989). Incomplete recrystallization during an early, high-grade blueschist to eclogite-facies metamorphism is recorded in the IIDK (upper Sesia) unit (Lardeaux et al. 1982, Pognante et al. 1988, Vuichard 1989). Relics of early high-pressure metamorphism were also observed in the upper Dent-Blanche unit, within the Valpelline series (Kiénast & Nicot 1971), the Arolla series (Ballèvre & Merle 1993) and the Mesozoic cover (Ayrton et al. 1982, Höpfner 1995), which terminates with Upper Jurassic levels (Fig. 8H).

### 3.3.2 Penninic UHP-HP metamorphism

HP-LT recrystallization widely affected the units of the European plate and the juxtaposed ophiolitiferous nappes (Fig. 6, 7). The metamorphic grade ranges from blueschist-facies in the inner Briançonnais (Acceglio zone, Vanoise-Ambin, Mont-Fort nappe) and external Piemont-Ligurian units (western Queyras, upper Val di Susa, Tsaté nappe) to eclogite-facies in the ICM (except Money and Pinerolo windows) and adjoining Ligurian units (M. Viso, Rocciavre, Zermatt–Saas Fee zone), see Saliot (1978), Caby et al. (1978), Goffé & Chopin (1986), Dal Piaz & Lombardo (1986), Deville et al. (1992), Ballèvre & Merle (1993) and references therein. Ultra-HP, coesite-bearing assemblages are found in one of the polymetamorphic basement units of southern Dora-Maira (Chopin 1984, Henry 1990, Chopin et al. 1991, Schertl et al. 1991, Kienast et al. 1991) and in the Zermatt metaradiolarites (Reinecke 1991).

In the Middle Penninic, inner Briançonnais units, the age of metamorphism is bound to postdate 45 Ma, and possibly 40 Ma, since HP recrystallization overprints locally fossiliferous, Middle Eocene (Lutetian) sediments ("flysch noir" from Vanoise-Ambin and Acceglio: Ellenberger 1958, Lefèvre & Michard 1976; "scisti calcarei" of the Brianzonese ligure: Messiga et al. 1983, Goffé 1984, Cabella et al. 1991). K/Ar ages from 49 to 40 Ma are recorded by white micas throughout the Briançonnais cover, while white micas from the basement yield a wide range of ages between 449 and 42 Ma (Bocquet et al. 1974), suggesting the presence of an excess argon component. $^{40}Ar/^{39}Ar$ step-heating studies on micas and glaucophane from Ambin-Acceglio basement rocks demonstrated that they have been overprinted by an Eocene HP-LT event at about 350–400 °C, 10 kbar (Monié 1990). Glaucophane results indicate a maximum age of 54 ± 1 Ma for Alpine overprinting and trapping of excess argon. Phengites from the overlying cover record a plateau-age of 37 ± 1 Ma (Monié 1990), related to the thermal peak of metamorphism, namely the "Lepontine event" that is recognized throughout the Western Alps. The inner Briançonnais units cooled before 30 Ma, at least in the Ligurian Alps where conglomerates from the Lower-Middle Oligocene transition unconformably overlie the Alpine metamorphic units (Sturani 1975, Lorenz 1986, Vanossi 1990).

Dating of HP metamorphism in the Upper Penninic (ICM) and South-Penninic (ophiolitiferous) units is more controversial. For a while, following Ellenberger's (1952, 1958, 1960–63) master work in the Vanoise massif, it has been admitted that everywhere in the Penninic domain, metamorphism was related to the Late Eocene-Oligocene nappe tectonics (e.g. Michard 1967). Then, Cretaceous K/Ar and Rb/Sr isotopic datings were obtained, not only from the Austro-Alpine Sesia zone, but also from the Penninic realm, where they are found together with Tertiary ages (Vialette & Vialon 1964, Hunziker 1970, 1974; Bocquet et al. 1974, Frey et al. 1974, Dietrich et al. 1974). The observation that HP-LT assemblages are overprinted by greenschist-facies minerals everywhere led to the idea that HP-LT early assemblages are systematically Eoalpine in age, while the HT-LP are of Mesoalpine age (Dal Piaz et al. 1972). On stratigraphical grounds, this opinion was challenged for the Briançonnais domain (Michard 1977, Caby et al. 1978, Goffé & Chopin 1986, Dal Piaz & Lombardo 1986). Unfortunately, post-Liassic fossils were unknown in the Piemont domain until recently and, by contrast, Cretaceous isotopic results became more and more abundant in the area (Chopin & Maluski 1980, Hunziker & Martinotti 1984, Chopin & Monié 1984; Monié 1985, Hunziker et al. 1989, Hurford

& Hunziker 1989). Accordingly, the concept of an Eoalpine HP-LT event, in the range 110–80 Ma, was generally accepted for the Upper and South Penninic units. The ages in the range 40–30 Ma for the same units were referred to the Mesoalpine "Lepontine" stage and to the further cooling history. Most isotopic datings first obtained from the Dora-Maira UHP-HP rocks appeared to be consistent with this overall timing (U/Pb/zircon, Rb/Sr and Sm/Nd: Paquette et al. 1989; $^{40}Ar/^{39}Ar$/phengites: Scaillet et al. 1990, Monié & Chopin 1991).

