Alpine metamorphism of the Western Alps : I. Middle to high T/P metamorphism

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Alpine metamorphism of the Western Alps: I. Middle to high T/P metamorphism

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Über allen Gipfeln ist Ruh (Goethe)
(On all summits there is peace)

Abstract

After an overview of the structure and lithology of the Western Alps, the Alpine metamorphic imprints of middle to high T/P ratio are examined. These are found in three different rock groups.

(1) In post-Variscan ophiolitic rocks, ocean-floor metamorphism predate all Alpine regional orogenic imprints. The facies of the ocean-floor metamorphism range from granulite to very low grade.

(2) In the more external zones, illite crystallinity shows the distribution of diagenetic, anchizonal and epizonal domains to be broadly parallel to the basement massifs, but with highs around these massifs and duplexes produced by thrusting. There is evidence of several stages, from Late Cretaceous to Miocene.

(3) In the Internal zones, ubiquitous greenschist facies can locally be subdivided into two main imprints, the first of middle pressure and characterized by phengite, the second being characterized by high T/P trend and not associated with a new foliation. The greenschist facies are replaced by amphibolite facies toward the Central Alps. Both greenschist facies are younger than the H-P/T Alpine imprints discussed in the companion paper. The age of the first greenschist imprint is Middle Eocene, the second is probably Oligocene in age. Both are interpreted within the geodynamical frame of the assumed continental magmatic arc of Eocene-early Miocene age.

Keywords: Western Alps, ocean-floor metamorphism, Alpine metamorphism, greenschist facies.

1. Introduction

Subsequent to the still ill-known Permian thermal metamorphism (see DESMONS et al., 1999a, this volume) the Alpine orogenic events have been accompanied by a polyphase metamorphic evolution. Widespread Alpine H-P/T assemblages characterize the Western Alps. They have been succeeded by Barrovian assemblages, themselves followed by L-P/T assemblages of very low to low grade. (In this paper, grade is used as equal to temperature, according to the recommendations of I.U.G.S. Subcommission for the Nomenclature of Metamorphic Rocks, unpubl.)

The relationships of the metamorphic facies with the main thrusting and the main structural boundaries allow a general subdivision of the Alpine metamorphic assemblages. Thus, the

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H-P/T assemblages occur only in some structural zones, some of them being earlier than the thrusting event and others coeval with it, whereas the distributions of the Barrovian and the L-P/T assemblages cross-cut main structural boundaries.

A chronological criterion has been used for 20–25 years to separate in the Western Alps Cretaceous (called eo-Alpine) from Tertiary (called meso- and neo-Alpine) metamorphism, the first being of H-P/T character, and the second of M- to L-P/T character. This has been done in the first edition of the map and its explanatory notes. However, as it will be explained below, the map is now finalized at a time when new radiometric results challenge previous chronological conclusions. This is why when drawing the map a prudent compromise had to be found between the chronological approach generally adopted for the map – an approach successful in the Eastern Alps – and a facies series basis. Because of the conflicting chronological data from the Western Alps the latter is adopted in this and its companion paper.

After a general outline of the structure and the lithology of the Western Alps, and after brief treatments of the ocean-floor metamorphism and the Meso-Cainozoic contact metamorphism, this *first paper* deals with very low- and low-grade, medium- and low-pressure type facies, that is, with incipient metamorphism and with greenschist facies, which are early Cainozoic (Paleogene) in age. Incipient metamorphism is seen in the External zones and the transition to the epizone intersects the boundary between External and Internal zones (the so-called Penninic front or thrust). The very low- to low-grade facies will be the subject of the first chapter. It will be followed by an overview of all greenschist facies in the Internal zones, and this first paper will end with the treatment of the Canavese and the Sestri-Voltaggio zones.

The *companion paper* (DESMONS et al., 1999b, this volume) will deal with H-P/T facies including UH-P, which are everywhere earlier than the incipient and greenschist-facies metamorphism but, in part, are also of lower Cainozoic age. The younger, greenschist, imprints are discussed before the older, H-P/T, ones in order to bring each facies to light, even the lowest-grade facies.

In order to shorten the papers, many references already included in the explanatory notes to the first edition of the map (BOCQUET [DESMONS] et al., 1978) or in comprehensive papers mentioned in the text will not be repeated.

2. General structure and lithologic units of the Western Alps

Readers unfamiliar with Alpine geology and wishing more extensive treatments of the structure and stratigraphy of the Western Alps are referred to the following papers: TRÜMPY (1960, 1980, 1998), DEBELMAS (1982, 1983), DEBELMAS and KERCKHOVE (1980), DEBELMAS et al., 1983, DESMONS (1989), DEBELMAS and DESMONS, 1997, DAL PIAZ (1992), ESCHER et al. (1988), VANOSSI, (1980, 1991), VANOSSI et al. (1984–86), CORTESOGNO et al. (1993), and references cited in these papers, as well as to the local geological maps and their explanatory notes.

2.1. EXTERNAL ZONES

The *Jura mountains* are a small arcuate belt of Mesozoic sedimentary material, folded and partly underthrust in the Miocene, together with Subalpine units, by the European basement.

The *External zones* include: (i) the Dauphinois (or Dauphiné)-Helvetic zone comprising basement massifs bearing an Upper Carboniferous to Mesozoic tegument (i.e. adhering cover), and Mesozoic-Cainozoic cover sequences separated into massifs called Subalpine; and (ii) the Ultradauphinois (or Ultradauphiné)-Ultrahelvetic zone, parautochthonous in the N–S oriented sector of the Western Alps which suffered less from the N-S displacements, allochthonous from the Morcles nappe northeastward.

The protoliths of the basement rocks and their metamorphism have been described by VON RAUMER et al. (1999, this volume). Summarizing, at the turn from the Palaeozoic to the Mesozoic the basement in the future External zones mostly consisted of middle- to low-grade schists and gneisses, of Variscan metamorphic age, with a few traces of a pre-Variscan history, intruded by Variscan and Late Variscan granitoids. Upper Carboniferous siliciclastic rocks include coal-bearing layers. Permian and, S of the Argentera massif in the Dôme de Barrot structure, Permo-Triassic rocks also occur.

The Triassic carbonate, often dolomitic, sequence contains spilites. The thickness of the calcareous and pelitic Jurassic and Cretaceous sequence is highly variable. Upper Cretaceous layers are locally discordant on older layers (due to the so-called pre-Senonian folding phase). The Middle Eocene to Oligocene transgressive sediments become more clastic upwards. The Taveyannaz and Champsaur epiclastic greywackes, of Oligocene age, are especially useful in order to evaluate metamorphic facies and conditions, their volcaniclastic portions being favourable to grow zeolites and other key metamorphic minerals. The appearance of very coarse deposits (blocks and olistoliths) signalled the arrival of the nappes. From the Lower Miocene the clastic sedimentation moved toward the external parts of the belt; the Miocene sequence consists of abundant deltaic and marine molasse deposits.

The post-orogenic sequences of the Upper Eocene to Miocene Piemonte-Liguria basin (not to be confused with the reconstructed Mesozoic oceanic basin of similar name, see below under 2.2.6.) consist of continental breccias and conglomerates with localized lignite deposits, molasse and arenaceous-marly deposits. They were affected by brittle deformation and hydrothermal alteration accompanied by the growth of indicative mineral assemblages.

Bounded towards the External zones by the Penninic front, a composite thrust plane reworked as an extensional feature, the *Internal zones* comprise: (1) Penninic zone, (2) zones called lower Austroalpine, (3) zones of mixed or uncertain origin, now resting on Dauphiné-Helvetic zones and preserved in gaps between, and beyond, the External Crystalline massifs, (4) the Canavese zone and, between the Ligurian Alps and the Apennine, the Sestri-Voltaggio zone.

2.2. PENNINIC ZONE

The *Penninic* zone comprises six zones. Their paleogeographic location is considered by some of the present authors as being in the margin of the future Southern Alps (Apulian-Dinaric block, Gondwanian plate). However, for others they represent the paleo-European margin.

2.2.1. Subbriançon zone

The *Subbriançon zone* (or Subbriançonnais) consists of Mesozoic-Lower Cainozoic sedimentary deposits without basement. Northward it gives place to the Valais zone, which increases in extent in Switzerland. In the Ligurian Alps it is poorly represented, possibly in part covered by the Flysch nappes.

2.2.2. Valais zone

The *Valais zone* consists of Upper Carboniferous to Upper Mesozoic sedimentary rocks including (in the Versoyen zone near the Ptt St-Bernard pass) Liassic marls, now transformed into calcschists, underlying [Middle Cretaceous to] Turonian pelites with sills and minor lava flows. In the Versoyen, French Valais zone, they are associated with a small (600 m thick) plutonic basic body of debated tectonic origin, the Clapey occurrence, and with rare slices (the Pte Rousse felsic gneiss, and serpentinite bodies). To the ENE, in the Central Alps, it gives place to a more external Valais zone overlying the Infrapenninic basement units of the Lepontine area

2.2.3. Briançon – Gd St-Bernard

The *Briançon* (Briançonnais)-*Gd St-Bernard* is a supernappe, in that it itself consists of several independent nappes. In Switzerland the Gd St-Bernard supernappe is composed of the following independent tectonic units, from bottom to top: the Houillère zone, the Pontis nappe, the Siviez-Mischabel nappe, and the Mont Fort nappe. The Houillère zone may be followed for more than 200 km to the Briançon area in France, and it reappears with a large extent in the Ligurian Alps. The Pontis nappe is a narrow, discontinuous zone in the frontal part of the Gd St-Bernard supernappe and consists of a polymetamorphic pre-Westphalian basement (from the western Ruitor zone to the eastern upper Stalden zone) and a monometamorphic Carboniferous to Triassic metasedimentary cover. The Siviez-Mischabel nappe forms a huge recumbent anticline with a core of pre-Westphalian basement and a metasedimentary cover of Upper Carboniferous to Lower Triassic age in the west and of Triassic to Eocene age in the east (Barrhorn series). This nappe is well represented in the eastern part of the Gd St-Bernard supernappe but is considerably reduced towards the west. The Mont Fort fold-nappe is made up of a basement core (Métaillier series) overlain by Mesozoic metasediments. Metabasites are the main rock type of the Métaillier series.

