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Deformation migration in an orogen-scale shear zone array: an example from the Basal Briançonnais Thrust, internal Franco-Italian Alps

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Abstract – Combined structural, geochemical and isotopic studies have allowed an understanding of the timing and nature of an orogen-scale fault array. The results indicate that the deformation loci within the internal western Alps, during the Alpine collision, occurred as a foreland propagating thrust sequence. The east to west deformation migration within the internal zones is apparently in-sequence in relation to the external zones. Rb–Sr white mica dating of syn-kinematic greenschist-facies mineral assemblages from the Basal Briançonnais Thrust indicate that thrusting ceased between 27 and 32 Ma, several million years after shearing in the hinterland and several million years prior to shearing in the foreland. The Briançonnais Domain, which constitutes the hanging wall to the Basal Briançonnais Thrust, preserves two major shearing episodes. The first, with a top-to-the-northwest overshear, has been tentatively dated at 45 Ma. The second, a very pervasive, east–west orientated, greenschist-facies event was previously dated at 34 Ma on the hinterland margin of the Briançonnais Domain and has know been dated at 27–32 Ma on the foreland margin of the Briançonnais Domain. The period between 34 and 27 Ma apparently dates the migration of deformation through the relict European passive margin, represented by the Briançonnais Domain. This is believed to be in response to overthrusting of Adria/Africa and its associated subduction complex. Structural mapping indicates that the present Basal Briançonnais Thrust in the Col du Petit St Bernard region, Franco-Italian Alps, is a break-back thrust which cuts through an already imbricated pile. Geochronological evidence suggests that the early imbrication of the Briançonnais stratigraphy occurred prior to full interaction of the European and Adria/African plates, that is, during subduction, docking and escape from the subduction complex under Adria. Therefore, although the present Basal Briançonnais Thrust is a break-back thrust in terms of local structural geometries, it is an in-sequence foreland-propagating structure. Geochronological, micro-structural and micro-chemical data indicate that the Briançonnais Domain in the Col du Petit St Bernard zone is formed from granitoid material which intruded and cooled at approximately 320 Ma. During the Alpine event, deformation and metamorphism were insufficient to affect the Sr isotopic system. This suggests that this portion of the Briançonnais Domain was probably subducted to much shallower depths and underwent much less pervasive deformation than the other internal European basement material.

1. Introduction

Many collision mountain belts consist of shear zones which, on the the scale of the orogen, slice together and duplicate sections of the continental crust (see, e.g. Coward *et al.* 1988; Jadoon, Lawrence & Lillie, 1992). Attempts have been made to interpret these crustal-scale structures in terms of thrust belt tectonics (e.g. Soper & Barber, 1982; Coward & Butler, 1985; Treloar *et al.* 1992), with the implication that deformation migrated through time to involve progressively outlying lithosphere which is accreted into the evolving orogen. Stratigraphic studies of syn-tectonic sediments in the outer parts of mountain belts commonly record this outward migration of deformation (e.g. De Graciansky *et al.* 1989; Jadoon *et al.* 1997). However, such stratigraphic approaches are not appropriate for the internal parts of mountain belts where deformation occurred at significant depths in the crust and has subsequently been exhumed. Regional patterns of deformation have been investigated using the isotopic dating of metamorphic cooling after deformation (e.g. Hunziker, Desmons & Martinotti, 1989), locally tied to intrusions. Thus, deformation is bracketed to specific periods, an analytical approach which can erroneously lead to the belief that deformation was pulsed or sporadic (e.g. Hunziker, Desmons & Martinotti, 1989; Marchant & Stampfli, 1997). With the development of analytical strategies to date directly the crystallization of rock-forming minerals within shear zones (Getty & Gromet, 1992) it is now possible to examine the migration of deformation within the internal parts of mountain belts, thereby testing large-scale models of orogenic evolution. Here we apply these methods to an array of shear zones within the internal part of the northwest Alps.