However, part of the Eoalpine ages reported from the Upper and Southern Penninic units (those older than ca. 80 Ma) conflict with the occurrence of *Globotruncanidae* in various metasediments from these domains, either from the ICM cover units or the Schistes lustrés (see preceding section and Fig. 8D–F). Now, if the K/Ar and $^{40}Ar/^{39}Ar$ apparent ages in the range 110 (120)–80 Ma are discarded on these stratigraphic grounds, then they suggest the presence of an excess argon component, which in turn raises suspicions even concerning the youngest Cretaceous ages. As a matter of fact, several isotopic studies, particularly in the last couple of years, yielded Tertiary ages for the UHP-HP metamorphism of both the continental and oceanic Penninic nappes. K/Ar measurements on phengite separates from Queyras calcschists yielded a best age of 50 ± 1 Ma (Liewig et al. 1981). Phengites from eclogitic metagabbros in the Monte Viso ophiolite provided concordant $^{40}Ar/^{39}Ar$ ages of 48–51 Ma (Monié & Philippot 1989). Sm/Nd isotopic analyses of an essentially unretrogressed eclogitic metabasalt from the Zermatt-Saas ophiolite yielded a 52 ± 18 age (Bowtell et al. 1994). First ion microprobe (SHRIMP) data on the latter area are in line with such a Paleocene–Early Eocene age (Gebauer 1995). In the continental UHP unit of southern Dora-Maira, five zircon samples from the pyrope megacrysts and pyrope-bearing quartzites yielded a higher intercept age at 38 ± 1.4 Ma in a Pb concordia diagram (Tilton et al. 1991), at variance with the less precise result of Paquette et al. (1989). Tilton et al. (1991) also obtained a 38 Ma age from Sm-Nd data for pyrope crystallization in Dora-Maira. These results are consistent with ion microprobe ages of 35 Ma obtained in the same area by Gebauer (1995). Similarly, the Alpe Arami and Cima di Gagnone eclogites from the Middle Penninic Cima Lunga-Adula nappe (Central Alps) yielded SHRIMP ages in the range 42–35 Ma for their early, UHP-HP recrystallization stage (Gebauer et al. 1992, 1995). Sm-Nd dating of garnet peridotites and an eclogite from Cima Lunga also yielded consistent mineral ages of ca. 40 Ma (Becker 1993). From all this, it now becomes apparent that the Early to Late Cretaceous "ages" obtained from the continental and oceanic Penninic units by Ar-data, Rb-Sr or conventional, multigrain zircon data do not actually correspond to any protracted HP-LT event. On the contrary, UHP-HP metamorphism is likely to have occurred about 50 Ma ago in the Ligurian units, and still later, about 40–35 Ma ago, in the European margin units. Similarly, von Blankenburg & Davies (1995) conclude from their compilation of ages from the Penninic HP rocks that subduction of the European crust took place at ca. 55–40 Ma, followed by uplift at 40–30 Ma.

### 3.3.3 Subduction setting of Western Alps metamorphism

HP-LT metamorphism, and *a fortiori* UHP metamorphism, implie some sort of subduction tectonics. Since Dal Piaz et al. (1972), it was assumed that HP-metamorphism in the Sesia rocks results from their carrying to depth in a southeast-dipping subduction zone which was active during Early-Late Cretaceous along the SE border of the Ligurian Tethys (e.g. Schmid et al. 1987, Stampfli & Marthaler 1990, Ballèvre & Merle 1993). The onset of flysch sedimentation and the widespread occurrence of Mid-Cretaceous, coarse conglomerates with Insubrian material in the Ligurian and Canavese sequences (Fig. 8, F, I) can be also related to the development of this southeastern subduction zone. An accretionary prism (Tsaté prism, Marthaler & Stampfli 1989) would have formed in front of the Austro-Alpine leading edge as early as during Aptian times (Stampfli & Marthaler, 1990, Stampfli & Marchant in press). Laubscher & Bernoulli (1982) suggested that the Austro-Alpine subduction would have been bivergent, steeply-dipping at depth and involving obduction of ophiolitic slivers onto the Apulian margin, at least along the Eastern Alps (see also Winkler 1988). However, Froitzheim et al. (1994) emphasize the westward migration of the Eoalpine orogenic wedge through the Austro-Alpine domain of the Eastern Alps. They consider that the subduction of the South-Penninic oceanic lithosphere under the Apulian plate was not the direct cause for the Eoalpine orogeny in the Austro-Alpine units of eastern Switzerland and Austria. This orogeny is to be ascribed to a continental collision east or southeast of the Austro-Alpine realm, i.e. along the Vardar-Hallstatt ocean (see Thöni & Jagoutz 1993, Stampfli & Marchant in press, fig. 5).

