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The Combin Fault: compressional reactivation of a Late Cretaceous–Early Tertiary detachment fault in the Western Alps

Dedicated to the memory of Ugo Pognante, Torino 1954 – Mont Blanc du Tacul 1992

by *Michel Ballèvre*[1] and *Olivier Merle*[1]

Abstract

Eclogite-facies rocks, including coesite-bearing ones, which now outcrop at the surface in the internal domain of the Western Alps were buried to a depth of 80–100 km during the eo-Alpine event (i.e. 90–100 Ma). This major collisional stage led to the overstacking of crustal slabs of the Adriatic palaeomargin and the oceanic lithosphere onto the European palaeomargin. It was followed by a period of crustal extension during which most of the exhumation of ultra-HP rocks took place along crustal detachment faults. This period of crustal extension coeval with widespread deep-marine flysch deposits predate the Tertiary tectonic events which started in the Middle to Upper Eocene (i.e. the meso-Alpine and neo-Alpine events).

This paper illustrates such a tectonic history that can be inferred from the metamorphic and structural analysis of the Combin Fault in the northern part of the Western Alps. The tremendous pressure change which can be seen on both sides of the Combin Fault can in no way be explained by the thrusting movement which occurred during the meso-Alpine event (i.e. Upper Eocene). It is the result of a detachment faulting stage which exhumed the footwall HP rocks of the Combin Fault. These HP rocks are now juxtaposed with shallower levels along the Combin detachment fault. Puzzling remnants of Sesia rocks located below the Combin Fault (e.g. the Etirol-Levaz unit) can be explained as parts of the Sesia zone that were cut off by the detachment fault and left behind during the eastward motion of the hangingwall Sesia zone. Meso-Alpine thrusting displacement (i.e. 5–10 km) along the Combin Fault can be considered as minor in regard to the displacement achieved during the earlier detachment faulting stage (i.e. 30 km).

Keywords: eclogite facies, Adriatic plate, detachment fault, tectonic evolution, Combin Fault, Western Alps.

Introduction

Since the well-known synthesis of ARGAND (1911), one could think that the knowledge of the overall structure of the Alps did not evolve as a whole. However, much new data in later studies in the Alps forced researchers to rethink some key points concerning the structure of the chain. This is exemplified in the concept of fold-nappe which appears to have now been abandoned in the Western Alps (e.g. STUTZ and MASSON, 1938 for the Dent Blanche nappe) as well as in the Central Alps (e.g. MILNES, 1974). In addition, great progress has been made in the past twenty years in the understanding of the metamorphic history of the Alps. Surprisingly, these metamorphic data are still poorly integrated into the tectonic evolution of the chain despite stimulating attempts (e.g. DAL PIAZ et al., 1972; ERNST, 1973; CABY et al., 1978; GILLET et al., 1986; PLATT, 1986). This paper is devoted to a discussion of the most relevant features of the metamorphic history of the northern part of the Western Alps, east of the Penninic Thrust, in order to show that they can be combined with structural data into a unified kinematical model.

We would like to stress that petrological studies have brought to light two different major points. Firstly, the maximum pressure to which a given unit was submitted is the only record of the

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maximum depth to which this unit was buried. This maximum pressure is usually different from that prevailing during the main ductile deformation of that unit. Secondly, a pressure (or temperature) gap between two neighbouring units implies that the contact between these two units postdates the time of the higher-P (or higher-T) metamorphism. In the case of the northern part of the Western Alps, these points are important for the understanding of the tectonic evolution from the Cretaceous to the present time.

In the beginning of this century, the "schistes lustrés" were believed to be a rather homogeneous mass of Mesozoic sediments. Lithological (BEARTH, 1967), then petrological studies (DAL PIAZ et al., 1972; KIÉNAST, 1973), established that they are cut by a major contact which was totally unexpected at the time of Argand. As a matter of fact, this contact separates a lower "schistes lustrés" unit containing early eclogite-facies assemblages from an upper "schistes lustrés" unit which never suffered eclogite-facies metamorphism. The lower unit is referred to as the Zermatt zone whereas the upper unit is referred to as the Combin zone. A general agreement on the metamorphic difference between these two units has now been reached among geologists (e.g. DAL PIAZ, 1976; CABY et al., 1978; BECCALUVA et al., 1984; MARTINOTTI and HUNZIKER, 1984; BALLÈVRE et al., 1986; DAL PIAZ and LOMBARDO, 1986). Despite this general agreement, the dramatic tectonic consequences have not been fully explored yet.

Following an early attempt (MERLE and BALLÈVRE, 1992), the goal of this paper is to show (i) that the contact separating the Zermatt and Combin zones is the most important one exposed in the Western Alps at the present time, (ii) that this fault was a large-scale detachment fault of upper Cretaceous–lower Tertiary age, (iii) that this detachment fault was reactivated as a thrust during the meso-Alpine event (i.e. upper Eocene), and (iv) that it was folded due to continued compression during the neo-Alpine event (i.e. Oligocene-Miocene). A new model for the tectonic evolution of the Alps is then proposed in which most of the exhumation of the high-pressure to ultra-high-pressure rocks is ascribed to a period of plate divergence between eo- and meso-Alpine events.

Geological setting

The structural map of the study area (Fig. 1) and the composite cross-section (Fig. 2) are constructed on the basis of previous regional syntheses (e.g. DAL PIAZ et al., 1972; DAL PIAZ, 1976; CABY et al., 1978; DAL PIAZ and ERNST, 1978; GOSSO et al., 1979; BALLÈVRE et al., 1986; DAL PIAZ and LOMBARDO, 1986; ESCHER, 1988; ESCHER et al., 1988 and 1993) as well as recent detailed studies which will be referred to below. Main units of the study area are shown on figures 1 and 2 and are described briefly below.

The *Grand Saint Bernard nappe* is now subdivided into several independent nappes (BEARTH, 1963; BURRI, 1983a, 1983b; ESCHER, 1988): the "Zone houillère", the Pontis and the Siviez-Mischabel nappes, and the Mont Fort nappe.

The "Zone houillère" mainly consists of Carboniferous and Permo-Triassic siliceous detrital sequences, with occasional Middle to Upper Triassic platform carbonates. The Alpine deformation is associated with a greenschist-facies metamorphism.

The Pontis and Siviez-Mischabel nappes (ESCHER, 1988; ESCHER et al., 1993) mainly consists of a pre-Alpine basement containing abundant relics of an amphibolite-facies metamorphism (e.g. BOCQUET, 1974; BURRI, 1983a and 1983b; WÜST and BAEHNI, 1986; DESMONS, 1992) (Fig. 3) as well as rare relics of an eclogite-facies metamorphism (THÉLIN et al., 1990; RAHN, 1991) (Fig. 3). Upper Palaeozoic monometamorphic rocks are also present. The sedimentary cover (Barrhorn and Toûno series) has been known for a long time as being of Vanoise or "Préalpes médianes rigides" affinity (ELLENBERGER, 1952) and has been fully documented recently (MARTHALER, 1984; SARTORI, 1987 and 1990). The Alpine tectonic history of the Pontis and Siviez-Mischabel nappes is polyphased. Early high-pressure assemblages include high-grade blueschists in some internal units only (Ruitor: BOCQUET, 1974; CABY et al., 1978; CABY and KIÉNAST, 1989; DESMONS, 1992) (Fig. 4). Other basement units preserve relics of blue amphiboles (BURRI, 1983a and 1983b) (Fig. 4) suggesting that low-grade blueschist conditions prevailed in the beginning of the Alpine history. The Mesozoic cover also contains low-grade blueschist assemblages (SARTORI, 1990). The retrograde history, which is associated with the last penetrative ductile deformation, took place under greenschist-facies conditions (e.g. WÜST and BAEHNI, 1986) (Fig. 5).

The Mont Fort nappe is composed of a monometamorphic basement of presumed Upper Palaeozoic age (the Métailler zone of BEARTH, 1963) and a sedimentary cover of Mesozoic age (MARTHALER, 1984; ALLIMAN, 1987 and 1989). The Alpine metamorphic history is characterized by the extensive development of mineral assemblages of high-grade blueschist facies (WOYNO, 1908;