The Basal Briançonnais Thrust is one of the major tectonic breaks of the western Alps. It is particularly visible on seismic reflection profiles acquired through the French (Nicolas *et al.* 1990) and Swiss Alps (Valasek *et al.* 1991), dipping at about 30° beneath the internal zones of

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the chain. Its hanging wall is the Briançonnais Zone, a tract of Permo-Carboniferous and older basement together with a Triassic–Eocene passive margin cover sequence (Ellenberger, 1958; Debelmas, Escher & Trumpy, 1983; Lemoine *et al.* 1986). Its footwall is the Valais Zone, an orogenic tract which, in Switzerland, has oceanic affinities (see, e.g. Ramsay, 1963). These relationships have been used to infer that the Basal Briançonnais Thrust was either a subduction zone or is a structure which masks one (e.g. Marchant & Stampfli, 1997). It clearly is an important shear zone for studies of Alpine orogenesis.

The purpose of this paper is to apply new geochronological methods to date the age of shearing along the Basal Briançonnais Thrust and to tie the structure to adjacent shear zones within the Alpine orogen, thereby establishing its broader significance. We present an integrated field- and micro-structural, metamorphic and isotopic study that yields a history of progressive deformation and crystallization within the western Alpine orogen (Fig. 1). We address the fundamental question of the timing of deformation within a major shear system by chemically identifying the minerals that can be linked to geometrically and thermally distinct episodes of deformation. Determining isotopic ages of those minerals by the Rb–Sr system allows resolution of activity on different structures within a complex polyphase history. The patterns of migrating deformation revealed by these data are then related to the orogen-scale models of shear system timing and linkage, and possible relationships to plate tectonic controls are explored.

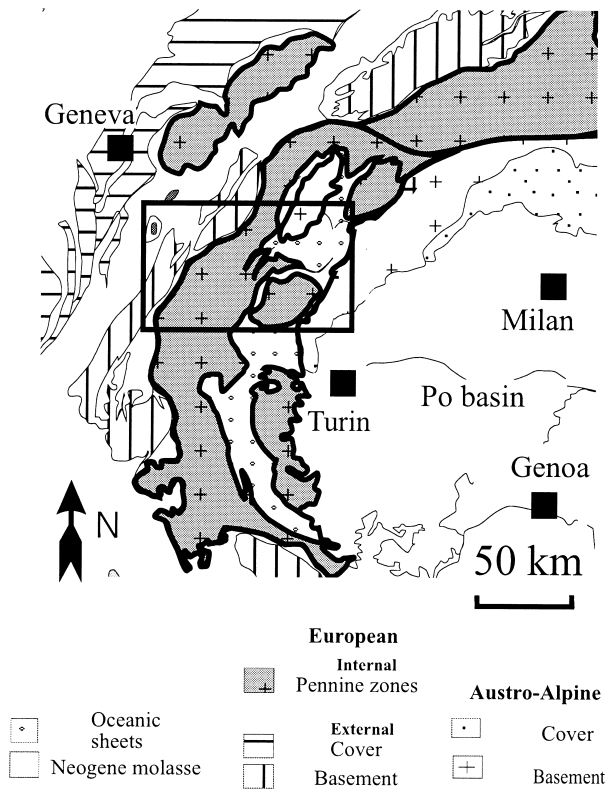


Figure 1. Geological map of the western and central Alps. The diagram shows the major litho-tectonic provinces. The square shows the area covered in Figure 2.

2. Regional geology

2.a. The western Alps

The western Alpine orogen (Fig. 1) is a west–northwest directed thrust belt which on a crustal scale, seismic reflection data show (e.g. Nicolas *et al.* 1990), dips eastwards beneath the lithospheric mantle of the Italian (Apulian) block. The simple, single direction of subduction of the lower and middle crust is not mirrored by upper crustal structure. Here the large-scale structure is directed outwards so that folds face away from the orogenic interior (see, e.g. Debelmas, Escher & Trumpy, 1983). This tectonic divergence is centred on the Briançonnais Zone (Fig. 2). The Basal Briançonnais Thrust on the northwest side of the zone dips moderately towards the east, generally with down-dip stretching lineations and top-to-the-northwest/west–northwest shear senses (Bertrand *et al.* 1996). The thrust is sub-parallel to the Pennine Front, a similar thrust which traditionally is interpreted as the metamorphic edge of the orogen (e.g. Ramsay, 1963). Structural studies show the Pennine Front to be a foreland-directed thrust (Butler, 1992a) and, like the Basal Briançonnais Thrust, it is prominent on seismic reflection profiles (Nicolas *et al.* 1990).