The plate tectonic explanation of the UHP-HP metamorphism in the Penninic units raises still more questions. An early, and most logical proposal was to ascribe the Penninic high-pressure metamorphism to the operation of the same subduction zone as that of the Austro-Alpine (Dal Piaz et al. 1972, Frisch 1979, Homewood et al. 1980, Laubscher & Bernoulli 1982, Goffé & Chopin 1986). Recrystallization would have progressed from the oceanic into the continental Penninic units, due to the eastward consumption of the Ligurian lithosphere beneath Apulia. HP-recrystallizations in the European continental margin were thus referred to an early collisional stage of the Alpine history. However, as soon as an Eoalpine age was accepted for these recrystallizations (in relation with the prevailing isotopic evidence), the preceding interpretation resulted in contradictions with the occurrence of Late Cretaceous marine sediments in the metamorphic, Piemont-Ligurian and ICM cover units, progressively unraveled (Marthaler 1981, Lemoine et al. 1984, Deville 1986, Mathaler et al. 1986). The presence of a continuous Mid-Cretaceous-Paleocene or Eocene sedimentation in the Ligurian-related Helminthoid Flysch nappes *sensu lato* (section 3.2.1) was also hardly reconcilable with the Ligurian ocean being closed as early as in Senonian times.

Stampfli & Marthaler (1990) suggested that the Piemont-Ligurian Late Cretaceous sedimentation would have occurred onto and at the front of a detached accretion prism overlying the Mont-Fort (Ultra-Briançonnais) domain. However, at the turn of the 80's, many authors thought that the only way to reconcile the isotopic and stratigraphic data in the frame of the Argandian reconstruction of the Alps was to imagine the operation of two roughly coeval subduction zones, both easterly dipping, i) one along the Apulian-Ligurian margin, responsible for the Sesia metamorphism and part of the Ligurian, and ii) the other one within the Ligurian ocean itself, progressively encroaching onto the European margin and responsible for its UHP-HP recrystallization (Mattauer et al. 1987, Lagabrielle 1987, Lemoine 1900, Deville et al. 1992, Avigad et al. 1993). Now, as discussed above, the concept of a Late Cretaceous subduction of the European margin is discarded by the latest and more reliable isotopic datings which support the idea of a Tertiary subduction of this margin. The old timing of the Penninic metamorphism (Ellenberger 1958)