From the Aosta valley to the upper Stura di Demonte valley, that is, in the N–S trending part of the Alpine belt, the basement of the Briançon supernappe is dissected into massifs, a result of E–W cross-cutting structural directions. From N to S these are the Ruitor massif, the Pourri-Bellecôte (or Northern Vanoise) massif and its NE prolongation the Zona Interna (both equivalent to the above-mentioned Mont Fort nappe), the Chasseforêt (or Southern Vanoise) massif, the Sapey zone bordering both Vanoise massifs, the Ambin massif and, in the Acceglio zone, the Sagnères and Pelvo d'Elva slices. Various cover sequences more or less internal in origin are distributed into thrust sheets. One of these sheets is characterized by so-called "prépiémontais" (Pre-Piemonte) sedimentary facies, i.e., thick monotonous Liassic siliceous marly deposits which somehow recall the sedimentary sequence of the Combin zone (see below under 2.2.6.). The tegument on the Briançon basement mainly consists of Permian and Permian-Lower Triassic rocks of the Pourri-Bellecôte massif and its prolongations.

From the Middle-Upper Triassic up to lowermost Eocene, the cover rocks, mostly deposited in marine environments, are carbonate or silico-carbonate rocks in their larger majority. The dolostones have been less sensitive to metamorphism than calcareous rocks. Besides, Al-rich alteration rocks such as bauxites have been the sites of metamorphic transformations which would otherwise have not been suspected.

In the cover of the Penninic, Briançon and Internal Penninic massifs, the Triassic deposits are of Alpine type, that is, they were deposited in warm, tropical, waters, contrasting with the External zones and also with Sardinia, which are closer to the Germanic Triassic type. This indicates that both areas at the time were far from each other, i.e. separated by an already existent Tethys ocean.

The *Ligurian Briançonnais domain* is a stack of internal over external units. The tectonically highest and innermost units are formed by a pre-Namurian polymetamorphic basement, the Savona and Bagnaschino massifs, which up to Triassic times were possibly adjacent to the Calizzano basement which is now rather considered as a "Piemonte" basement (equivalent, in the Ligurian Alps, to the Internal Penninic massifs of the north and central Western Alps, see below). There is recent evidence of primary stratigraphic relationships between the Calizzano basement and cover rocks, the C. Tuberto unit. Remnants of the Permian-Carboniferous tegument are locally preserved. The basement in the underlying units mainly consists of Variscan metarhyolites, and only some of the lowest units, such as the Mallare unit, comprise basement rocks. Upper Palaeozoic volcanic and sedimentary sequences, sometimes very thick, characterize many of the units, e.g. the Ormea Unit. The abundant, mostly acidic, volcanic rocks include andesites which easily registered metamorphic transformations. The Mesozoic to Eocene cover is everywhere marked off by a stratigraphic gap, the extent of which decreases from the internal to the external sectors, i.e. from the higher, Castelvecchio-Cerisola, to the lower units, e.g. M. Carmo, but it includes at least the Lower Jurassic.

The Briançon rocks originate in the South-Alpine block (Apulian-Dinaric plate), not in the European plate nor the Iberian-Corsican-Sardinian plate, which was a third plate interplaying in the Alpine and Apennine orogens, but which arrived on the stage later than Briançon emplacement. However, the Ligurian Briançon basement has much in common with the Sardinian basement as concerns the metamorphic evolution, Variscan magmatism and Permo-Carboniferous history, thus suggesting a possible closeness in pre-Mesozoic times of the Ligurian Briançon units and a history possibly different from the northern and central parts of the Briançon zone.

2.2.4. Internal Penninic massifs

The *Internal Penninic massifs* (Monte Rosa, Gran Paradiso + the small Brusson, and Dora-Maira massifs) are composite basement nappes bearing thin sporadic Lower Mesozoic tegument rocks. The Dora-Maira bears a Triassic-Cretaceous cover resembling Briançon cover and Combin ophiolite-free sheets with which it is infolded.

In the Ligurian Alps, the *Torrente Visone* and *Calizzano units* are analogous slabs of pre-Namurian crystalline basement, which are here ascribed to a "Piemonte" domain, sometimes less properly called "Pre-Piemonte", here considered to correspond to the paleo-continental margin and to its transition to the oceanic basin.

The protoliths and the metamorphism of the *Internal basement rocks* have been described in another paper (DESMONS et al., 1999a, this volume). As in the External zones, the basement rocks of the Internal zones had been exhumed before Alpine times to near-surface level. One type of the future basement – the ancient basement – consisted of mostly amphibolite facies gneisses and schists, of pre-Variscan (Pan-African?) metamorphic age (according to some authors, reworked during Variscan times under high- to low-grade conditions), intruded by Ordovician and/or Variscan granitoids. Late Variscan magmatic bodies are present in the Siviez-Mischabel nappe and in the Ligurian Alps. A second type of basement, the younger basement, consists of clastic rocks, unmetamorphosed or metamorphosed under very low-grade conditions, intruded by, and interlayered with, mafic and felsic magmatic rocks. These rocks probably are lower Palaeozoic in age.

Within the *cover rock sequences* lying upon the basement massifs in the Internal zones there is no Upper Carboniferous. Kilometre-thick Upper Carboniferous-Lower Permian clastic rocks, deltaic, coal-bearing, collectively shown as molasse on the map, form the Houillère zone. Here, lava flows and dykes are now useful as having preserved Alpine metamorphic assemblages otherwise absent from the metasedimentary rocks. As stated in the paper on the pre-Alpine metamorphism of the Internal zones (DESMONS et al., 1999a, this volume), the undated rocks of the Pinerolo unit in the Dora-Maira massif, if they are not Upper Carboniferous as generally thought until now, might belong to the younger type of basement (Lower Palaeozoic in protolith age), in spite of reversed tectonic positions with respect to the basement of the Internal Penninic Massif.

Most acidic plutonic rocks of the Dora-Maira massif, now orthogneiss, metagranitoid, metadiorite and dioritic gneiss, are likely derivatives of Permian magmatic rocks. A small amount of the augengneisses, called "gneiss amygdalaires", which contain polycrystalline augen, are derived from older igneous protoliths, since they are cut by Permian granitoids.

2.2.5. Zermatt, Voltri and Lonza zones

The ophiolitic *Zermatt-Voltri zone* is a composite nappe thrust upon the Internal Penninic massifs and farther west, cropping out in the Zermatt-Saas-Val d'Aosta area, the periphery of the Gran Paradiso, west of the Dora-Maira massif (the Rocciavrè and Viso massifs) and in the Ligurian Alps (the Voltri massif). It consists of several thrust sheets. Ultramafic rocks volumetrically predominate over gabbros and basalts. The ultramafic part of the sequence was exposed on the oceanic floor of the basin before the basaltic volcanism took place. In the Zermatt area there is radiometric evidence of a Jurassic age of the gabbros, which possibly are much younger than the ultramafic rocks.

In the *Voltri zone*, ocean crust slices are piled up together with subcontinental mantle-derived ultramafites and minor slices of platform carbonate rocks. The different tectonic units are likely folded together within a complex recumbent ultrafold verging towards NW, and re-folded during four main phases. They are distinguished on the basis of primary lithological associations, lithospheric belonging and metamorphic overprint. The widest Beigua-Ponzema unit consists of predominant ultramafites intruded by gabbro and Fe–Ti oxide gabbro; in the W and NW sector the primary relationships are in part preserved between the ultramafic basement, represented by ophicarbonate rocks (= serpentinite carbonate breccias) and the volcaniclastic cover consisting of pelitic sedimentary rocks associated with basalts. In most cases, however, the volcaniclastic cover is allochthonous on the ophiolitic sequence.

The *Lanzo massif* is a large body of mantle spinel-peridotite, partly re-equilibrated in the plagioclase-peridotite field, surrounded by serpentinized peridotite and sheared antigorite serpentinite. The geothermometers give two T-clusters around ca. 1300 and 900 °C, most likely corresponding to re-equilibration in the spinel and in the plagioclase stability fields respectively. Peridotites preserved in the massif core contain gabbros, locally showing H-T, most likely oceanic, shearing and recrystallization, cut by undeformed basaltic dykes. For many authors the Lanzo ultramafic rocks likely derive from the ocean-floor of the Zermatt basin, but they were juxtaposed to the continental gneissic Sesia zone before, or during the early stages of, the Alpine eclogite facies metamorphism.

2.2.6. Combin zone and Cravasco-Voltaggio-Montenotte zones

In the Aosta valley, the *Combin zone* has since decades been defined and differentiated from the tectonically underlying Zermatt zone. Around the Dent Blanche, part of it received from Swiss geologists the name of Tsaté zone. The Montgenèvre-Cesana-Chisone area and the Queyras region are assimilated to Combin on the basis of structural position and metamorphic evolution. The Combin zone actually is a stack of thrust sheets. It consists of Upper Triassic to Upper Cretaceous carbonate rocks, mainly calc-schists (Schistes lustrés, Calcescisti, Bündnerschiefer in the three Alpine languages) interlayered with ophiolitoid rocks: metamorphosed lava flows, sills, tuffs, collectively called prasinites, mainly Cretaceous in age. In the Queyras region SW of Briançon, the lower thrust sheet contains typical ophiolitic rocks (alternatively interpreted as olistoliths, remnants of paleohighs of an oceanic floor, tectonic slices, nappe remnants), dated as Upper Jurassic by radiolarian faunas preserved in overlying cherts. Marly rocks from the "Schistes lustrés" easily recorded greenschist-facies or anchimetamorphic transformations, and pelitic layers thereof are also liable to show subtle metamorphic differences. The Combin zone overlies most of the other Penninic zones: the Internal Penninic massifs, the Zermatt zone, and the Briançon zone including the Houillère zone.