The hinterland margin of the Briançonnais Zone is a major back-fold and backthrust system which includes some of the most prominent antiforms in the Alps (Dal Piaz, 1928; Caby, 1968) (Fig. 2). Recent studies show these structures to repeat tectonic stratigraphy and to be east–southeast directed thrusts (Butler & Freeman, 1996). Similar structures are found within the Briançonnais Zone itself (see, e.g. Caby, 1968; Platt *et al.* 1989). On the scale of the orogen the Briançonnais Zone forms a composite 'pop-up' (in the sense of Butler, 1982). Since the various thrusts and folds within the Briançonnais do not involve rocks deeper than a few kilometres beneath the basement–cover contact, Butler (1986) interpreted these structures as splaying from a detachment in the upper crust of the Alps. This interpretation is consistent with seismic reflection data (Nicolas *et al.* 1990; Valasek *et al.* 1991).

Our aim is to examine the evolution of the Briançonnais shear zone array, building on our earlier studies of the Entrelor Thrust on the hinterland side of the array (Butler & Freeman, 1996; Freeman *et al.* 1997). First it is necessary to review briefly the metamorphic setting of these rocks.

2.b. High pressure metamorphism–subduction of continental crust

The internal parts of the western Alps are the restacked remnants of a subduction system. It is associated with widespread high pressure metamorphism recorded at outcrop (Chopin, 1987) and inferred at depth (Butler, 1986).

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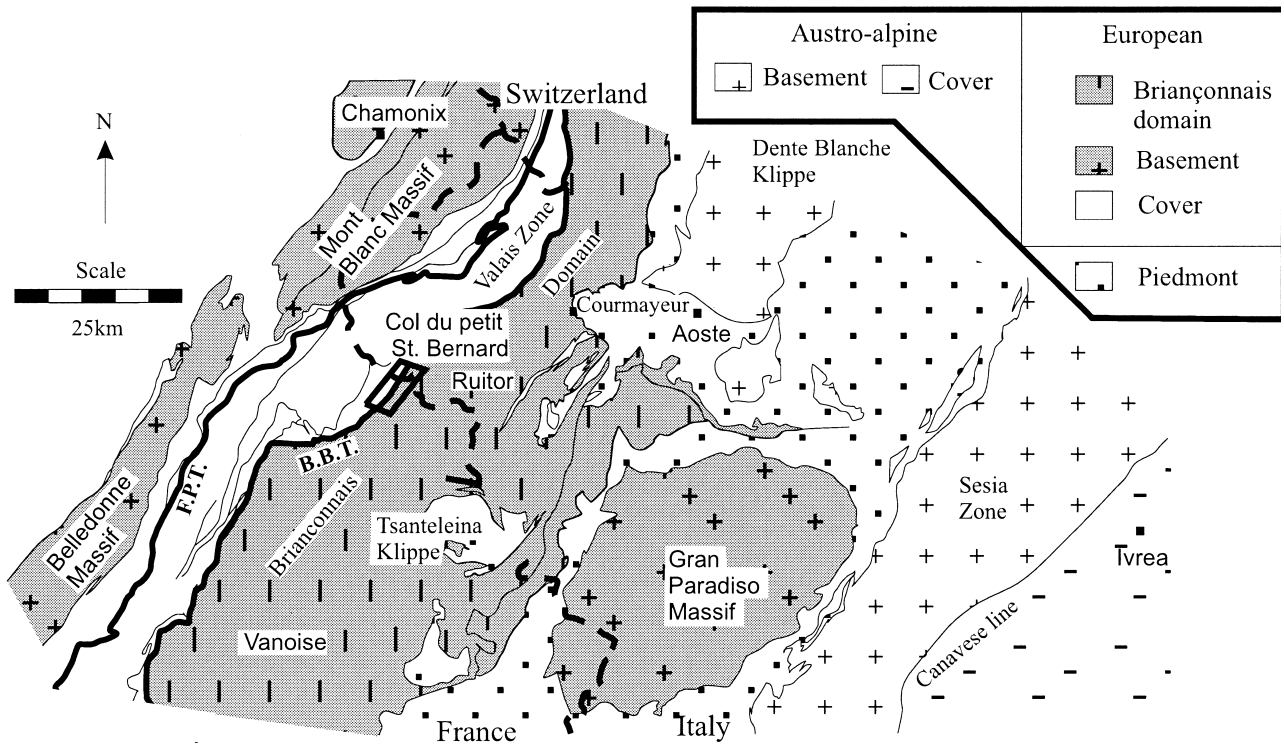