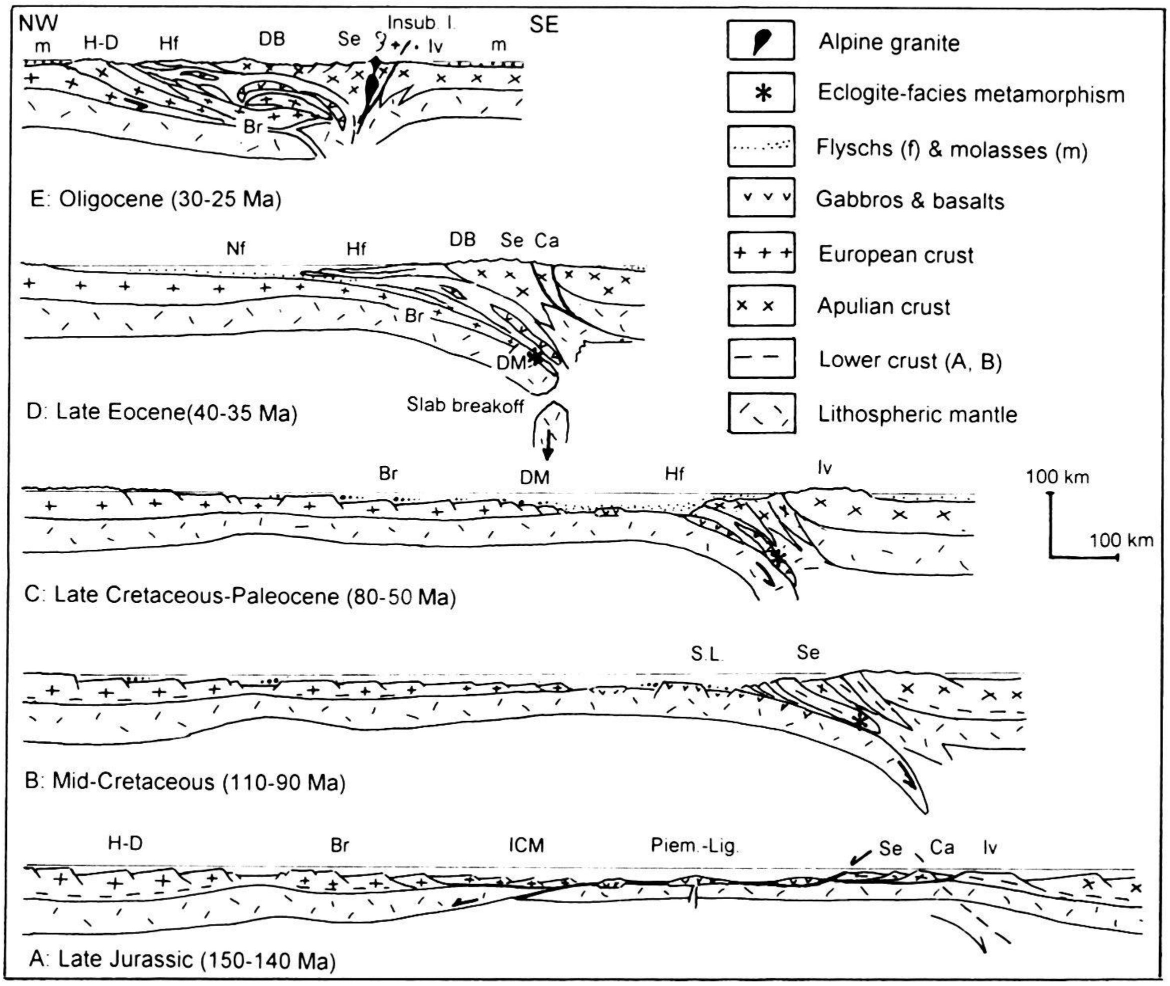



Fig. 9. Tectonic model for the evolution of the Western Alps. Br: Briançonnais. Ca: Canavese. DM: Dora-Maira. H-D: Helvetic-Dauphinois. Hf: Helminthoid flysch. ICM: Internal crystalline massifs. Iv: Ivrea. m: molasses. Nf: Nummulitic flysch. PF: Penninic front. P-Lig: Piemontais-Ligurian ocean; Se: Sesia. SL: Schistes Lustrés basin.

is re-actualized, and a model with only one subduction can again be hypothesized (Fig. 9).

At the early stage (Mid-Cretaceous) of the Europe-Adria convergence in the Western Alps (Fig. 9B), i.e. during the Eoalpine recrystallization of the Sesia nappe, the Benioff zone is likely to have been steep, in connection with the high density of the Ligurian lithosphere, about 50 Ma-old and including a relatively thin basaltic-gabbroic crust. The shortness of the subducting slab, together with its steepness would explain the lack of associated arc magmatism. No relic of the oceanic lithosphere subducted then has been exhumed afterward. The highest grade, UHP-HP eclogitic ophiolites (Zermatt-Saas, Monte Viso) were probably brought into the subduction zone during Mid-Cretaceous times (apparent end of their sedimentary sequence, Deville et al. 1992), while they reached their maximum burial depth during the Paleocene (Fig. 9C). This would correspond to a subduction rate of 0.25–0.30 cm/year. Within the narrowing Ligurian ocean, fault scarps are evidenced by chaotic breccias (Lagabrielle 1987, Deville et al. 1992). The compressional origin of these faults is not demonstrated, and there is no reason to take them as emerging intra-oceanic subduction zones. The latter hypothesis is hardly reconcilable with the narrow width of the Ligurian ocean (about 400 km, see next section). Sedimentation went on in the western part of the Piemont-Ligurian basin and upon the ICM at least until the Senonian, probably until Paleocene-Eocene times (cf. Helminthoid Flysch nappes). After consumption of the rest of oceanic lithosphere, the thinned continental crust of the European margin (ICM material) was in turn involved in the subduction zone during Eocene and reached its deepest burial depth (about 100 km) about 35 Ma ago (Fig. 9D). This would correspond to a convergence rate in the order of 1 cm/year, quite consistent with the geodynamic data for Paleocene-Eocene times along this transect (Le Pichon et al. 1988).

