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S. MARTIN, P. TARTAROTTI and G.V. DAL PIAZ

THE MESOZOIC OPHIOLITES OF THE ALPS: A REVIEW

Abstract. The Alpine ophiolites are mostly concentrated in the inner and axial sectors of the Penninic nappe pile, as slices of the Jurassic Tethyan ocean (Piedmont-Ligurian and Penninic basins) which opened between the Europe and Adria (Africa) passive margins. The ophiolitic units consist of Mesozoic metasediments (calcschists-schistes lustrés-Bündnerschiefern) and mafic-ultramafic rocks, also including slices of subcontinental mantle. Metasediments with flysch affinity and huge ophiolites are predominant in the external and internal sectors of the Penninic zone, respectively. These sequences display polyphase metamorphic signatures and related deformations of Cretaceous-Tertiary age, i.e., a subduction-related eclogite (locally coesite-bearing) to blueschist facies imprint and/or a greenschist to amphibolite facies re-equilibration, recording the collisional thermal relaxation. Mineral and/or textural relics of pre-Alpine protoliths are locally preserved in low-strain domains, hence a comparison with equivalent sections of the oceanic lithosphere from the present-day oceans may be attempted. Despite the pervasive transposition and the absence of a convincing sheeted dyke complex, the ophiolites record place to place a complete lithological sequence that is typical of the world's best preserved ophiolitic complexes. Nevertheless, the ophiolitic associations frequently display anomalous stratigraphic features. Most mafic rocks are derived from tholeiites with normal-MORB type affinity, which likely originated in mature ridges, even if intra-plate geochemical signatures were also locally reported. A first group of supposedly syn-rift gabbros and/or subcontinental mantle lherzolites may be locally associated with the oceanic sequences. Contrasting models of paleogeographic restoration and kinematic evolution proposed so far for the ophiolitic units in the Alpine nappe pile are summarized in this paper. In contrast with traditional models that interpretate the Alpine ophiolites as sutures of two or even more oceanic gaps alternating with continental blocks, other models also suggest that the Alpine ophiolites are dismembered records of a single oceanic basin, which were tectonically scattered at different structural levels of the nappe pile during the Alpine convergence.

INTRODUCTION

Mesozoic ophiolites crop out along the entire Alpine chain as metamorphic units which are scattered at different structural levels of the present nappe pile, from the uppermost Platta-Arosa unit to the lowest and external Versoyen and Valais units (Figs. 1, 2). Since the famous work by Argand (1911, 1916), all ophiolites from the Western-Central Alps have been referred to the Penninic zone (refs. in Dal Piaz and Dal Piaz, 1984). In other words, these ophiolites were envisaged as members of both the Penninic nappe belt and the Penninic paleogeographic domain through palinspastic restoration back to Jurassic times. Moreover, the whole Penninic zone was envisaged as a continental domain in which the ophiolites were emplaced as ensialic igneous bodies. In the modern literature the term Penninic is generally still adopted for defining the pile of basement and cover nappes located between the Penninic front and the Austroalpine basal thrust, as well as the Swiss and Austrian ophiolitic units and their inferred sources. Nevertheless, this definition may generate confusion between the present geometrical setting

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Dipartimento di Geologia, Paleontologia e Geofisica, Università degli Studi, via Giotto 1, 35137 Padova, Italy.

and a restoration of the oceanic lithosphere that, as discussed later, cannot be unequivocally defined.

The Alpine ophiolites are currently interpreted as tectonic units sliced during the Alpine convergence from the oceanic lithosphere of the western Tethys, that opened in the Late Jurassic between the Europe and Adria (Africa) passive margins. In the Alpine Tethys the occurrence of either a single Piedmont ocean or of two separated oceanic channels (South-Penninic/Piedmont and North-Penninic/Valais basins) has been traditionally envisaged in the French-Italian Alps (e.g. Sturani, 1973, 1975; Dal Piaz, 1974; Beccaluva et al., 1984) and from the Valais to the Eastern Alps, respectively (e.g. Trümpy, 1980; Koller and Hock, 1987; Schmid et al., 1990). Other interpretations have been recently suggested by applying modern orogenic wedge models to the Alpine evolution (Platt, 1986; Marthaler and Stampfli, 1989; Polino et al., 1990; Stampfli and Marthaler, 1990).

The distribution of ophiolite associations throughout the Alps displays a noticeable lithological zonation, as shown in maps and schematic sections across the Alpine chain (Figs. 1, 2, 3, 9). For instance, most of the mafic and ultramafic bodies are concentrated in the inner and axial sectors of the chain (Piedmont/South-Penninic zone), whereas metasedimentary sequences of flysch affinity englobing minor ophiolitic slices mainly occur in the external North-Penninic (Valais) zone. This heterogeneous distribution has been attributed in the past to different paleogeographic settings, as different basins or different sectors of the same basin. The Alpine ophiolites are also characterized by an orogenic metamorphic imprint that, by contrast, is missing in the Northern Apennine ophiolites (Ligurian nappe; Abbate et al., 1988). The orogenic metamorphism is recorded throughout the Alps by polyphase signatures acquired under subduction-related high- P/low T conditions to continental collision-related Barrovian conditions operating from the Cretaceous-Paleocene (Eoalpine stage) to Eocene-Lower Oligocene (Mesoalpine) (e.g., Frey et al., 1974; Niggli, 1978; Hunziker et al., 1989; Koller and Hock, 1990, 1992; Polino et al., 1990).

It is well known that the Alps are a classic site for eclogite facies metamorphism. Typical eclogites and related rocks were discovered in ophiolite and basement units during the last century (inventory and refs. in Frey et al., 1974; Compagnoni et al., 1977; Niggli, 1978; Droop et al., 1990; Dal Piaz et al.,1993) and they were described early from the petrographic and geochemical point of view also (for the Western Alps: Franchi, 1895, 1897, 1902; Novarese, 1985; Zambonini, 1906; Dal Piaz, 1928). Up to some decades ago, these eclogites and related rocks were interpreted as a burial effect of Tertiary nappe stacking (e.g., Niggli, 1970). The Cretaceous-Paleocene (Eoalpine) age of the high-P metamorphism in ophiolites and basement rocks was first documented in the Western Alps through radiometric methods (Dal Piaz et al., 1972; Bocquet et al., 1974; Hunziker, 1974) and later throughout in the Alpine mountain chain, although with some debated exceptions (Hunziker et al., 1989; Polino et al., 1990; Droop et al., 1990; Dal Piaz et al., 1993, and refs.therein). The Eoalpine radiometric data are corroborated by detrital ophiolitic material from the Walsertal Upper Cretaceous flysch, which displays clastic blueschist facies minerals (glaucophane, crossite, lawsonite) deriving from exotic high-P ophiolites early exposed at the erosion surface (Winkler and Bernoulli, 1986).

The high-P ophiolitic units from internal to external sectors of the Alps exhibit contrasting metamorphic peak conditions. In the Western Alps, the Piedmont zone shows a zoned distribution of the Eoalpine metamorphism, grading from eclogite to blueschist facies conditions from the inner side towards the foreland (Bearth, 1967, 1974; Bocquet, 1974; Dal Piaz, 1974; Frey et al., 1974; Dal Piaz et al., 1978; Kienast, 1983; Pognante et al., 1983; Messiga et al., 1983; Desmons, 1986, 1989; Goffé and Chopin, 1986; Droop et al., 1990; Dal Piaz et al., 1993). Blueschist relics are recorded also in the more external Versoyen-Valais ophiolites yielding a few Cretaceous-Eocene radiometric ages (Bocquet et al., 1974; Schurch, 1987).

By contrast, the high-P metamorphism is highly scattered and more scarcely recorded in the Central Alpine ophiolites. It is represented by sporadic occurrences of blue amphiboles (e.g., the Mesocco zone and Platta nappe; Oberhänsli, 1978, 1986; Philipp, 1982; Hunziker et al., 1989) and locally also by classic eclogites, mainly in the structurally lower Adula-Cima Lunga nappe (Evans and Trommsdorff, 1979; Heinrich, 1986). A few blue amphiboles from the Central Alps yield radiometric ages of about 100 Ma (Deutsch, 1983; Deutsch and Steiger,

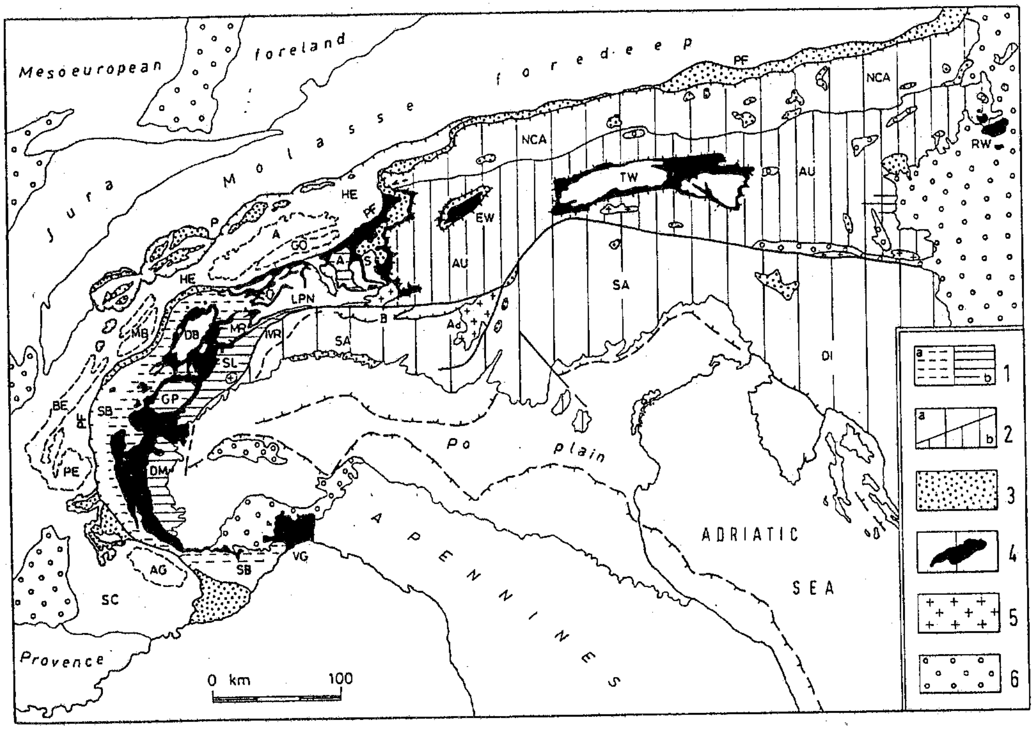

Fig. 1 - Tectonic map of the Alps and distribution of Mesozoic ophiolites and related flysch units (modified from Polino et al. 1990). 1) Eoalpine, locally Eocene (e.g., Briançonnais, Adula) blueschist (a) and eclogite facies (b) metamorphism in continental units of the Western and Central Alps; 2) Eoalpine very low grade (a) to greenschist and amphibolite facies metamorphism (b) in the Eastern Austroalpine, including the supposedly Eoalpine eclogites from Koriden; 3) Cretaceous-Eocene flysch nappes and Gosau beds; 4) undifferentiated ophiolitic units with eclogitic-blueschist or Barrovian imprint (VG: Voltri Group); 5) major Tertiary plutons along the Periadriatic fault system (B: Bergell, Ad: Adamello); 6) Oligocene and/or Miocene basins. Tectonic units from the foreland to the interland: Prealpine klippen (P); Helvetic-Dauphinois (HE), Aar-Gotthard (A-GO), Mont Blanc (MB), Belledonne (BE), Pelvoux (PE) and Argentera (AG) external massifs and cover units, including the Subalpine chains (SC); Penninic frontal thrust (PF), lower Penninic nappes (LPN), including the Adula nappe (A); Gran St.Bernard (Briançonnais) system (SB), upper Penninic Monte Rosa (MR), Gran Paradiso (GP), Dora Maira (DM) and Suretta (S) nappes; Western Austroalpine Sesia-Lanzo (SL) and Dent Blanche (DB) nappes.; Eastern Austroalpine (AU) cover and basement nappes, including the Northern Calcareous Alps (NCA); Southern Alps (SA), Ivrea zone(IVR) inclusive; Dinarides (DI); Engadine (EW), Tauern (TW) and Rechnitz (RW) windows.

1985; Hunziker et al., 1989). Younger ages have alternatively been suggested for the eclogitic metamorphism (Gebauer et al., 1992). In the Eastern Alps, an Eoalpine eclogitic and/or blueschist imprint at around 90-60 Ma is documented for the Tauern and Rechnitz ophiolites (Miller, 1974; Raith et al., 1978, 1980; Holland, 1979; Frank et al., 1987; Droop et al., 1990; Dal Piaz et al., 1993).

The Mesoalpine overprint is characterized by greenschist to amphibolite facies assemblages with temperature increasing down across the collisional nappe pile (e.g. Frey et al., 1974; Niggli, 1978).

This review briefly summarizes geographic distribution and tectonic allocation of the Mesozoic ophiolitic units from the Western Alps to the Central and Eastern Alps, as well as the main lithological associations, geochemical features and metamorphic evolution. Emphasis is addressed to Italian occurrences and the related literature. A detailed tectonic base is provided by the Structural Model of Italy at 1:500 000 scale (sheets 1 and 2, CNR 1990). This paper also discusses classic models and alternative interpretations of the paleogeographic restoration, generation and kinematics of the Alpine ophiolitic nappes.

THE WESTERN ALPS

Most of the Western Alpine ophiolites are concentrated in the Piedmont "Greenstone" (=Pietre Verdi) zone. The Mesozoic age of the associated calcschists was definitely established by Franchi (1898). The Piedmont zone overrides the upper and middle Penninic basement nappes (Monte Rosa-Gran Paradiso-Dora Maira and Grand St.Bernard) and in turn is capped by the Austroalpine Dent Blanche and Sesia-Lanzo system. Minor ophiolites are associated with the predominating metasediments of the external (lower) Penninic domain from the Versoyen to the Valais zone (e.g., Trümpy, 1960, 1980; Dal Piaz, 1974; Dietrich et al., 1974; Dietrich, 1980; Polino et al., 1990 and refs. therein).

The North-Western sector

The Piedmont ophiolite zone of the Aosta valley may be regarded as a structurally composite nappe system sandwiched between the Austroalpine and the underlying Upper and Middle Penninic nappes (Figs. 2, 3). On the basis of contrasting lithology and metamorphic evolution the Piedmont ophiolites from the northern Aosta Valley and southern Valais are usually divided into two main tectonic units: the greenschist facies-dominated Combin zone and the eclogitic Zermatt-Saas zone (Bearth, 1953, 1959, 1967, 1973, 1974; Dal Piaz,1965, 1974, 1976, 1988, 1992; Dal Piaz and Ernst, 1978; Bearth and Schwander,1981; Kienast,1983; Ballèvre et al., 1986; Sartori, 1987). North of the east-west-trending Aosta-Ranzola fault the Combin Zone is structurally composite, including the Combin (Dal Piaz, 1988) or Tsaté (Marthaler, 1984; Vannay and Allemann, 1980) ophiolitic nappe and the underlying to internally sandwiched ophiolite-free Permian-Cretaceous décollement unit, known either as the Pancherot-Cime Bianche-Bettaforca unit (Italian side; Dal Piaz, 1988) or as Mont Fort-Frilihorn unit (Swiss side; Marthaler, 1984; Escher et al., 1986; Sartori, 1987). The Combin zone is the tectonic sole of the northern Austroalpine Dent Blanche-Mt.Mary-Pillonet klippen, altogether forming a couple of eclogite-free nappes which experienced the same metamorphic evolution (Ballèvre et al., 1986; Polino et al., 1990). The Combin zone is locally separated from the underlying Zermatt-Saas nappe through the eclogitic Austroalpine Etirol-Levaz slice (Kienast, 1983; Ballévre et al., 1986; Dal Piaz, 1992).

South of the Aosta-Ranzola fault, the Piedmont ophiolites exhibit lithological and metamorphic features similar to the Zermatt- and Combin-type sequences (Dal Piaz Gb., 1928; Elter, 1971, 1972), but their tectonic repartition is a matter of debate. The occurrence of a unique eclogitic Zermatt-type ophiolitic nappe has been postulated by Ballèvre et al. (1986) as the footwall of the eclogitic southern Austroalpine klippen (Fig. 3). By contrast, several tectonic units with eclogite or greenschist facies signatures have been envisaged here within the Piedmont zone (Elter, 1971; Dal Piaz and Nervo, 1971; Nervo and Polino, 1977; Dal Piaz et al., 1979; Battiston et al., 1984; Benciolini et al., 1984). Eclogite and garnet-glaucophanite bodies are widely seen between the Austroalpine klippen and the Gran Paradiso nappe, mainly deriving from Fe-Ti-rich gabbro and basalt protoliths (Dal Piaz and Nervo, 1971; Ballèvre et al., 1986; Benciolini et al., 1987, 1988; Martin and Tartarotti, 1989). These ophiolites are locally covered by manganiferous metacherts with classic high-P assemblages (Martin, 1982; Mottana, 1986; Martin and Kienast, 1987).