Figure 2. Geological map of the western Alps. The map shows places named within the text. The box shows the area shown in Figure 3. After Debelmas (1980).

Until recently, high pressure metamorphism in the western Alps was ascribed to a so-called 'Eo-alpine' orogeny in the Cretaceous, largely on the basis of sparse geochronological data from Switzerland. However, new work presents a different story. It is probably wrong to assume that high pressure metamorphism in the western Alps is ubiquitously of Cretaceous age (see discussion by Gebauer, 1996, p. 325) .

Most recent studies of high pressure metamorphism in the Sesia zone (Adrian side of the orogen) suggest peak high pressure conditions occurred between 60 and 70 Ma (discussed by Inger *et al.* 1996). Other units from the palaeo-oceanic Zermatt-Saas Zone consistently yield Tertiary ages of high pressure conditions (Bowtell, Cliff & Barnicoat, 1994; Rubatto & Gebauer, 1996), while the structurally lower Internal Basement Massifs of European affinity have yielded ages for high pressure metamorphism in the 38–40 Ma range (Tilton, Schreyer & Schertl, 1991; Gebauer *et al.* 1993; Becker, 1993; Gebauer, 1996; Chopin & Maluski, 1980).

The Internal Basement Massifs form the most eastern parts of the European continental margin within the Alpine subduction zone and record the highest pressure conditions of metamorphism (Chopin, 1987; Fig. 1). Prior to orogenesis these massifs formed the outer edge of the northern margin of Tethys (Lemoine *et al.* 1986). The Briançonnais Zone forms the next part of the continental margin (Cannic *et al.* 1995; Fig. 2). However, high pressure assemblages are only locally recognized, largely because rock compositions are unsuitable. Glaucophane-bearing assemblages with a prominent southeast–northwest stretching lineation assumed to be of 'Alpine' age are recorded from some basement rocks of the Briançonnais (e.g. Bocquet *et al.* 1978; Platt & Lister, 1985). The best evidence for high pressure conditions comes from the far eastern parts of the Briançonnais Domain in Switzerland. The Adula-Cima Lunga nappe complex preserves eclogites and garnet pyroxenites. Metamorphic overgrowths on zircons have yielded U–Pb SHRIMP ages of *c.* 33 Ma (Rubatto & Gebauer, 1996), thought to date peak high pressure conditions. However, the large spatial separation of this site from the Western Alps, together with the amount of documented diachroneity in, for example, the age of metamorphism in the internal basement massifs, renders correlation unwarranted at present.

To the northwest of the Briançonnais Zone, the calc-mica schists of the Valais Zone, in the footwall to the Basal Briançonnais Thrust, include rare eclogitic assemblages together with ferro- and magnesiocarpholite with chloritoid (Cannic *et al.* 1996). The assemblages appear to have formed early in local deformation chronologies and have been considered to be Cretaceous in age (e.g. Hunziker & Martinotti, 1984; Marchant & Stampfli, 1997). Until our study the only modern geochronological study of the Valais Zone in the Western Alps was directed at obtaining cooling rather than deformation ages (Seward & Mancktelow, 1994). However, stratigraphic correlation of distinctive sediments within the upper part of the Valais Zone with units in the less-deformed parts of the Briançonnais of the western Alps suggests that deposition here continued into middle Eocene times (*c.* 45–48 Ma: Barfety, Tricart & de Grissac, 1992).