## 4. Discussion

### *4.1 Two different oceanic domains*

Comparing the geology of the Oman-Makran and Western Alps regions makes apparent major differences in the orogenic evolution of these Tethyan transects. Plate convergence begins at about the same time (Mid-Cretaceous) in both areas, but the oceanic realms between the converging continents are strongly distinct. In the Oman transect, the oceanic lithosphere was more than 2000 km wide (Le Pichon et al. 1988, Dercourt et al. 1993). The maximum width of the Ligurian ocean (Tithonian) can be estimated from the relative path between Africa and Eurasia during the Jurassic, accepting that Apulia was a northern promontory of the African plate at that time (Lowrie 1986). Dewey et al. (1989) suggested a 600-km-wide Ligurian ocean, but a width of ca. 400 km could be a better estimate (Lemoine et al. 1987, Le Pichon et al. 1988, Ziegler 1988, Dercourt et al. 1993, Stampfli & Marchant in press). Drastic differences also occur in the nature and age of these oceanic lithospheres, which will control their tectonic behaviour during convergence. The Tethyan ocean between Oman and Makran was partly underlain by an old, Permian-Jurassic lithosphere and partly by a young lithosphere formed along a Cretaceous, fast-spreading oceanic ridge (Nicolas 1989). When convergence began, the fragment of oceanic lithosphere which constituted the upper plate of the shallow-dipping intra-oceanic subduction (detachment) was less than 5 Ma old (Boudier et al. 1988, Michard et al. 1991). In contrast, the Ligurian ocean was about 50 Ma old when convergence began, and was underlain by a largely peridotitic lithosphere with a thin and irregular crustal section (Lemoine et al. 1987, Vissers et al. 1991, Trommsdorff et al. 1993). The low density and limited thickness of the Cretaceous lithosphere east of Oman allowed it to become the upper plate of a converging plate system, while the high density and great thickness of the Ligurian lithosphere caused its subduction. The large, ca. 15 km-thick ophiolitic slab of Oman was detached at the lithosphere-asthenosphere boundary. The restricted, usually less than 1 km thick ophiolitic slivers of the Western Alps were sampled at the uneven top of the subducting lithosphere (Auzende et al. 1983, Tricart & Lemoine 1988).

### *4.2 Two versus one subduction zone(s) s. l.*

The large width of the Tethyan ocean in the Oman transect allowed two subduction zones *sensu lato* to form when plate convergence began. During Late Cretaceous, HP-LT metamorphism occurred there in two subduction zones of distinct types, i) an eastern, Benioff-type subduction along the Iranian margin, and ii) an obduction-continental subduction zone along the Arabian margin. In the latter zone, the high P/T conditions replaced low P/T conditions which had prevailed during the intra-oceanic stage of lithospheric duplication. The record of the pre-HP, initial stage of the obduction is preserved in the HT-LP metamorphic sole of the ophiolite.

In the Western Alps transect, by contrast, the small width of the Ligurian ocean would hardly permit the formation of two coeval subduction zones. In fact, there is no metamorphic sole at all beneath the Western Alps ophiolites to attest for such an intra-oceanic detachment as in Oman. Careful examination of the stratigraphic data (e.g. Marini 1988, Deville et al. 1992), as well as the most recent and accurate isotopic results (Monié & Philippot 1989, Tilton et al. 1991, Gebauer et al. 1992, Becker 1993, Gebauer 1995), strongly suggest that HP-LT metamorphism occurred in a single, westward retreating subduction zone. During the Late Cretaceous, blueschist- and eclogite-facies recrystallizations affected the Austro-Alpine margin involved in a possibly bivergent subduction zone (Laubscher & Bernoulli 1982). The Ligurian lithosphere was progressively consumed in this subduction zone. Sedimentation went on in the western part of the oceanic basin and the adjoining European margin (ICM) until the Paleocene and locally even the Eocene (cf. Flysch nappes sequences). Some sheared slivers escaped definitive subduction and were exhumed from various depths. In the Monte Viso ophiolite, sampled from ca. 45 km depth at 50 Ma (Monié & Philippot 1989), the diversity of the syn- to post-eclogitic metamorphism is interpreted as the result of syn- to post-subduction shearing of the oceanic lithosphere (Nisio et al. 1987). The European margin was, in turn, involved in the subduction process during late Eocene and then affected by UHP-HP metamorphism, the grade of which decreases from the distal part of the paleomargin (now the ICM) to the proximal part (Briançonnais). Hence, the metamorphism of the European margin initiated in a collisional settting, about 50 Ma after the initiation of the Austro-Alpine, subduction-related metamorphism. The collision process not only triggered UHP-HP metamorphism in the lower plate, but also initiated the obduction of some ophiolitic slivers, part of them already metamorphosed (e.g. Monte Viso).