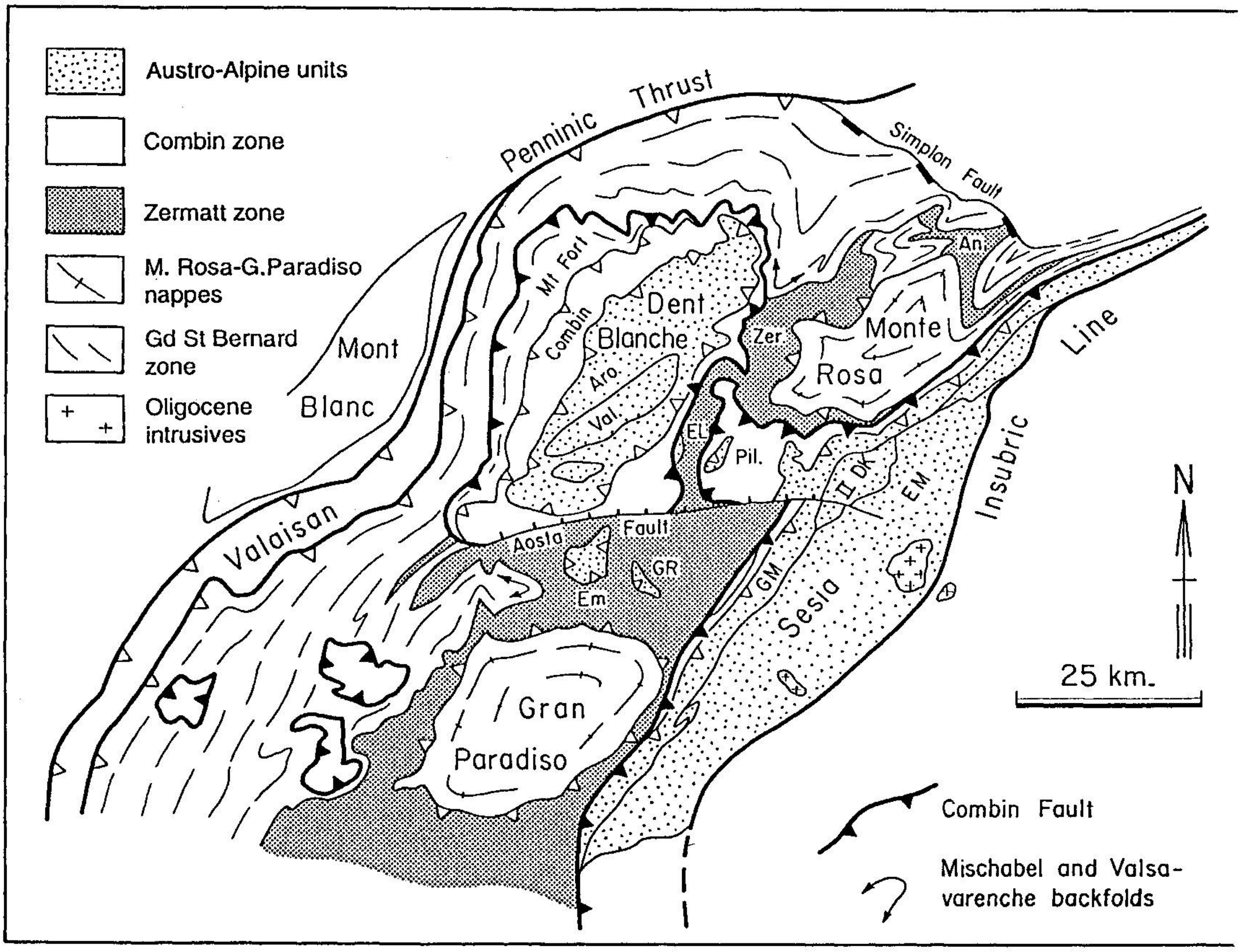

Fig. 1 Simplified geological map of the northern part of the Western Alps. An: Antrona; Aro: Arolla; EL: Etirol-Levaz; EM: Eclogitic micaschists; Em: Emilius; GM: Gneiss Minuti; GR: Glacier-Raffray; Pil: Pillonet; IIDK: Seconda zona dioritico-kinzigitica.

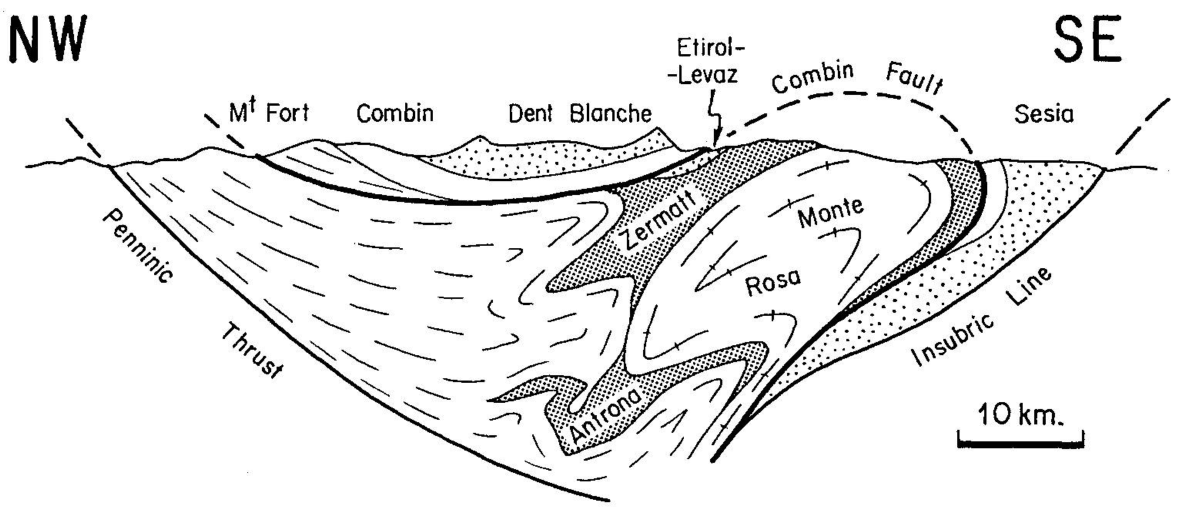

Fig. 2 Simplified and composite geological cross-section of the study area (modified from Escher et al., 1988). Same symbols as for figure 1.

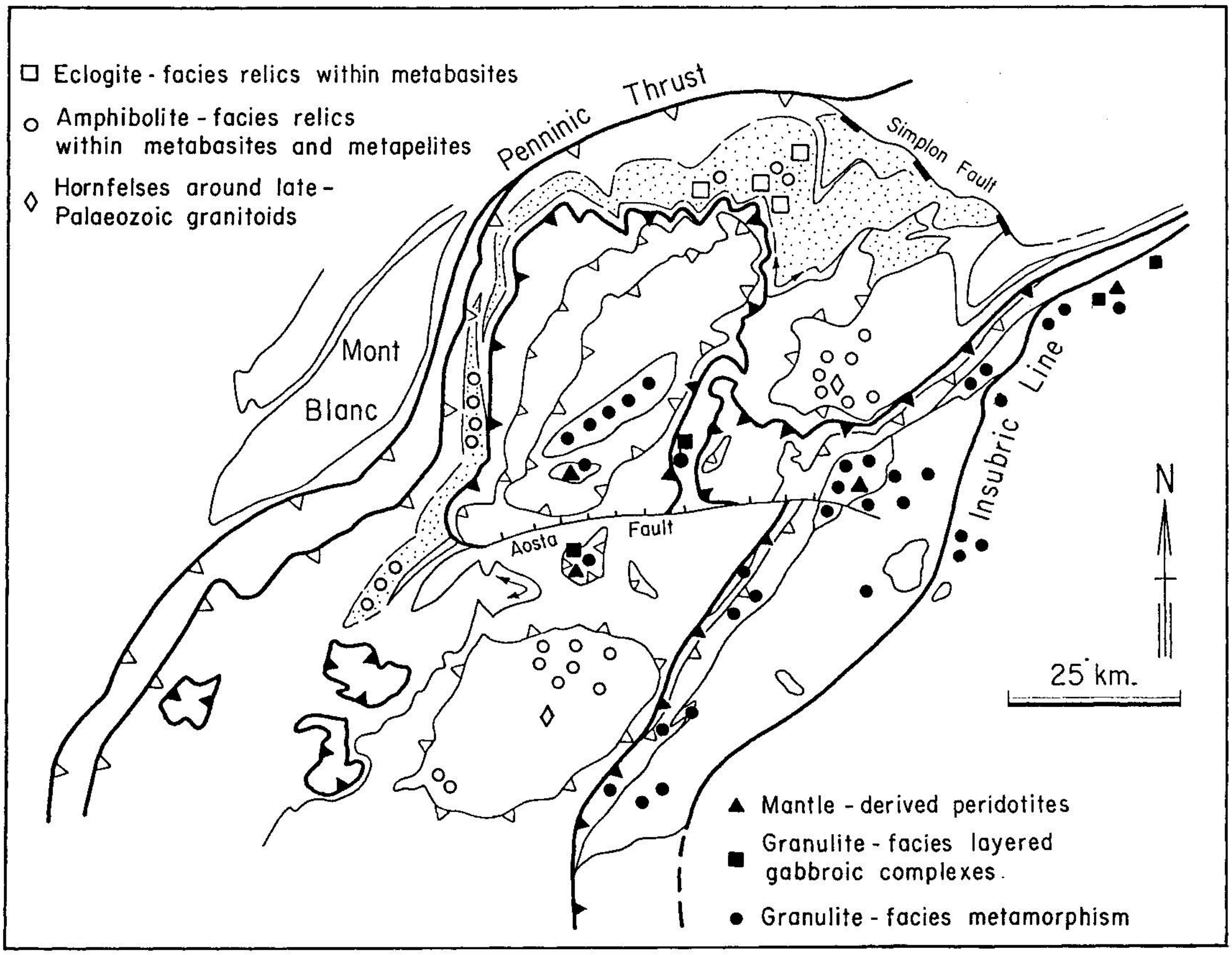

Fig. 3 Pre-Alpine metamorphism. Note that granulite-facies relics and mantle-derived peridotites are only found in Austroalpine and Southalpine units whereas amphibolite-facies metamorphism is observed in European units. Sources of data in text.

TSCHOPP, 1923; SCHÜRMANN, 1953; SCHAER, 1959; BEARTH, 1963; WÜST and BAEHNI, 1986) (Fig. 4). The retrograde history took place under greenschist-facies conditions (Fig. 5).

The *Monte Rosa and Gran Paradiso nappes* are made of pre-Alpine continental basement with relics of amphibolite-facies metamorphism (Monte Rosa: BEARTH, 1952; DAL PIAZ, 1966; LADURON, 1976; DAL PIAZ and LOMBARDO, 1986; Gran Paradiso: CALLEGARI et al., 1969; COMPAGNONI and PRATO, 1969; COMPAGNONI et al., 1974; DAL PIAZ and LOMBARDO, 1986; BALLÈVRE, 1988) (Fig. 3). Monometamorphic graphitic and conglomeratic metasediments of presumed Upper Palaeozoic age can be seen in the Gran Paradiso (Money unit: COMPAGNONI et al., 1974; BALLÈVRE, 1988). A Mesozoic cover has been reported in some places (ELTER, 1971; DEVILLE, 1989).

The Alpine tectonic history is characterized by an early eclogite-facies metamorphism (Fig. 4). Relics of this event are best preserved in mafic rocks that are transformed into eclogites (Monte Rosa: FRANCHI, 1903; DAL PIAZ and GATTO, 1963; DAL PIAZ, 1966; WETZEL, 1972; KLEIN, 1978; DAL PIAZ and LOMBARDO, 1986; Gran Paradiso: COMPAGNONI and LOMBARDO, 1974; BATTISTON et al., 1984; BENCIOLINI et al., 1984; DAL PIAZ and LOMBARDO, 1986; BALLÈVRE, 1988; BIINO and POGNANTE, 1989). It can also be observed in rocks of Mg-rich bulk composition derived either from peculiar sediments or from metasomatism within early-Alpine ductile shear zones (DAL PIAZ, 1971; WETZEL, 1972; COMPAGNONI and LOMBARDO, 1974; CHOPIN, 1981; CHOPIN and MONIÉ, 1984; DAL PIAZ and LOMBARDO, 1986; BALLÈVRE, 1988). The main ductile deformation of the Monte Rosa and Gran