The Combin-type ophiolite sequences chiefly consist of carbonate to terrigenous flysch-type metasediments (calcschists *sensu lato*), commonly including multiple interleavings of tabular greenschist facies metabasalts (prasinites; Novarese, 1895) and minor serpentinite slices. Major ophiolitic bodies with a few manganiferous quartzites and other oceanic metasediments may locally predominate in the top section of the Combin nappe (Dal Piaz et al., 1979; Baldelli et al., 1983; Sartori, 1987; Vannay and Allemann, 1990). Classic Combin ophiolites have been described, for instance, in various southern Valais localities (Iten, 1948; Bearth, 1953,1967; Zimmermann, 1955; Mazurek, 1986; Sartori 1987; Wust and Silverberg, 1989), in the Valtournanche-Alagna area (Dal Piaz,1965, 1974, 1976, 1988, 1992; Dal Piaz and Ernst,1978; Le Goff, 1986; Vannay and Alleman, 1990), in the Ollomont valley (Diehl, 1938; Diehl et al., 1952), and in the Graian Alps (Elter, 1971, 1972). The geochemical signature of these ophiolites generally does not differ from the Zermatt-Saas ones (Dal Piaz et al., 1981; Beccaluva

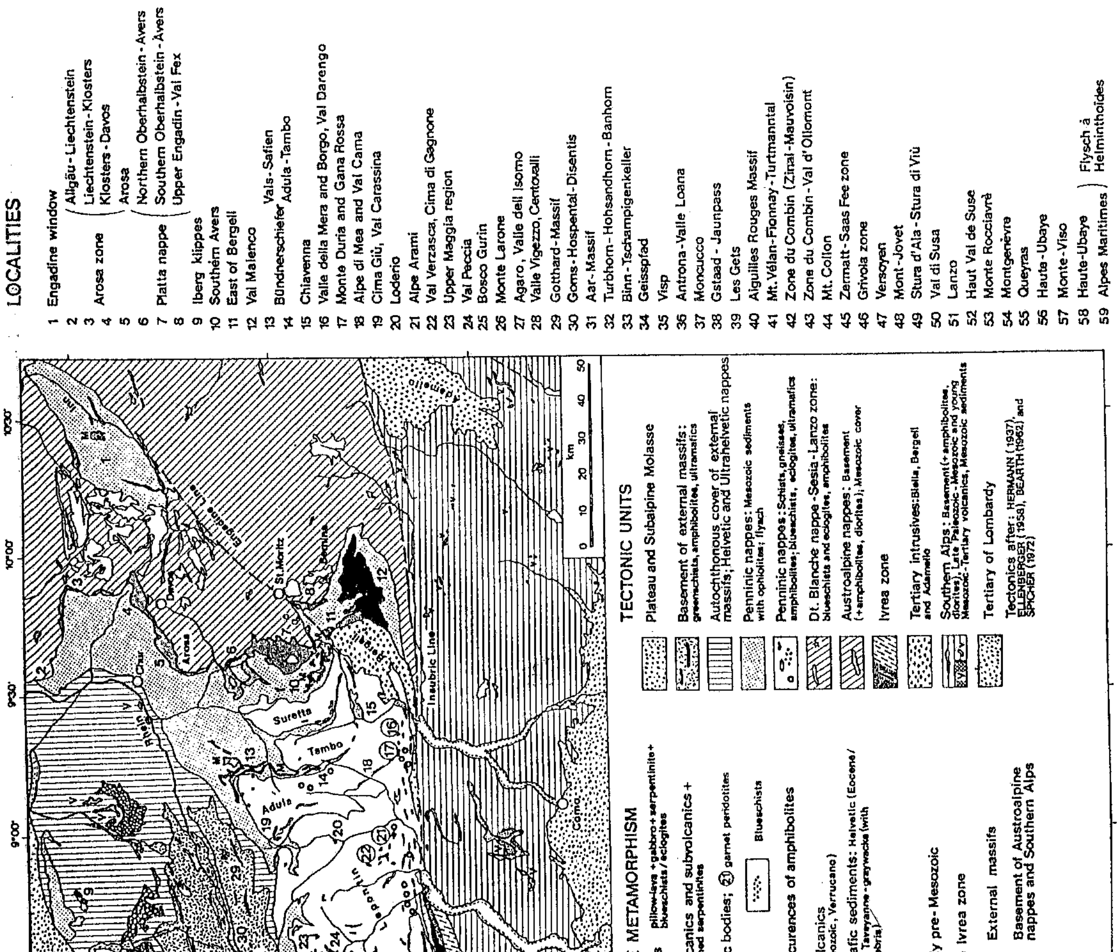

Fig. 2 - Inventory of Mesozoic ophiolites and other older mafic-ultramafic bodies from the Western and Central Alps (after Dietrich et al., 1974).

et al., 1984).

The Combin-type units display a peculiar greenschist facies imprint of mid-Tertiary (mesoalpine) age (38 Ma, according to Hunziker, 1970; Frey, et al., 1974; Delaloye and Desmons, 1976), that undoubtedly predates the transecting Oligocene (31-30 Ma) andesite-lamprophyre dykes (Dal Piaz et al., 1979; Diamond and Widenbeck, 1986). In addition, relatively high-P relics have been found locally in metabasalts, metagabbros and metasediments, represented by blue amphiboles, paragonite, phengite, garnet and rutile (e.g., Kienast, 1973; Dal Piaz, 1976; Ernst and Dal Piaz, 1978; Baldelli et al., 1983; Ballèvre et al., 1986; Sperlich, 1988; Vannay and Allemann, 1990). These relics may be related to a subduction event of unknown age, but probably late Eoalpine (Paleocene) or early Eocene, due to the occurrence of capping Upper Cretaceous flysch protoliths within the Combin (Tsaté) calcschists sequence (Marthaler, 1984; Sartori, 1987).

The Zermatt-Saas-type ophiolites are mainly composed of huge mafic and ultramafic bodies with minor metasedimentary cover remnants. Alpine transposition and metamorphic reworking have essentially effaced the stratigraphic relations and primary features. Therefore, only tentative reconstructions may be done by comparison with classic, less tectonized ophiolitic complexes described elsewhere and with present-day oceanic lithosphere. Nevertheless, scattered magmatic textures and structures have been locally observed, especially in the Zermatt region (Bearth, 1967; Barnicoat and Fry, 1986), in the Aosta valley (Dal Piaz, 1974; Tartarotti, 1988; Martin and Tartarotti, 1989) and in the Palavas-Monviso area (Bearth et al., 1975; Lombardo et al., 1978), where the shape of flattened pillow lavas or pillow breccias and gabbro-cumulate peridotite textures are still preserved.

The Zermatt-Saas-type ophiolites consist of the following main lithology which clearly exhibits oceanic affinity: 1) serpentinized mantle peridotites, usually cut by rodingite dykes, often grading into ophicalcites (Bearth, 1967; Dal Piaz, 1969; Dal Piaz et al., 1980; Driessner, 1993); 2) Mg-rich and minor Fe-rich metagabbros and related ultramafics, locally associated with more evolved rocks such as metatrondhjemites (Novo et al., 1989); 3) metamorphic massive basalts, pillow lavas, pillow breccias and hyaloclastites; 4) Mn-rich metacherts (Bearth, 1967; Dal Piaz et al., 1979; Mottana, 1986) capped by a sedimentary cover with oceanic affinity, consisting of marbles and minor calcschists. In spite of transposition and high-P metamorphic imprint, the supra-ophiolitic cover recalls the sequence of radiolarian cherts- Calpionella limestones- "Palombini" shales capping the ophiolites (Ligurian nappe) from the Northern Apennines (e.g., Tartarotti et al., 1986; Busconi et al., 1987). Furthermore, the metasedimentary cover locally includes an eclogitic subduction melange, known as the Rifelberg-Garten complex (Dal Piaz, 1965, 1988; Bearth, 1967; Dal Piaz et al., 1979).

Although a reliable volume proportion of these lithologies is difficult to establish, a large part of the Zermatt-type ophiolites is represented by serpentinites. Gigantic mantle-derived antigorite serpentinite bodies occur on both sides of the Aosta valley (M. Avic and M.Rosso di Verra-Breithorn massifs). These bodies are often mantled by thick ophycarbonates (Driessner, 1993) or locally by thin rodingitic reaction zones along the contact with the country mafic ophiolites and metasediments (Dal Piaz, 1969). Serpentinite bodies also include boudinaged rodingitic dykes, mainly of gabbro composition, locally so abundant that they may be regarded as a sort of "sheeted complex" within the lithosferic mantle (Dal Piaz, 1969; Dal Piaz and Ernst, 1978).

A normal-MORB geochemical affinity is reported for the mafic rocks (e.g., Dal Piaz et al., 1981; Beccaluva et al., 1984; Pfeiffer et al., 1989). On the basis of lithological association and geochemical signature, the Zermatt-Saas-type ophiolites have been so far unanimously interpreted as slices of oceanic lithosphere.

All these lithologies are affected by an eclogitic imprint of Cretaceous age (100-90 Ma, according to Bocquet et al., 1974; Hunziker, 1974; Chopin and Maluski, 1980; Desmons, 1986, 1989; Hunziker et al., 1989; Polino et al., 1990). Eclogitic asemblages are particularly well preserved by the classic Allalin body (Bearth, 1967; Fry and Barnicoat, 1987; Droop et al., 1990) and by differentiated Fe-gabbros from the Italian side (Baldelli et al., 1985; Benciolini et al., 1987, 1988, and quoted refs.). Mineral equilibria in eclogitic gabbros suggest metamorphic conditions of T=450°-650°C for minimal pressures of 1.0-1.5 GPa (Chinner and Dixon, 1973;

Ernst and Dal Piaz, 1978; Benciolini et al., 1988; Martin and Tartarotti, 1989). Representative T-values obtained from mafic eclogites of the Aosta valley are shown in Fig.4. In the northern Aosta valley, temperature for the eclogitic climax seems to attain higher values than in the southern ophiolites (Ernst and Dal Piaz, 1978; Oberhänsli, 1980, 1986). Coesite has been recently found within Mn-Mg-rich garnets from Cignana (Valtournanche) metacherts of the Zermatt-Saas nappe, showing metamorphic peak conditions of T=590°-630°C and P=2.6-2.8 GPa (Reinecke, 1991). Also the titanian clinohumite-olivine- clinopyroxene assemblage and further signatures found in the Zermatt-Saas-type antigorite serpentinites are representative of the high-P event (Dal Piaz et al., 1980). Predominant reactions observed in Chatillon ophycarbonate breccias suggest conditions of approximately 500°C and 1.2 to 0.8 GPa (Driessner, 1993, and refs. therein). These assemblages are missing in mantle ultramafics of the Combin zone.

Post-eclogitic evolution of the Zermatt-Saas ophiolites is typically characterized by multiple re-equilibration under decreasing pressure, at first under high T-blueschist facies conditions (glaucophane-garnet-bearing assemblages) and later under Ab-amphibolite and greenschist facies conditions (e.g., Bearth, 1967; Ernst and Dal Piaz, 1978; Kienast, 1983; Ballèvre, 1988). A scheme for the decompressional evolution is shown in Fig. 5, according to selected examples from the Aosta valley ophiolites (e.g. Martin and Tartarotti, 1989; Reinecke, 1991).

The Antrona ophiolites (Ossola valley) display lithological association and metamorphic evolution similar to those recorded in the Zermatt-Saas nappe, even if the Antrona ophiolites correspond to a lower structural level (below the Monte Rosa nappe) of the Piedmont zone (Blumenthal, 1952; Bearth, 1954, 1957, 1958; Laduron, 1976; Beccaluva et al., 1984; Colombi and Pfeiffer, 1986; Pfeiffer et al., 1989; Dal Piaz, 1992). The Antrona nappe consists of predominant massive to pillowed metabasalts and minor metagabbros with polyphase metamorphic imprint evolving from eclogite facies to Barrovian conditions. A transitional-MORB affinity is supported by geochemical signatures (Beccaluva et al., 1984; Pfeifer et al., 1989). The nappe also encompasses some antigorite serpentinite bodies, rodigitic dykes and some remnants of the post-volcanic sedimentary cover represented by thin layers of impure quarzites, marbles and calcschists, broadly transposed within the ophiolitic sequences.

Relict pods of eclogites record metamorphic conditions of T=550°-600° C and P=1.8-2.4 GPa (Colombi and Pfeifer, 1986; Ladeuze, 1988). The Mesoalpine re-equilibration increases from the greenschist facies conditions (first event: T=450°-500° C and P= 0.6-1.0 GPa, Ganguin, 1988; second event: 450°-500°C and P=0.3- 0.5 GPa) on the western side, to amphibolite facies conditions (T=770°C, Pfeifer et al., 1989 and refs.) on the eastern side. Compared with the Zermatt-Saas ophiolites, the amphibolite facies conditions attained by the Antrona ophiolites during the post-nappe Mesoalpine metamorphism depend on the different structural depths within the collisional nappe pile and Lepontine thermal doming.

The Lanzo peridotite body

The Lanzo massif is a huge and well preserved mantle peridotite body of the Piedmont zone lying between the Viù- Locana ophiolitic strip and the eclogitic basement of the Austroalpine Sesia-Lanzo zone (Fig. 6). Peridotites are largely preserved in the massif core, whilst in the border zone they are completely serpentinized. These ultramafics mostly consist of spinel-lherzolite, partly re-equilibrated under plagioclase facies conditions (Nicolas and Jackson, 1972; Nicolas, 1984; Beccaluva et al., 1984; Ottonello et al., 1984). Minor cpx-poor lherzolite and harzburgite types have also been locally observed, as well as discontinuous layers of dunite inside the lherzolites (Boudier and Nicolas, 1972; Boudier, 1976). Pyroxenite layers, gabbro and diabase dykes are contained in the peridotites. Dykes exhibit chemical compositions quite similar to that of the mafic dykes reported from the Ligurian-Piedmont ophiolites (e.g., Beccaluva et al., 1976; Venturelli et al., 1981).

The origin of the Lanzo massif is still a matter of debate. It was first interpreted as an ophiolitic mushroom extrusion within an ensialic Piedmont basin (Nicolas, 1966) and then as a slice of the Southalpine continental mantle (Nicolas et al., 1972; Boudier, 1976). Successively the Lanzo peridotites were referred to the oceanic lithosphere (Lombardo and Pognante, 1982; Pognante et al., 1986; Bodinier et al., 1986; Bodinier, 1988) and this interpretation has been accepted

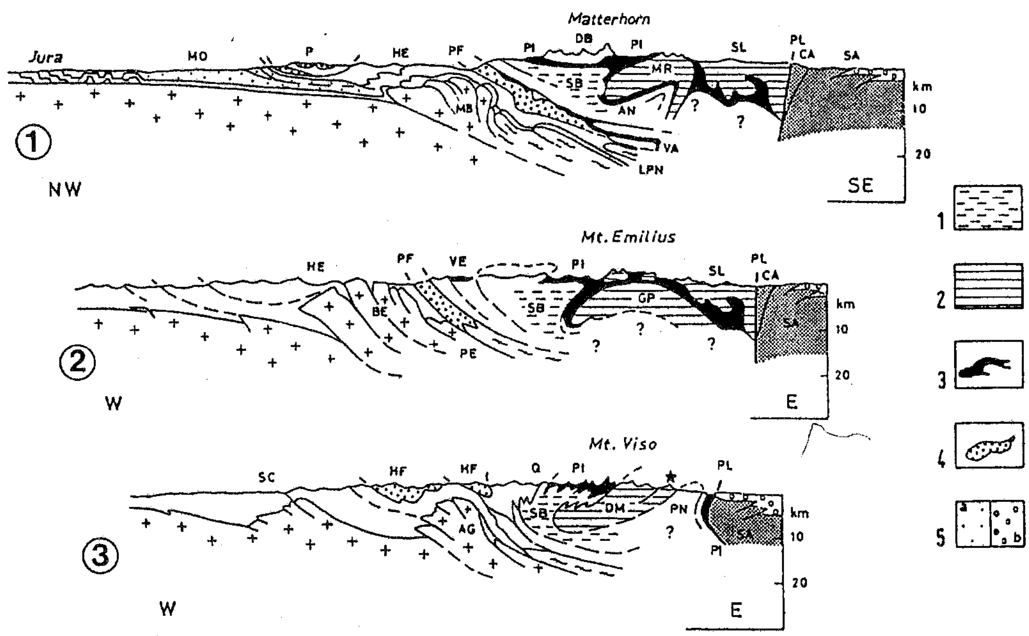

Fig. 3 - Schematic cross-sections through the Pennine Alps (1), Graian Alps (2) and Cottian Alps (3). Legend: 1) blueschist and 2) eclogite facies imprint (b) in Penninic and Austroalpine basement and cover units; 3) undifferentiated ophiolites and related metasediments; 4) Flysch décollement units, mostly Cretaceous; 5) external molasse (a) and Po plain Tertiary deposits (b). Tectonic units: Prealpine klippen (P); Helvetic-Dauphinois (HE) basement (MB: Mont Blanc, BE: Belledonne, PE: Pelvoux and AG: Argentera) and cover units; Penninic frontal thrust (PF); Versoyen (VE)-Valais (VA) ophiolitic and flysch units; Penninic basement nappes: Ossola-Tessin lower Penninic nappes (LPN), SB, PN: middle Penninic Gran St. Bernard-Briançonnais (SB) and Pinerolese (PN) nappes; upper Penninic Monte Rosa (MR), Gran Paradiso (GP) and Dora Maira (DM) nappes; Antrona (AN) and Piedmont (PI) ophiolitic units (Q: Queyras, HF: Helmintoid flysch); Western Austroalpine Sesia-Lanzo (SL) and Dent Blanche (DB) nappes; Periadriatic (Canavese) fault system (PL); Canavese zone (CA); Southern Alps (SA).

also by Boudier and Nicolas (1985) who pointed out the similarities between the Lanzo body and the Mesozoic ophiolites of the Ligurian-Piedmont nappes. There were also discussions regarding the genetic significance of the included gabbro and diabase. According to the subcontinental mantle model, the mafic dykes are the product of mantle partial melting due to viscous heating during the mantle slice emplacement (Boudier, 1976, 1978). According to the ophiolitic model, the gabbro and basalt dykes may be linked to differentiation processes within small magma pockets inside the rising sub-oceanic mantle (Lombardo and Pognante, 1982; Pognante et al., 1985; Lemoine et al., 1987; Lagabrielle et al., 1987; Bodinier et al., 1986).

In our opinion, this debate on the Lanzo body neglects the basic problem of whether a formerly subcontinental lithospheric mantle could be actually denudated during the Mesozoic rifting and emplaced in the Piedmont ocean floor, as hypothesized for instance for the Malenco area (Trommsdorff. et al., 1993). Later, it may have been covered by basaltic flows and/or pelagic sediments, as suggested by the occurrence of metasediment and metabasite remnants presently transposed inside the Lanzo ultramafics, which may be seen as relics of their own oceanic cover (Lagabrielle et al., 1989). From this perspective, the Lanzo peridotites can no longer be interpreted as the source of the associated basalt dykes.