### *4.3 From burial to exhumation tectonics*

Both burial and exhumation tectonics differ for the lower plate units of the Oman Mountains and Western Alps. The Saih Hatat blueschist- and eclogite-facies metamorphism developed in the Arabian margin when the latter was buried (subducted) beneath a wedge of oceanic material overloaded by a virtually unmetamorphosed ophiolitic slab (obduction). The Penninic continental crust recrystallized under high- and ultra-high-pressure conditions when it was underthrust (subducted) beneath a continent-continent collisional wedge. This tectonic wedge involved metamorphic slivers of both oceanic and continental (Sesia) origin and was overloaded by a virtually unmetamorphosed, continental slab. However, all the documented prograde P-T paths (Fig. 10) show similar shapes,

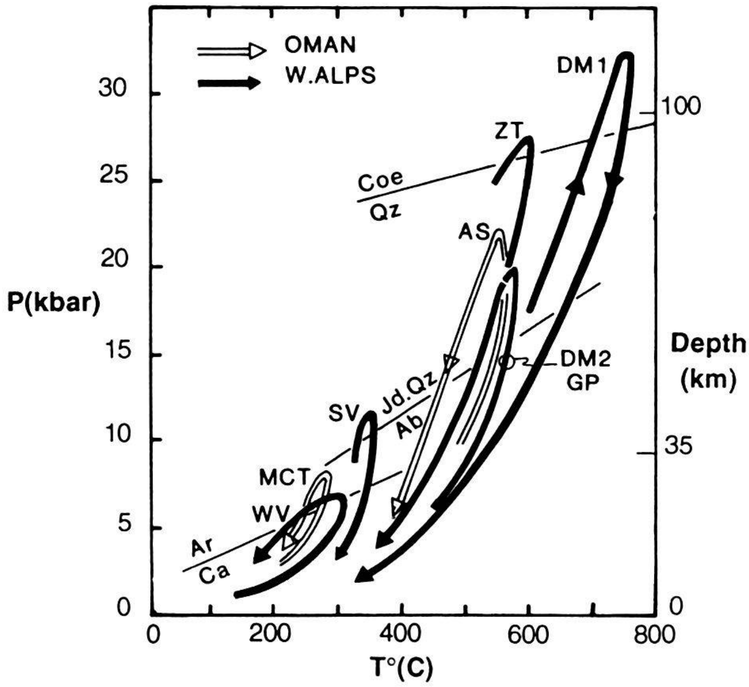



Fig. 10. Pressure-temperature paths for the Oman Mountains (Saih Hatat) and the Western Alps (Penninic nappes). AS: As-Sifah (Wendt et al. 1993). DM1, 2: Southern Dora-Maira units (Schertl et al. 1991, Chopin et al. 1991, Michard et al. 1993). GP: Gran Paradiso (Ballèvre & Merle 1993). MCT: Muscat nappes (Goffé et al. 1988). SV, WV: Southern and western Vanoise (Gillet & Goffé 1988). ZT: Zermatt meta-radiolarites (Reinecke 1991).

concave toward the P axis (e.g. blueschist-facies units: Muscat nappes and Vanoise; eclogitic units: As-Sifah and Dora-Maira UHP unit). This is consistent with the results of P-T modelling for subducting crustal slabs (e.g. Ruppel & Hodges 1994, Hacker & Peacock 1995).

The retrograde parts of the reported P-T paths also show broadly similar, concave shapes in the Saih Hatat and Penninic units. The available P-T data fail to unravel any difference in the exhumation mechanisms of the discussed areas, particularly due to the lack of extensive data in the low temperature range. In contrast, the temperature-time (T-t) paths indicate distinct cooling rates in the range 250–50 °C (Fig. 11). In Oman, the initial cooling of the low- and high-grade units occurred at a rate around 70–50 °C/Ma, while the last part of the cooling evolution, from about 250 °C to the subsurface temperature, occurred with a rate lower than 10 °C/Ma (Saddiqi et al. 1995 a, b). Even a weak reheating is recorded by the fission tracks in apatite, which is related to the Paleogene subsidence of the metamorphic belt (Saddiqi 1995). In the Western Alps, the initial cooling rate of the Gran Paradiso eclogitic unit is in the same range as that of the As-Sifah unit, but in contrast, the end of the exhumation history still shows a relatively high cooling rate, close to 15–20 °C/Ma (Hurford & Hunziker 1989).