Because of lithological similarities of ophiolites and calc-schists in both zones, and because the Combin zone is directly overlying the Zermatt zone, both zones have for a long time been considered as forming only one zone, deriving from a single basin called Piemontese-Ligurian or Liguro-Piemontese, respectively from its margin and from its centre. This cannot be, since the Combin zone starts with Triassic deposits and since the metamorphic evolution of both zones has been different (without and with eclogite); Zermatt and Combin have to be considered as two separate basins.

In the Ligurian Alps, all ophiolite-bearing units are, however, so far considered as deriving from such an assumed single Piemontese-Ligurian basin. These units are, from east to west, the *Monte Figogna unit* (in the Sestri-Voltaggio zone, see below) associated with the Busalla Flysch, the Voltri Group described above, the *Cravasco-Voltaggio* (also in the Sestri-Voltaggio zone) and *Montenotte* units. Upper Cretaceous to Eocene flysch units occurring in the western Ligurian Alps, devoid of their original basement, are considered as also derived from the same domain. The Montenotte unit is represented by an ophiolite sequence where serpentinite, gabbro and Fe–Ti oxide gabbro, commonly cut by basaltic dykes, are covered by a sedimentary sequence of ophiolitic breccia, thin banded chert, siliceous limestone and pelite. The Montenotte unit tectonically overlies the Triassic-Liassic carbonate units, the Savona basement, the Voltri meta-ophiolitic unit, and locally the Oligocene basal conglomerates. The thrust surfaces in part developed during the Apennine-verging phase. The *Montaldo unit*, a marly Cretaceous sequence containing ophiolitic olistoliths, was possibly the closest to the ocean and might represent a lateral equivalent of the basal thrust sheet(s) of the Combin zone. Among the other units, the *Arnasco-Castelbianco unit* deserves a special mention because of the huge volumes of Liassic megabreccias that distinguish its stratigraphic sequence. The Triassic to Liassic unit in the Sestri-Voltaggio zone (see below), the quartzites and dolostones associated with the calcschists in the Voltri-Rossiglione unit, as well as other Triassic-Liassic units to the west, likely also originated from the "Piemonte" domain. The same origin is also possible for the Arenzano basement and its Triassic to Liassic cover.

Associated with the ultramafic and mafic rocks (gabbro, and pillowed or brecciated basalt) are uncommon felsic rocks (Queyras, Montenotte-Cravasco). The initial almost anhydrous composition of the basic rocks contrasts highly with the sedimentary rocks. Hence, they reacted differently to the same set of metamorphic conditions and the resulting different assemblages must not be interpreted as indicating different metamorphic conditions.

2.3. LOWER AUSTROALPINE

The Sesia zone, the Dent Blanche nappe and the many outliers (klippen) in the Val d'Aosta between Dent Blanche and Sesia form three complexes commonly called *Lower Austroalpine*. This means that they rest on top of Penninic nappes, but it does not imply that their emplacement was geodynamically related to the Austroalpine nappes of the Eastern Alps. As can be deduced from their superposition order and their individual tectono-metamorphic evolution, the emplacement of these bodies occurred in more than one event (Sesia + klippen; Dent Blanche + Second Diorito-Kinzigitic zone; and earlier superposition within each). The internal part of the *Sesia zone*, the Eclogitic Micaschist complex, is a folded polymetamorphic basement zone. The contact between the Eclogitic Micaschist complex and the more external Gneiss Minuti complex is marked by discontinuous and narrow sheets (from a few metres to a few kilometres) of the *Second Diorito-Kinzigitic zone* infolded at kilometric scale (named Vasario subunit), which towards the northeastern sector between Val d'Aosta and Val Sesia forms a large nappe overlying the two complexes. The gneiss of the Gneiss Minuti complex, the Arolla series and some Austroalpine outliers, are considered to have derived from Permian magmatic rocks.

To the South, the Sesia zone also includes a subunit, defined as "Rocca Canavese thrust sheets", interposed between the Sesia zone and the Southern Alpine Canavese zone, close to the Canavese Line. This composite subunit consists of pre-Alpine granulite, mostly serpentinized lherzolitic tectonite, basic schist, and leucocratic gneiss probably derived from late Paleozoic granitoids. The different thrust sheets are separated by H-P/T metamorphic blastomylonites of Alpine age.

A possible cover of the Sesia Eclogitic Micaschist complex has recently been found, which includes a thin carbonate sequence considered as Triassic-Jurassic in age, and quartzites bearing Mn-nodules. The basic rocks, first considered as ophiolitic rocks, have recently been dated as Early Carboniferous in age.

The *Dent Blanche nappe* forms a great klippe of predominantly basement rocks, measuring about 50 by 20 km, overlying Combin rocks and Zermatt rocks (respectively N and S of the Aosta valley). The basement is subdivided into the Arolla series (mainly granitoids of presumed Late Variscan age, paragneiss including Verrucano-type conglomerate, and minor Permian gabbro) and the overlying Valpelline series (pre-Alpine high-grade amphibolite-facies rocks with well preserved granulite-facies relics). A thin Mesozoic cover comprises Triassic to Jurassic sedimentary rocks.

Among the *outliers*, the Emilius, Glacier-Rafrey, Pillonet, Tour Ponton, Eaux Rousses, Etirol-Levaz, Santanel, Grun, Châtillon, and Perrière klippen are related to the Eclogitic Micaschist complex, whereas the Mont Mary, i.e. the Dent Blanche s.str., is related to the Gneiss Minuti complex.

In the Ligurian Alps, there is no unit equivalent to the Sesia and Dent Blanche zones.

The Late Alpine igneous rocks, which include the intrusive stocks of Biella and Traversella, the lamprophyric and andesitic dykes, and the andesites of the Sesia zone, have not been affected by metamorphism, but the plutons induced contact metamorphic effects.

2.4. ZONES OF UNCLEAR ORIGIN

The *Helminthoid Flysch nappes* (Embrunais-Ubaye and, in the Maritime Alps, the Sanremo, Moglio-Testico and Borghetto nappes) consist of two thrust sheets of Upper Cretaceous to Paleocene flysch deposits, with, at the base, Subbriançon and minor Briançon olistoliths from kilometre to metre in size, associated with a minor ophiolite-bearing sheet. They overlie the External zones in the gap between Ecrins-Pelvoux and Argentera massifs and south of the latter.

South and SE of the Leman Lake, also resting on the Dauphiné-Helvetic zone, here in the gap between the Mt Blanc-Aiguilles Rouges and Aar-Gotthard massifs, the *Prealps* tectonically rest on Dauphiné-Helvetic zone. They consist of various Penninic-type cover material, distributed in thrust sheets topped by flysch sheets bearing ophiolitic and basement olistoliths (Simme and Gets; see FREY and FERREIRO MÄHLMANN, this volume).

2.5. CANAVESE AND SESTRI-VOLTAGGIO ZONES

The *Canavese zone* consists of a crystalline basement and a cover. The basement includes low-grade amphibolite-facies para- and orthogneiss, intruded by large bodies of granite to diorite with very minor gabbro. The cover consists of Permian volcanic rocks (rhyolites) and volcaniclastics, and a Mesozoic mainly carbonate sequence.

The *Sestri-Voltaggio zone* is a thin, N–S elongated, belt which is exposed just east of the Voltri meta-ophiolites and consists of three main parallel units with subvertical dip. The two westernmost units, the Triassic-Liassic unit of M. Gazzo-Isoverde and the Cravasco-Voltaggio unit are folded together by early deformation phases and are likely attributed to the "Piemonte" domain. The Cravasco-Voltaggio unit also contains ophiolite sequences analogous to those of the Montenotte unit in their primary and metamorphic features. The third unit, the M. Figogna unit, which dips to the east below Helminthoid flysch, is formed by an ophiolitic sequence including serpentinite, basalt and a thin cover of chert, siliceous limestone and pelite, which likely grades to a marly and arenaceous flysch, equivalent to the Val Lavagna Flysch of the Northern Apennine.

3. Ocean-floor and related metamorphism

Ocean-floor metamorphism affected the ophiolite sequences present in the Zermatt, Combin, Valais, Voltri and other ophiolite-bearing zones in the Ligurian Alps. The basic rocks interlayered with the sedimentary rocks also have undergone transformations such as hydration before the Alpine regional metamorphic phases. Early alteration also affected the Lanzo peridotites. For graphical reasons the ocean-floor and related metamorphism has not been represented on the map.

From stratigraphic-paleontological data on the accompanying metasediments and from radiometric data on ophiolitic gabbros (see section 6.2) we know that the age of this metamorphism ranges from at least Late Jurassic (?) in the Zermatt zone and lowermost Late Jurassic in the Combin zone, to Cretaceous for the prasinites in the Combin and the Valais zones. High-T oceanic metamorphism along shear zones in gabbro and peridotite appears to be older than the typically low-T ocean-floor metamorphism of basaltic lavas.

Peridotites are fresh only in the undeformed core of the Lanzo massif, its satellite bodies of Balangero, Moncuni, Monte S. Giorgio and Piossasco, and very locally in the Zermatt zone at Locana in the Orco valley, Mont Roz and Larsinaz near Cogne, and Mont Avic in the upper Champorcher valley, Aosta. The serpentine minerals developed during the hydrothermal oceanic alteration of peridotite were mainly lizardite and chrysotile. Only microstructural relationships can differentiate antigorite and lizardite formed during the same process from those deriving from Alpine phases of orogenic metamorphism. In the Lanzo massif, pre-Alpine ocean-floor lizardite veins cut by Alpine antigorite veins have been described by BOUDIER (1971). In the Voltri unit, chrysotile, antigorite and lizardite developed during polyphase ocean-floor metamorphism and later re-equilibration phases (CORTESOGNO et al., 1994).

In spite of the ocean-floor metamorphism, the magmatic plagioclase has been preserved in some large gabbro bodies such as the Allalin gabbro near Zermatt and the Rocciavrè gabbro in the Susa valley, as well as in gabbro and peridotite of the Lanzo massif. Elsewhere it recrystallized to plagioclase of lesser anorthite content according to the metamorphic grade. In gabbro and basalt, the igneous clinopyroxene was locally replaced by brown hornblende, green hornblende or chlorite; in peridotite the orthopyroxene was replaced by "bastite", i.e. a one to one serpentine pseudomorph, which preserves the original pyroxene shape and exhibits a single optical extinction under crossed polarizers. Pre-Alpine assemblages are also well preserved in metabasites of the Erro-Tobbio unit (SCAMBELLURI et al., 1991).