The metamorphic evolution of the Lanzo ultramafic-mafic complex is characterized by a pre-Alpine event occurring under amphibolite facies conditions (Compagnoni et al., 1984). It may be referred to either intra-oceanic deformations (e.g., Mével et al., 1978) or rifting-related lithospheric attenuation (Dal Piaz, 1993). This event was followed by a subduction-induced high-P overprint which is best recorded by mafic eclogites (Compagnoni and Sandrone, 1979; Pognante and Kienast, 1987). Eclogitic mafic dykes and ophiolites with Eoalpine imprint (95 Ma according to Carpena et al., 1986) from the western border of the Lanzo massif exhibit a subsequent greenschist facies equilibration (Bente and Lensch, 1981). In conclusion, the Alpine evolution of the Lanzo body is comparable to that recorded by the Mesozoic ophiolites of the Piedmont zone.

The Versoyen zone

A minor ophiolite unit, known as Versoyen zone, occurs in the Northwestern Alps within the external Penninic domain (Figs. 2, 3). The unit is located between the Mont Blanc massif and the Petit St.Bernard pass, near the French-Italian border (Elter and Elter, 1965; Dal Piaz, 1992). The Versoyen zone was early considered as the frontal part of the supposedly calcschists-bearing Dent-Blanche nappe, deeply involved beneath the middle Penninic units (Hermann, 1938). Later it was generally referred to the Tarentaise Breccias décollement nappe of the external/lower Penninic (Valais) zone (Barbier, 1948, 1951; Trümpy, 1951, 1955, 1958, 1960; Zulauf, 1963; Elter and Elter, 1965; Loubat, 1968; Antoine, 1971, 1972; Dal Piaz, 1974).

The Mesozoic ophiolites mainly consist of serpentinites and massive to pillow metabasalts, while the supra-ophiolitic cover is represented by flysch-type calcschists, phyllites and marbles of Mesozoic age. The décollement nappe also includes some basement slices (or olistoliths?) of Late Paleozoic micaschists and impure quartzites (Schoeller, 1929; Zulauf, 1963; Elter and Elter, 1965). The mafic rocks display MORB-type and island-arc tholeiitic affinity. The early metamorphic imprint is characterized by oceanic alteration through metasomatic reactions between hot hydrothermal fluids and the mafic rocks. The polyphase Alpine overprint is recorded by blueschist facies mineral assemblages of debated Eoalpine age and by a Mesoalpine greenschist facies re-equilibration (Bocquet, 1974; Lasserre and Laverne, 1976; Schurch, 1987; Polino et al., 1990).

The Versoyen ophiolites have been recently interpreted as a cedar-shaped volcanic edifice consisting of subhorizontal basalt sills overlain by pillow lavas (Schurch, 1987). Serpentinites were attributed either to the igneous suite (Loubat, 1975; Schurch, 1987) or to the pre-Triassic crustal basement (Elter and Elter, 1965). A paleogeographic setting very similar to the marginal basins of the California Gulf transform system has been envisaged by Loubat (1975), Schurch (1987) and Cannic et al. (1993), due to the geochemical and lithological affinity of the Versoyen ophiolites with those of the present Guaymas basin. Alternatively, derivation of the Versoyen ophiolites from the Piedmont basin was envisaged by Bocquet (1974) and Lasserre and Laverne (1976). In our opinion, the exotic origin of the Versoyen ophiolites is convincingly supported by the occurrence of the high- P imprint, which is missing in the surrounding décollement units of the Tarantaise Breccias zone.

The South-Western sector

The most important ophiolite bodies from the southern sector of the Western Alps are the eclogitic Orsiera-Rocciavrè and the Monviso complexes cropping out on the inner side of the Piedmont nappe system (Figs. 2, 3, 7). Minor eclogite-free ophiolites are scattered in external sectors of the Cottian Alps where they are associated with calcschists regionally reported as "Schistes lustrés" of the "internal Piedmont series" or Chabrière-type series (Bourbon et al., 1979; Fig.7). Typical blueschist facies assemblages are recorded throughout the area (Bocquet, 1974; Bocquet et al., 1974; Frey et al., 1974; Dal Piaz and Molin, 1978; Dal Piaz et al., 1978; Goffé and Chopin, 1986; Lagabrielle, 1987, and therein refs.), as well as primary cumulate to volcanic structures (Bearth et al., 1975; Bertrand et al., 1982). Among these external Piedmont ophiolites, only the classic, poorly reworked Monginevro body will be described.

The Orsiera-Rocciavrè ophiolitic complex tectonically overlies the Dora-Maira basement nappe which is the southern homologue of the Monte Rosa and Gran Paradiso nappes (Figs. 3, 7). This ophiolitic complex has been subdivided into two tectonic units (Pognante, 1979). The former extends to the south-east and exhibits typical oceanic character. It consists mainly of metagabbros (a few hundreds meters thick) which frequently preserve coarse-grained igneous structures and mineral assemblages. The gabbros contain eclogitized Fe-Ti gabbros occurring

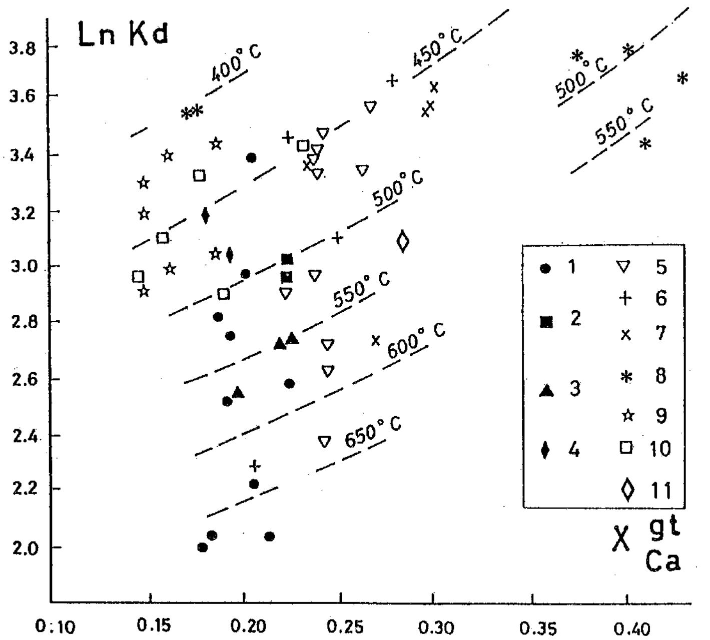

Fig. 4 - LnKd vs. XCa(gt) diagram (Ellis and Green, 1979) of clinopyroxene-garnet pairs from eclogitic ophiolites of northern (full symbols) and southern (open symbols) Aosta valley. Temperature values are obtained at nominal P = 1.0 GPa. Sample location: 1) Zermatt (Oberhänsli, 1980); 2) St. Jacques-Valtournanche (Dal Piaz and Ernst, 1978); 3) Breuil (Dal Piaz and Ernst, 1978); 4) Verres (Baldelli et al, 1985); 5) St. Marcel (Martin-Vernizzi, 1982; Martin and Tartarotti, 1989); 6) Acque Rosse-Urtier, east of Cogne (unpubl. data from the authors); 7) Cogne (Benciolini et al., 1988); 8) Soana valley (Benciolini et al., 1988); 9) Orco valley (Pognante et al., 1986); 10) Locana (Benciolini et al., 1988); 11: Mt) Avic (unpubl. data from the authors).

as layered sequences (ilmenite-magnetite gabbros and gabbro-norites). This sequence is similar to other gabbro- ferrogabbro series occurring elsewhere in the Piedmont zone (Monviso, Zermatt-Saas nappe, lower Susa valley), although it extensively displays the unusual occurrence of magmatic orthopyroxene (Pognante, 1979; Lombardo and Pognante, 1982).

The whole gabbro section overlies huge serpentinite slices. Rodingitic reaction zones are commonly developed in between (Bortolami and Dal Piaz, 1970). The basal serpentinites underline the tectonic contact of the Piedmont zone with the Dora-Maira basement nappe.

The other unit of the Orsiera-Rocciavrè ophiolites consists of tectonic slices of calcschists, garnetiferous quartzites and mafic ophiolites. Thin interleavings of Triassic platform marbles and gt- free quartzites may be interpreted as imbricated décollement remnants of the Dora-Maira sedimentary cover (Pognante, 1979). The metabasites are represented by banded rocks with blueschist- greenschist mineral assemblages and scarce relics of omphacite and garnet. In places these rocks contain lenses of fine-grained eclogites. Relics of original breccia and pillow structures have been locally preserved.

The early metamorphic evolution of the Orsiera-Rocciavrè ophiolites are characterized by an oceanic alteration producing brown hornblende and cummingtonite replacing magmatic pyroxene in the gabbro section. The rocks were successively affected by the polyphase Alpine metamorphism characterized by early high-P mineral assemblages which are particularly widespread in the metagabbro section of the south-eastern unit.

The Monviso complex is one of the widest ophiolitic body out cropping in the Western Alps (Figs. 2, 3, 7). To the east, it is tectonically juxtaposed over the Dora Maira nappe whilst, to the west, it is overlain by the Piedmont calcschists and ophiolitic units of the French Queyras region. On the basis of structural setting and lithological composition, the Monviso complex has been subdivided into several units (Lombardo et al., 1978, Fig. 8): 1) the Vallanta unit; 2) the Costa Ticino series; 3) the Colle del Viso serpentinites and the Passo Gallarino complex; 4) the Viso Mozzo metabasalts; 5) the Lago Superiore unit; 6) the lowermost serpentinites.

1) The Vallanta Unit (VU in Fig. 8) forms the western face and the Monviso summit. It consists mostly of 200-300m thick massive or layered metabasalts which exhibit well preserved eclogitic mineral assemblages. In spite of the Alpine deformations, relict pillow lava structures are still preserved. The tectonic boundary with the underlying Costa Ticino series is marked by serpentinite slices.

2) The Costa Ticino series (CT in Fig. 8) is formed by an overturned sequence of metagabbros, in places preserving magmatic textural relics, and of massive to pillow metabasalts locally covered by oceanic metasediments consisting of quartz-bearing micaschists and quartzites. Serpentinized peridotites containing metagabbro layers occur locally. Even if metamorphic basalt dykes are very common in the metagabbro section, the occurrence of a true sheeted dyke complex cannot be argued. This unit shows well preserved eclogitic assemblages, partly reequilibrated by a blueschist and greenschist facies overprint.

3) The serpentinites of Colle del Viso from the tectonic boundary between the Costa Ticino unit and the Passo Gallarino complex. These serpentinites preserve relics of clinopyroxene, and contain slices of eclogite and eclogitic Fe-gabbro. The ophiolitic sequence at Passo Gallarino consists of eclogitic Fe- gabbros and omphacite-bearing metagabbros which are partly characterized by a pervasive mylonitic foliation.

4) The Viso Mozzo unit is separated from the overlying Passo Gallarino unit by a transposed layer of calcschists capping the underlying banded metabasites. These mafic rocks locally preserve relics of pillow and breccia structures. Also this unit displays an eclogitic imprint.

5) The Lago Superiore unit (MG in Fig. 8) is mostly formed from Cr-omphacite-bearing metagabbros. Well preserved magmatic textures show that these metagabbros derive from plagioclase- clinopyroxene $\pm$ olivine-bearing Mg-gabbros. These rocks are locally cross-cut by fine-grained eclogitic basalt dykes. Pervasive Alpine deformation and metamorphic retrogression produced flaser textures and Ab-bearing mylonites. The Lago Superiore unit is underlain by a tectonic multilayer consisting of calcschists, mylonitic metagabbros and eclogite lenses associated with gabbros or surrounded by serpentinites.

6) The lowermost serpentinites (BS in Fig. 8) occur at the base of the previously described units. These ultramafics consist of antigorite-bearing serpentine-schists preserving clinopyroxene relics, and are transected by rodingitic mafic dykes. Metagabbro slices are interlayered in the upper section of the serpentinites.

Although the Monviso ophiolites were dismembered into several tectonic units, they ecompass the complete classic lithology of the modern oceanic lithosphere, from gabbro and ultramafic cumulates to massive volcanic rocks, pillow lavas and mafic dykes. The oceanic basement is capped by a sedimentary cover which locally preserves the original stratigraphic contact with the volcanic rocks.

The ophiolitic units from the Monviso area show, as a whole, the same metamorphic evolution (Lombardo et al., 1978). During the opening of the Mesozoic Ligurian-Piedmont basin, these rocks suffered early oceanic alteration, producing brown and green hornblende at the expense of magmatic pyroxenes. The alpine metamorphic imprint is characterized by polyphase signatures. The early event of debated Eoalpine or mid-Tertiary age (Monié and Philippot, 1989) produced high-P mineral assemblages such as Na- clinopyroxene, almandine, chloritoid, epidote, rutile in the mafic rocks (Lombardo et al., 1978; Kienast and Messiga, 1987; Pognante and Kienast, 1987). Physical conditions of $T = 400°C \pm 50°C$ and P (minimal) $= 1.0$ GPa have been suggested for the metamorphic climax (Nisio and Lardeaux, 1987). Different physical conditions in undeformed and mylonitic eclogites have been estimated by Philippot and Kienast (1989), ranging from $T = 333°C$-$468°C$ to $T = 236°C$-$513°C$, for a minimal pressure of 1.0 GPa, in the low strain and high strain domains respectively. However, as suggested by same authors, the wide temperature range may be influenced by the composition of the garnet-clinopyroxene pair.

The post-eclogitic evolution is first recorded by relatively high-T blueschist facies assemblages,

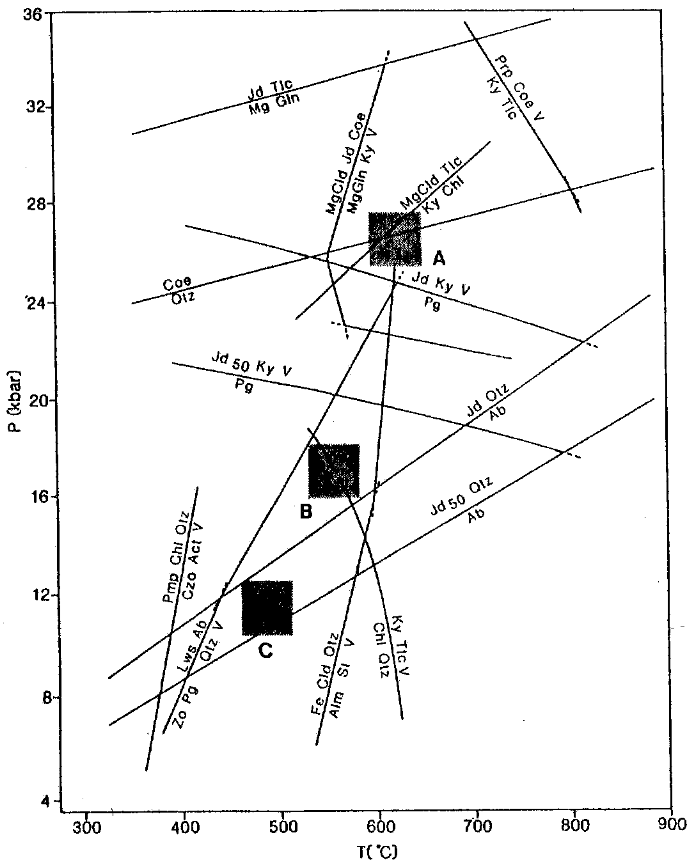

Fig. 5 - P-T grid showing inferred metamorphic conditions during the climax of the Eoalpine event in some eclogitic ophiolites of the Piedmont zone (modified after Droop et al., 1990): A) Cignana (Valtournanche) coesite-bearing metacherts (Reinecke, 1991); B) Zermatt ophiolites (Oberhänsli, 1986); C) St. Marcel ophiolites, southern Aosta valley (Martin and Tartarotti, 1989).

such as glaucophane ± crossite and Fe-talc in mafic rocks (Nisio and Lardeaux, 1987). Temperature values of 400-350°C for P=0.2-0.3 GPa have been suggested on the basis of the observed paragenesis (Nisio and Lardeaux, 1987). Successively, a greenschist facies overprinting generated mineral assemblages such as albite, Fe-epidote, Fe- chlorite, barroisite and tremolite-actinolite amphiboles.

Contrasting post-eclogitic evolutions have been reported for the Passo Gallarino and the Lago Superiore units which record the same Eoalpine imprint (Nisio, 1985; Nisio et al., 1987): while the Passo Gallarino eclogites suffered a complete re-equilibration under progressively decreasing pressure, the Lago Superiore eclogites are completely preserved. This contrasting metamorphic evolution has been tentatively explained by decoupling of the oceanic crust during the closure of the Ligurian-Piedmont basin.