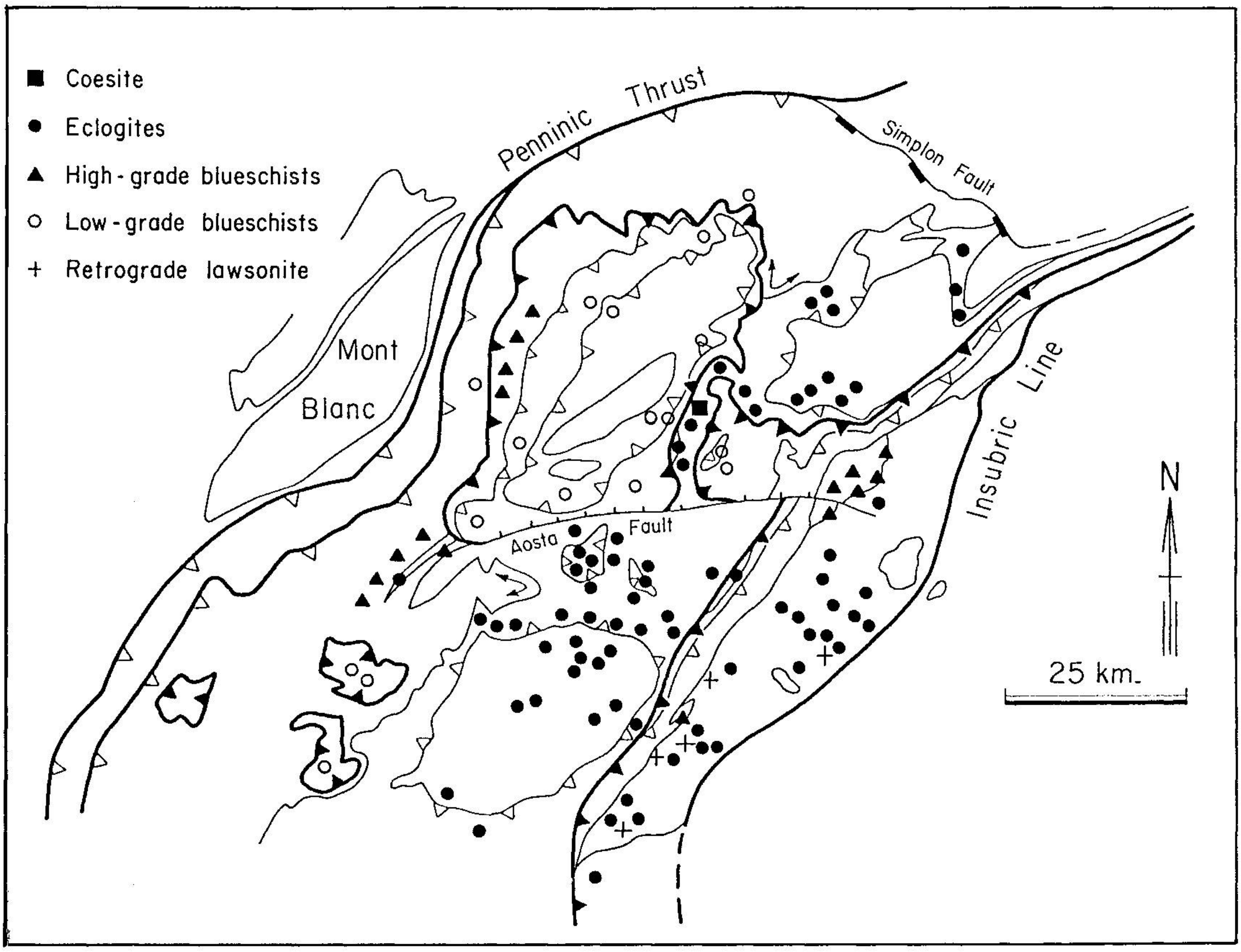

Fig. 4 Peak P-T conditions for the Alpine metamorphism. Sources of data in text.

Paradiso nappes took place under albite-epidote amphibolite facies (VEARNCOMBE, 1984; BALLÈVRE, 1984 and 1988; LE GOFF and BALLÈVRE, 1990) (Fig. 5). The P-T history of these units is mainly characterized by a nearly isothermal decompression (see continuous line on Fig. 5 after BALLÈVRE, 1988). Recent works suggest a two-stage evolution (BORGHI et al., 1992), with a late increase in temperature at low pressures (dashed line on Fig. 5).

The *Antrona and Zermatt zones* are composed mainly of metaperidotites, metagabbros and metabasalts (e.g. BEARTH, 1967; COLOMBI, 1989). In addition, some metasediments from the Zermatt zone contain Mn or Fe-sulfide mineralizations (e.g. DAL PIAZ et al., 1979a; SALIOT et al., 1980; BEARTH and SCHWANDER, 1981; TARTAROTTI et al., 1986; MARTIN and KIÉNAST, 1987). Chemical studies reveal a tholeiitic trend for the mafic rocks and a close similarity with transitional to normal MORBs (BECCALUVA et al., 1984; PFEIFER et al., 1989). Peridotites are relatively undepleted lherzolites. No significant difference exists between the Zermatt and Antrona ophiolites. Thus, they are interpreted as part of an oceanic lithosphere formed in a slow-spreading environment (BARRETT and SPOONER, 1977; BOUDIER and NICOLAS, 1985; LAGABRIELLE and CANNAT, 1990).

The Zermatt zone is well known for its high-pressure parageneses (e.g. BEARTH, 1967; ERNST and DAL PIAZ, 1978; BARNICOAT and FRY, 1986) and maximum P-T conditions are well defined. The discovery of coesite inclusions within garnet in some Mn-bearing metasediments in the Valtournanche area suggests that peak metamorphic conditions were of the order of 26–28 kbar, 590–630 °C (REINECKE, 1991) (Figs 4 and 7). The assemblage jadeite + quartz has been found within metatrondhjemites (NOVO et al., 1989). The main deformation took place at pressures signifi-

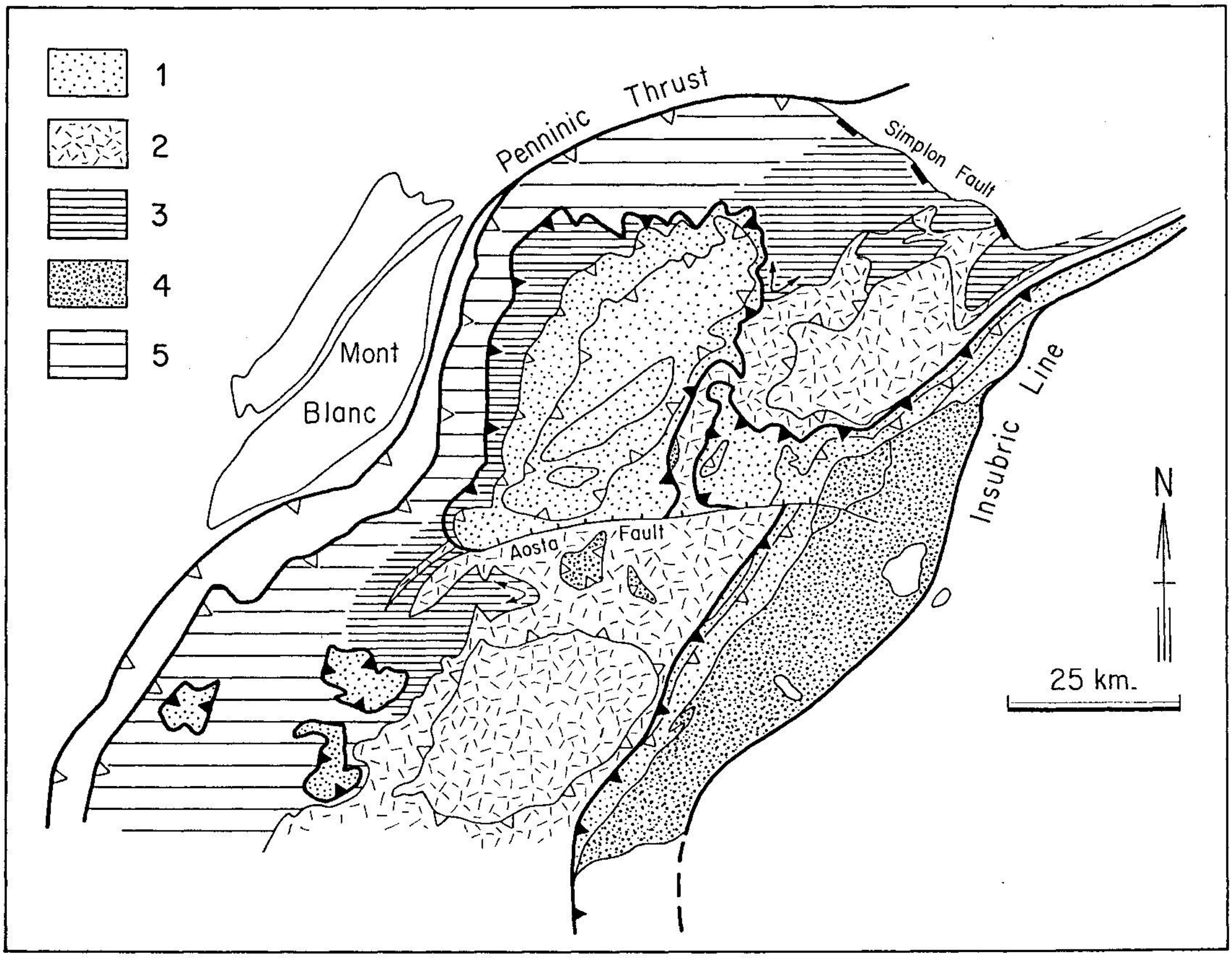

Fig. 5 P-T conditions of the last pervasive ductile deformation. 1: greenschist facies of meso-Alpine age; 2: eclogite facies to albite-epidote-amphibolite facies of eo-Alpine age; 3: high-grade blueschist facies of presumed eo-Alpine age; 4: eclogite facies of eo-Alpine age; 5: low-grade blueschist or greenschist facies of meso-Alpine age. Sources of data in text.

cantly lower than the peak pressure but at slightly lower temperatures (Figs 5 and 7), i.e. at the transition between the eclogite facies and the albite-epidote amphibolite facies.

The Antrona zone also displays relics of eclogite-facies parageneses (LADURON, 1976; COLOMBI and PFEIFER, 1986) despite a strong overprint in amphibolite facies, up to sillimanite grade (COLOMBI, 1989).

Two main types of units are recognized within the *Combin zone*.

The first type is essentially composed of marly metasediments (i.e. calcschists) and greenschists deriving mainly from basaltic volcanics but also from gabbros. Some serpentinites are also found. In addition, manganiferous quartzites as well as Cu–Fe bodies (DAL PIAZ et al., 1979a; CABY, 1981; BALDELLI et al., 1983; LE GOFF, 1986) are recognized. These lithologies can be ascribed to oceanic units (the Tsaté nappe according to SARTORI, 1987; ESCHER et al., 1988; SARTORI, 1990) and are found in the top upper part of the Combin zone. It has been proposed that the Tsaté unit was part of an accretionary prism (MARTHALER and STAMPFLI, 1989; STAMPFLI and MARTHALER, 1990).