The ocean-floor metamorphism is often connected to shear deformation and tectonic melange (e.g. MEVEL et al., 1978). Gabbro was transformed into flasergabbro, commonly predating the diabase intrusions. The basalts were spilitic before the Alpine regional orogenic metamorphisms (e.g. BEARTH and STERN, 1971, 1979) and hydrated pillow rims have often been described (e.g. MEVEL, 1975; OBERHÄNSLI, 1982).

The ocean-floor metamorphism appears to have covered the whole facies range. Part of the higher-grade transformation may actually have been late-magmatic. Where the ocean-floor transformation is still clearly visible under the Alpine regional metamorphism, as in many portions of the Ligurian Alps, three different ocean-floor imprints can be distinguished. (i) In gabbro and ultramafite, high- to medium-grade transformation, under temperature ranging from the subsolidus to the amphibolite-greenschist transition, has been associated with extensional and subsequent viscous-plastic deformation. The inferred temperature decreases from the lower to the upper units, and with time. These transformations are especially well recognizable in the Montenotte and Cravasco-Voltaggio units, less so in the Voltri units. (ii) The basaltic members and part of the associated sediments, mainly breccias, have been affected under mostly greenschist-facies conditions. (iii) An essentially hydrothermal alteration can be interpreted as an off-axis process.

Augen (flaser) and banded textures in gabbro and Fe–Ti oxide gabbro cut by basaltic dykes, are found in the Montenotte-Cravasco-Voltaggio meta-ophiolites where relics of ocean-floor metamorphic clinopyroxene, Ti-pargasitic, tschermakitic and Mg-hornblende are common. Also in the Voltri meta-ophiolites, and in particular in the Palmaro-Caffarella unit, H-T ocean-floor textures are preserved.

In the Combin and Valais zones and the Lanzo massif the following has been found. Kaersutite brown amphibole, and green Mg-hornblende and Fe-pargasite overgrown by barroisite are found in Montgenèvre-Chenaillet (MEVEL et al., 1978; BERTRAND et al., 1984), in the Combin zone below the Dent Blanche nappe (AYRTON et al., 1982), in the Clapey body and in the thickest sills of the Valais-Versoyen zone (LASSERRE and LAVERNE, 1976). Mg-taramite has been described from Queyras (DESMONS, 1990) and brown hornblende from the Lanzo massif (COMPAGNONI et al., 1982, 1984). MEVEL et al. (1978) mention values of 700–800 °C and ~1.2 kbar for the amphibolite-facies metamorphism.

In the prasinites of the Combin zone some green amphibole, albite, chlorite and titanite may have existed already before the Alpine regional metamorphic phases. Some ovardites from the type locality (Torre Ovarda in Val di Viù) may have been derived from palagonitized basalts (LEARDI et al., 1986).

Metasomatism occurred at the contact between ophiolitic and sedimentary rocks. Essentially Na, K and Si ± Mg ± Fe have been leached out (e.g. BARNICOAT and BOWTELL, 1995). In the Arc valley, in an iron-ore occurrence at the contact between prasinites and calc-schists, skarn rocks contain an assemblage of diopside, garnet ($Adr_{\sim 95}$) and actinolite (DESMONS, 1990).

Ophispherites, i.e. rounded fragments of gabbro included in ultramafic rock and metasomatized during the serpentinization, have been described from the Chenaillet massif (VUAGNAT and PUSZTASZERI, 1964; BERTRAND et al., 1980).

Rodingites are known from all ophiolite-bearing zones and from the Lanzo massif. They consist of the usual range of Ca-minerals: Ca-rich garnet with grossular-hydrogrossular-andradite solid solutions, vesuvianite, epidote group minerals, prehnite, pumpellyite, etc. Some rodingites predate the Alpine metamorphisms and, thus, resulted from ocean-floor transformations, but they were later re-equilibrated during the Alpine phases.

The ocean-floor transformations contributed to hydrate the ocean-floor rocks and made them reactive to the Alpine orogenic metamorphism. The stable isotope ratios acquired were later retained, at least during the H-P metamorphism (BARNICOAT and CARTWRIGHT, 1995).

4. Contact metamorphism of Alpine age

Contact metamorphism of Alpine age occurs around the so-called peri-Adriatic magmatic stocks. The outer limit of the contact metamorphism around the Oligocene Biella (also called Valle del Cervo) and Traversella plutonic bodies intruded into the Sesia zone is marked by the first appearance of biotite. The intrusions were shallow, as suggested by the presence of sharp contacts and the occurrence of andalusite. In the Biella body, andalusite is present only on the internal side of the contact metamorphic aureole facing the Canavese Line, in agreement with the original location of the magmatic body which has been significantly modified by the post-Oligocene tilting of the internal part of the Sesia zone (e.g. LANZA, 1984).

In black schists around Cretaceous sills of the Versoyen zone, pseudomorphs of andalusite or cordierite have been ascribed to a thermal effect. Contact metasomatic rocks such as adinoles (pelites which have been albitized at the contact with Na-rich mafic rocks) have also been described from the Valais zone (LOUBAT and DELALOYE, 1984, and ref. therein; SCHÜRCH, 1987).

5. Anchizone and transition to the epizone in the western regions

The illite crystallinity method (IC) has been used in order to differentiate the anchizone from diagenesis and epizone domains defined according to KÜBLER (1964). The anchizone as determined by IC corresponds to the middle-pressure – very-low grade domains shown on the map, and the epizonal domains as determined by IC have been included in the greenschist areas (for more details on the question, see FREY, 1987). The diagnosis may be corroborated by additional criteria: the presence of newly-formed minerals such as pyrophyllite, paragonite or chloritoid, the coalification grade, the homogenization temperature in fluid inclusions, the presence of pyrrhotite and the isotopic reworking.

5.1. DISTRIBUTION OF THE METAMORPHIC GRADE

The general trend of the metamorphic grade, increasing from W to E, from diagenesis through anchizone to epizone, is perturbed by tectonic duplexes and the proximity of the basement. Moreover, there is a S–N increase as shown by the epizone-greenschist grade which in the S is not reached before the internal slope of the Mercantour massif, whereas in the N it already appears west of the Mont Blanc-Aiguilles Rouges massifs. A zeolite facies, more precisely a Lmt-Pmp-Prh subfacies, essentially with rock-forming Lmt, and Pmp-Prh in veins, is found in Taveyannaz-type greywackes: in the Champsaur area (S of the Pelvoux massif), the Thones syncline (between northern Belledonne and Annecy) and the Platé massif (W of the Mt Blanc massif and N of the Arve valley). As the Priabonian schists and limestones associated with the greywackes exhibit diagenetic IC values, the Lmt, Pmp and Prh occurrences shown on the map plot inside the diagenetic area.

Examples of higher-grade bodies thrust on lower-grade ones are found in the areas of the Digne thrust, the Helminthoid Flysch nappes, and Ultradauphiné zone (between Arc and Isère valleys, E of Belledone) (see also the Ultrahelvetic klippen, in FREY and FERREIRO MÄHLMANN, 1999, this volume). Other disturbances of the general pattern, even if more discreet, are (see details in APRAHAMIAN, 1988) :

– The offset of the anchizone by the Beaufin fault (near the northern termination of the Digne thrust against the Pelvoux massif).

– The offset of the Pelvoux basement by the Valgaudemar fault cutting the anchizone (S of the Beaufin feature).

– In the Champoléon area, S of the Pelvoux, both overturned basement + anchizonal Liassic strata, overlapped by Priabonian layers showing zeolite facies and diagenetic IC.

– An isolated anchizone outlier in the Platé massif, resulting from contact heating by the Helvetic nappes.

Close to, and around, the External Crystalline massifs, especially Belledonne, the metamorphic zones display a more or less symmetrical arrangement. West of the massif, a thin anchizone rim is observed, but the greenschist facies locally extends beyond the western border of the massif, especially in its southern inlier, the La Mure massif. Crossing to the eastern slope of the Belledonne massif and the satellite bodies of Rocheray and Gdes Rousses, one finds epizone, then again anchizone and, in a limited area, diagenesis only under the Ultradauphiné nappe. Beyond the Penninic thrust eastward the grade increase resumes toward greenschist. The lay-out around the Belledonne and La Mure massifs might be interpreted as an "effet de socle" (basement effect, due to higher heat conductivity in crystalline rocks, FONTEILLES and GUITARD, 1964), but the uprise of the External Crystalline massifs may also have played a part in this structure.

North of the Pelvoux massif a similar pattern is found, whereas south and southwest only the anchizone extends beyond the basement outcrops, and this discontinuously. The limit between diagenesis and anchizone runs under the Helminthoid Flysch nappe and its attitude suggests its continuation within the basement. Lack of IC data owing to absent suitable rock type precludes a diagenetic field from being demonstrated in that place. On the map the Tertiary greenschist facies shown for the whole Pelvoux massif thus results from rough extrapolation and lack of precise facies delimitation. Eastward, the greenschist facies is again found in the tiny basement windows of Haute Salce and Dormillouse, and their aureoles.

The widest greenschist-facies field is seen along the La Mure-S Belledonne-N Pelvoux transect. This results from the tectonic assembly of these huge thrust basement bodies. Likewise, just N of the Pelvoux and E of the Gdes Rousses massif, the wide extent of the anchizone is due to the duplex formed by the Ultradauphiné thrust (and also to a particular structural position in a kind of "gulf" sheltered from crushing).

5.2. CHRONOLOGICAL DATA

5.2.1. Geological and cartographical data

The relationships between metamorphism and tectonics, such as in the case of transported metamorphism, are chronologically interesting in so far as the tectonic event is dated by the youngest involved formations and/or the oldest overlying formations. Examples are the post-Priabonian Helminthoid Flysch thrust and the Ultradauphiné thrust, which transported Paleogene meta-sedimentary rocks onto layers of similar Nummulitic age, but here the nappe overload was essential to obtain the necessary pressure and temperature. This case is evidence of at least two metamorphic episodes, both post-Priabonian in age, one pre- and one post-thrusting.