The Monginevro Massif

The Monginevro ophiolites are located in the Cottian Alps, north-west of the Monviso massif. They represent the uppermost Piedmont unit in this sector of the Alpine belt, tectonically overlying other Piedmont ("Chabrière series", Lemoine, 1971) and continental margin-derived cover units ("pre-Piedmont" Gondran zone, Lemoine et al., 1970; Lemoine, 1971). The Monginevro ophiolites escaped the pervasive deformation and metamorphic imprint of the Alpine orogeny, so that they may be easily compared with the poorly reworked Ligurian ophiolites of the Apennines

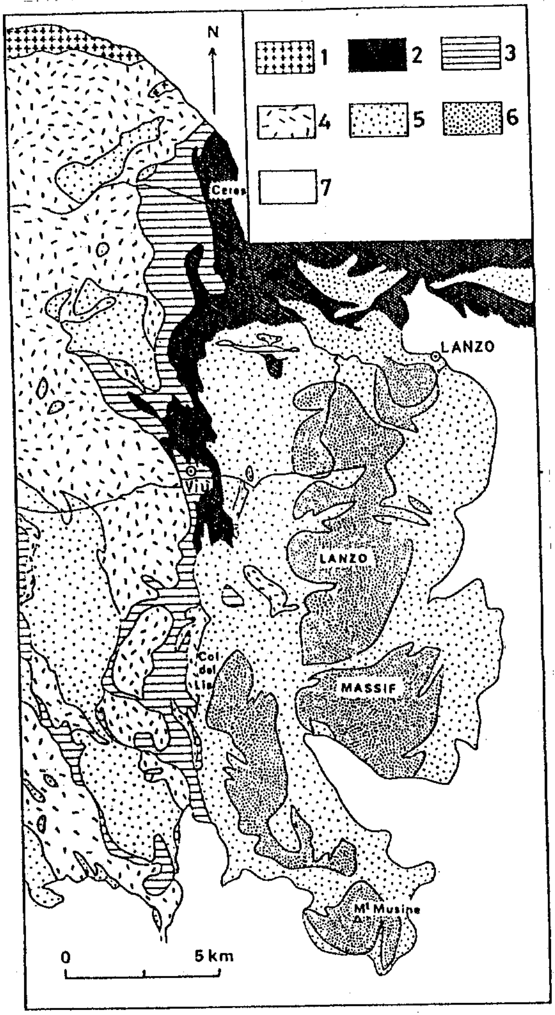

Fig. 6 - Geological sketch map of the Lanzo peridotite body and surrounding units (after Bodinier, 1988). 1) Gran Paradiso basement nappe; 2) Sesia-Lanzo zone; Piedmont zone: 3) Mesozoic metasediments, 4) mafic ophiolites, 5) Lanzo serpentinites and 6) Lanzo peridotites; 7) Recent deposits.

and the present oceanic crust.

Until a few years ago, the Monginevro ophiolites were considered as unique tectonic unit (e.g., Bertrand et al., 1982). More recently, two distinct ophiolitic units have been recognized on the basis of different Alpine evolutions (Polino, 1984; Polino and Lemoine, 1984;). They are known as the Lago Nero unit and the overlying Chenaillet unit.

1) The Lago nero unit consists of an ocean-type basement and a metasedimentary cover. The oceanic basement is represented by serpentinites and minor ophicalcites. The sedimentary cover is one of the most complete supra-ophiolitic sequence of the Western Alps. It encompasses basal sedimentary breccias, with serpentinite and basalt components, and the classic "Chabriére series". The latter consists of radiolarian metacherts of Upper Oxfordian-Middle Kimmeridgian age (De Wever and Caby, 1981; Martin and Polino, 1987), marly limestones (Tithonian-Neocomian), limestones and clayey schists ("Replatte formation", Lower Cretaceous), black or green schists ("black shales"-facies, Middle Cretaceous), and then of arenaceous limestones grading upwards to feldspar-rich arenaceous limestones alternating with clayey schists (Upper Cretaceous, Lemoine, 1984). A further peculiar feature of the metasedimentary series is the

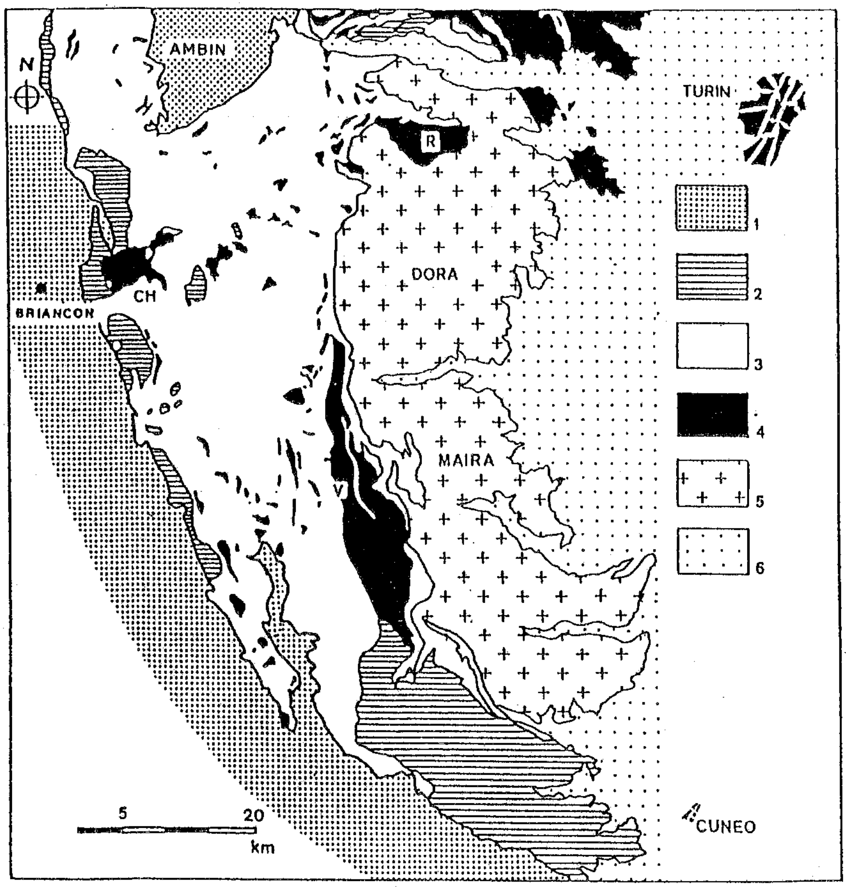

Fig. 7 - Sketch map of the French-Italian Cottian Alps (Lagabrielle, 1987) and location of ophiolites described in the text: Rocciavré (R), Chenaillet (CH), Mt. Viso (V). Legend: 1) Grand St.Bernard-Briançonnais system; Piedmont zone: 2) "external" units, 3) undifferentiated calcschists (schistes lustrés) and 4) ophiolites; 5) underlying Dora Maira nappe; 6) Po plain Recent deposits.

occurrence of detritus from both oceanic and continental sources (Polino, 1984; Lagabrielle and Polino, 1985). The oceanic detritus is chiefly represented by serpentinite and chlorite sandstones, ophiolite breccias and olistoliths, the continental by siliceous schists, carbonaceous breccias and dolomitic olistoliths.

2) The overlying Chenaillet unit (Fig. 7) consists of mantle peridotites and mafic-ultramafic cumulates, mainly consisting of troctolites, plagioclase-bearing dunites and pyroxenites (Bertrand et al., 1982, and refs. therein). They are associated with massive to pillowed basalts, pillow breccias, hyaloclastites and basaltic dykes. The gabbroic sequence well preserves the igneous structure and shows an oceanic metamorphism (Mével et al., 1978; Steen et al., 1980). The sedimentary cover is only poorly represented by some serpentinite breccias, schists and ophiolitic sandstones.

Low temperature-blueschist assemblages occur in the Lago Nero unit (Martin and Polino, 1987), while they seem to be missing from the Chenaillet unit (Bertrand et al., 1982).

The Voltri Group

The ophiolitic Voltri massif is located at the south-western end of the Alpine chain, near the gulf of Genoa (Fig. 1). It consists mainly of two groups of lithological associations derived from the oceanic lithosphere of the Mesozoic Piedmont-Ligurian basin: a) mantle-derived antigoritic serpentinites and metagabbros, b) metabasalts with ophiolitic detritus and metasediments,

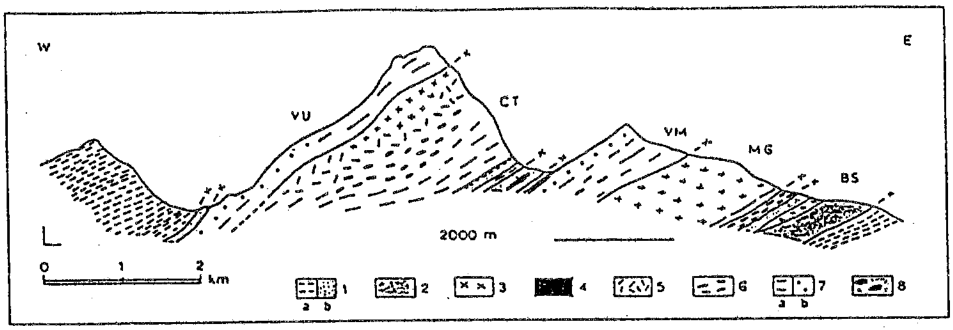

Fig. 8 - Schematic cross-section through the central Mt. Viso ophiolitic massif (Lombardo et al., 1978). 1) Transposed cover metasediments: calcschists (a) and quartz-rich micaschists (b); 2) antigorite serpentinites; 3) metagabbros; 4) eclogitic gabbros and eclogites; 5) massive metabasalts; 6) pillowed metabasalts and basaltic breccias; 7) banded (a) and brecciated (b) metabasites; 8) shear zone between basal serpentinites and metagabbros. Tectonic units: Vallanta (VU), Costa Ticino (CT), Viso Mozzo (VM), Lago Superiore (MG), basal serpentinites (BS).

consisting essentially of prasinites and calcschists. These ophiolites belong to the Piedmont zone and represent an Alpine composite nappe system overridding, together with the blueschist facies Montenotte nappe, the Penninic basement and cover units of the Maritime Alps (Vanossi et al., 1986, and refs.therein).

In places the antigoritic serpentinites display clinopyroxene and spinel relics. The Al-Na depleted clinopyroxene composition and the high Cr content of spinel suggest depletion of peridotite during partial melting processes. Therefore this peridotite may be interpreted as a slice of the depleted asthenosphere formerly underlying the Piedmont-Ligurian oceanic lithosphere.

The serpentinite-gabbro complex is in turn capped by the Erro-Tobbio peridotite unit, consisting of partially serpentinized mantle lherzolite, minor harzburgite, dunite and pyroxenite layers. These ultramafics are intruded by gabbroic and basalt dykes with MORB affinity. The Erro-Tobbio peridotites may be interpreted as fragments of subcontinental lithospheric mantle derived from the Adria southern plate (Piccardo, 1984; Piccardo et al., 1989). Further details are shown in a specific paper of this volume.

The Voltri and Erro-Tobbio ophiolites display traces of oceanic alteration and a polyphase metamorphic history during the Alpine orogeny from subduction high-P/low-T conditions to greenschists facies reequilibration (Chiesa et al., 1977; Messiga et al., 1983; Piccardo, 1984; Messiga, 1987; Morten et al., 1989; Messiga and Scambelluri, 1991). Petrological data mainly refer to the serpentinites, metasediments and mafic rocks of the Voltri Group (Chiesa et al., 1977; Cimmino et al., 1981; Messiga et al., 1983; Messiga, 1987, and refs. therein). Petrological studies on the associated eclogite-metagabbro bodies suggest T=450C° and P=1.3-2.0 GPa for the eclogitic peak (Messiga et al., 1983; Messiga and Scambelluri, 1991). A comparable Alpine metamorphic evolution has been recognized also in the Erro-Tobbio peridotites which exibit titanian clinohumite-antigorite assemblages (Piccardo et al., 1988).

THE CENTRAL ALPS

The ophiolites of the Central Alps mostly occur in the Malenco, Grisons and Engadine areas (Figs. 1-2), as thin slices sandwiched between the Austroalpine and the Penninic basement and cover nappes.

The ophiolitic Platta-Arosa and Malenco-Avers nappes consist predominantly of serpentinized mantle peridotites, mafic ophiolites and supraophiolitic cover sequences (Pasquarè, 1975; Trümphy 1980; Ferrario and Montrasio 1976; Montrasio and Trommsdorff, 1985; Peretti 1985; Weissert and Bernoulli, 1985; Schmid et al.,1990). By contrast, the underlying and external Valais zone mainly consists of calcareous, shaley and sandy flysch-type calcschists

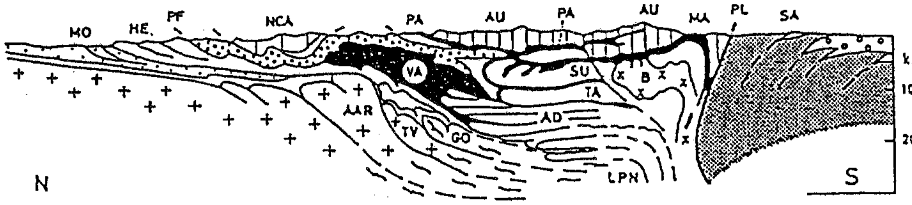

Fig. 9 - Schematic cross-section of the Central Alps (Polino et al., 1990). Symbols as in Fig. 1 and 3. Tectonic units: Helvetic-Ultrahelvetic Aar, Tavetsch (TV) and Gotthard (GO) basement units and related cover nappes (MO: Morcles, HE: Ultrahelvetic); Penninic frontal thrust (PF); Ossola-Tessin lower Penninic nappes (LPN), including the Adula nappe (AD); middle-upper Penninic Tambò (TA) and Suretta (SU) nappes; Valais (North-Penninic) ophiolitic and flysch units (VA); Malenco-Avers (MA) and Platta-Arosa (PA) ophiolitic nappes; Eastern Austroalpine basement system (AU) and décollement nappes (NCA: Northern Calcareous Alps). B: Oligocene Bergell pluton; PL: Periadriatic (Insubric) fault system; SA: Southern Alps.

(Bündnerschiefern) with only a few ophiolitic bodies (Pantic and Gansser, 1971; Favre and Stampfli, 1992; Faupl and Magreich, 1992 and refs.therein).

According to classic restorations of the Alpine Tethys, the upper and internal Platta-Arosa and Malenco-Avers ophiolites belong to the Piedmont-Ligurian (South- Penninic) ocean, whilst the external and lower ones may be allocated to the Valais (North-Penninic) basin, which was floored by oceanic to thinned continental crust. A North-Penninic restoration has traditionally been suggested also for the eclogitic ophiolites and related garnet peridotites of the Adula-Cima Lunga nappe inside the Ossola-Tessin window (Evans and Trommsdorff, 1978; Evans et al., 1979, 1981; Heinrich, 1983; Schmid et al., 1990), although alternative solutions cannot be excluded (Hunziker et al., 1989; Polino et al., 1990).

The South and North Penninic basins are thought to be separated by a continental strip corresponding to the Brianççnnais- Tambo structural high (e.g. Dietrich et al. 1974; Trümpy 1980; Schmid et al. 1990). As discussed later, the Jurassic opening of the Swiss-Austrian segment of the oceanic Tethys has been related to a left-lateral transtensional regime (Weissert and Bernoulli, 1985; Faupl and Wagreich, 1992).

Structural setting

Numerous ophiolitic units are exposed at different structural levels along the Malenco valley-Grisons cross-section and within the Engadine window (e.g., Dietrich et al., 1974; Trümpy, 1980; Schmid et al., 1990; Figs. 9, 10, 11). From top to bottom and from the internal to the external side the ophiolites occur as follows:

1) The Platta-Arosa nappe is sandwiched between the overlying Austroalpine Bernina-Err-Silvretta nappes, and the underlying lowermost Austroalpine Margna nappe and some Penninic décollement units (Trümpy; Bernoulli and Weissert, 1985; Ring et al. 1990). The ophiolites are represented by serpentinized lherzolites, ophicalcites and pillow metabasalts (Peters, 1963; Ishiwatari, 1985). The sedimentary cover comprises Jurassic radiolarian cherts, pelagic limestones, dark marls, shales and capping siliceous black shales which recall the supra-ophiolitic cover of the Ligurian nappes from the Northern Apennines (Bernoulli and Weissert, 1985). The black shale deposition marks the onset of the Tethyan basin closure.

2) The Malenco-Avers nappe includes the Malenco-Forno and the Lizun-Avers units which are separated by the Engadine fault. The Malenco serpentinized lherzolites and associated ophicarbonates form a 200 km-wide and 2 km-thick body (Trommsdorff and Evans, 1980; Montrasio and Trommsdorff, 1985; Trommsdorff et al., 1993) which is interbedded between the Margna nappe and the underlying Upper Penninic Suretta nappe. Lower massive olivine-antigorite-magnetite-bearing serpentinites and upper foliated antigorite-clinopyroxene-chlorite-magnetite-olivine ± talc-bearing serpentinites are recorded in the Malenco area (Bigioggero et al., 1990). They also include layers of pyroxenite and dykes of metarodingite. These ultramafics have recently been interpreted as the huge remnants of the Adria subcontinental mantle which was denudated by extensional unroofing in a late stage of the Permian-Mesozoic asymmetric

rifting (Trommsdorff et al., 1993). Lithosphere extension and asthenosphere partial melting facilitated also the generation of ensialic gabbros, such as the Fedoz body (Dal Piaz, 1993; Trommsdorff et al., 1993).

The Forno and Lizun ophiolitic units consist of metamorphic gabbros, pillow lavas, basaltic tuffs, Mn-Fe-rich quartzites, carbonaceous- and calcschists of flysch affinity (Montrasio, 1973; Ferrario and Montrasio, 1976; Peretti, 1985; Peretti et al., 1992; Sciesa and Montrasio, 1992; Sidler and Benning, 1992). The oceanic sequences were developed over the Malenco mantle sole after the onset of the drifting stage during the Late Jurassic. The Lizun ophiolites are tectonically coupled with the Avers Bündnerschiefern (=Grisons calcschists) which also contain ophiolitic slivers. The Lizun and Avers units are both infolded within the Suretta basement nappe. Ring et al. (1990) point to the lithological equivalence between the Malenco-Forno and the overlying Platta nappe. Southwards, the Bagni di Masino ophiolites are exposed within a small window cutting the Bergell body, probably representing the southern prolongation of the Malenco- Forno unit. East of the mid-Oligocene Bergell pluton, the Suretta nappe is locally exposed beneath the Malenco ultramafics within narrow tectonic windows, known as the Lanzada-Scermendone zone (Montrasio, 1984; Peretti et al., 1992; Sciesa and Montrasio, 1992). The window core consists of basement rocks and of Permian-Mesozoic siliceous to carbonaceous metasediments, scarp breccias and calcschists. The cover sequence recalls the Permian-Cretaceous décollement units of continental affinity associated to the Combin zone, Western Alps.