In fact, exhumation of the Saih Hatat and Penninic nappes occurred in contrasting tectonic settings. In Oman, a mechanism of denudation of the metamorphic core complex type (cf. Hodges & Walker 1992, Hill et al. 1992), involving the inversion of the obduc-

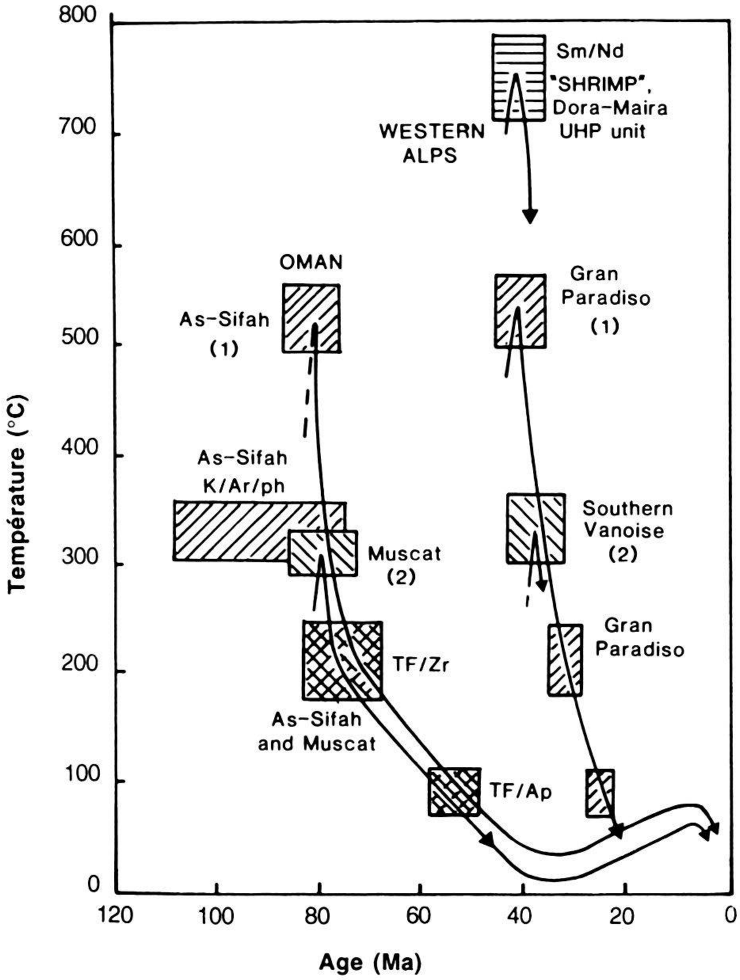



Fig. 11. Temperature-time paths for HP-LT continental units from Oman (Saih Hatat) and the Western Alps. (1): boxes constrained by geological interpretation (ages) and mineralogical thermometry (temperatures). (2): boxes constrained by stratigraphy (ages) and mineralogical thermometry (temperatures). Other boxes from isotopic geochronology or fission track dating, as indicated. Data sources: i) Oman: stratigraphy: Béchennec et al. (1992); K/Ar and $^{40}Ar/^{39}Ar$ ages from Montigny et al. (1988) and El-Shazly & Lanphere (1992); fission track ages from Saddiqi et al. (1995a, b) and Saddiqi (1995). ii) Western-Alps: Dora-Maira (DM) after Tilton et al. (1991) and Gebauer (1995); Gran Paradiso peak temperature after Ballèvre & Merle (1993) and fission track cooling ages after Hurford & Hunziker (1989); Vanoise box after Gillet & Goffé (1988).

tion thrust contact, is supported by the top-to-the-north kinematic indicators associated with unloading (Michard et al. 1994). Unroofing of the Saih Hatat metamorphic window would have occurred in the frame of a regional plate divergence event related to the activity of the proto-Carlsberg ridge east of Oman (Mountain & Prell 1990). In the Western Alps, some authors also suggested that the exhumation of the ICM could be referred to an extensional event during Late Cretaceous times (Butler 1986, Deville 1990, Ballèvre & Merle 1993). However, this way of thinking is strongly jeopardized as soon as a Late Eocene age is accepted for the Penninic peak metamorphism. It would be unlikely to imagine a plate divergence event in the Western Alps during the Oligocene, which in contrast corresponds to a time of advanced Alpine collision (Fig. 9E; e.g. Schmid et al. 1987, Stampfli & Marchant in press) connected to ongoing convergence between Europe and Africa, at a rate higher than 1 cm/year (Le Pichon et al. 1988). The mechanism of exhumation hypothesized for Saih Hatat cannot be applied to the ICM. The metamorphic structure of the latter massifs is characterized, at least in Gran Paradiso and Dora-Maira, by a contractional thrust at their base (UHP-HP eclogitic units over blueschist units, cf. Figs. 6, 7; Michard et al. 1993, Fig. 4), which is unknown in Oman. Exhumation by transverse extension of the thickened wedge (Avigad 1992) is structurally documented for the latest stages of unloading (Philippot 1990, Henry et al. 1993). In contrast, the uplift of the UHP-HP units up to crustal depth is entirely open to speculation. A forced flow mechanism as suggested by Wheeler (1991) and Michard et al. (1993), but with a less protracted timing, would be reconcilable with the ongoing convergence setting. The outward migration of the thrust front would account for the collapse and thinning of the inward metamorphic pile (cf. Jolivet et al. 1994, Froitzheim et al. 1994). Slab breakoff is likely to have affected the subducting plate at ca. 40–30 Ma (Fig. 9D) and would be responsible both for the Periadriatic magmatism (e.g. Biella granite, Fig. 9E) and for the buoyant rise of released crustal slices from mantle depths (von Blankenburg & Davies 1995).