The Beaufin fault, which offsets the Belledonne-La Mure higher-grade aureoles, is considered to be an old, repeatedly activated, feature.

The Valgaudemar offset is possibly post-tectonic. Geographical proximity and geometrical similarity point to the same age and interpretation as in the Champoléon case. Here, the disappearance of the anchimetamorphic aureole may be ascribed to basement gliding while overturned, during its uprising. In the Champoléon case, both the upright setting of the basement rocks and the anchizone imprint itself (also overturned) are certainly pre-Priabonian since layers of that age unconformably rest on both basement and cover, and probably pre-Senonian. In the Priabonian sandstones, a zeolite-facies imprint (Lmt) constitutes a second, very-very low-grade metamorphic imprint (diagenetic IC values), thus younger than Priabonian. This post-Priabonian imprint is likely to have resulted from the thrusting of the now near-by Flysch nappe. As evidenced by other Priabonian outcrops it increases in grade eastward up to anchizone and even greenschist, the latter around the tiny basement windows of Dormillouse and Haute Salce.

In conclusion, all over the area considered at least three very low- to low-grade metamorphic imprints have been identified. Two of them are younger than Priabonian, one is older than Priabonian and possibly even older than Senonian. It is as yet unknown whether this possibly pre-Senonian imprint was strictly limited to the southern Pelvoux, or was more widespread, or if it was replaced northward by a fourth, pre-Priabonian but still Cainozoic, metamorphic episode.

5.2.2. Radiometric data

Numerous radiometric data concerning both basement and cover have been obtained through K–Ar and Rb–Sr measurements on bulk rocks, mineral separates and clay fractions, and by fission-track measurements. Moreover, thermopaleomagnetic data have recently been tentatively applied to geochronology (CROUZET, 1997; CROUZET et al., 1997): the temperature of magnetic inversions is very precise, but the correlation with the time-scale remains a delicate matter. However, the significance of most of these data is debatable, and practically none of them dates a metamorphic peak.

In cover rocks, apparent K–Ar ages are younging from diagenesis to the anchizone fields, and are changing from older to younger than the stratigraphic age (BONHOMME et al., 1980, and pers. comm. to J.A.). In the basement, apparent Alpine ages mostly ranging from 45 to 35 Ma, but possibly with older and younger episodes (DEMEULEMEESTER, 1982; DEMEULEMEESTER et al., 1986), are associated with shear zones which are possibly local and very late features. Moreover, K–Ar dating generally is vitiated by excess Ar. Ages of 26 ± 2 Ma (Late Oligocene) yielded by upper Liassic slate samples from E Belledonne seem to be reliable (NZIENGHI, 1993). Imprints older than Priabonian (cf. geological data) would thus not have left any datable trace.

Northward the obtained apparent ages are younger: east of Belledonne in the Arc valley and at col de la Madeleine, K–Ar ages are 21 to 18 Ma (NZIENGHI, 1993); white mica and adularia from veins of the Mont Blanc massif yielded Rb–Sr and K–Ar ages of 18 to 13 ± 2 Ma (LEUTWEIN et al., 1970).

Available apatite fission track data are in the range of 9 Ma in the Argentera massif (BOGDANOFF et al., 1993), 12 and 4 Ma in the Champsaur area, and 11 to 3 Ma in the Pelvoux massif where zircon fission track ages are of 38 Ma (SEWARD et al., 1997), 9 to 5 Ma between Belledonne and Gdes Rousses (Bourg-d'Oisans area, SABIL, 1995), 7 to 1.5 Ma in the northern Belledonne, Mt Blanc and Aiguilles Rouges massifs where zircon fission track ages are in the range of 17.5 to 10.5 Ma (SOOM, 1990; SEWARD and MANCKTELOW, 1994). These data are useful to constrain the cooling between ~225 (or, according to recent estimates, 250–300 °C) and ~110 °C.

The northward younging of these dates can be interpreted as denoting a higher-grade rather than a younger metamorphic peak. Deeper erosion due to stronger uprise probably also induced a longer cooling time because of the higher attained temperature. Cooling rates corresponding to this hypothesis are in the range of 30 to 5 °C/Ma, consistent with those calculated by CROUZET (1997).

5.3. P-T-t EVOLUTION

Until quite recently the external sector of the Western Alps was considered as almost untouched by Alpine metamorphism. Retrogression of basement rocks was interpreted as a kind of alteration and the fissility in the lower cover rocks as a mechanical effect. Experimental and microscopic evidence is now available which supports the presence of metamorphic transformations, suggesting unexpected T and P values.

The organic matter in La Mure is anthracite, as in the Briançon zone. The corresponding temperature is not accurately known. The laumontite subfacies, which is associated with diagenetic IC values from adjacent sedimentary rocks, would correspond to a temperature of about 200 °C (205–215 °C in the Taveyannaz greywackes according to STALDER, 1979). The pumpellyite-prehnite assemblage, which is associated with anchizone IC values, would correspond to 245–255 °C. On the basis of the Si-content of phengites from basement rocks of the Pelvoux massif, SALIOT (1978) proposed P-T conditions of 1.8 kbar for 335 °C or 2.6 kbar for 400 °C. According to MULLIS et al. (1993) the transition from diagenesis to anchizone would occur at about 230 °C, and the anchizone to epizone transition at 300 °C. From fluid inclusions in vein minerals of the Champsaur area SALIOT et al. (1982) obtained 170–310 °C and 0.2–0.65 kbar.

Another temperature estimate is derived from the presence of the 2M white mica-illite polytype which is said to be the only one stable above 350 °C. In basement massifs the following T and P values have been proposed on the basis of fluid inclusions: in the Mont Blanc massif, 400–420 °C, 2–3 kbar (POTY et al., 1974) or 350–400 °C and 2.5 kbar (VON RAUMER, 1987); in the Pelvoux massif near Bourg-d'Oisans, 360–375 °C and 2.35–2.5 kbar in the basement and 280 °C, 1.4 kbar in the cover (POTY et al., 1974) or 260–360, 1.8–2.6 kbar (GRATIER et al., 1973).

The various mineral associations including cookeite, pyrophyllite, chloritoid, margarite, paragonite (JULLIEN and GOFFÉ, 1993; LEIKINE et al., 1983) found from N and E Belledonne to N of the Pelvoux massif, would correspond to (north) 340 < T < 370–400 °C or (south) 270 < T < 340 °C for P ranging from 2 to 5 kbar.

An assemblage of sphalerite and arsenopyrite in an undeformed Alpine vein of northern Belledonne would indicate 400 °C and 3.5–4.5 kbar (NEGGA, 1984) and from thermoremanent magnetic measurements in cover rocks near Bourg-d'Oisans conditions of 350–320 °C for 2.3 kbar have been inferred (CROUZET et al., 1997).

Summarizing the data, we obtain for the External Crystalline massifs and their cover maximum temperatures higher than 300 °C, likely about 350 °C, or even 400 °C in the northern sector, and pressures ranging between 2 and 5 kbar. In a classical petrogenetic grid these values belong to the greenschist facies although such a facies is commonly denied to the rocks examined, because the minerals are too small to be identified as neoformed, and because characteristic minerals such as actinolite-tremolite and epidote are absent.

It seems difficult to question T and P values which have been consistently obtained with a variety of techniques. Before assuming that the values have been systematically overestimated, we must envisage that greenschist-facies conditions have indeed been attained. Possible explanations can be suggested in order to explain why the greenschist facies is not identified in the rocks in question. The time of peak temperature might have been too short for the minerals to largely crystallize and for equilibrium to be attained, or the very fine grain of the rocks was inherited from the sediment, or the minerals have been prevented from reacting by water saturation in the attending fluid. Moreover, peculiar chemical compositions may have played a part as concerns the cookeite (Li), pyrophyllite (Al), and paragonite (Na)-bearing rocks, as well as P_{H_2O}, P_{CO_2} or other fluid pressure in confined systems: high P_{CO_2} might have prevented actinolite-tremolite and epidote formation, or induced their breakdown.

The presence of chloritoid, cookeite, margarite or paragonite has been considered in the map as indicative of greenschist facies, but not necessarily pyrophyllite and interlayered paragonite-muscovite. Paragonite and/or interlayered muscovite-paragonite which are often associated with muscovite, are responsible for the broadening of the characteristic 10 Å peak. Many IC measurements are likely to be biased in this way, thus leading to narrowing the epizone (as defined by IC) for the benefit of the anchizone. Spectrum analysis will make it possible to differentiate the phases reflecting close to 10 Å, thus allowing the true crystallinity of illite (muscovite) to be measured from its "extracted" peak, to follow the associations as a function of the IC, and to validate, or invalidate, the hypothesis made above concerning the distribution of chloritoid, paragonite and paragonite-interlayered.

Cooling rates can be calculated on the basis of the recently obtained thermopaleomagnetic data (CROUZET, 1997) as correlated with the magnetostratigraphic scale. The results indicate that the cooling has been fast during the 26.5 to 21.5 Ma period (an average of ~ 20 °C/Ma), slowed down to 7 °C/Ma between 20 and 10 Ma and quickened again afterwards.

Such differences imply tectonic denudation, erosion and tectonic re-adjustments. It must be recalled that the fluid inclusion data imply a significant tectonic thickening with 8–12 km of overburden; even if nappe thrusting has been rapid, the thermal adjustment also must have been rapid (and this could explain the very fine grain size mentioned above).

As a conclusion it must be pointed out once more that there is no "one limit" between diagenesis and metamorphism. The success of the anchizone is due to the fact that one can use it without deciding whether it corresponds to diagenesis or metamorphism or still another category, which is both an advantage and a source of confusion. However, as usual in such cases of enlarging a limit to a new category, there are now two limits to be defined instead of only one. In fact, the limit is transitional, corresponding to a range of T and P, and differently located in the T-P space according to the selected criterion.