3) The Chiavenna and Misox ophiolitic units are located at a lower structural level, being sandwiched between the Tambo and the underlying Adula basement nappes. The former ophiolite occurs internally, extending from the Bergell to the Bondasca valleys, and is formed by mafic and ultramafic rocks with minor marbles (Bigioggero et al., 1990). The external Misox (Mesolcina) zone includes ophiolites and oceanic metasediments (Gansser, 1937; Trümpy, 1980; Teutsch, 1982). Northwards, supposedly equivalent ophiolitic units occur around the front of the Adula nappe, as well as in the Areua-Bruschghorn to the Martegnas zones, forming a discontinuous strip which bounds the front of the Schams décollement nappe (Schmid et al. 1990, and refs. therein).

4) The Adula-Cima Lunga nappe displays most of the Alpine eclogite facies assemblages from the Central Alps and includes a few imbricated ophiolitic slices which probably yield a Mesozoic age (Heinrich, 1986). They consist of severely transposed mafic eclogites, metarodingites and calcschists, as well as of the garnet-lherzolite bodies from Alpe Arami, Cima di Gagnone and M. Duria (Evans and Trommsdorff, 1978, and therein refs.). Further ophiolites are reported from the Paina zone (Santini, 1991).

The Engadine window (Fig. 11) is located within the Eastern Austroalpine nappe system at the Swiss/Austrian border. It displays, from top to bottom, the capping Arosa (=Platta) ophiolitic nappe, the Tasna cover and basement nappe, the Ramosch and Roz- Champatsch-Prutz units which include an ophiolitic strip and flysch sequences; the lowermost Pfunds zone displays Mesozoic calcschists and minor ophiolites (Trümpy 1972, 1980; Hock and Koller, 1987; Koller and Hock, 1992 and refs. therein). The Idalp ophiolites from the Arosa nappe (Daurer 1980; Hock and Koller 1987, 1989) may be derived from the South-Penninic (Piedmont) domain (Trümpy, 1972; Oberhauser, 1980) and be correlated with the Totalp (Davos) bodies of the Platta nappe (Weissert and Bernoulli, 1985). These ophiolites consist mainly of serpentinites with boudinaged rodingitic dykes, ophicalcites, sliced gabbros with diabase dykes, pillow basalts alternating with massive lava flows and hyaloclastites. The supra-ophiolitic cover includes radiolarian cherts and schists. The Ramosch (Vuichard, 1984a) and Piz Mundin (Pfunds zone; Heugel, 1975) ophiolitic sequences are referred, on the contrary, to the North-Penninic (Valais) domain (Hock and Koller, 1987). The Ramosch ophiolites consists of serpentinized lherzolites with rodingitic basalt and gabbro dykes, ophicarbonates and serpentinite breccias, and of pillow basalts and minor mafic sills (Hock and Koller, 1989).

Metamorphic signatures

The ophiolitic units of the Central Alps are poorly to pervasively reworked by the Alpine metamorphism. They display contrasting metamorphic signatures of the subduction event, reaching

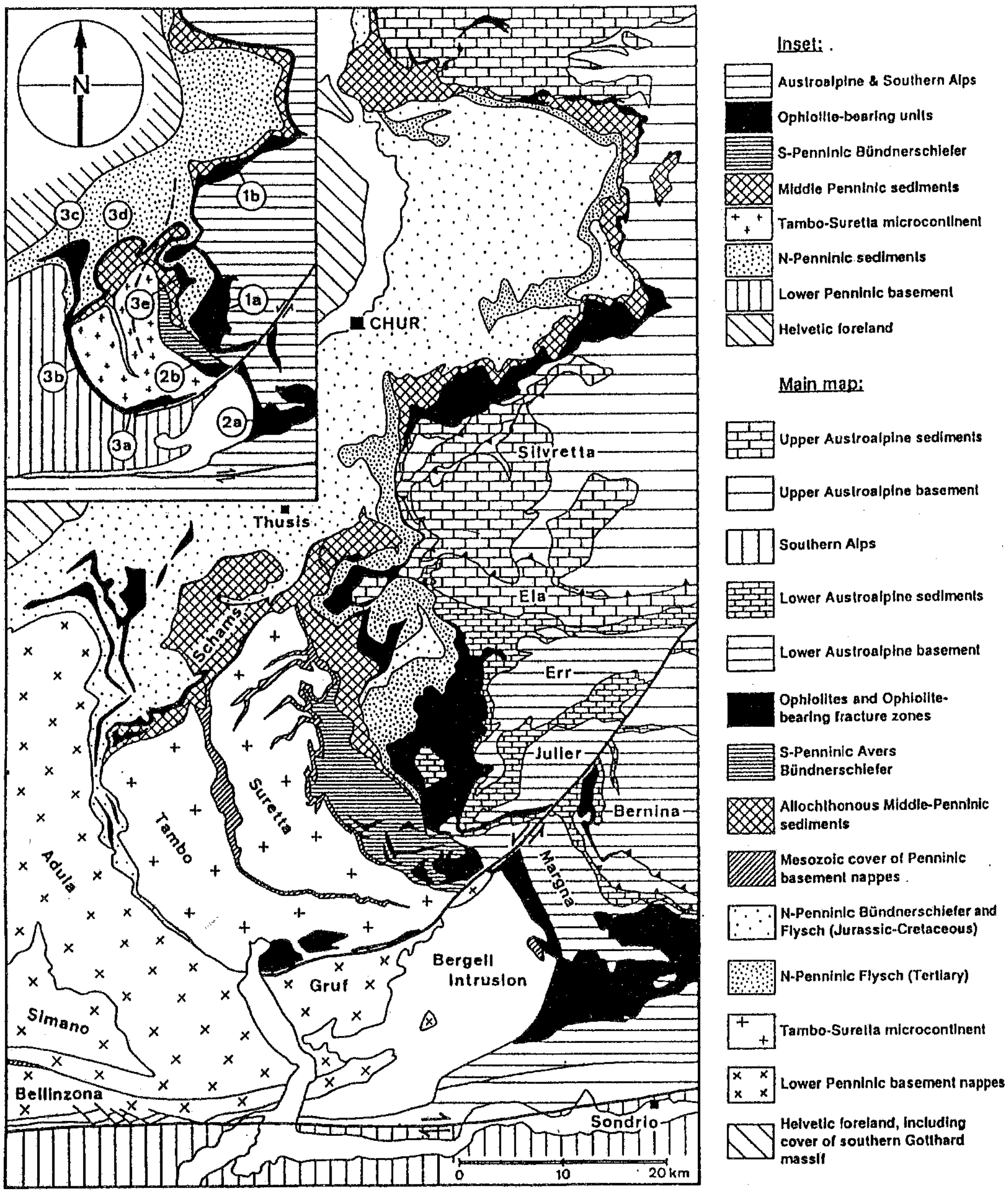

Fig. 10 - Tectonic map of Grisons and surrounding areas, Central Alps (Schmid et al., 1990). Inset: 1a) Platta, 1b) Arosa, 2a) Malenco), 2b) Lizun and Avers, 3a:) Chiavenna, 3b) Misox, 3c: frontal Adula nappe, 3d) Areua-Bruschghorn, 3e) Martegnas ophiolites.

eclogitic signatures only in the Misox zone and especially in the Adula-Cima Lunga-Paina nappe (Fig. 1; Dietrich et al., 1974).

The occurrence of barroisite in the ensialic Fedoz metagabbros (Margna nappe) suggests a pressure of between 0.4 and 0.7 GPa (Guntli and Liniger, 1989).

In the underlying Malenco serpentinites, Mellini et al. (1987) estimated a T range of 390-465 °C. These conditions are consistent with the stable antigorite-olivine-diopside-titanian clinohumite assemblage and with the fluid inclusions described in the western part of the Malenco ophiolites by Trommsdorff and Evans (1972), Peretti et al.(1992), Trommsdorff et al. (1993).

Blueschist facies mineral assemblages are recorded in some ophiolitic units in the southwestern Grisons and in the Engadine window (Dietrich et al., 1974; Oberhänsli, 1986; Sciesa and Montrasio, 1991). In the southern Grisons, the Bundnerschiefern units of the western side (Adula and W-Avers) display stable glaucophane, whereas on the eastern side (E-Avers, Platta, Malenco-Forno-Lizun) only Mg-riebeckite is recorded (Oberhänsli, 1986). Geothermobarometers and specific assemblages in the Avers unit (blue and green amphiboles, zoned spessartine to almandine-rich garnet and chloritoid) suggest T of about 350- 400°C and P ranging from 0.9 to >1.2 GPa.

High-grade blueschist conditions (P=0.8-1.0 GPa and T=400- 450°C) are estimated by Sciesa and Montrasio (1991) for the Lanzada-Scermendone unit beneath the Malenco ultramafics. Crossite occurs in the metasedimentary cover of the Suretta nappe (Giere, 1985), where Ring (1992) estimated T=380-420°C and P=0.8 $\rightarrow$1.0 GPa. Glaucophane is reported from epidote-rich zones of the underlying Tambo basement nappe (Gansser, 1937). A P-T range of 400-480°C and 0.8-1.0 GPa is suggested by Ring (1992) for the northern Tambo nappe.

The partly retrogressed glaucophane-bearing eclogites from the Misox ophiolites record P>1.2 GPa and T=460-560 °C (Santini, 1991; Ring, 1992). On the northern side of the underlying Adula nappe, metamorphic conditions of P=0.6-0.8 GPa and T=380-450°C have been estimated by Löw (1987). Data obtained from the Adula white-schists indicate T=600 °C and P > 1.5 GPa (Ring, 1992).

Mafic-ultramafic slices are imbricated in the frontal sector of the Adula-Cima Lunga nappe. They consist of garnet peridotites, metarodingites and/or eclogites with tholeiitic affinity (Bocchio et al., 1978; Evans and Trommsdorff, 1978; Evans et al.1979, 1981; Trommsdorff, 1991). P-T estimates of the high-P metamorphism range from T=800 °C and P > 2.0 GPa in the Cima di Gagnone and Alpe Arami bodies to T=500°C and P=1.2 GPa in the Northern Adula basement (Trommsdorff, 1991, and refs.therein).

The blueschist facies imprint in the Central Alps yields some Eoalpine radiometric ages (100-60 Ma: Jaeger, 1973; 120-80 Ma: Hanson et al., 1969; Dietrich and Oberhänsli, 1976). The eclogitic metamorphism in the Adula-Cima Lunga nappe has been referred either to the Eoalpine event (Hunziker et al., 1989) or to a much younger but unconvincing age (28 Ma; Gebauert et al., 1992). The Adula-Cima Lunga nappe has been overprinted by a Barrovian Mesoalpine metamorphism characterized by greenschist facies (T=400-450°C, P=0.3 GPa, Mellini et al., 1987) to amphibolite facies assemblages in the eastern and western Grisons respectively (Frey et al., 1974; Niggli, 1978). The Mesoalpine (Lepontine) metamorphism yields radiometric ages at around 38-35 Ma (K-Ar, Rb-Sr: Hunziker, 1969) and 47-31 Ma (40Ar/40Ar: Santini, 1991). Moreover, the Malenco-Forno ophiolites and the underlying Suretta nappe in the Sissone valley record a thermal imprint due to the Bergell intrusion during the Oligocene (30 Ma; Trommsdorff and Nievergelt in Dal Piaz, 1985).

The tectonic units exposed inside the Engadine window display a pumpellyite-actinolite (T=350°C and P > 0.3 GPa; Koller and Hock, 1987) to greenschist facies imprint. Riebeckite has also been found in mafic metavolcanics and in the associated metasediments (Oberhänsli, 1986). The occurrence of lawsonite has been reported by Leimser and Purtscheller (1980). Despite the Alpine imprint, some metagabbro bodies of the Idalp zone preserve oceanic signatures recorded by brown-green hornblende partly replaced by actinolite.

Summing up, the Platta nappe and the underlying ophiolitic units of the SW Grisons experienced during the subduction episode different metamorphic conditions with respect to some presently associated basement nappes. The supposedly South-Penninic (Piedmont) and North-Penninic (Valais) units show slightly higher P (up to or even more than 1.2 GPa) signatures than the Suretta and Tambo basement nappes (0.8-1.0 GPa, Ring, 1992a). By contrast, a pronounced pressure jump (0.3-1.0 GPa) occurs between the Bündnerschiefern and the Adula nappe. A top-to-the-west unroofing (Ring, 1992a; Durr, 1992) may have facilitated the initial exhumation of the high-P rocks, whilst tectonic coupling between the Bünderschiefern and the Adula nappe may have occurred at a structural level characterized by blueschist facies conditions (Ring, 1992a). Since the onset of continental collision, the rate of subduction noticeably slowed. Mid-Tertiary evolution of the collisional belt was characterized by thermal relaxation and related

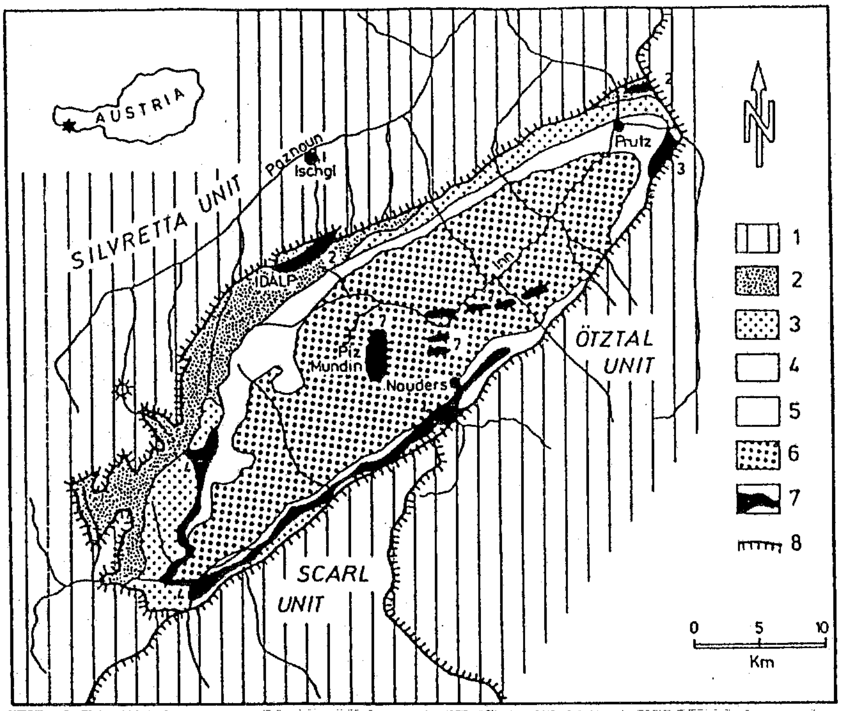

Fig. 11 - Geological map of the Engadine window (Hock and Koller, 1987). 1) Austrolapine units; 2) Arosa nappe; 3) Tasna zone; 4) Ramosch Zone; 5) Roz-Champatsch-Prutz zones; 6) Pfunds zone; 7) ophiolites; 8) main thrusts and tectonic lines.

Lepontine regional metamorphism.

THE EASTERN ALPS

The ophiolitic units exposed in the Eastern Alps mainly occur within the Tauern window and the Rechniz window group (Fig. 1). The former extends in the central part of the Eastern Alps over an area of about 170 by 40 km (Fig. 12). There, the ophiolitic units are sandwiched between the capping Austroalpine nappe system and the underlying Penninic cover and basement nappes (Zentralgneiss). The second group is located about 100 km south of Vienna, where the collapsed easternmost edge of the Alpine mountain belt is buried beneath the Neogene-Recent deposits filling the Vienna and Pannonian basins. This group comprises the small ophiolites of the Rechnitz, Bernstein, Moltern and Eisenberg windows. As in the Central Alps, these ophiolitic units are thin slices of Mesozoic oceanic lithosphere including serpentinized mantle peridotites, metamorphic mafic-ultramafic cumulates, basalts, cherts and abundant calcschists (Bündnerschiefern), mainly thought to derive from Cretaceous flysch sequences. During the Alpine orogeny these protoliths experienced a composite metamorphic evolution, evolving from an Eoalpine high-P event to a mid-Tertiary greenschist-amphibolite facies overprinting (e.g., Frey et al., 1974; Niggli, 1978; Raith et al., 1978, 1980; Miller, 1986; Frank et al., 1987; Koller and Hock, 1992).

Minor ophiolites are locally reported also in the external Penninic Rheno-Danubic flysch and Klippen zone, underlying and bordering the Austroalpine Northern Calcareous Alps: they consists of slivers or detrital fragments of radiolarian cherts, pillow and breccia lavas, gabbros and serpentinites (Plöchinger, 1973; Bickle and Pearce, 1975; Dietrich, 1980; Kirchner, 1980; Winkler et al., 1985; Egger, 1990; Schnabel, 1992, and refs. therein). Lastly, an ophiolite is reported from the Reckner zone, south of Innsbruck. This body displays an unusual tectonic location, as it overrides the Austroalpine Innsbruck quartz-phyllite unit, not far from the northwestern corner of the Tauern window (Dingeldey and Koller, 1990). The ophiolite association consists of serpentinized mantle lherzolites, gabbros and ultramafic cumulates, covered by thin ophicarbonates, cherts and basalts of supposedly Mesozoic age. These rocks show a pervasive oceanic alteration, orogenic blueschist facies signatures and a greenschist facies overprint.

The internal Tauern and Rechnitz ophiolites are generally envisaged as sliced remnants of the South Penninic (Piedmont) ocean, whilst those from the external Rheno-Danubic flysch zone are referred to the North-Penninic (Valais basin). The Engadine, Tauern and Rechnitz ophiolites display roughly similar stratigraphic settings and genetic characteristics. Most metabasalts display a normal-MORB affinity (Hock, 1983) and may be classified as high-Ti ophiolites according to Beccaluva et al. (1983). Nevertheless, minor metabasalts with off-axis and within-plate signatures are also locally reported from the Tauern and Rechnitz windows: they range from tholeiite to mid-alkaline basalt composition, with Nb, Zr and LREE enrichment (T-type to E-type MORB, Koller, 1985). The variable Zr/Y ratio has been referred to different degrees of partial melting (Hock and Koller, 1989).