The second type contains a sedimentary succession beginning with siliceous conglomerates (Permian), quartzites (lower Triassic) and dolomitic marbles (Middle-Upper Triassic). These are sometimes followed by calcareous sequences of presumed Jurassic age. These sediments were deposited on a continental crust and are now located as discontinuous slices at the base of the Combin zone. Their palaeogeographic origin is still not clear. According to some authors, they belong to the Mont Fort nappe (SARTORI, 1987; ESCHER et

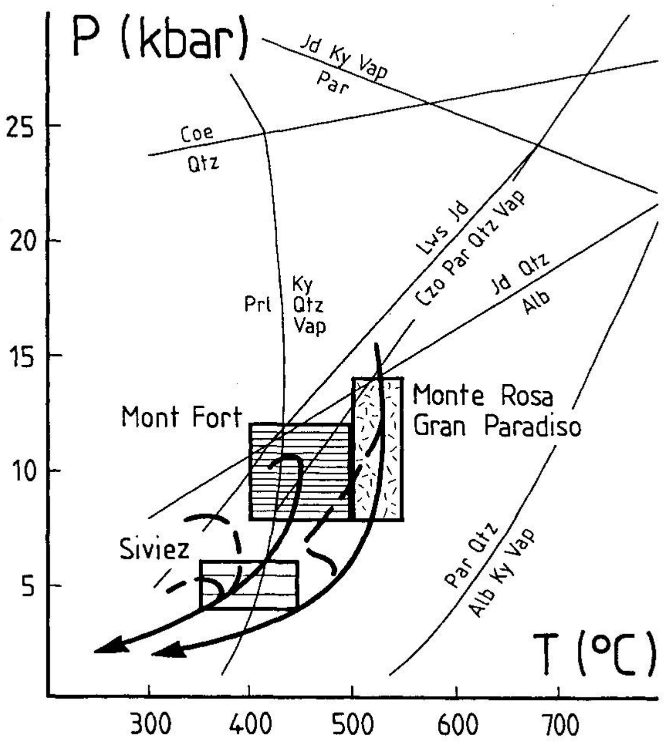

Fig. 6 P-T paths for European units. Equilibrium curves are calculated using the data set of BERMAN (1988). Alb: Albite; Coe: Coesite; Czo: Clinozoisite; Jd: Jadeite; Ky: Kyanite; Lws: Lawsonite; Par: Paragonite; Prl: Pyrophyllite; Qtz: Quartz; Vap: Vapor. Boxes show estimated P-T conditions for the last pervasive ductile deformation. Patterns within boxes are identical to those shown on figure 5.

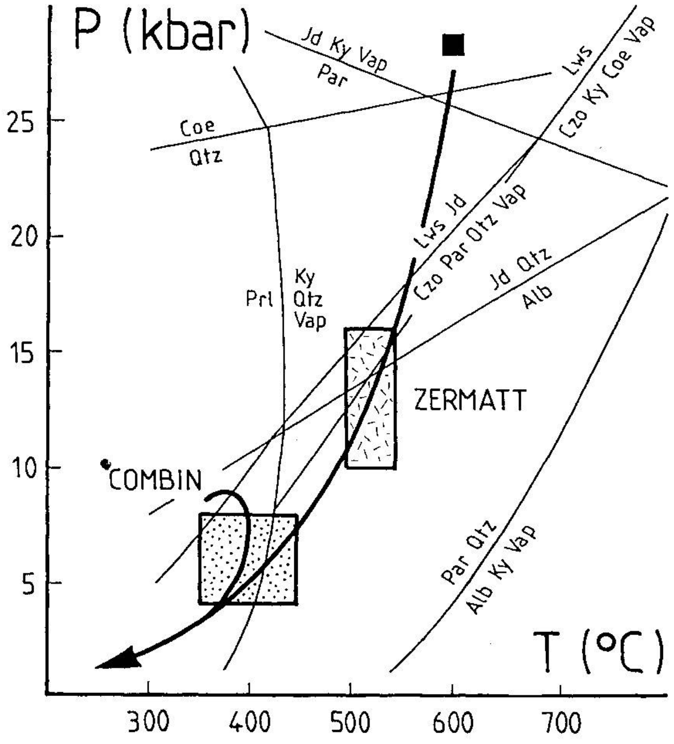

Fig. 7 P-T paths for oceanic units. Same equilibrium curves as in figure 6. Boxes show estimated P-T conditions for the last pervasive ductile deformation. Note that peak P-T conditions for the Zermatt zone (square) are significantly different from those prevailing during the last penetrative event (box). Patterns within boxes are identical to those shown on figure 5.

al., 1988; STECK, 1989; SARTORI, 1990). Other workers suggested that they belong to an independent unit (STAMPFLI and MARTHALER, 1990).

The metamorphic history of the Combin zone is characterised by an early high-pressure event. However, maximum P-T conditions within the Combin zone are still poorly documented. Relics of blue amphiboles are found in some metabasites, whereas some quartzites reveal Mn-rich garnet + sodic amphibole assemblages (DAL PIAZ, 1976; DAL PIAZ and ERNST, 1978; DAL PIAZ et al., 1979a; CABY, 1981; AYRTON et al., 1982; BALDELLI et al., 1983; SPERLICH, 1988). Carpholite pseudomorphs have been reported by PFEIFER et al. (1991). This suggests greenschist-blueschist transition, that is pressure around 6–8 kbar for temperatures of about 400 °C (Figs 4 and 7). Synkinematic assemblages in the Combin zone belong to the greenschist facies (Figs 5 and 7).

Three main types of *Austroalpine units* are classically distinguished (e.g. COMPAGNONI et al., 1977).

The first type of units (*"eclogitic micaschists"* from the Sesia zone, Etirol-Levaz, Emilius, Glacier-Raffray, Torre Ponton, Santanel) consists of a pre-Alpine basement (Fig. 3). This basement is made of paragneisses with relics of intermediate to low-pressure granulite-facies metamorphism (LARDEAUX and SPALLA, 1991). Mantle-derived peridotites have been recently discovered in the Emilius klippe (BENCIOLINI, 1989). In addition, layered gabbroic complexes with granulite-facies parageneses are known in the Etirol-Levaz (KIÉNAST, 1983; BALLÈVRE et al., 1986) and the Emilius (PENNACCHIONI, 1991) units. The pre-Alpine basement was later intruded by granitoids of Permian age (PAQUETTE et al., 1989). Sedimentary cover of Mesozoic age is rarely present (VENTURINI et al., 1991).

The Alpine history is characterized by an early eclogite-facies metamorphism (e.g. COMPAGNONI, 1977; POGNANTE et al., 1980; LARDEAUX et al., 1982; DAL PIAZ et al., 1983; KOONS, 1986; VUICHARD and BALLÈVRE, 1988; POGNANTE, 1989;

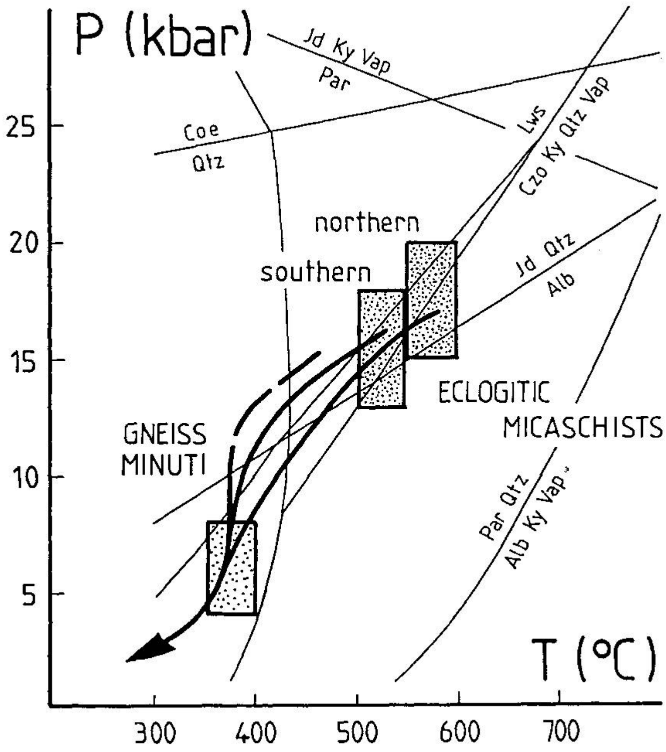

Fig. 8 P-T paths for Austroalpine units. Same equilibrium curves as in figure 6. Note the convex shape of the P-T paths that is believed to result from the structural position of the Austroalpine units with respect to the Combin Fault (see explanation in text). Patterns within boxes are identical to those shown on figure 5.

CASTELLI, 1991). The P-T history of the "eclogitic micaschists" is characterized by a strong decrease in temperature at the beginning of the decompression (Fig. 8). This is shown by the growth of chloritoid at the expense of garnet and kyanite (POGNANTE et al., 1980; VUICHARD and BALLÈVRE, 1988) as well as by the development of lawsonite porphyroblasts, in equilibrium with jadeite and quartz, in the southern part of the Sesia zone (POGNANTE, 1989) (Fig. 4). Greenschist-facies retrogression is only observed in some localised ductile shear zones, especially in the internal part of the "eclogitic micaschists" but is more developed towards the East, near the contact with the "gneiss minuti". As a whole, the main ductile deformation in the "eclogitic micaschists" took place at or near peak P-T conditions (Figs 3 and 4).

The second type of units (*"seconda zona dioritico-kinzigitica"* (IIDK) of the Sesia zone and *"Valpelline series"* of the Dent Blanche klippe) displays pre-Alpine basement which is quite distinct from that of the first type, especially because of the lack of late-Variscan granitoids (Fig. 3). Granulite-facies parageneses are well preserved there (IIDK: DAL PIAZ et al., 1971; VUICHARD, 1987; POGNANTE et al., 1988; Dent Blanche: ARGAND, 1934; NICOT, 1977). In addition, mantle-derived peridotites are described in the Sesia zone (BECCALUVA et al., 1979) as well as in the Dent Blanche klippe (CESARE et al., 1989). This type of units is very similar to the Ivrea zone (e.g. CARRARO et al., 1970), i.e. to the lower crust of the South-Alpine domain (e.g. RIVALENTI et al., 1984; ZINGG et al., 1990; RUTTER et al., 1993).