In the Ligurian Alps, lower anchizone conditions characterize the external flysch (BONAZZI et al., 1987 a and b). According to stratigraphic evidence this transformation is of Tertiary age. On the map the metamorphic conditions in the external Briançonnais units have been shown as anchizonal. The occurrence of palygorskite as centimetre layers likely replacing volcaniclastic products is in accordance with very low temperatures during the metamorphism. However, the very local occurrence of lawsonite and carpholite (GOFFÉ, 1984–86) may indicate a H-P facies.

In the Montenotte meta-ophiolite units (and also in the Cravasco-Voltaggio, see below), where the H-P/T metamorphism has been comparatively low, as well as in the internal Ligurian Briançon units, evidence of retrogressive events lies in the destabilization of Lws, Na-Am and Na-Ca-Cpx, replaced by Ab, Chl and Ep, and by decreasing celadonite contents in white micas. As pumpellyite is unaltered, and actinolite found only as a relic from the higher-pressure conditions, the metamorphic facies is considered as being sub-greenschist.

6. Greenschist facies (Barrovian to H-T)

The transition from the anchi- to the epizone, from W to E, has been detailed in the preceding section. In the following we shall study the full blossom of the greenschist facies with its various mineral assemblages as far as its transition toward medium-grade facies in the Monte Rosa area. Albite-epidote-amphibolite associations are included in the greenschist facies although they are considered by some authors as belonging to a separate facies.

6.1. METAMORPHIC ASSEMBLAGES AND THEIR DISTRIBUTION

The greenschist-facies assemblages overprinted and obliterated the pre-Alpine assemblages of the basement units, as well as the earlier H-P/T Alpine assemblages which will be studied in the companion paper. From the boundary line with very low-grade rocks east- or northeastward, the most apparent variations in greenschist-facies rocks are the coarsening of the grain size, the appearance of biotite and the replacement of actinolite by hornblende and barroisite. A few examples of the assemblages are as follows (see also Fig. 1).

In the Gd St-Bernard supernappe basement, the common metamorphic assemblage includes (abbreviations according to BUCHER and FREY, 1994; in addition Wm for white mica): Wm (Phe to Ms) + Chl + Ab + Qtz ± Ep ± Act ± Kspar ± Cal ± Ttn ± Stp (BURRI, 1983 a and b; BURRI and MAR-

RO, 1993; SCHAER, 1960; SARTORI and THÉLIN, 1987). In marly rocks near Briançon the assemblage is: Cal + fine-grained sheet silicates (sericite) ± Qtz ± Ab ± Chl ± opaque minerals (BOCQUET [DESMONS], 1974).

In the granitoid Arolla gneiss, which forms the largest part of the Dent Blanche nappe, a typical Alpine metamorphic assemblage is Phe + Bt + Chl ± Act + Ep ± Stp + Grt + Ab + Kfs + Qtz ± Ttn (AYRTON et al., 1982; MAZUREK, 1986; DE LEO et al. 1987; PENNACCHIONI and GUERMANI, 1993). Therefore, the Dent Blanche nappe is shown in greenschist facies on the Alpine metamorphic map although there are also traces of high-pressure metamorphism at the base of the Arolla gneiss and at the contact between Valpelline and Arolla series (see the companion paper DESMONS et al., 1999b, this volume).

In white mica, Si-contents as high as 3.6 have been mentioned (SALIOT and VELDE, 1982, in Chasseforêt basement rocks, etc.), but these high values may correspond to an earlier, H-P/T, metamorphic event. Muscovite compositions are more widespread in the NE and E parts of the Western Alps, e.g. in the Siviez-Mischabel nappe and in the Monte Rosa (THÉLIN et al., 1994). In the Combin zone the Si-content in phengite ranges from 3.3 to 3.5, depending on the structural site (LIEWIG et al., 1981).

In the Ruitor, Sapey, Pourri-Bellecôte, Chasseforêt and Ambin massifs two main greenschist assemblages can be recognized. They could not be differentiated on the map. The first (GS_I) is characterized by medium pressure, with phengitic white mica and Chl_I, the second (GS_{II}) by Ms + Bt + Chl_{II}. The growth of poikiloblastic albite, which is not associated with a foliation, can be ascribed to the thermal climax of GS_{II} (PLATT and LISTER, 1985). This albite poikiloblastic growth was ubiquitous, locally abundant as in the Barneuza zone of the Siviez-Mischabel nappe (SARTORI and THÉLIN, 1987). Oligoclase rims around porphyroblastic albite occur from the Ruitor massif northeastward and, in a general way, in the most internal zones. In the Monte Rosa massif oligoclase entirely replaces albite in metagranitoid rocks and an oligoclase isograd can be drawn.

Vitrinite reflectance values of 6.0–8.3%R_{max} have been measured in the Houillère zone in Valais (KÜBLER et al., 1979), indicating lower greenschist-facies conditions. Southward, the organic matter in the Houillère zone currently is anthracite (4–5 wt% of volatile matter). In the Pourri-Bellecôte massif as well it is not graphite (FABRE, 1961; SCHADE in GUILLOT, 1987).

Actinolite often predominates over chlorite in the more internal parts of the Combin, the Zermatt zone and in the Internal Penninic massifs. Assemblages observed in the Susa valley (LEARDI et al., 1986) are: in ovardite, Ab + Chl (together making > 60%) ± Am ± Ep ± carbonate minerals ± Ttn ± Qtz ± Rt ± Ap ± Wm ± Bt ± opaque minerals; and in prasinite, Ab + Act + Chl + Ep in equal amounts. The transition from actinolite to hornblende, together with the appearance of almandine-rich garnet and clinopyroxene, also occurs in the Monte Rosa metabasic rocks (COLOMBI, 1989, and references therein). In the Montgenèvre area the imprint of the greenschist facies indicated on the map actually is not pervasive.

Garnet occurs in the northern and eastern parts of the Western Alps, mostly in the Internal

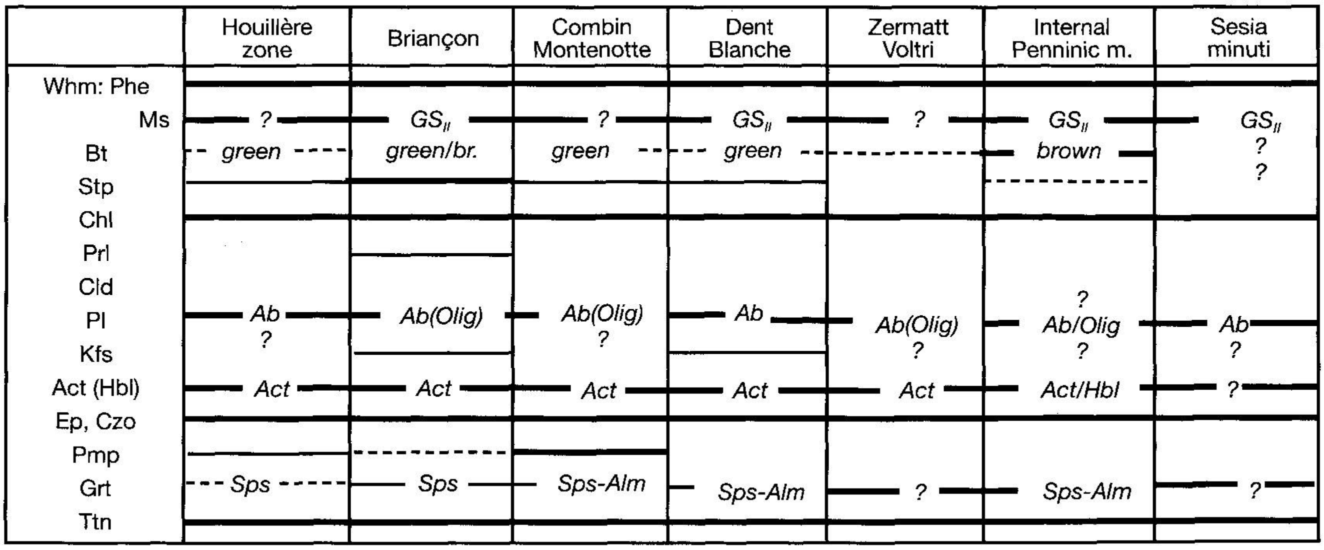

Fig. 1 Greenschist-facies minerals present in the Western Alps (± Qtz ± Cal ± accessory minerals). Dashed lines represent very rare occurrences, thin lines uncommon occurrences.

Penninic massifs, and in the Combin zone in the Aosta valley. It is spessartine-rich in the Ruitor massif (DESMONS et al., 1977; LADURON and DESMONS, 1981). Locally in the Combin zone spessartine-rich cores are surrounded by clear-cut rims richer in almandine. In the Houillère zone garnet occurrences are rare (DESMONS, unpubl. data).

Green Ti-poor biotite is present in most Briançon basement massifs (Acceglio, Ambin, Chasseforêt, Sapey), but it is rare in the Houillère zone. Its colour gets brownish from the Ruitor northeastward. It is present in the Versoyen, the Combin zone, the Arolla series in the Dent Blanche nappe and the Sesia Gneiss Minuti complex. It is very rarely mentioned in the Zermatt zone. In the Internal Penninic massifs it is brown. In the UH-P Brossasco-Isasca unit of the southern Dora-Maira massif, phlogopite may be a retrograde phase, resulting from a reaction involving pyrope garnet, phengite and kyanite. Biotite is absent from the cover rocks in the Briançon area itself and southwards (except in the Acceglio zone), as is stilpnomelane. In Briançon basement rocks, stilpnomelane is rather common, it occurs also in the Houillère zone and is generally described as a late mineral.

In addition to participating in the H-P/T assemblages, chloritoid is a greenschist-facies mineral in the internal parts, the distinction being made on the basis of mineral relationships.

The Ti-mineral is titanite, from which tiny needles of undetermined Ti-oxide grew as a late phase.

Pumpellyite-actinolite assemblages occur in Montgenèvre, in various places of the Combin and Briançon zones and locally in the Sapey gneisses. An assemblage characteristic of the external part of the greenschist area of the Western Alps (FREY, 1986), it is also commonly found in the anchizone such as in the Ligurian Alps (see below).