The Tauern window

The Tauern window exposes a complete cross-section through the ophiolitic and Penninic basement nappes beneath the Austroalpine system (Fig. 12). The Mesozoic ophiolitic units are traditionally reported in the Austrian literature as Upper Schieferhülle (e.g. Morteani, 1974; Lammerer, 1986; Miller, 1986; Frank et al., 1987; Hock and Miller, 1987; Selverstone et al., 1992). The underlying Penninic basement and cover nappes are formed by Alpine metamorphic derivatives from Late-Paleozoic granitoids and pre-granitic paraschists (Zentralgneis), and from minor Permian-Mesozoic clastic to carbonatic sequences (Lower Schieferhülle).

The ophiolitic associations consist of predominant Jurassic to Early Cretaceous calcareous schists and flysch units (Bündnerschiefer complex) and of ophiolitic bodies. Differences in sedimentary facies allowed identification of the Brennkogel, Glockner and Fusch series (Thiele, 1970; Hock and Miller, 1987). The mafic and ultramafic bodies are concentrated in the central part of the Tauern window, especially along the southern and north-eastern sectors, where they have been reported as units I and II-III respectively (Fig. 12, Hock and Miller, 1987). These units override a sequence of calcschists which grade to the underlying Brenkogel facies clastic metasediments. The basal section of the restored stratigraphic sequence (Hock and Koller, 1989) comprises serpentinitic slices (from mantle harzburgite and lherzolite), locally associated with metamorphic ultramafic cumulates and gabbros and with mylonitic derivatives (tremolite-chlorite-antigorite schists). Coarse-grained metagabbros occur within the cumulate sequence in the northern unit (II) or as tectonic slices within the calcschists in the southern unit (I); they show evidence of two crystallization stages. The maximum thickness of the intrusive section does not exceed 1000 m, as in the Western Alps (Dietrich, 1980; Hock, 1980, 1983; Hock and Miller, 1980, 1987). The overlying volcanic sequence (200-600 m thick) comprises metamorphic pillow basalts, hyaloclastites and breccias which only rarely preserve the primary texture. The metavolcanics are capped by a sedimentary cover (up to 400 m thick), consisting of rare quartzites, calc-micaschists and black phyllites with some basal volcanic interbeddings. The ophiolitic metabasalts from units I and II display, as a whole, a tholeiitic composition with normal-MORB affinity and include high-Ti types (Hock, 1983; Koller and Hock, 1987 and refs. therein). By contrast, mafic metatuffite and metabasalt layers within the Brennkogel facies metasediments of the unit III (Hock and Miller, 1987) record a wider compositional variety, ranging from MOR to within-plate basalts.

In the Italian south-western corner of the Tauern window the calcschist nappe overthrusts the Paleozoic-Mesozoic Greiner nappe. The former displays basal carbonaceous calcschists, a hectometric body of metabasalts with minor Mn-quartzites and marbles, capped by flysch-type calcschists with minor mafic interbeddings (Dal Piaz Gb., 1934; De Vecchi and Piccirillo, 1968; Baggio, 1969; De Vecchi, 1989). Scarce serpentinites are reported in the Vizze valley.

The Tauern window is regionally characterized by a pervasive greenschist to amphibolite facies metamorphism of mid-Tertiary age which overprints subduction-related Eoalpine signatures. The Eoalpine metamorphism is characterized by two groups of high-P assemblages. The first

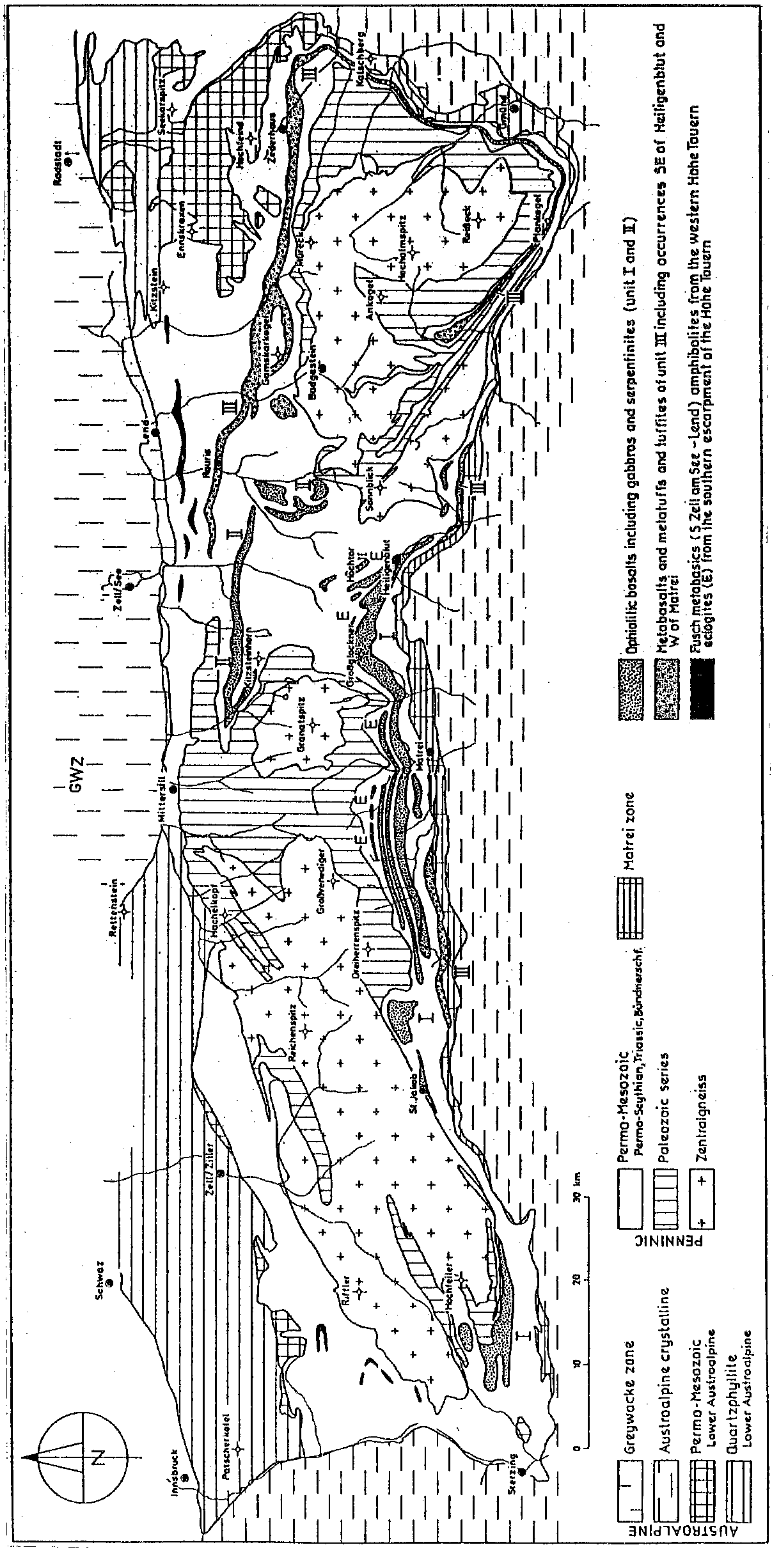

Fig. 12 - Geological map of the Tauern window (Hock and Miller, 1987).

group is recorded by eclogites occurring in a narrow ophiolitic zone from the southern side of the Tauern window (Figure 13; Miller, 1974, 1977,1986; Hock and Miller, 1987; Selverstone et al., 1992). The mineral assemblages are represented by omphacite-quartz, talc-kyanite ± Mg-cloritoid ± quartz, kyanite ± zoisite ± quartz (T=600°C, P about 2 GPa; Holland, 1979; Frank et al., 1987; Selverstone et al., 1992 and refs.therein). The second group is characterized by a blueschist facies imprint (T=350-450 °C and P=0.7-0.9 Gpa) widely recorded in mafic ophiolites and calcschists by pseudomorphs after lawsonite, and by mineral assemblages including glaucophane-garnet ± omphacite, glaucophane-phengite-paragonite-clinozoisite, epidote-garnet-carbonate. Na-amphiboles yield K-Ar ages ranging from 90 to 60 Ma (Raith et al., 1978). The mid-Tertiary overprint ("Tauern Kristallization") is represented by greenschist facies assemblages mainly reported from bordering and structurally higher sections of the Tauern window, and by amphibolite facies assemblages from the central and lower sectors (T=400-550°C and P=0.4-0.6 GPa). The ophiolitic units from the Italian side (SW Tauern) display only epidote-oligoclase or Ab-amphibolite facies assemblages of Tertiary age (De Vecchi and Piccirillo, 1968; De Vecchi, 1989).

The underlying continental basement nappes preserve scarce and partly re-equilibrated high pressure relics (T=450-500° C; P=1.0-1.2 Gpa and ; Zimmermann and Franz, 1989).

The Rechnitz window group

The eastern windows of the "Rechnitz group" expose beneath the Austroalpine basement some ophiolitic bodies and related metasedimentary cover sequences (Koller, 1985; Koller and Hock, 1990). Basal serpentinites and ophicarbonates are associated with minor metamorphic derivatives from cumulate ultramafics, Mg-rich gabbros and overlying Fe-Ti gabbros, plagiogranites and Fe- diorites. The intrusive bodies reach a maximum thickness of 60-70 m only. Dykes and lenses of rodingitic gabbro inside serpentinites display relict diopside, hydrogrossular or Mg-rich pumpellyite and chlorite. The volcano-sedimentary cover consists of massive metabasalts and Fe-Ti-rich varieties, radiolarian metacherts and calcschists with tuffitic interbeddings. A relatively high-T oceanic alteration is recorded in the gabbro bodies by Mg- hornblende and pargasite replacing magmatic clinopyroxenes. Generation of Cr-bearing andradite in some ophicarbonates may be related to the oceanic event (Koller, 1985). An early stage of the Alpine metamorphic evolution is characterized by Mg-pumpellyite from leuco-gabbros and by Na-clinopyroxene, Fe-glaucophane, stilpnomelane, hematite, rutile from Fe-gabbros and some plagiogranites. Lawsonite (as pseudomorph only) is reported from some greenstones. The blueschist facies assemblage records P-T conditions in the range of 0.6-0.8 GPa and 330-370°C (Koller, 1985). The second metamorphic stage is characterized by greenschist facies assemblages (T=390-430°C and P > 0.3 GPa). Alkali pyroxene from Fe-rich rocks is replaced by blue amphiboles with Mg-riebeckite or riebeckite composition; new biotite growths on stilpnomelane; rutile and Ti-rich hematite are replaced by idiomorphic magnetite and fine-grained sphene.

DISCUSSION

Palegeographic restoration models

Palinspastic restoration procedures used so far in the Alps have been strongly influenced by simple unfolding of a supposed cylindrical fold-nappe pile from its present-day position (Argand, 1916). This method leads to a unique paleogeographic configuration for the Alpine area, as the nappe stack is seen as final product of the incremental ductile shortening of the Alpine Tethys through a physically continuous sequence of antiformal folds and synformal basins evolving up to a pile of recumbent fold- nappes. There are implicit doubts on this restoration procedure since nappe generation and emplacement processes were assumed to be governed by shear mechanisms (e.g., Hermann, 1925; Stutz and Masson, 1938; Staub, 1958; Milnes et al., 1981). Doubts applied especially to the idea that a nappe stack may only be de-imbricated as a similarly ordered sequence of paleogeographic domains. This notwithstanding, most thrust sheet restoration models through "card-deck" de-imbrication continued to recall the paleotectonic picture imposed by the unfolding method: this was probably due to the traditional approach based on horizontal

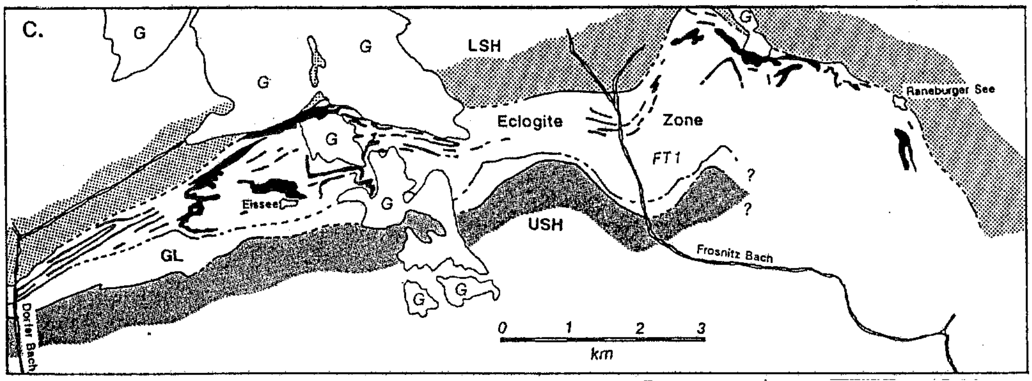

Fig. 13 - Detailed map of the eclogite zone in the southern side of the central Tauern window (Selverstone et al.,1992). The ophiolitic units (black: massive mafic eclogites; white: metasediments and banded mafic eclogites) are sandwiched between the Lower Schieferhülle (LSH) and the overlying "Glimmerschiefern" Lamella (GL) and the Upper Schieferhülle (USH). G: glaciers.

basin analysis which underevaluated or ignored metamorphic constraints from the nappe trajectories through deep structural levels (discussion in Polino et al., 1990).

Since the ensialic igneous emplacement previously suggested for generation of the Alpine ophiolites has been replaced by plate tectonic models, the occurrence of one, two or even more oceanic gaps in the Alpine Tethys between the Austroalpine and Helvetic domains has become a matter of debate. Further questions regard the allocation, width, paleostructural characteristics and opening mechanism of the Jurassic ocean.

Focusing on these problems, it can first be pointed out that interpretation of the opposite-vergent Ligurian and Piedmont ophiolitic nappes from the Apennines to the French-Italian Alps allowed the generally accepted reconstruction of a single Ligurian-Piedmont ocean (Sturani, 1973; Dal Piaz, 1974; Beccaluva et al., 1984; Lemoine, 1984; Lemoine et al., 1986; Abbate et al., 1988, and refs.therein; Fig. 14). The surrounding European and African (Adria) passive margins were restored in accordance with the sense of tectonic transport, and strictly following the vertical sequence of nappes from the present setting back to the Late Jurassic lateral juxtaposition. Accordingly, the following paleogeographic domains were envisaged from the Po plain hinterland towards the French foreland (Figs. 14, 15, 16): 1) the Adria passive margin, including the Southern Alps, the Canavese strip and the Western Austroalpine zone (Sesia-Lanzo and Dent Blanche); 2) the Piedmont-Ligurian ocean; 3) the European passive margin, including the Penninic and the Helvetic-Dauphinois zones (e.g. Dal Piaz, 1974; Sturani, 1975; Trümpy, 1980; Beccaluva et al., 1984; Lemoine et al., 1986; Vanossi et al., 1986). The Penninic zone was in turn divided into the ocean-facing Pre-Piedmont domain (Monte Rosa-Gran Paradiso-Dora Maira) and the Briançonnais-Subbriançonnais domain (Grand St.Bernard) which, south of the Aosta valley, is directly juxtaposed on the external Helvetic-Dauphinois zone (e.g., Elter, 1960; Sturani, 1975). This western segment of the Alpine belt does not record traces of other ophiolitic units supporting the existence of further oceanic gaps. The trench-derived flysch nappes of Cretaceous age, presently occurring near the Penninic-Dauphinois boundary, are commonly restored within the closing Piedmont-Ligurian basin.

Northwards, the Jurassic Piedmont ocean was thought to be limited by an east-west trending transcurrent/transtensional zone parallel to the Central-Eastern Alpine border of the Adria continent, displacing eastwards some short ridge segments (Beccaluva et al., 1984; Weissert and Bernoulli, 1985; Trümpy, 1988; Dal Piaz and Polino, 1989; Schmid et al., 1990; Faupl and Wagreich, 1992). The prolongation to the Central-Eastern Alps of the Piedmont ocean, alternatively named South-Penninic basin, may be inferred here and there from the Antrona, Malenco-Avers, Platta-Arosa, Tauern and Rechnitz ophiolitic units (e.g., CNR 1990). Nevertheless, the Antrona and Platta-Arosa ophiolites presently occur outside the structural level of the nappe pile commonly envisaged as suture of the Piedmont (South-Penninic) basin: the former is sandwiched between the Monte Rosa and Grand St.Bernard nappes, the latter between the Eastern Austroalpine system and the underlying Margna nappe, which in turn overrides the Malenco-Avers ophiolites (=Piedmont zone). Alpine convergence or older transpression may be responsible for the anomalous tectonic location of the supposedly Piedmont-derived Antrona and Platta-Arosa ophiolites (Hawkesworth et al., 1975; Polino et al., 1990). Alternatively, these units could also be envisaged as sutures of further oceanic branches separating the Monte Rosa and/or Margna microcontinents (e.g., Pasquarè, 1975; Trümpy, 1980; Platt, 1986; Ring, 1992a) or of transtensive basins (Schmid et al., 1990).

Further ophiolites are exposed at a noticeably lower structural level with respect to the Piedmont nappe, from the upper Aosta valley (Versoyen zone) to the east, along the Penninic border throughout the Valais and Grisons area (Figs. 1, 2, 14, 15). Ophiolites are embodied within predominating calcschists and Cretaceous flysch units that underlie the frontal Grand St. Bernard system or are imbricated within the Lower-Middle Penninic basement and cover nappes of the Ossola-Tessin window (e.g., Elter and Elter, 1965; Dietrich et al., 1974; Dietrich, 1980; Schmid et al., 1990; Figs. 2, 3, 9). These décollement units are reported as the Sion-Courmayeur, Valais or North-Penninic zone and are generally correlated with the Rheno-Danubic border flysch (Fig. 1). According to some models, the North-Penninic zone also includes the lower ophiolitic calcschists from the Engadine window (Trümpy, 1980; Fig. 11). The occurrence of metagabbro and pillowed metabasalt bodies with MORB affinity (e.g., Dietrich and Oberhänsli, 1975; Oberhänsli, 1978) indicates that the source area was at least locally floored by oceanic crust. These lower ophiolitic units, as a whole, are traditionally interpreted as a suture of the Valais (North-Penninic) basin that, in turn, is restored between the northern Briançonnais-Tambo continental promontory and the Helvetic domain (e.g., Trümpy, 1980; Beccaluva et al., 1984; Schmid et al., 1990; Stampfly and Marthaler, 1990; Stampfly, 1993, and refs.therein; Figs. 13, 15 and 16). An east-west trending left-lateral transtension system has been recently proposed for reconstructing the Central Alpine paleostructure during the Jurassic, including the north-eastern edge of the Piedmont ocean and the Valais trough (Schmid et al., 1990). Conversely, this megashear zone may have prevented the southward extension of the Valais basin towards the French-Italian sector of the Western Alps. In conclusion, while the Western Alpine Tethys is characterized by the unique Piedmont-Ligurian ocean, the traditional scenario of the Central-Eastern Alpine segment during the Late Jurassic is based on a two-basin model accounting for all ophiolites and related flysch units.