## 5. Conclusion

R. Trümpy (1991) remarked that the greatest danger for the comprehension of a chain like the Alps is parochialism. It was inviting to avoid this danger by going 6000 km-far from the parish churches of the Alps and looking, after many others, to the geology of the Alps from the minaret of an Oman mosque. The point was that probably the pre-collisional stage still preserved in the Oman-Makran transect should roughly represent the state of the Alpine transect during Late Cretaceous time. We hope to have demonstrated that the reality is quite different. The initial conditions were very distinct from one transect to the other, when the Africa-Eurasia convergence began and, additionally, the successive orogenic processes were also completely distinct.

In Oman, the youngest part of the oceanic lithosphere, formed a couple of Ma ago, was able to form the upper plate of an intra-oceanic, shallow-dipping subduction zone (detachment). The oldest (Permian-Jurassic) parts of the ocean were subducting along the Makran active margin during Late Cretaceous, at the same time as the detached young oceanic lithosphere encroached on the Arabian margin (obduction). This margin subducted for a while under a thick oceanic wedge. HP-LT metamorphism occurred at the same time in both subduction zones (sensu lato).

In the Western Alps, the Ligurian oceanic lithosphere was narrow, old and not much supplied by basaltic-gabbroic crust. This high-density lithosphere was bound to subduct under the Austro-Alpine collisional wedge formed earlier (Mid-Cretaceous) and to the East. The small ophiolitic remnants of this narrow ocean are slivers sampled on top of the Ligurian lithosphere during its subduction. They were obducted onto the European margin when the continent-continent collision began, during Late Eocene time. UHP and HP-LT metamorphism was triggered at the same time in the subducting European margin. This Penninic metamorphism post-dates that of the Austro-Alpine by about 40 Ma.

The comparison between the Oman Mountains and Western Alps required a reappraisal of the Alpine tectonic model. The Argandian position of the oceanic suture, as actualized by C. Sturani (1973), is still valid and strongly supported by the contrast between the Penninic and Austro-Alpine/South-Alpine pre-Alpine relics. By contrast, the Eoalpine (Late Cretaceous) isotopic ages introduced during the 70's-early 80's for the Upper- and South-Penninic eclogitic metamorphism are to be re-interpreted, in particular in terms of excess Ar. This is required by the recent and reliable isotopic datings, as well as by the careful examination of the stratigraphic data. It is worth noting that also in Oman, the eclogite-facies rocks yielded apparent K/Ar and Ar/Ar ages older than the coeval, lower-grade rocks.

In spite of their different geotectonic setting, the Oman and Penninic HP-LT belts present similarities. In both cases we deal with a subducting thinned continental crust. The recorded P-T paths are similarly concave in both cases. However, cooling below 250 °C was slower in Oman than in the Penninic. Exhumation of the Oman HP-LT core complex occurred in a regionally divergent setting, while exhumation of the Penninic units occurred in a strongly convergent setting and from the depth of a migrating orogenic wedge.

The main difference between the Western Alps and Oman-Makran transects does not lie in the (provisional) absence of collision in the latter, but in the contrasting pre-collisional evolutions. The pre-collisional setting, which is preserved in Oman almost in its Late Cretaceous state, was never realized in the Western Alps. In other words, even after the total consumption of the Gulf of Oman by the Makran subduction, the resulting belt will be distinct from the present Alpine collision belt. This could be checked by looking at the geology of Zagros and Taurus (e.g. Michard et al. 1984, Okay 1989, Yazgan & Chessex 1991, Yilmaz 1993).

## Acknowledgments

Reviews by E. Gnos and G. Stampfli helped considerably improve a first version of this paper. Some of the strong objections formulated by G. Stampfli turned out to be particularly fruitful. We thank S. Schmid and N. Froitzheim for their stimulating support and thorough review of the final paper. F. Mion and L. Mamou helped us with the figures. We dedicated this paper to the memory of C. Sturani for the 20th anniversary of his death in a fossiliferous quarry of the Montferrato. One of us (A.M.) benefited from many friendly discussions with him, both in the field and in the laboratory, concerning the Europe-Adria relationships in the Western Alps and northern Apennines.

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Manuscript received April 14, 1995
Revision accepted November 11, 1995