In Mn-rich rocks associated with ophiolites, parsettensite and rhodochrosite have been mentioned (e.g. in Cesana, MARTIN and POLINO, 1984–87).

In the Ligurian Alps, greenschist-facies assemblages are widespread within the tectono-metamorphic units that had been affected by eclogite to blueschist facies, the Voltri and Torrente Visone units (CHIESA et al., 1975, 1977; CIMMINO et al., 1980, and pers. data). The stability of subcalcic hornblende and green, but rarely brown, biotite is typical of the transformation from H-P/T through albite-amphibolite facies to the greenschist assemblages characterized by the post-kinematic growth of helicitic albite. Greenschist-facies assemblages (Phe, green Bt, Chl, Cal, Ep) predominate in most metasedimentary sequences of the Voltri Group such as, for example, the Voltri-Rossiglione unit, where H-P/T relics are only few. Spessartine-rich garnet is common within quartzschists, where widespread chloritoid became metastable during the latest retrograde phases, supporting the decompressional evolution.

In Voltri meta-ophiolites affected by conditions of a grade lower than eclogite to blueschist during the H-P/T metamorphic peak as well as in the "Piemonte" Montaldo unit, a greenschist overprint also developed, but the textural re-organization is generally poor. Only the Varazze meta-ophiolite unit south of the Voltri Group, where any H-P relic is lacking, shows greenschist-facies assemblages. In metagabbros largely re-equilibrated to Ab + Act + Chl + Czo + Ttn ± Phe typical assemblages, augite relics are preserved. The associated ultramafic rocks contain Atg, Mag, Chl, Di, Tr and Hu.

Greenschist-facies assemblages are also found in mylonitic bands developed in, or between, the meta-ophiolitic units of the Voltri Group, along shear zones and fractures in the metabasites of the Arenzano basement and its metasedimentary cover. As said above, the retrogressive metamorphism in the Cravasco-Voltaggio and Montenotte meta-ophiolite units and in the internal Ligurian Briançon units is considered to have been of sub-greenschist facies.

A remarkable feature is shown by the Sesia zone. In its internal part, consisting of the Eclogitic Micaschist complex, the major Alpine imprint is of H-P/T. Greenschist-facies assemblages are pervasively developed mainly along shear zones and at the border zones with the more external Gneiss Minuti complex, as well as in the northern part toward Monte Rosa and the Central Alps. In the Second Diorito-Kinzigitic zone any greenschist or lower-grade imprint is very weak and concentrated close to the basal thrust plane. In the Gneiss Minuti complex the greenschist facies is pervasive, similar to that of the Arolla Gneiss Minuti in the Dent Blanche nappe. It may be assumed that both the Eclogitic Micaschist complex and the Second Diorito-Kinzigitic zone (the parts of them that are now exposed) already were too shallow to be affected by the greenschist-facies metamorphism.

6.2. CHRONOLOGICAL DATA

6.2.1. Geological data

From the stratigraphic age of the lowermost post-thrusting clastic strata it has been concluded that the nappe thrusting of the cover started during Ypresian (lowermost Eocene), with a climax in Lutetian, protracted to, or repeated, in Bartonian depending on the zones considered (DEBELMAS et al., 1988). Since the distribution of the greenschist metamorphic imprints cross-cuts the main structural boundaries, both greeenschist imprints clearly appear to be post-Middle Eocene in age. In the Ligurian Alps also, the greenschist facies imprint post-dates the piling-up of the Beigua-Ponzema, Erro-Tobbio and Palmaro-Caffarella units in the Voltri Group. Evidence as to the upper age limit is derived from clastic rocks of Upper Eocene-Lower Oligocene age, the Bracce di Costa Cravara, which contain fragments bearing greenschist facies assemblages. Furthermore, Lower Oligocene (lower Stampian) sediments of the Pianfolco unit, deposited upon the Voltri Group, are devoid of any greenschist imprint (LORENZ and REBORA, 1982), although there is evidence of tectonic transport during (Mid ?) Oligocene accompanied by zeolite growth.

The minerals of the first greenschist assemblage (GS_1) lie in a foliation plane which is a major S, younger than the nappe emplacement. It is the S_2 of most structural geologists, a schistosity related to folding, which post-dated D_1, which itself, according to the authors, post-dated or accompanied the main nappe thrusting (e.g. PLATT and LISTER, 1985a; LEMOINE and TRICART, 1993; DEBELMAS and DESMONS, 1997, and references therein).

Structural data from the Chasseforêt massif (PLATT and LISTER, 1985 a, b), corroborated by data from other areas, indicate that the evolution after S_2 proceeded through a thermal peak marked by the growth of albite poikiloblasts, to extensional backfolding still under greenschist conditions (D_3), and finally to differential uplift under lower-grade conditions, mainly evidenced as vein mineral crystallization. A similar sequence of events has been assumed in the Dent Blanche nappe (OBERHÄNSLI and BUCHER, 1987). In Queyras the structural evolution comprised thrusting associated with an S_1, followed by S_2 and S_3 folding, and finally backfolding (GIDON, 1994; BARFÉTY et al., 1995).

In the Ligurian Alps three main, possibly composite, deformation events associated with penetrative foliation or cleavage define the evolution of the Voltri meta-ophiolites and the Torrente Visone units, where greenschist assemblages essentially developed post-kinematically with respect to the second deformation event. At microscale it can be observed that microhinges of the phases 1 and 2 are included in helicitic albite.

6.2.2. Radiometric data

This section gives a summary of the available data. Discussion of the methods can be found in some of the recent papers. The interpretation of the present authors is explained in the next section on the P-T-t evolution and in the conclusions (see also the companion paper, DESMONS et al., 1999b).

Radiometric measurements obtained with the K–Ar, $^{40}Ar/^{39}Ar$ and Rb–Sr methods (see compilation in HUNZIKER et al., 1992; previous references are not repeated) had been summarized as 38 ± 2 Ma, that is late Bartonian or Priabonian depending on the chosen time scale. However, only $^{40}Ar/^{39}Ar$ dates evaluated according to the mineral, the grain size, the cooling rate, etc., may be relied upon as likely giving the closure age below the specific temperature. The K–Ar and many $^{40}Ar/^{39}Ar$ data of the low-grade metamorphic minerals of the Western Alps must thus be regarded with criticism, even if they represent plateau ages and even if they seem geologically acceptable.

Some of the radiometric data recently acquired with the $^{40}Ar/^{39}Ar$, Rb–Sr and U–Pb methods again are close to 38 Ma (41 to 36 Ma ± various errors). They were obtained from Sesia and Combin rocks retromorphosed in greenschist facies (RAMSBOTHAM et al., 1994; WHEELER et al., 1995; INGER et al., 1996) and by synkinematic micas of both basement and cover rocks of the Siviez-Mischabel nappe (MARKLEY et al., 1998). BARNICOAT et al. (1995) give evidence of an early cooling at ~ 44 Ma of Zermatt rocks under the Mischabel fold and of possibly protracted cooling until 36–32 Ma within the Mischabel rocks themselves. Rb–Sr measurements of white mica from mylonites of the Aosta valley and from the area between the Gd St-Bernard zone and the Gran Paradiso massif (BUTLER et al., 1997; FREEMAN et al., 1997) indicate diachronous thrusting, from ~ 45–42 Ma for the Dent Blanche, Zermatt and Sesia zones, through 38–36 Ma for the Combin zone and the Gran Paradiso massif to 34 Ma for the backthrusting.

On the other hand, zircon SHRIMP dating (GEBAUER et al., 1997; RUBATTO, 1998) gives ages around 35 Ma for eclogite metamorphism in the Gran Paradiso massif and for the ultra-high pressure metamorphism in the Dora-Maira massif. According to the latter authors, in these zones at least, the greenschist assemblages thus would be younger than 35 Ma; this point will be discussed again in the general conclusions of the companion paper (DESMONS et al., 1999b).

In the Siviez-Mischabel Pb-isotope measurements yielded ages of 25 Ma (ZINGG, 1989), and in the Monte Rosa massif ages of 29 Ma (ROMER et al., 1996, and references therein) may reflect the thermal culmination, i.e., GS_{II}, which must have taken place at some time that from geological evidence can only be situated between Lower Oligocene and the beginning of the Miocene (RADELLI and DESMONS, 1997). This time may be more precisely constrained by data from the Central Alps and, farther beyond, the Tauern window, and their thermal culmination. According to GEBAUER (1999, this volume) the preferred value for the T-peak in the Lepontine area is 30 Ma.

As shown by fission track ages of zircon (HURFORD et al., 1991; HEJL and WAGNER, 1991; and references therein), at around 33 Ma (between 39 and 32 Ma) central Sesia was already below a temperature of ~ 225 °C (or 300–250 °C according to recent estimates). The Internal Penninic massifs passed through this temperature in Late Eocene (30 Ma for Dora-Maira according to GEBAUER et al., 1997, but other published values are mentioned in the companion paper; also 30 Ma for the Gran Paradiso massif). Fission track zircon ages as early as 26 Ma to 16 Ma have been obtained from the Briançon zone, and 17 Ma from the External Crystalline massifs (see above, 5.3.2.). These fission track dates concern the cooling path after GS_{II}.

6.3. P-T-t EVOLUTION

Maximum P and T conditions estimated for the first greenschist facies, GS_I, are 250 to 400 °C and 2.5 kbar in the External Crystalline massifs (see above), about 300 °C and ~ 3–4 kbar in the Houillère zone, 350 ± 50 °C and ~ 3–4 kbar in the Chasseforêt massif, ≤ 400 °C in the Dent Blanche nappe, and 450–500 °C and < 5–6 kbar in the Internal Penninic massifs. The pressure, comparatively high during GS_I as suggested by the high celadonite content of phengite, may have been very low during the GS_{II} imprint in most parts of the Western Alps, as indicated by the presence of muscovite + chlorite ± biotite, and by fluid inclusion data (POTY et al., 1974). From the presence of biotite and the lack of associated deformation, the GS_{II} facies can be considered as a regional thermal metamorphism, characterized by geotherms of 40 to 53 °C/km (DESMONS, 1992; BORGHI et al., 1994, 1996). In the northeastern parts of the Western Alps, the oligoclase rim around porphyroblastic albite, commonly associated with biotite, indicates a T increase of 60 to 80 °C. In the Second Diorito-Kinzigitic zone, modeling of partly rejuvenated pre-Alpine micas suggests that in the nappe the temperature never exceeded ~ 300 °C (REDDY et al., 1996).