Discussions on lithological and geochemical features of the ophiolitic units lead to consistent or contrasting implications on the debated questions. In spite of the absence of a true sheeted dyke complex, the lithological association and MORB affinity of the Apennine and Alpine ophiolites fit well a model in which the Jurassic formation of the Ligurian-Piedmont ocean is referred to drifting processes from an active slow-spreading axis (e.g. Beccaluva et al., 1984; Pfeifer et al., 1989 and refs. therein). When the dismembered Alpine ophiolites are reconstructed place to place, they record the lithological association of the classic ophiolitic sequences and modern homologues, from mantle peridotites (generally converted to antigorite serpentinites) to ultramafic-mafic cumulates, Mg- and Fe-rich gabbros, massive and pillow basalts and to supra-ophiolitic oceanic covers capped by orogenic trench-derived flysch. However, these sequences display a very reduced thickness in comparison with the normal oceanic crust, never exceeding 1-2 km (sliced mantle inclusive). This may be due to anomalous stratigraphic features, tectonic sampling or thinning during the Alpine orogeny. For instance, some mantle serpentinites from the Ligurian and Alpine units are covered by thick ophicarbonates (ophicalcites) or, together with gabbros, by ophiolitic sandstones and tectono-sedimentary breccias. These deposits are covered by negligible amounts of basalt, or directly by Jurassic radiolarian cherts when the mafic volcanics are missing (Abbate et al., 1980; Auzende et al., 1983; Tricart and Lemoine, 1983; Vuichard, 1984a, 1984b; Bernoulli and Weissert, 1985; Lagabrielle, 1987 and refs. therein). In other cases, gabbros are lacking or volcanics are noticeably thicker than the plutonic section.

Lateral variability of the Tethyan oceanic crust composition is suggested by the lithological assemblages of the Alpine ophiolites. Similar features are observed in the present-day oceanic crust generated in slow or very slow spreading centers, where deep rocks directly covered by

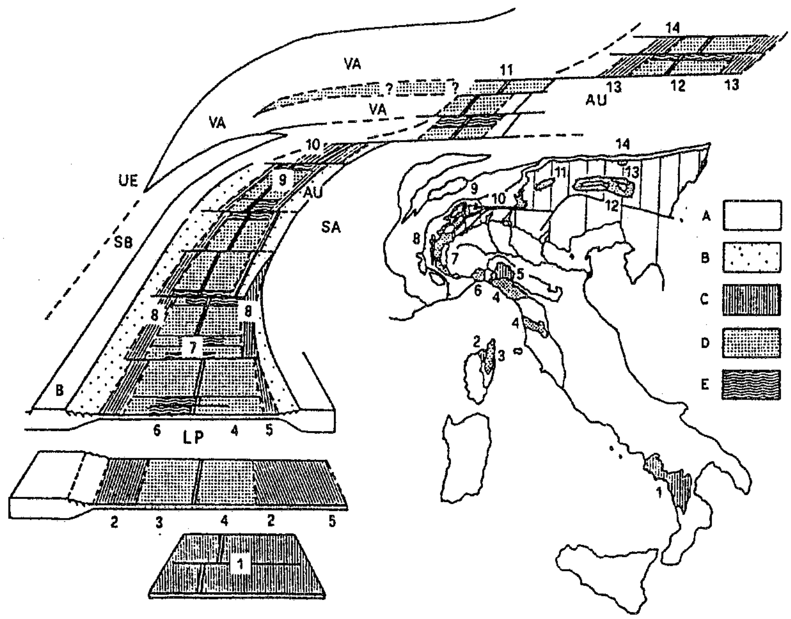

Fig. 14 - Ophiolite distribution in the Alpine-Apennine system and restoration of the oceanic western Tethys (Beccaluva et al., 1984). Ophiolites with transitional to normal MORB affinities: 1: Calabria; 2: Balagne; 3: Inzecca; 4: Internal Ligurides; 5: External Ligurides; 6: Voltri Massif; 7: Monviso; 8: Mongenèvre; 9: Zermatt-Saas and Combin; 10: Antrona; 11: Engadine window; 12, 13, 14: Tauern window; Strobl area, Austria. Ornaments: A) continental crust, B:) transitional crust, C) oceanic crust with transitional-MORB affinity, D) oceanic crust with normal-MORB affinity, E) fracture zones. Inferred paleostructural domains: Southalpine (SA), Austroalpine (AU), Ligurian-Piedmont (LP), Internal Pennine and Briançonnais (B), Subbriançonnais (SB), Valais (VA), Ultrahelvetic (UE).

volcanic flows or breccias are exposed on the ocean floor (e.g., Atlantic ocean in Karson et al., 1987; Mevel et al., 1989, 1991). Thin crustal sections (2-3 km) are reported, for instance, from ridge-transform intersections in slow spreading environments, such as the Mid-Cayman rise and the Kane fracture zone (Cormier et al., 1984; Karson and Dick, 1984). A similar setting could explain the predominantly ophiolitic association from Eastern Liguria, the Cottian Alps, Arosa, Ramosch, Reckner and Rechnitz (e.g., Auzende et al., 1983; Abbate et al., 1984; Vuichard, 1984a, 1984b; Weissert and Bernoulli, 1985; Hock and Koller, 1989; Dingeldey, 1990; Koller and Hock, 1992). Moreover, the multiple interbeddings of tabular metabasalts within Mesozoic calcschists from the Tauern window (Fusch and Brennkogel facies), the Combin zone and other areas, if not clearly related to Alpine transposition, have modern homologues, such as in the Gulf of California. There, during the last rifting stage or early ocean opening, basaltic magmas derived from a relatively enriched asthenospheric mantle were generated and injected into the sedimentary cover as thick coarse-grained sills (Einsele et al., 1980).

The width of the oceanic Tethys in the Alpine-Apennine area is a matter of further discussion. A very small ocean, as also supported by the lithological features reported above, was suggested in some of the classic models quoted before, through pure restoration of ophiolitic nappes and related flysch units. By contrast, a larger oceanic lithosphere was indirectly envisaged in other models necessitating a long-lived subduction for generating the steady-state thermal low recorded by the Cretaceous-Paleocene eclogite and blueschist facies assemblages (Polino et al., 1990, and refs.therein). The latter hypothesis has been further supported by discussion on sampling mechanisms of the ophiolitic units from the closing ocean. The Apenninic and Alpine ophiolites could correspond only to thin fragments of topographic highs (e.g., horsts, rises in fracture zones, seamounts, diapirs, aseismic rises and so on) in a relatively wide ocean, being truncated by collision with the active margin when entering the trench and the subduction gate (Polino et al., 1990). These delaminated slices were subsequently sampled from different depths of the subduction zone, as shown in the Alps by high-P metamorphic signatures, and incorporated within an accretionary wedge growing at the active margin of the upper plate as a pre-collisional thrust belt (e.g. Treves, 1984; Polino et al., 1990 and refs.therein). It thus appears that Alpine and Apennine ophiolites mainly record the discontinuous and anomalous structures of a wide ocean that were scraped off the subducting oceanic lithosphere. Conversely, the sectors of normal oceanic crust devoid of topographic irregularities were hardly able to produce ophiolitic bodies, being completely consumed at depth. If the comparison with modern settings is valid, the Alpine and Apenninic ophiolites cannot absolutely account for the original width of the Piedmont-Ligurian ocean (Polino et al., 1990).

Distribution of high-P metamorphism in the Alpine nappe pile is a further useful tool for constraining the restoration procedures (Laubscher and Bernoulli, 1982; Hunziker et al., 1989; Polino et al., 1990). As previously reported, the axial sector of the Alps displays multiple alternances of ophiolitic and continental nappes with contrasting subduction signatures. When this early Alpine thrust belt is regarded as a pre-collisional orogenic wedge, two groups of restoration models may replace the classic paleogeographic reconstructions: all basement nappes of the wedge are to be allocated within the lower oceanic plate as a proper number of microcontinents (e.g. Platt, 1986), or alternatively on the active margin of the continental upper plate (Polino et al., 1990). In the former case, accretionary processes dominate the growth of the orogenic wedge, through tectonic mass transfer from the lower to upper plate and progressive stacking of basement and ophiolitic units derived from the colliding microcontinents and the intervening oceanic channels. In the second case, the continental slices may be pulled away from the underside of the highly fragmented upper plate front (tectonic erosion), when they are indented with prominent topographic structures of the subducting ocean floor. Moreover, a third solution is represented by isolated "allochtons" or "ribbon continents" of an extensional upper plate that evolved, during asymmetric rifting, to an advanced denudation of subcontinental mantle and associated gabbros (Dal Piaz et al., 1991; Oogerdnijng Strating, 1991; Dal Piaz, 1992, 1993). In any case, stacking of continental and ophiolitic nappes is expected to have been produced early by tectonic underplating at the base of the orogenic wedge (Platt, 1986). Change in reciprocal position of the nappes then extensively occurs along the upward trajectories, through ductile infoldings, decoupling and further coupling processes. It thus appears that the classic "nappe emplacement" becomes a long history, even if it is only discontinuously recorded by metamorphic and deformation signatures.

In conclusion, the ophiolitic units that are scattered at different structural levels of the Alpine nappe pile can no longer be unequivocally interpreted as sutures of two or even more oceanic gaps alternating with continental blocks. Most probably the Alpine ophiolites are dismembered records of a unique oceanic basin that has been sutured between the front of the pre-collisional orogenic wedge of the active margin and the colliding passive margin of the Europe lower plate (Polino et al., 1990). Since the onset of the continent-wedge collision, this major suture cannot necessarily be outlined by continuous ophiolitic records. Accordingly, the ocean suture may be roughly allocated along the Penninic frontal thrust (Hunziker et al., 1989).

After having summarized the main features and restoration problems of the Alpine ophiolites, we can examine some still debated questions about rifting and oceanization of the Alpine Tethys. Recognition of the overall driving forces constitutes a topical point because some events of the Western Tethys fit well the opening and changes in spreading history of the Atlantic ocean, whereas other signatures of the Piedmont-Ligurian basin seem to contrast with the Atlantic evolution in terms of timing or structural features (Dal Piaz and Polino, 1989, and refs. therein). The Western Tethyan ocean is the result of the sinistral movement along the Europe/Africa plate margin that is related to the Central Atlantic opening, or alternatively to older independent events (e.g., Biju-Duval et al., 1977; Bernoulli and Lemoine, 1980; Dercourt et al., 1986; Savostin et al., 1986; Abbate et al., 1988). Accordingly, the Piedmont-Ligurian ocean has been envisaged either as a northern segment of the Mid-Atlantic ridge diplaced eastwards by the Newfoundland-Gibraltar fault, or as a strike-slip basin independently formed through passive extensional mechanisms. In the latter case, attenuation processes of the continental lithosphere progressed up to complete crustal splitting and mantle denudation (De Candia and Elter, 1969; Piccardo, 1977, 1983; Tricart and Lemoine, 1983; Lombardo and Pognante, 1984; Weissert

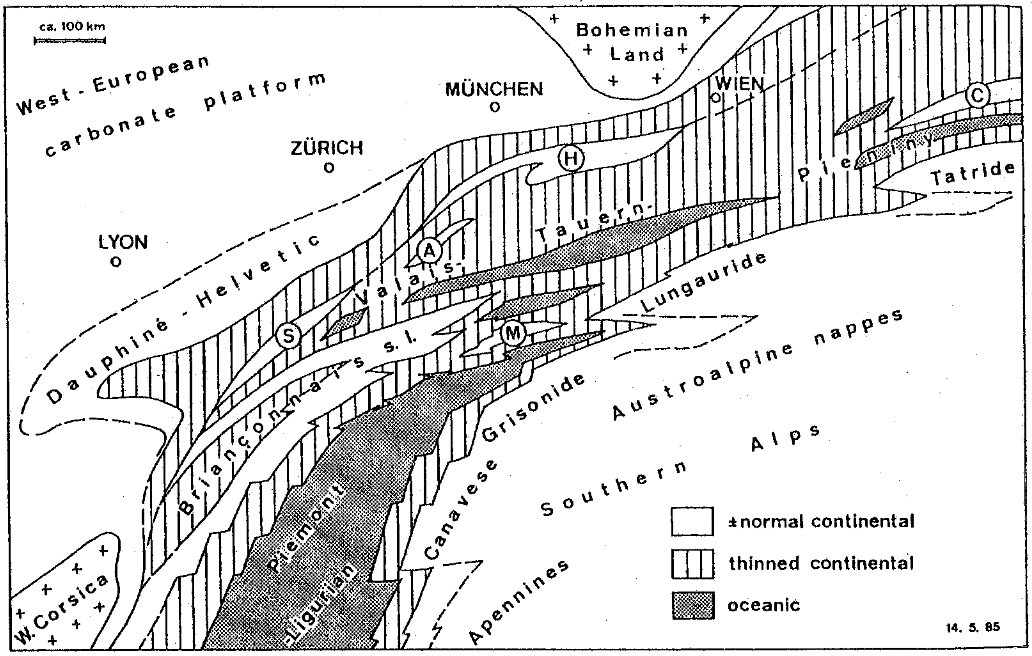

Fig. 15 - Paleogeographic domains in the Alpine Tethys during the Jurassic (Trümpy, 1980, 1988). Structural highs in thinned continental crust: S) marginal high and Cordillère Tarine, A) Adula rise, M) Margna rise, H) Hochstegen rise; C) Czorstyn rise.

and Bernoulli, 1985; Lemoine et al., 1987; Abbate et al., 1988; Dal Piaz and Polino, 1989; Piccardo et al., 1990, 1993; Vissers et al., 1991; Trommsdorff et al., 1993).

The former interpretation is supported by the lithological inventory and normal MORB affinity of mafic ophiolites, capped by Oxfordian-Callovian cherts, as well as by the oceanic metamorphism, including ocean-type Mn and Cu-Fe hydrothermal mineralizations. Transform fault systems as in modern oceans may be a reliable source of most Alpine and Apennine ophiolites.

The latter model is substantiated by the absence of a true sheeted dyke complex and especially by the occurrence of mantle-derived ultramafics with fertile lherzolitic composition from the Western-Central Alps and Apennines, probably recording slices of the Adria subcontinental mantle (Abbate et al., 1984; Ishiwatari, 1986; Piccardo et al., 1988, 1990, 1993, this volume; Trommsdorff et al., 1993). By contrast, depleted harzburgites are reported only occasionaly in the Eastern Alps (Hock, 1983). Further support is provided by continental igneous underplating of gabbro batholiths widely recorded since the early rifting stage, as well as by the occurrence in numerous gabbro bodies of a ductile flaser texture demonstrably older than the extrusion of basalts and sedimentary cover deposition (Dal Piaz and Ernst, 1978; Lombardo and Pognante, 1982; Dal Piaz and Polino, 1989; Dal Piaz, 1993; Piccardo et al., 1993; Trommsdorff et al., 1993). This view is corroborated by Triassic-Lower Jurassic fission- track age determinations on zircons from Piedmont metagabbros in the Cottian Alps (Carpena and Caby, 1984; Dal Piaz and Lombardo, 1985).

Most of the conflicting points which influenced these reconstructions may be convincingly overcome when the contrasting signatures are properly evaluated as the record of diachronous processes. In consequence, the supposedly alternative interpretations reported above may be integrated within a single evolutionary model of the Piedmont-Ligurian basin. Major steps in the evolution may be briefly summarized as follows (Dal Piaz, 1993; Trommsdorff et al., 1993, and refs. therein): 1) asymmetric attenuation of the continental lithosphere under perturbed thermal conditions, coupled with generation of a first group of gabbro batholiths emplaced from Permian times in the thinning upper plate (Adria); 2) ductile deformation and dynamic recrystallization of the gabbro bodies at decreasing pressure along a low-angle detachment zone, as recorded for instance in the Fedoz body (Trommsdorff et al., 1993); this tectonothermal evolution is shown also in gabbros from ophiolitic associations as flaser texture and relics of an "oceanic- type" metamorphism characterized by high temperature (pargasite and Mg-hornblende; e.g. Mevel et al., 1978; Abbate et al., 1988) that possibly developed as a syn-rift event; 3) further thinning and dismembering of the upper plate up to extensional unroofing of the subcontinental mantle and related gabbros which become exposed on the Jurassic ocean floor; there they may even be altered and covered by ophicarbonates or ophiolitic sandstone and breccia deposits; 4) generation of the basaltic flows with normal-MORB affinity, coupled with a second group of shallower gabbro bodies, hydrothermal activity and deposition of the oceanic sedimentary cover.

Moreover, it can be shown that the last magmatic event may predominate in some segments of the Piedmont-Ligurian ocean, and be coupled with the rise of depleted asthenospheric bodies to the lithospheric level; by contrast, in other segments this process may be poorly developed. This contrasting scenario is to be expected, especially when low spreading and transtensional mechanisms operate along different segments of the Apennine-Alpine Tethyan ocean.

Kinematic models

This chapter concentrates on the Eoalpine (Cretaceous-Paleocene) history of the Alps and on some controversial questions regarding the generation and uplift of the high-P basement and ophiolitic nappe belt. The historical basis of the birth and establishment of the nappe theory (e.g. Dal Piaz and Dal Piaz, 1984) is neglected here, as is the relatively less controversial matter of the Tertiary-Present collisional evolution.