The Alpine history is still poorly known, especially for the Valpelline series. Incomplete recrystallizations during an early, high-pressure event are observed in most of the rocks from the IIDK (DAL PIAZ et al., 1971; LARDEAUX et al., 1982; POGNANTE et al., 1988; VUICHARD, 1989). These parageneses crystallized at slightly lower P-T conditions than those within the "eclogitic micaschists" and are ascribed to the high-grade blueschist facies. Such parageneses have also been reported from the Valpelline series (KIÉNAST and NICOT, 1971).

The third type of unit (*"gneiss minuti"* of the Sesia zone, *"Arolla series"* of the Dent Blanche and Pillonet klippen) consists mainly of granitoids of presumed Variscan age, with a minor amount of pre-Alpine country-rocks (Fig. 3). Mesozoic sediments have been identified in the Sesia zone (FUDRAL and DEVILLE, 1986; VUICHARD, 1989) as well as in the Pillonet (DAL PIAZ, 1976) and the Dent Blanche (DAL PIAZ, 1976; AYRTON et al., 1982; BALLÈVRE et al., 1986; BALLÈVRE and KIÉNAST, 1987) klippen.

The Alpine history is mainly defined by an extensive greenschist-facies metamorphism (e.g. LATTARD, 1974; MAZUREK, 1986) (Figs 5 and 8). Because of this latter, relics of high-pressure parageneses of presumed eo-Alpine age are difficult to recognize. Moreover, it cannot be precluded that part of the greenschist-facies metamorphism in the Dent Blanche nappe could be of eo-Alpine age as the Dent Blanche klippe represents the frontal part of the eo-Alpine nappe stack. In the "gneiss minuti", relics of sodic amphiboles, sodic pyroxenes and the high Si content of the white micas suggest an early higher pressure event (Dent Blanche: AYRTON et al., 1982; OBERHÄNSLI and BUCHER, 1987; CESARE et al., 1989; CANEPA et al., 1990; Pillonet: DAL PIAZ, 1976; VOGLER, 1984; DAL PIAZ and MARTIN, 1986) (Fig. 4). Moreover, the assemblage jadeite + quartz has been reported within some orthogneisses from the "gneiss minuti" (POGNANTE et al., 1987; POGNANTE, 1989; SPALLA et al., 1991) (Fig. 8). In the Mesozoic cover of the Dent Blanche klippe, some quartzites display garnet + blue amphibole assemblages, but their formation is related to the Mn-rich bulk-composition of the sediment (BALLÈVRE and KIÉNAST, 1987).

Structural relationships between the Austroalpine units are still a matter of controversy but can be considered as a key element in the understanding of the kinematical evolution of the Western Alps. According to STUTZ and MASSON (1938) for the Dent Blanche klippe and CARRARO et al. (1970) for the Sesia zone, most authors makes a clear distinction between a "lower element" (i.e. Arolla series, gneiss minuti and eclogitic micaschists) and a "upper element" (i.e. Valpelline series and IIDK). The upper element was thrusted over the lower element in the early stages of the Alpine history as the contact between the IIDK and the eclogitic micaschists is a high-pressure mylonitic zone (DAL PIAZ et al., 1972; LARDEAUX et al., 1982; RIDLEY, 1989; VUICHARD, 1989). This implies that the mylonitic zone observed at the contact between the Arolla and Valpelline series in the Dent Blanche klippe is probably also of eo-Alpine age.

The subsequent stages of the tectonic evolution of the Austro-Alpine units are still unclear. Most authors now agree that the threefold division of the Sesia zone results from a late, heterogeneous, greenschist-facies deformation (e.g. WILLIAMS and COMPAGNONI, 1983; VUICHARD, 1986; RIDLEY, 1989; SPALLA et al., 1991). According to SPALLA et al. (1991), this late deformation overprinted almost completely the early, eclogite-facies deformation in the external part of the Sesia zone. Large-scale folding of nappe boundaries within the Dent Blanche klippe (STUTZ and MASSON, 1938; CANEPA et al., 1990) could have occurred also during this period. Nevertheless, one must stress that none of the previous models explains the location of some Austroalpine units (e.g. the Etirol-Levaz unit) below the Combin zone. We return to this problem later.

The *Tertiary magmatism* is represented by a few plutons of dioritic to syenitic composition (Fig. 1), numerous lamprophyric dykes and a few remnants of a volcano-sedimentary cover (NOVARESE, 1943; ZINGG et al., 1976; DAL PIAZ et al., 1979b; VENTURELLI et al., 1984). This magmatism is only known in the vicinity of the Insubric Line, especially in the Austroalpine units (Traversella and Biella plutons) but also within Southalpine units (Miagliana pluton) (Fig. 1). The age of this magmatism is well established at around 30–33 Ma (i.e. Lower Oligocene) (KRUMMENACHER and EVERNDEN, 1960; CARRARO and FERRARA, 1968; HUNZIKER and BEARTH, 1969; HUNZIKER, 1974; SCHEURING et al., 1974; ZINGG et al., 1976; DIAMOND and WIEDENBECK, 1986).

Major discontinuities resulting from the Alpine collision

This paper deals with the structural evolution of the internal part of the northwestern Alps (i.e. between the Penninic Thrust and the Insubric Line). Following earlier syntheses in which several types of nappe boundaries are distinguished from their deformation and metamorphic history (e.g. CABY et al., 1978; GOSSO et al., 1979), three main classes of faults can be recognized in the study area (Figs 1 to 4).

The first class consists of faults separating units with similar (or slightly different) P-T evolutions. In the field, three main faults are part of this group. Firstly, the IIDK of the Sesia zone was thrusted over the eclogitic micaschists during an early, high-pressure event as shown by mylonitic zones with synkinematic high-pressure assemblages (DAL PIAZ et al., 1972; LARDEAUX et al., 1982; RIDLEY, 1989; VUICHARD, 1989). Secondly, the Austroalpine units were thrusted over the Zermatt zone during the eo-Alpine event as can be inferred from the structural position of the Etirol-Levaz, Emilius and Glacier-Raffray units. An intense eclogite-facies ductile deformation is associated to the thrusting of the Austroalpine units over the Zermatt zone. This contact was strongly folded before greenschist-facies overprint (PENNACCHIONI, 1991). Thirdly, the Zermatt zone was thrusted over the Gran Paradiso and Monte Rosa European units (ELTER, 1971). Contrary to the two previous ones, this contact may have been reactivated later, i.e. after the high-pressure eo-Alpine metamorphism as it is marked by mylonitic zones developed during greenschist-facies metamorphism (CARPENA and CABY, 1983; BALLÈVRE, 1988).

The second class of faults corresponds to faults where higher-P parageneses are observed in the footwall than in the hangingwall indicating a tremendous pressure gap on both sides of the fault. The most striking fault of this class is a large-scale fault, referred to here as the Combin Fault, that separates two groups of units (e.g. KIÉNAST, 1973; DAL PIAZ, 1976; CABY et al., 1978; BALLÈVRE et al., 1986). The Siviez-Mischabel, Monte-Rosa, Zermatt and Etirol-Levaz units are located in the footwall of the Combin Fault. From west to east, the hangingwall displays the Mont Fort, Combin and Dent Blanche–Sesia units. The pressure gap is best shown in the Valtournanche area where peak pressures in the footwall are of the order of 26–28 kbar (REINECKE, 1991). In contrast, maximum pressures in the hangingwall may be of the order of 6–8 kbar.

The third class of faults crosscuts the entire nappe stack as well as isograds of the amphibolite-facies in the Central Alps (i.e. the Lepontine metamorphism). Two such faults are on a regional scale. The Simplon Fault (BEARTH, 1956; MANCKTELOW, 1985 and 1990; MANCEL and MERLE, 1987) is a low-angle fault with a top to the south-west displacement. The main movement along the Simplon Fault occurred during the Miocene (18–12 Ma according to MANCKTELOW, 1990). The Aosta Fault (STELLA, 1905) is a late subvertical fault revealing a downthrow of the northern block with respect to the southern block. Movement along the Aosta Fault occurred in the range from 30 Ma to 25 Ma according to structural relationships and geochronological data (DIAMOND and WIEDENBECK, 1986; HURFORD et al., 1991).

It should be stressed that the Combin Fault is largely ignored in most models on the tectonic evolution of the Western Alps (e.g. GILLET et al., 1986; MATTAUER et al., 1987; LACASSIN et al., 1990). These models dealt with the tectonic history of faults of class I only. Recent models deal with the tectonic significance of faults of class III (e.g. HUBBARD and MANCKTELOW, 1992).

The Combin Fault

GEOMETRICAL FEATURES

In the western and central parts of the Combin Fault, the contact between overlying (i.e. the Combin unit, the Dent Blanche and the Pillonet klippen) and underlying units (i.e. the Zermatt zone) is flat-lying. To the east, the Combin Fault is rotated into a vertical position and roots in front of the Sesia zone. The shape of the Combin Fault is the result of later folding and faulting (Figs 1 and 2). We stress the following structural features.

1. The Combin Fault is folded into a huge antiform around the top of the southern termination of the Monte Rosa nappe (see detailed map of the contact between the Zermatt and Combin zones in BEARTH, 1967; DAL PIAZ, 1976; DAL PIAZ et al., 1979a; DAL PIAZ, 1988). This southern termination can be interpreted as the place where the Vanzone antiform (ARGAND, 1911; BEARTH, 1957; LADURON and MERLYN, 1974; LADURON, 1976; KLEIN, 1978; MILNES et al., 1981) dies out along strike.

2. The Combin Fault is cut off by the late steeply-dipping Aosta Fault. The cross-cutting relationships can be traced out over a few tens of kilometres to the east of Châtillon-Saint Vincent (e.g. ELTER, 1988). To the West, the Aosta Fault can be traced along much of the Dora valley in accordance with jumps in metamorphic grade (Figs 3 and 4) and patterns of exhumation rates as revealed by fission-track data (HURFORD et al., 1991).

3. The Combin zone was largely eroded to the South of the Aosta fault as the southern block of this fault was uplifted with respect to the northern block. Nevertheless, one can note that the "schistes lustrés" klippen of the Vanoise area might well be remnants of the frontal part of the Combin zone (CABY et al., 1978; BALLÈVRE et al., 1986) as they present similar lithological, structural (ELLENBERGER, 1958) and petrological (BOCQUET, 1974) characteristics.