The P-T-t path defined by these values and the ages given in the preceding chapter are illustrated in figures 2 and 3 of the companion paper (DESMONS et al., 1999b. It is interesting to note that the P-T path from GS_I to GS_{II}, characterized by T increase and moderate P decrease, is also recorded in the zoning of the small idioblastic garnet included in porphyroblastic albite of the Internal Pennnic massifs (BORGHI et al., 1996); it has, therefore, been used to constrain this part of the P-T path.

In the Penninic zones the greenschist imprint has been very pronounced in the northern and northeastern parts, especially in cover sedimentary rocks which before the metamorphism still contained much of their fluids. But also in little hydrated rocks such as the basement and ophiolite rocks the Alpine deformation has been strong enough as to allow some fluid influx, the appearance of a new foliation and the breakdown of former minerals. However, O and H isotopic measurements in the Monte Rosa massif and the Sesia zone (FREY et al., 1976; DESMONS and O'NEIL, 1978) demonstrate that the fluids circulating in basement rocks have not been pervasive and to a large extent have been inherited from the precursor rocks.

It may be assumed that the paleo-isotherms were gently rising towards E and NE as far as the western boundary of the Sesia Eclogitic Micaschist complex (with respect to present-day surface; actually they have been uplifted and tilted by the tectonic culmination of the Central Alps).

7. Very low- to low-grade facies in the Canavese and Sestri-Voltaggio zones

All over the *Canavese zone* the most significant mineral assemblage is Prh + Pmp ± Act ± Chl ± Fe-rich to Fe-poor Ep ± Stp ± Ab ± Wm + Tnt (BIINO and COMPAGNONI, 1989). On the basis of the coexistence of pumpellyite and actinolite, this assemblage has been shown as greenschist facies on the map. However, the presence of prehnite may indicate a very low grade, whereas IC values (ZINGG et al., 1976) are intermediate between anchizone and epizone. This Alpine metamorphic imprint is mainly localized along the shear zones and the tectonic contacts of both basement (granitoids and metamorphic rocks) and cover (mainly Permian volcanic rocks). The age of this imprint would be 60–76 Ma according to K–Ar measurement of mylonitic micas (ZINGG and HUNZIKER, 1990), that is, around the age of the Sesia eclogites. It is noteworthy that whereas the Insubric Line (here named the Canavese Line) is marking a sharp jump in metamorphic grade between the Sesia and Canavese rocks, the tectonic boundary between Canavese and Ivrea zones, named Internal Canavese Line by BIINO and COMPAGNONI (1989), is not associated with any Alpine metamorphic break, the same overprint being present on both sides.

Farther east, in the Torino Hills and beyond, the rocks though folded bear no metamorphic imprint and they have been marked accordingly on the map.

In the *Sestri-Voltaggio zone* of the Ligurian Alps, the mutual stratigraphic relationships between the meta-ophiolite Monte Figogna unit and the Busalla Flysch seem to be preserved. In metabasalt Ab + Pmp + Act + Chl + Ttn assemblages are widespread, whereas lawsonite associated with albite is restricted to aggregates pseudomorphing the largest plagioclase phenocrysts. On the map these areas have been shown as anchizonal. However, the metamorphic conditions have been estimated at 300 °C < T < 350 °C and P = 3–4 kbar (LUCCHETTI et al., 1990).

Illite crystallinity data in ophiolitic cover and associated flysch rocks are mutually consistent and agree with the data obtained from the eastern Liguria Val Lavagna flysch (CORTESOGNO and VENTURELLI, 1978; LEONI et al., 1996). The age of these transformations is considered, on stratigraphic evidence, as Late Cretaceous to Paleocene, as is the H-P facies of the nearby Cravasco zone, that is earlier than in the external flysch zone. It has not been possible to indicate this age difference on the map.

8. Conclusions

1. The *general picture* of the anchizonal to greenschist facies in the Western Alps is that *of metamorphism rather regularly distributed* over all structural zones. In a general way, the inferred metamorphic grade increases towards the Internal Penninic massifs, from anchizonal through chlorite-white mica to biotite-muscovite assemblages. In the external sectors this increase is only disturbed by duplication and other tectonic effects, and by a sort of "basement effect" around the External Crystalline massifs. At the time of the Cainozoic metamorphisms, the whole area broadly reacted as a single zone. Even late movements at the contact between Internal and External zones determined only minor gaps in the distribution of metamorphic facies all over the Western Alps. Such a regular distribution of the greenschist facies seems to indicate that, before the greenschist imprints, the units to be affected had reached a similar crustal level and that their spatial relationships were not far from those of today.

The major exception lies in the metamorphic jump along the Insubric line. After the greenschist-facies imprints, the Internal zones have been uplifted with respect to the downthrown South-Alpine and Canavese zones. Also, on the internal slope of the Alps, the inferred grade decreases to low greenschist and anchizonal conditions. Most likely the exceptional preservation of the eclogite-facies assemblages of the Sesia Eclogitic Micaschist complex is due to its rapid exhumation and shallower tectonic position at the time of the greenschist imprints, a position which has later been modified by post-Oligocene clockwise southeastward tilting of the most internal portion.

2. A first greenschist imprint, GS_I, has been succeeded in the Internal zones by a thermal regional metamorphism, GS_{II}, which increased in grade and is the main detectable imprint northeastwards, i.e. towards the Lepontine thermal high of the Central Alps. The greenschist-amphibolite facies transition is reached within the Monte Rosa massif. During GS_{II} green and brown biotite were formed, as well as poikiloblastic albite or, in the most internal regions, oligoclase.

The best evidence of such a succession and its relationship with deformation is found in those areas of the Briançon and Combin zones that are intermediate between the more internal regions where GS_{II} has been of higher grade and locally obliterated GS_I (or has been the only one greenschist imprint) and regions farther away from the Lepontine thermal high where GS_{II} has been very weak. It is important to stress that GS_I and GS_{II} refer to assemblages, to metamorphic imprints, and not to tectono-metamorphic phases: deformation phases succeeded each other, but the assemblages have not been each time modified.

3. At the time of GS_I the *pressure* was medium in most of the Western Alps, as shown by the presence of phengite, pumpellyite-actinolite, laumontite and the absence of wairakite. In the Houillère zone it is even unclear whether the facies has not been a lawsonite-chlorite-albite facies, in which lawsonite did not regionally form due to too low a CaO content in the Carboniferous metaclastic rocks. The same may also have been the case of the Queyras area during GS_I. On the contrary, during GS_{II} the pressure was low. Fluid inclusions data of vein minerals from the External Crystalline massifs indicate an Abukuma-type facies series. Thus, the early medium T/P gradient turned to higher values of about ≥ 40 °C/km during GS_{II}.

4. On the basis of geological evidence, the most likely *age* for GS_I is ~ 38 Ma, i.e. late Bartonian, and for GS_{II} between 30 and ~ 23 Ma (Early

Oligocene to Early Miocene). Radiometric evidence points to around 30 Ma for GS_{II}. Repeated deformation phases occurring between, and after, these imprints cannot be differentiated on the basis of the metamorphic assemblages: the differentiated deformation phases are more numerous than the metamorphic imprints. In the areas affected by anchimetamorphism, as explained in the relevant chapter, there are indications of polyphase anchizonal conditions, from Late Cretaceous to Miocene.

5. In the Internal zones, greenschist minerals are assumed to have crystallized from about the time of the D_2 deformation onwards. The high *P/T ratio* due to the D_1 thrusting phase had started decreasing. When GS_{II} assemblage formed, the T/P ratio had become high. In these zones the greenschist facies is often considered as a sort of unescapable continuous alteration of higher-P assemblages during their exhumation, an alteration that could have occurred diachronously, i.e. at any times in the different zones. However, two lines of evidence stand against this opinion. (i) If some sluggish alteration may have occurred in the wake of the H-P/T transformations, such a thermal metamorphism as GS_{II} was related to a specific tectonic setting, imposed to the whole Alpine belt at the same time. (ii) The greenschist-facies assemblages were associated with, or intercalated between, deformation phases. The deformation phases, in the present state of knowledge, seem to be consistent all over the Penninic zones. Moreover, in many places the greenschist-facies assemblages post-date glaucophane-schist or epidote blueschist assemblages and, as it will be explained in the companion paper (DESMONS et al., 1999b), these H-P/T assemblages at least in many zones formed at the time of the thrusting phase, at an already rather shallow crustal level (shallower than the often assumed ≥ 20 km). Finally, both GS_I and GS_{II} indeed are present in the P-T paths inferred by BORGHI et al. (1996) for the Dora-Maira and Gran Paradiso massifs, and they are quite similar to the Vanoise P-T path (see Fig. 2-c in the companion paper).

6. According to the commonly adopted view, the greenschist-facies metamorphism resulted from the relaxation of the geotherms during and after the exhumation following subduction of Tethyan basins and subsequent collision of Europe with an "Adria" or "Apulia" plate. However, it has been argued (RADELLI, 1995) that the collision had already occurred in lower Mesozoic and that the Eocene-Oligocene *tectonic environment* of the Alps was that of A-subduction of the European lithosphere forced below the Alpine zones by N-Atlantic spreading. Melting of the mantle underlying these zones in a supra-subduction zone setting entailed partial melting (and the so-called peri-Adriatic magmatism eventually), as well as metamorphism within the neighbouring crust. The piled-up units at first imposed a low T/P gradient. Afterwards an extensional regime was accompanied by a strong uprise of the geotherms and a high T/P regime, resulting in anatexis locally in the Central Alps and, farther away from the Lepontine thermal high, in the growth of biotite and albite. According to this view, the Eocene-Oligocene metamorphism is unrelated to continental collision.

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