Most kinematic reconstructions of the Alpine orogeny suggested in the last twenty years have only partially been influenced by plate tectonics. Since the first applications of modern subduction models to the re-interepretation of the high-P metamorphism (Ernst, 1971; Dal Piaz, 1971; Dal Piaz et al., 1972; Hunziker, 1974), major questions have focused on mechanisms that may have governed the generation of the Eoalpine belt, the thermal screen effect and the exhumation of high-P basement and ophiolitic units. Contrasting solutions were proposed for each of these problems, whilst there was a nearly general agreement only on the basic assumption that Adria and Europe acted as upper and lower plates respectively. Interpretations of the Eoalpine orogeny may be grouped into two main groups, i.e., the syn-collisional and pre-collisional orogenic wedge models.

Syn-collisional models

Despite the generally accepted evidence of subduction metamorphism as in modern convergent plate margins, the Alps were envisaged firstly as a collisional thrust belt since the Eoalpine stage. This was mainly due to the influence of the classic restoration of the Alpine Tethys, still constrained by Argand's palinspastic procedures, and to the misleading occurrence of subduction signatures within basement nappes supposedly derived from the continental crust of the European lower plate. The development of high-P ophiolitic and basement nappes was referred, although at different dates, to a subducting collisional megashear zone which had operated since the ocean closure and progressed towards the foreland. Accordingly, the colliding Adria and Europe continental margins were sliced, coupled with ophiolitic units, dragged down and stacked as nappe sequences within the subduction zone, consistently following the presumed order of the paleogeographic domains (e.g., Ernst, 1971, 1973; Dal Piaz et al., 1972; Martini, 1972; Hunziker, 1974; Hawkesworth et al., 1975; Dietrich and Franz, 1978; Ballèvre et al., 1986; Gillet et al., 1986; Mattauer et al., 1987).

A Late Cretaceous closure of the Tethyan ocean was suggested by models which evaluated the Eoalpine eclogitic signatures of the Penninic Monte Rosa-Gran Paradiso basement nappes, and which accepted their classic Pre-Piedmont allocation at the toe of the European passive margin (e.g., Dal Piaz et al., 1972; Hunziker, 1974, and so on). From this perspective, the European margin was already colliding with the active margin when it entered the subduction zone during the Cretaceous. This conclusion also results from models which hypothesized the

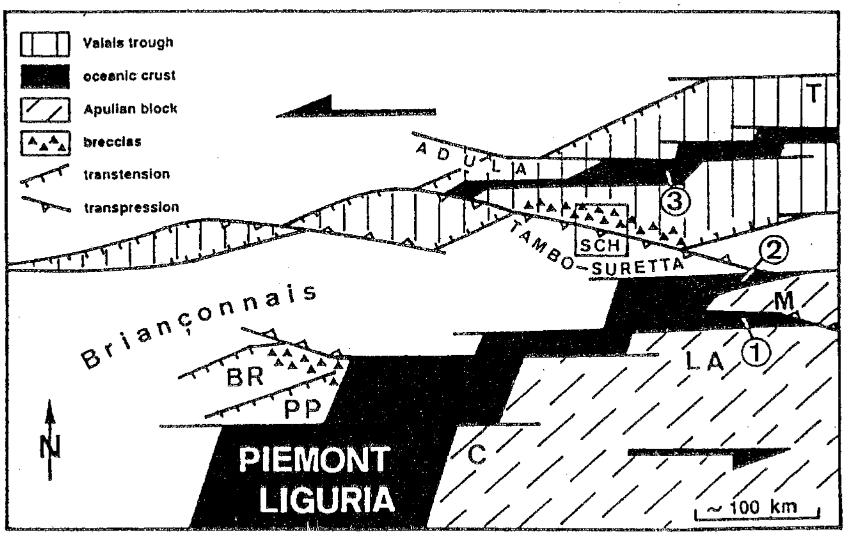

Fig. 16 - Jurassic paleostructural reconstruction of the Central Alps envisaging two oceanic basins in a transtensional system (Schmid et al., 1990). 1) Platta unit, 2) Malenco-Lizun-Avers unit, 3) Chiavenna-Malenco-Areua-Martegnas unit, BR) Breccia nappe, PP) Northern Pre-Piedmont zone, SCH) Schams nappe, T) Tauern Bündnerschiefern, M) Margna nappe, LA) Lower Austroalpine (Grisons), C) Canavese.

unusual European restoration of the eclogitic Sesia-Lanzo zone (Mattauer et al., 1987). Other reconstructions disregarded the problem of the high-P metamorphism and, following the traditional timing of the Alpine orogeny, suggested an Eocene age for the ocean closure and for onset of continental collision (e.g., Trümpy, 1973; Laubscher, 1974). Moreover, most models envisaged a single subduction zone, involving a unique Piedmont-Ligurian ocean (e.g., Dewey and Bird, 1970; Dal Piaz et al., 1972; Martini, 1972; Laubscher, 1974; Hawkesworth et al., 1975; Dietrich and Franz, 1978) or a lower oceanic plate with one or more microcontinents (e.g., Ernst, 1971; Pasquaré, 1975; Frisch, 1978; Schmid et al., 1990; Ring , 1992a, 1992b). A few models suggested also the occurrence of multiple subduction zones (e.g., Desmons and Radelli, 1989; Ring, 1992b).

Exhumation from undercrustal to superficial levels of basement and ophiolitic nappes with eclogitic and blueschist mineral assemblages is recorded to a different extent by a sequence of mineral transformations under decreasing pressure (e.g., Compagnoni, 1977; Compagnoni et al,.1977; Ernst and Dal Piaz, 1978; Kienast, 1983; Droop et al., 1990, and refs. therein). Modelling of the exhumation history of high-P bodies has been extensively carried out during the last 15 years: both tectonic mechanisms (e.g., Cowan and Silling, 1978; Pavlis and Bruhn, 1983; Platt, 1987) and related thermal consequences are treated (e.g., Richardson and England, 1979; England and Thompson, 1984; Davy and Gillet, 1986). In the Alps, this process was formerly referred to purely buoyant uprise (e.g., Ernst, 1971; England and Holland, 1979), or to a forced reversal of imbricated slices above and parallel to the collisional subduction zone operating before the mid-Tertiary thermal restoration (Dal Piaz et al., 1972). When the P-T-t paths of eclogitic and blueschist nappes were improved with further petrologic and geochronologic data and evaluated by thermal modelling, the upward rise of the high-P nappes was interpreted as isostatic uplift and erosion of a thick continental crustal duplex above the subduction zone (Rubie, 1984; Gillet et al., 1986; Chopin, 1987) or as a process facilitated by collisional wedging (e.g., Mattauer et al., 1987). Moreover, the uplift of the eclogitic bodies was estimated to occur in different units under roughly isothermal conditions (e.g., the Upper Penninic nappes: Gillet et al., 1986; Ballèvre, 1988; Pognante and Sandrone, 1989) or with decreasing or even increasing temperature (e.g., the Piedmont ophiolites and the Sesia-Lanzo zone: Thompson, 1967; Fry and Barnicoat, 1987; Pfeiffer et al., 1989; Pognante, 1989).

The thermal low assisting the development of the high-P assemblages was provided by the consuming cold oceanic lithosphere and may have been integrated by progressive underthrusting of continental slices, through outward migration of the thrust surface within the collisional subduction complex. This model may be supported by the radiometric age of the pressure peak, which seems to progressively decrease in the present nappe pile from the capping eclogitic Sesia-Lanzo zone, through the eclogitic Piedmont ophiolites and Upper Penninic basement nappes, to the lower and external blueschist basement and ophiolitic units (Bocquet et al., 1974; Hunziker, 1974; Oberhänsli et al., 1985; Gillet et al., 1986; Hunziker et al., 1989; Droop et al.1990, and refs.therein).

Orogenic wedge models

Whilst collisional models were still predominant in the Alpine area, exploration of the modern convergent plate margins had already provided well documented examples of wide and thick thrust belts, known as accretionary orogenic wedges, growing at the toe of active margins above a subducting oceanic lower plate (e.g., Karig, 1974; Dickinson and Seely, 1979; Treves, 1984; Polino et al., 1990 and refs. therein). These models explain ophiolite generation mainly as sliced fragments from topographic highs of the ocean floor entering the trench, as well as their association with tectonic mélanges and sedimentary prisms. These tectonic processes developed before the complete suturing of the ocean and the continental collision.

Despite the pioneering but generally unquoted suggestions by Roeder and Bögel (1978) and Treves (1984), application of orogenic wedge models to the Alpine belt was considered only when the classic restoration of the Penninic nappes along the European passive margin became a serious matter of criticism. Doubts about the European allocation of the Monte Rosa and other high-P Penninic nappes were expressed by Laubscher and Bernoulli (1982), Desmons (1986), Martinotti and Hunziker (1987), Hunziker et al. (1989), due to their Eoalpine subduction signatures. Nevertheless these authors did not suggest any correlations with modern orogenic wedges.

The major objection to any continent/continent collisional model as proper mechanism for generating the Eoalpine high-P thrust belt comes from the evidence that the Tethyan ocean was demonstrably open to deposition of the supra-ophiolitic flysch sequences when the Austroalpine and the Penninic basement nappes had already entered the subduction zone and experienced the eclogite facies metamorphism. In other words, these basement nappes were deeply subducted when the Piedmont-Ligurian ocean was still partly open.

As previously stated, the occurrence of basement units within an orogenic wedge may be referred either to the accretion of thin slices delaminated from lithospheric microcontinents of the lower plate (Platt, 1986) or to tectonic erosion of the active margin of the upper plate (Polino et al., 1990, and refs. therein). In the former model, major questions concern the microcontinent delamination for generating thin basement nappes and the elimination at depth of huge volumes of the remaining continental crust, especially when these continental blocks are supposed to be volumetrically equivalent to the present-day microcontinents. These difficulties can be overcome if alternatively we hypothesize that extensional denudation processes of the lithospheric mantle in the latest rifting stage may have occasionally preserved isolated fragments (allochtons, "ribbon continents") of the thinned continental crust. Even tectonic erosion may be facilitated by an extremely dismembered upper plate margin through a low-angle asymmetric detachment zone during the Permian-Early Jurassic rifting. This extensional process may generate allochtons volumetrically equivalent to the basement nappes stacked within the orogenic wedge, and some of them could have even been trapped inside the ocean as a consequence of mantle denudation processes (Polino et al., 1990; Dal Piaz et al., 1991; Ogerduijng Strating, 1991; Dal Piaz, 1992, 1993, and refs.therein).

Moreover, orogenic wedge models provide the more convincing explanation for the uplift of the high-P nappe stack from the rear sector through corner flow, deep flow, buttress effect or other mechanisms of tectonic transport, consistently assisted by extensional unroofing of the suprastructure during permanent convergence (Pavlis and Bruhn, 1983; Platt , 1986, 1987; Polino et al., 1990, and refs. therein). Translation along upward trajectories is marked by decompressional metamorphic signatures and post-nappe deformations under steady-state low thermal conditions.

However, whatever restoration models of the Western Tethys are hypothesized, all ophiolitic slices and related basement nappes constitute an Eoalpine nappe system generated at the Adria active margin before the complete closure of the Mesozoic ocean and its collision with the Europe passive margin.

CONCLUDING REMARKS

In spite of the Alpine tectonic reworking, the entire evolution of the Mesozoic Tethyan ophiolites from the rifting stage to basin closure and continental collision can still be recognized in different tectonic settings and reconstructed from place to place at different structural levels. The Jurassic oceanic stage is well expressed within the external slightly-metamorphosed units (e.g., Chenaillet and Monginevro ophiolites, Cottian Alps). The Eoalpine subduction event is particularly well recorded by the high-P units in the internal Piedmont nappe system (e.g., Zermatt-Saas unit) and in the innermost Tauern window. During the mid-Tertiary continental collision, the nappe pile recorded a thermal re-equilibration under greenschist to amphibolite facies metamorphic conditions, from the units located at the base of the Austroalpine system down to the lower Penninic Ossola-Tessin core.

Continental break-up and incipient rift magmatism

This signature is particularly well expressed in some mantle peridotite and/or gabbro units, such as from Erro-Tobbio (Voltri group), Lanzo, Matterhorn-Collon and related minor bodies (Dent Blanche and Sesia-Lanzo), the Ivrea zone, Malenco-Fedoz and from the Eastern Austroalpine basement. These bodies were emplaced under high-T and decreasing P-conditions which may be related to regional extension, upwelling of asthenospheric mantle and lithospheric thinning locally followed by mantle denudation. The model envisages partial melting of asthenospheric mantle and the generation of gabbro bodies from the Permian onwards, before the Tethyan ocean opening, through igneous underplating beneath the thinning continental crust. The origin and evolution of the ultramafic bodies which show a relatively fertile geochemical signature are discussed by Piccardo et al., this volume. At the superficial level, continental break-up is recorded by syn-rift sediments of mainly Upper Triassic-Lower Jurassic age.

Oceanic magmatism

The Jurassic oceanic magmatism is represented by volcanic extrusives and by a second generation of gabbroic intrusions, relatively younger and shallower than the continental syn-rift bodies. Evidence of a coeval sheeted dyke complex is missing, but this may even be an effect of the Alpine subduction. Most gabbros have a Mg-rich composition, although if Fe-gabbros and plagiogranites have been observed. Normal-MORB geochemical signatures, as expected for a mature oceanic ridge system, have been described for the ophiolites from the Piedmont nappe system to the Tauern and Rechnitz-Bernstein windows. Well preserved magmatic structures, such as pillow lavas and gabbros, are locally described in external slightly metamorphosed units (e.g. Chenaillet, Mongenevre, Platta-Arosa) and in blueschist to eclogitic units. Multiple interleavings of tabular greenschist facies metabasalts and calcschists occur extensively in some ophiolitic units, from the Combin zone to the Tauern window. These mafic rocks have been traditionally interpreted as sills and/or multiple basalt flows within flysch deposits. In our opinion, most of them may have been produced mainly by tectonic transposition inside Alpine shear zones.

Oceanic sedimentary cover

Stratigraphic relationships between all ophiolitic terms and the Upper Jurassic sedimentary cover are scatterly documented in small outcrops throughout the Alps. Gabbros and serpentinized mantle peridotites are sometimes directly covered by ophicalcites and/or ophiolitic metasedimentary breccias prior to basalt extrusions that locally may be missing. These occurrences have been interpreted as evidence of anomalous oceanic settings (e.g., fracture zones, seamounts and so on). The classic supra-ophiolitic cover commonly starts with Mn-rich cherts of Oxfordian-

Callovian age, as in the Apennine ophiolites. Mn-Fe-Cu mineralizations testify to hydrothermal activity during the oceanic spreading. Upwards, the sedimentary cover consists of pelagic marbles and of Cretaceous calcschists with flysch affinity.

Pre-orogenic high-temperature alteration

Evidence for syn-rift and/or oceanic alteration processes is well recorded in the Northern Apennines ophiolites and locally also in the Alpine metaophiolites. In the Alps, relics of such alteration occur in metagabbros of the Chenaillet massif and also in eclogitic (Rocciavrè) and blueschist (Rechnitz) units. Moreover, most of the serpentinized ultramafic bodies and related rodingites may be referred to similar hydrothermal processes.

Subduction

High-P/low-T metamorphic assemblages represent a basic signature for identifying the paleo-subduction tectonic events. In the Alpine ophiolites this event yelds radiometric ages ranging from Middle Cretaceous (100-90 Ma) to Early Eocene. High-P nappes of ophiolitic and continental composition were stacked during the Eoalpine orogeny. This early nappe pile may be interpreted as a pre-collisional orogenic wedge. As expected in wedge kinematics, the high-P metamorphic units may be exhumed to the surface through upward flow, buttress effect or tectonic underplating coupled with extensional unroofing. Upward mass transfer is mainly related to the inner wedge zone, the so-called "root zone" (steep belt) in the Western and Central Alps.

The kinematic evolution of the high-P nappes is recorded by a few P-prograde relics and by a sequence of mineral assemblages from the eclogitic, locally ultra-HP metamorphic climax towards the high-T blueschist and greenschist facies conditions. Incomplete eclogitic reactions in low-strain gabbro domains, such as coronitic microstructures and epitactic overgrowths, allow us to recognize the role of kinetics during the high-P metamorphism (e.g., Rubie, 1990). By contrast, only strongly deformed to high-T mylonitic domains exhibit equilibrium textures and complete eclogitic recrystallization (e.g., Pognante, 1985; Philippot and Kienast, 1989).

Post-eclogitic uplift

Metamorphic signatures of the post-eclogitic uplift trajectories up to the Mesoalpine amphibolite or greeschist facies overprint are recorded in most ophiolitic units of the Alps. The estimated P-T-time path varies from cooling or adiabatic decompression to substantial heating during exhumation (Dal Piaz et al., 1993; Spalla et al., 1994). This should correspond to differences in timing and space for the wedge evolution. Uplift during continuous subduction providing a "thermal screen effect" evolves through adiabatic decompression, while collisional uplift may provide a "cover effect" and heating (e.g., Davy and Gillet, 1986; Gillet et al., 1986). Moreover, the structural position of the Alpine ophiolitic slices with respect to the hypothetical "Lepontine thermal dome" of mid-Tertiary age (e.g., Thompson, 1976) may be taken into account (e.g., Pfeiffer et al., 1989) for the increasing temperature locally recorded during the late uplift stage.

This metamorphic evolution predates the Upper Oligocene, as it is constrained by not metamorphic calc-alkaline plutons and dykes (33-30 Ma) sharply cutting the collisional nappe pile and Mesoalpine isogrades. During the Neoalpine (Neogene) stage, the collided high-P nappe stack was affected by brittle-ductile and essentially brittle deformations and further uplift, while opposite-vergent thrusts developed widely within the Southalpine and Helvetic border domains (Polino et al., 1990; Schmid et al., 1990, and refs. therein).

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