4. The relationships with backfolds (and backthrusts) is well-known at a broad scale as the Mischabel fold and the Monte Rosa post-nappe fold can be observed below the sub-horizontal Dent Blanche unit (e.g. ARGAND, 1934). It can also be seen in the Täsch valley to the north (SARTORI, 1987) and in the Aosta valley to the south (CABY, 1981). Both in the north and in the south, backthrusting and backfolding are located in the footwall units underneath the flat-lying Dent Blanche unit. However, in the Täsch valley, large-scale east-vergent recumbent folds can be seen affecting the Combin Fault (SARTORI, 1987).

5. A few elongate Austroalpine bodies of Sesia affinity are located in the footwall of the Combin thrust, as can be established unambiguously for the Etirol-Levaz unit (CABY et al., 1978; SALIOT et al., 1980; BALLÈVRE et al., 1986) (Figs 1 and 2) or has been argued for the Emilius, Glacier-Rafray and Torre Ponton units (BALLÈVRE et al., 1986).

KINEMATICS

Only few data are available to unravel the kinematics of the Combin Fault. A cross-section through most of the units is exposed in the Valtournanche, including the Etirol-Levaz unit, and can thus be considered as a key area. Recent mapping in this area has enabled a better definition of boundaries between major units (LE GOFF, 1986; DAL PIAZ, 1988; VANNAY and ALLEMANN, 1990) as well as an improved understanding of their deformational histories (LE GOFF, 1986; VOGLER, 1987; STECK, 1989; VANNAY and ALLEMANN, 1990). Main results are briefly summarized here.

In the Zermatt zone, the last penetrative ductile deformation is characterized by a moderatly-dipping foliation (around 30 to 50° to the west or the north-west), parallel to the lithological boundaries. The stretching lineation is defined by the shape fabric of sodic pyroxenes and sodic amphiboles and by the orientation of pressure shadows

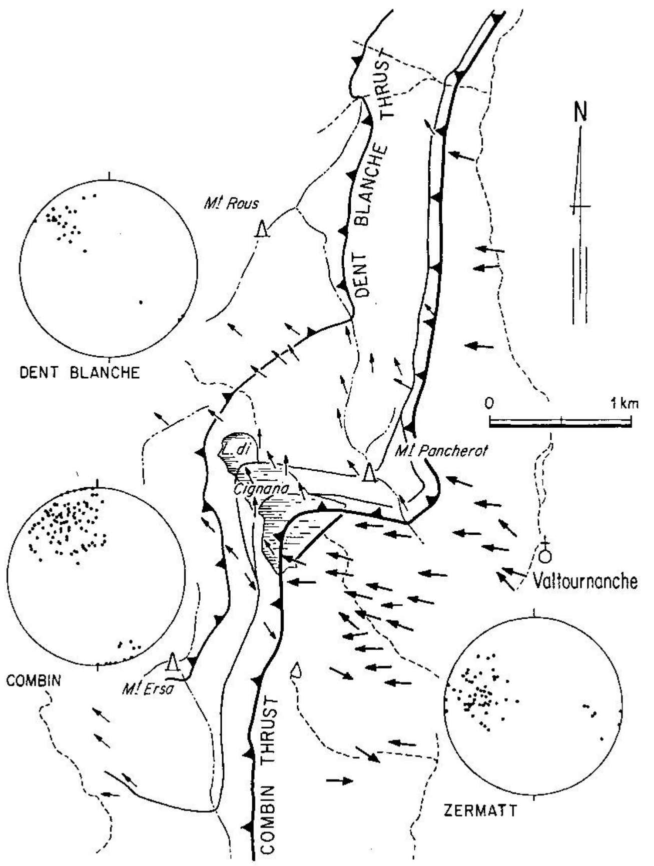

Fig. 9 Stereoplots (lower hemisphere) and trends of stretching lineations on both sides of the Combin Fault in the Valtournanche area (further explanation in text). Data from LE GOFF (1986) and unpublished data.

around garnet porphyroblasts and lawsonite pseudomorphs. The stretching lineation trends east-west (Fig. 9). Numerous shear bands can be observed within the metabasites. These frequently occur in conjugate sets and cannot be used to determine the bulk sense of shearing. Synkinematic minerals grew during the eclogite-facies metamorphism or at the transition between the eclogite facies and the albite-epidote amphibolite facies (see box in Fig. 7).

The Combin Fault truncates the lithological layering and the foliation of the Zermatt zone (LE GOFF, 1986). This is observed not only in the Valtournanche area, but also at a larger scale (CABY et al., 1978; CABY, 1981). In the Combin zone and at the base of the Dent Blanche nappe, foliation of the last penetrative ductile deformation is flat-lying (5–20° to the north-west) and is parallel to the thrust plane. Foliation planes bear a prominent stretching lineation, defined by strained pebbles within conglomerates of presumed Liassic age, or by mineral aggregates and pressure shadows around rigid objects. The stretching lineation consistently strikes NW–SE (Fig. 9). Consistent sense of shearing can be seen from shear bands within metasediments (i.e. calcschists) and from asymmetric quartz fabrics within Triassic quartzites (LE GOFF, 1986). They always indicate a top-to-the-NW sense of shear. Synkinematic minerals are coeval with the greenschist-facies metamorphism (see box in Fig. 7).

AGE OF THRUSTING

The age of the Combin thrusting is estimated using several independent methods.

Firstly, early Tertiary sediments are palaeontologically dated in the footwall of the Combin Thrust. The sedimentary cover of the Siviez-Mischabel nappe (Barrhorn and Toûno series) reveals Upper Cretaceous calcschists (ELLENBERGER, 1952; MARTHALER, 1984; SARTORI, 1987) followed by the mid-Eocene "Flysch noir" (SARTORI, 1990). This is consistent with the discovery of Upper Cretaceous microfaunas within the sedimentary cover of the Mont Fort nappe (MARTHALER, 1984) i.e. in the hangingwall of the Combin Fault.

Secondly, at outcrop scale as well as at map scale (Fig. 4), the base of the Combin Thrust is underlined by a thick zone of highly ductile deformation in greenschist-facies conditions. This shear zone includes not only the Combin zone but also the basal part of the Dent Blanche–Sesia (i.e. the "gneiss minuti") unit (MAZUREK, 1986; VUICHARD, 1986; STÜNITZ, 1989; SPALLA et al., 1991). This deformation overprints earlier high-pressure assemblages (see Figs 3 and 4 and previous discussion). Despite the fact that no detailed isotopic study is available so far in this sheared zone, greenschist-facies assemblages are thought to have crystallised at around 40 Ma. This widely-accepted interpretation (e.g. COMPAGNONI et al., 1977; MARTINOTTI and HUNZIKER, 1984; HUNZIKER et al., 1989) is based on a few K–Ar and Rb–Sr data on micas from the Combin zone (samples C.97 and C.99 of DELALOYE and DESMONS, 1976; sample KAW 1833 of AYRTON et al., 1982), from the external part of the Sesia zone (samples KAW 415 and 475 of HUNZIKER, 1969; HUNZIKER and BEARTH, 1969; HUNZIKER, 1974) or from the Mesozoic cover of the Arolla series (samples KAW 1527 and 1529 of AYRTON et al., 1982).

Thirdly, the Combin thrusting predates the emplacement of the lamprophyric dykes of Oligocene (30–33 Ma) age. Crosscutting relationships between dykes and main tectonic structures show consistently that Oligocene intrusions postdate all ductile deformations in this area (DAL PIAZ et al., 1971; GOSSO et al., 1979; STÜNITZ, 1989).

To sum up, the Combin thrusting took place during the early Tertiary, most probably at around 45–40 Ma.

INTERPRETATION (FIG. 10)

As stated above, greenschist-facies metamorphism in the hangingwall of the Combin Fault is coeval with the meso-Alpine thrusting whereas eclogite-facies metamorphism in the footwall is related to the eo-Alpine event. The meso-Alpine Combin thrusting is responsible for low-grade blueschist or greenschist-facies metamorphism in European units located below the contact. P-T conditions are in the range of 4–6 kbar and 350–450 °C in the Mesozoic sediments in the Barrhorn unit below the front of the Combin Thrust (SARTORI, 1990). This implies that the overriding plate (i.e. the Combin zone and the Dent Blanche nappe) was around 15–20 km thick at that time. In this sense, relics of high-grade blueschist-facies in the overriding plate (e.g. in the Mont Fort nappe) have occurred prior to the meso-Alpine event and have to be ascribed to the eo-Alpine event. Thus, there is a large pressure discontinuity on both sides of the Combin Fault. The meso-Alpine thrusting movement cannot account for this pressure discontinuity. As a matter of fact, we face a major problem as we are considering a meso-Alpine thrust which is rooted at great depth and apparently brings blueschist-facies rocks onto eclogite-facies rocks. As the Combin Thrust was not reactivated in the late Tertiary, it is the most likely that the Combin Thrust corresponds to the reactivation of an earlier fault that was responsible for the pressure discontinuity now observed in the field. Disregarding the late meso-Alpine thrusting, the metamorphic distribution on both sides of the fault is that of a huge normal fault which displaced the hangingwall eastward for several tens of kilometres.

Such an interpretation is enhanced by the elongate bodies of Sesia affinity scattered below the Combin Fault (i.e. the Etirol-Levaz unit and other similar units). They can be interpreted as pieces of the Sesia zone that were cut off by the detachment fault and remained on top of the Zermatt zone as part of the footwall. This probably results from the fact that the detachment fault smoothed a previously folded eo-Alpine contact (see the study of the contact between the Emilius klippe and the Zermatt zone by PENNACHIONI, 1991).

These remnants of Sesia rocks left behind below the decollement zone make possible to propose a minimum estimate of the displacement achieved along the detachment. The distance between the most external remnants and the Sesia zone is in the range of 20 to 25 km. Considering further the forward meso-Alpine thrusting movement, the displacement is likely to be in the range from 25 to 30 km. Assuming that the mean dip of

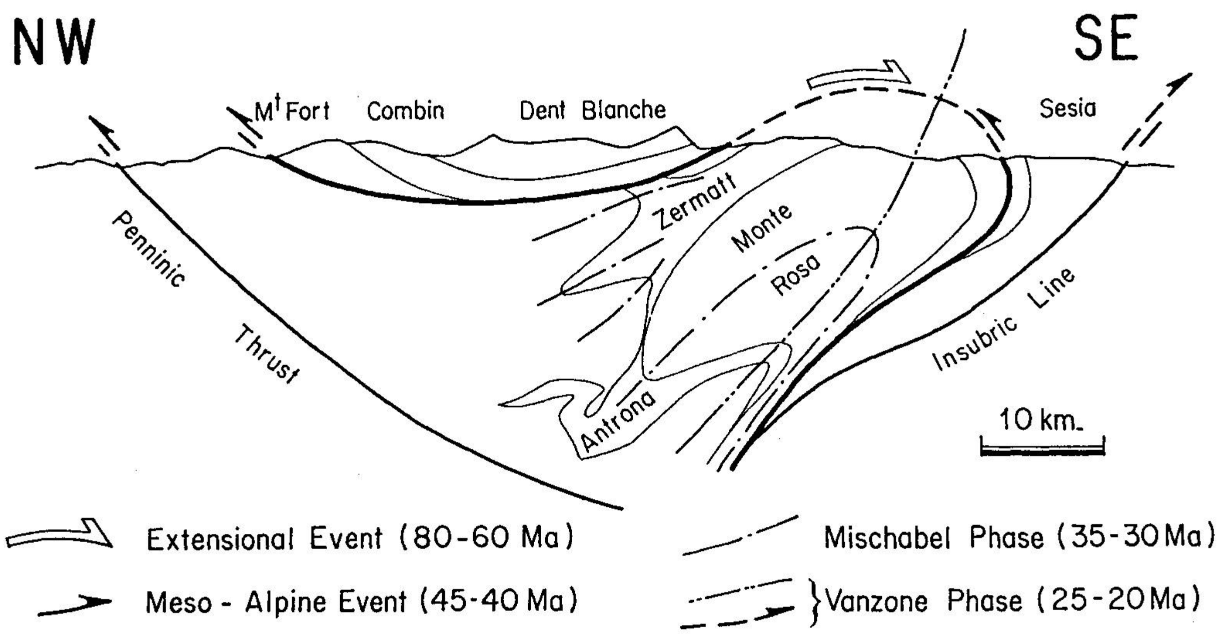

Fig. 10 Cross-section showing ages of displacement along major faults and traces of axial planes for Mischabel and Vanzone phases.

the normal fault was around 30° when active, the pressure discontinuity now observed on either side of the contact implies a displacement of about the same order of magnitude.

Tectonic evolution of the Western Alps

THE EO-ALPINE EVENT (FIG. 11)

The eo-Alpine event corresponds to a period of convergence between both tectonic plates, which led to the overthrusting of the oceanic lithosphere (i.e. the Zermatt zone) and of continental slices of the Adriatic plate upon the European palaeomargin. Pieces of continental crust (i.e. the Monte Rosa and Gran Paradiso massifs, the Sesia zone) and oceanic lithosphere (i.e. the Zermatt zone) were deeply buried to eclogite-facies conditions at that time.

Very little is known about the kinematics of this major continental collision. East-west striking HP stretching lineations in the Zermatt zone (Le Goff, 1986; Steck, 1989) (Fig. 9) are consistent with the top to the west shearing reported at the eastern margin of the Central Alps (e.g. Ring et al., 1988) and in the Eastern Alps (e.g. Ratschbacher et al., 1989). However, north-south eclogite stretching lineations in the Sesia zone (Choukroune et al., 1986; Vuichard, 1986) as well as in the Viso oceanic unit (Philippot, 1990; Ballèvre et al., 1990) make these kinematics uncertain.

The age of the eo-Alpine event can be established using three independent methods.

1. Palaeomagnetic studies reveal a major change of the relative movement between Africa and Europe between the magnetic anomalies M0 and 3434, that is between 118 Ma and 84 Ma (Savostin et al., 1986; Dewey et al., 1989). This time bracket corresponds to the period during which the convergence between both tectonic plates started. The precise timing of this major change cannot be dated more accurately from palaeomagnetic studies as no reversal of the Earth's magnetic field occurred during this period.

2. It has been known for a long time that the late Cretaceous and the early Tertiary in the Alps correspond to a period of widespread flysch sedimentation (e.g. Kerckhove, 1980; Caron et al., 1989). Flysch facies are typically deep-marine deposits. They are mainly due to the erosion of rocks of Austroalpine and Southalpine affinities. First occurrence of terrigenous beds is dated of the Coniacien-Santonien in the Helminthoid Flysch now preserved within Prealpine klippen (Caron et al., 1979). Flysch deposits from the Southalpine domain are also of the same age (Elter et al., 1966; Aubouin et al., 1970). This reveals the existence of an eroding mountain belt during the Upper Cretaceous. The eo-Alpine event which is believed to be the cause of the eroding relief is then slightly earlier or coeval with the beginning of flysch deposits.

3. The age of the eclogite-facies metamorphism is a matter of controversy in the Western Alps.

In the Austroalpine units, Rb–Sr whole-rock and mineral isochrons from the Sesia zone are used to argue an Early-Cretaceous age for the eclogite-facies metamorphism (i.e. 130–110 Ma: Oberhänsli et al., 1985). The significance of these ages is still a matter of debate. Early stages of cooling in the late Cretaceous (i.e. 80–70 Ma) are recorded by Rb–Sr and K–Ar data on micas (Hunziker, 1974; Oberhänsli et al., 1985; Stöckert et al., 1986). The final stages of the exhumation history are well-constrained by fission-track data on zircon and apatite (Hurford and Hunziker, 1985; Hurford et al., 1989).

Geochronological data in European units (Chopin and Maluski, 1981; Chopin and Monié, 1984; Monié, 1985) as well as in the Zermatt zone (Bocquet et al., 1974) were first considered as consistent with the Cretaceous age of the eclogite-facies event. However, recent studies emphasize scattered Tertiary ages which are very often used to reconsider the age of the eclogite-facies metamorphism and to relate it to the meso-Alpine event (e.g. Monié and Philippot, 1989; Barnicoat et al., 1991).

We believe that it is hardly a coincidence if data in European and oceanic units yield ages difficult to interpret whereas data from the Sesia zone are much more consistent with a late Cretaceous age for the eo-Alpine event. It is a matter of fact that the Sesia zone which is located in the hangingwall of the Combin Fault was not buried again when the meso-Alpine event occurred. On the contrary, units in the footwall of the Combin fault were buried a second time in the Tertiary as soon as the Combin Thrust was active. Ages from the footwall are therefore highly questionable as they are most probably partially or totally reset. For these reasons, we consider that data from the hangingwall (i.e. the Sesia zone) are probably the best geochronological estimate of the age of the eo-Alpine event in the Western Alps.

THE EXTENSIONAL EVENT (FIG. 11)

Following the eo-Alpine event, some crustal-scale thrusts were reactivated as detachment faults.

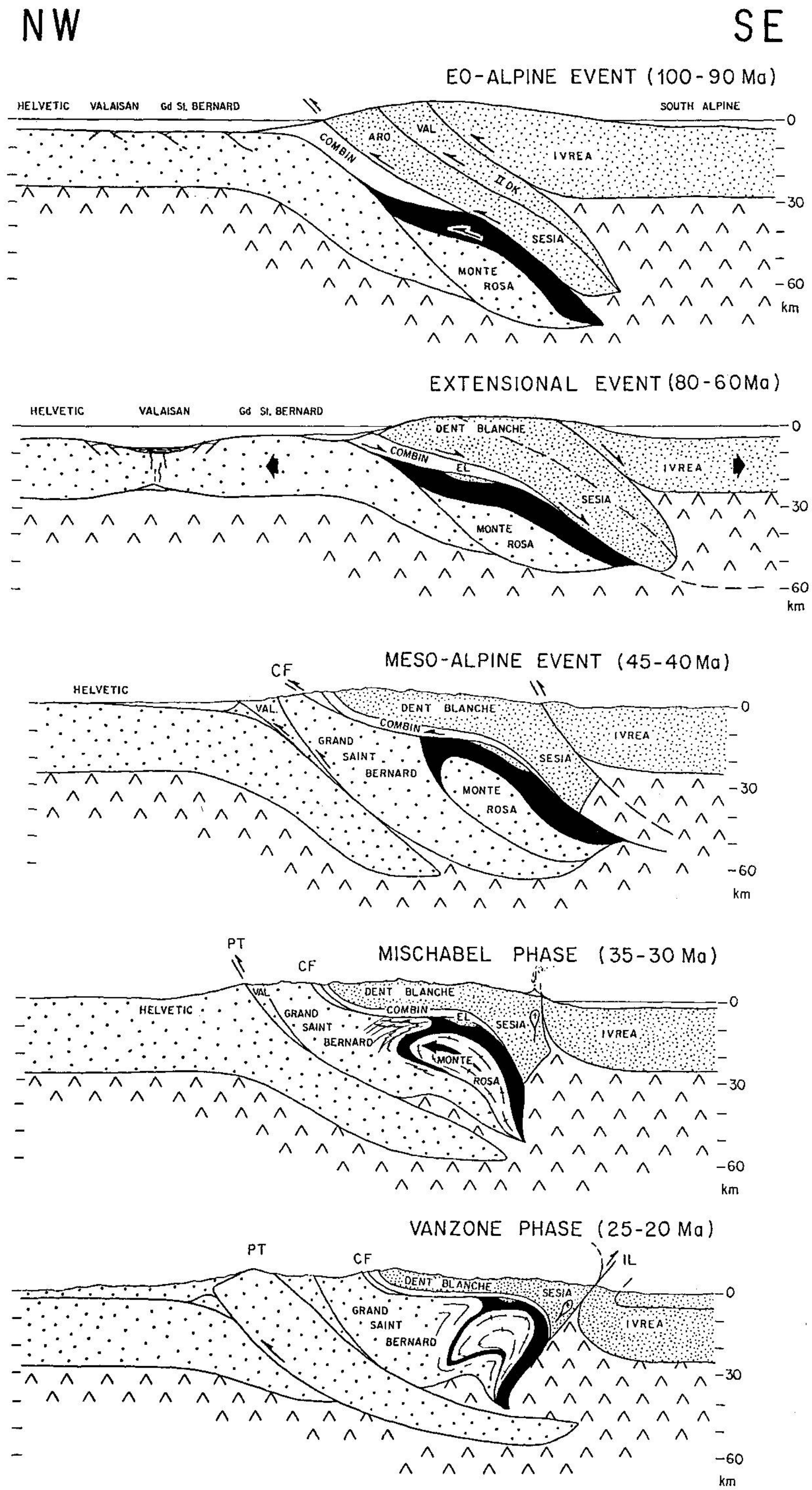

Fig. 11 Tectonic evolution of the northern part of the Western Alps (explanation in text). The contact within Austroalpine units is no longer active after the eo-Alpine event. CF: Combin Fault; IL: Insubric Line; PT: Penninic Thrust.

One of these detachment faults is the Combin Fault. This explains the exhumation process of high-pressure to ultra-HP rocks. As a result of the detachment faults, tectonic denudation occurred and footwall rocks were brought back towards the surface in a way similar to that described for the metamorphic core complexes in Northwestern America (e.g. DAVIS and CONEY, 1979; SPENCER, 1984; LISTER and DAVIS, 1989). In this sense, eo-Alpine eclogite-facies rocks were dragged out from beneath, along the decollement zone, and are now found to be juxtaposed with shallower levels. It must be noted, however, that eclogite-facies clasts from basement units have never been identified in the flysch deposits. This means that eclogite-bearing basement units did not reach the surface at that time but were brought to a level close to the surface only as inferred from the Upper Cretaceous cooling ages in the Sesia zone (HUNZIKER, 1974; OBERHÄNSLI et al., 1985; STÖCKERT et al., 1986).

The extensional event can be ascribed to two different tectonic settings. Firstly, detachment faults can occur during on-going convergence as already proposed by PLATT (1986). In this case, flysch deposits mainly occurred in flexural basins ahead of the frontally propagating thrust. Secondly, detachment faults can be contemporaneous of a plate-divergence stage. In that case, flysch deposition took place in basins created ahead of the major detachment fault.

We favour the second hypothesis for the following reasons :

1. This interpretation is consistent with the basic differences between Upper Cretaceous flysch and Tertiary molasse. The flysch basins were not deformed during sedimentation whereas the molasse basins were (HOMEWOOD and LATELTIN, 1988). This accords well with a plate divergence setting during flysch deposits. On the contrary, the subsidence of molasse basins is related to a flexure of the lithosphere ahead of an area of active tectonics and is thus an indicator of a plate-convergence setting (HOMEWOOD et al., 1986).

2. Plate divergence in the Alpine domain might have accommodated the relative displacement of Iberia with respect to Europe that took place during the Upper Cretaceous (SRIVASTAVA, 1990; MALOD and MAUFFRET, 1990). The opening of the Bay of Biscay induces a sinistral strike-slip movement of Iberia relative to Europe. This sinistral strike-slip movement is likely to induce extension to the north of the Pyrenean transform fault.

Large-scale detachment faults cutting across the entire thickened crust and coeval deep-marine deposits are unlikely to occur during ongoing convergence between both tectonic plates. Such a divergent tectonic setting at the end of the Cretaceous explains why Tertiary tectonic events are clearly separated in time and in character from Cretaceous tectonic events as already pointed out by TRÜMPY (1973 and 1980).

THE MESO-ALPINE EVENT (FIG. 11)

The meso-Alpine event resulted in the formation of two major thrusts i.e. the Combin Thrust and the Penninic Thrust (Fig. 6c). As discussed in detail above, the Combin Thrust reactivates an earlier detachment fault. Forward displacement along the Combin Thrust (about 5–10 km according to geological maps) can be considered as minor with respect to the displacement achieved earlier along the detachment fault. The Penninic Thrust emplaces the "Zone houillère" and the Siviez-Mischabel nappe with its Briançonnais cover over the external domain. It corresponds to the closure of the Valaisan domain. As the Combin Thrust, the age of the Penninic Thrust is also well-constrained by stratigraphic data, especially from the age of the olistostromic formation located on top of stratigraphic sequences in the external domain (Priabonian to early Oligocene) (DEBELMAS and KERCKHOVE, 1988; DEBELMAS et al., 1988).

The meso-Alpine event is probably responsible of the late re-heating of the Gran Paradiso nappe (BORGHI et al., 1992). One should note that the meso-Alpine metamorphism is different from the Lepontine metamorphism which prevailed in the Central Alps, east of the study area, at around 30–25 Ma (e.g. STEIGER, 1964; KÖPPEL and GRÜNENFELDER, 1975; DEUTSCH and STEIGER, 1985; VANCE and O'NIONS, 1992). In the eastern part of the Monte Rosa massif, the latter Lepontine amphibolite-facies metamorphism overprints much of the greenschist-facies metamorphism. According to MONIÉ (1985), the two superimposed metamorphisms can be distinguished from each other using geochronological data in the Monte Rosa area.

Again, one point should be emphasized about the time when eclogite-facies rocks cropped out at the surface in the Alps. The first record of eclogite pebbles in detritic sediments can be dated from the Oligo-Miocene conglomerates in the "Torino hill" (ELTER et al., 1966; POLINO et al., 1991). Sediments in the Torino hill record a progressive exposure of deeper Alpine domains following the meso-Alpine event. Rocks of South-Alpine affinity were first eroded during Lower to Upper Oligocene time, then followed by blueschist-bearing ophiolites in upper Oligocene and Aquitanian-Burdigalian time. Eclogite-bearing ophiolites and Penninic basement rocks become abundant in the Langhian. As already stated above, this strongly suggests that the detachment faulting episode brought eclogites to shallow levels but that they were not cropping out yet at the time when the meso-Alpine event occurred.

THE NEO-ALPINE EVENTS (FIG. 11)

The subsequent deformation of the internal part of the Western Alps is characterized (i) by backfolding and backthrusting at the boundary between the Siviez-Mischabel and more internal units, (ii) by a large-scale folding in the easternmost part of the studied area (Vanzone fold: MILNES et al., 1981), and (iii) backthrusting along the Insubric Line (SCHMID et al., 1987 and 1989). The relative timing of these late, post-nappe deformations can be established using the following criteria.

1. According to MILNES et al. (1981), the Gabbio fold (i.e. the synformal closure of the Monte Rosa nappe) is of the same age as the Mischabel backfold and its axial plane is deformed by the Vanzone fold. Thus, the "Mischabel phase" predates the "Vanzone phase".

2. According to SCHMID et al. (1987 and 1989), backthrusting along the Insubric Line is coeval with the Vanzone folding.

3. The only time marker in that part of the Alps – where there is lack of Tertiary sediments – is provided by the Oligocene magmatism. Structural studies show that the lamprophyric dykes are not deformed (DAL PIAZ et al., 1971; GOSSO et al., 1979; STÜNITZ, 1989), except in the mylonite zone along the Insubric line (SCHMID et al., 1987). Palaeomagnetic data reveal a clockwise rotation of the undeformed dykes (LANZA, 1977, 1979 and 1984; SCHMID et al., 1989). Taking these constraints into account, the tectonic model (Fig. 11) emphasizes the progressive evolution of the structures during continued compression.

The Mischabel phase. This phase can be interpreted in terms of a tectonic process recently proposed for the Central Alps (MERLE and GUILLIER, 1989; SCHMID et al., 1990; DÜRR, 1992; RING and MERLE, 1992). As a result of additional shortening across the suture zone, part of the deepest European basement (i.e. the Monte Rosa nappe) vertically extruded and laterally intruded overlying units (i.e. the Zermatt zone). Then, the Zermatt zone was wrapped around the intruding European basement and part of it is now overlain by the Monte-Rosa massif (i.e. the Antrona zone). Further, the forward motion of the Monte Rosa basement intruded also the Grand Saint-Bernard unit making both units as parts made to telescope. To some extent, this interpretation of the large-scale Mischabel backfold is close to that proposed by Argand many years ago (see dicussion about the "encapuchonnement" process in Argand, 1911, p. 19–20). This deformation took place below the Dent-Blanche unit which can be considered at that time as an orogenic lid (ARGAND, 1934, p. 182; LAUBSCHER, 1983).

Such an interpretation explains the structural position of the Monte Rosa unit with respect to the European palaeo-margin. The present-day location of the Antrona oceanic lithosphere underneath the Monte Rosa is the result of the Mischabel phase. It cannot be argued from a structural point of view that the Monte-Rosa massif was a microplate sandwiched between two oceans during the rifting period (e.g. LAUBSCHER and BERNOULLI, 1982). The Monte-Rosa is part of the European palaeo-margin and the Antrona and Zermatt zones were parts of the same ocean.

The Vanzone phase. Continued shortening across plate boundaries led to the rotation of the Sesia zone and Mischabel backfolds as demonstrated by palaeomagnetic studies in magmatic intrusions of Oligocene age. This rotation and tightening of Mischabel structures is coeval with the formation of the Insubric Line along which much of the recent exhumation took place (SCHMID et al., 1989).

Concluding remarks

The simplest tectonic model to explain the exhumation of HP rocks in the Alps assume that rocks slowly moved upward from the deepest levels to the surface as a result of erosion and isostatic adjustment alone (RUBIE, 1984; GILLET et al., 1986; HSÜ, 1991). On the other hand, more fashionable models deal with the exhumation of HP rocks during continued convergence between both tectonic plates. The first one invokes the flow pattern associated with the "corner-flow model" of COWAN and SILLING (1978) (WINKLER and BERNOULLI, 1986; SCHMID et al., 1987; POLINO et al., 1990). Alternatively, underplating at depth coeval with large-scale extensional faults would explain the exhumation of rocks which were deeply buried during the early stages of the collisional history (e.g. PLATT, 1986; AVIGAD, 1992).

To assess the tectonic setting during which exhumation of HP rocks took place, the following points have to be taken into account: (i) exhumation of HP rocks was permitted by large-scale detachment faults along which at least 30 km of displacement was achieved, (ii) the detachment faulting stage undisputably predates the Tertiary tectonic events, (iii) the detachment faulting stage is coeval with deep-marine deposits in the whole Alpine arc, (iv) tectonic activity cannot be shown to have occurred during flysch sedimentation.

The tectonic model for the Western Alps proposed in this paper is the only one which explains both (i) the structural position of eclogite-bearing Austroalpine units located below the Combin Fault (i.e. the Etirol-Levaz, Emilius and Glacier-Rafray units) and (ii) the distribution of eo-Alpine metamorphism on both sides of the Combin Fault. PLATT (1986, point 4, p. 1038) was the first to stress the great significance of pressure variations along major faults in Phanerozoic orogenic belts. In the case of the Alps, his kinematic interpretation fails as it does not yield a structural explanation of Austroalpine eclogite-bearing units located below the Combin Fault. In our model, emphasis is put on the exhumation of HP to ultra-HP rocks in the time gap separating eo-Alpine and Tertiary events. Salient characteristics during this period of time seem difficult to reconcile with the view of a syn-convergence exhumation of HP rocks in the Alps. They indicate a period of crustal extension which might well support a plate-divergence setting hypothesis (see also DEVILLE, 1993). Crustal thinning and plate-divergence is probably the easy way to explain widespread deep-marine deposits in the Alps during this time lapse. Flysch basins probably enlarged with time as a consequence of crustal thinning accommodated along crustal detachment faults.

Such a period of plate-divergence setting in the Western Alps, unknown until now, remains to be taken into account in global kinematic reconstitutions on the scale of the whole Mediterranean domain.

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