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The message from all these ages from rocks derived from the European continental margin now stacked in the western Alps is that subduction occurred during late Eocene times and not during the Cretaceous, as suggested by old data from Switzerland (Hunziker & Martinotti, 1984). Sedimentation continued across the external parts of the Alps, together with the Valais and Briançonnais zones, until middle Eocene times. Oceanic lithosphere now represented by the Zermatt-Saas ophiolites, and continental crust of the Sesia Zone, entered the subduction zone rather earlier. Nevertheless, our starting point for studies of the subsequent evolution of the western Alps is that the orogen initiated from a single, southeast directed subduction zone.

2.c. Greenschist-facies overprinting: the return to middle-crust conditions

Overprinting the high pressure assemblages of the European margin in the northwest Alps is a widespread greenschist-facies assemblage implying exhumation to relatively shallow crustal depths. In the northwest Alps the exhumation mechanism is not at all clear although it probably involves regional syn-orogenic extension (see, e.g. Inger & Ramsbotham, 1997). However, extensional structures are not obvious (cf. Cannic, Mugnier & Lardeaux, 1995; Wheeler & Butler, 1994; Inger & Ramsbotham, 1997). The Briançonnais Zone is restacked under greenschist-facies conditions by east–southeast directed faults, folds and shears which repeat tectonic units. The largest of these, the Entrelor Thrust, carries the eastern Briançonnais onto the structurally higher Zermatt-Saas Zone and is best explained as a backthrust (Butler & Freeman, 1996).

Traditionally, the greenschist-facies metamorphic overprint of the western and central Alps is related to the Lepontine thermal dome of Switzerland, apparently dated at about 38 Ma (Hunziker & Martinotti, 1984). There have been severe problems interpreting isotopic data related to greenschist-facies metamorphism in the Alps, particularly assessing equilibrium and the effects of fluid ingress. In earlier studies related to the theme of our contribution, polyphase recrystallization histories were established from local sites, linked to penetrative deformation. The earliest greenschist-facies ages from the northwest Alps all cluster at about 38–39 Ma (Inger & Ramsbotham, 1997; Freeman *et al.* 1997). However, younger ages are recorded in local shears as young as 34 Ma. The initial implication of these data are that rocks from the subduction system now found at outcrop within the Alps were already exhumed out of their deeply buried sites to reside in the upper crust by about 38 Ma. Subsequent deformation restacked these units.

There is little evidence within the Briançonnais Zone for large-scale and continuous faults and shears which operated at shallow crustal levels, specifically $< 300\,^{\circ}C$. Thus, differences in cooling age across the region, recorded by fission track analysis on zircons and apatites (Hurford, Hunziker & Stöckhert, 1991; Seward & Mancktelow, 1994), presumably reflect differential erosion through laterally heterogeneous thermal structures or are linked to greenschist-facies shear zones outside the sample site.

3. Geology of the Col du Petit St Bernard area

Our study focuses on the Basal Briançonnais Thrust (BBT) as exposed on the Franco-Italian border in the northwest Alps (Fig. 3). At outcrop the BBT dips 30–60° east but is subparallel to lithological variations and foliations in its footwall. We consider the dips to reflect post-kinematic tilting, presumably associated with uplift of the external basement massifs to the west. At the Col du Petit St Bernard the immediate footwall is the Petit St Bernard unit (PSB) (Elter, 1987), part of the Valais Zone and formed from a uniform carbonate graphitic schist. This unit is highly foliated for several hundred metres below the BBT (Fig. 3) and has a strong foliation defined by white mica, together with compositional banding defined by alternations of ‘calcite rich’ and ‘quartz-phyllosilicate rich’ layers. This foliation intensifies substantially towards the upper surface of the Petit St Bernard unit. A similar intensification of the mineral lineation up towards the BBT is also apparent. This lineation, defined by mineral aggregations of white mica, quartz and calcite, plunges east–southeast, down the dip of the foliation (Fig. 3). We consider these fabric elements to define the shear zone in the footwall to the BBT.

Although the calc-schists of the Petit St Bernard unit contain a single, intense, foliation, there are two linear fabrics. The most intense of these, as described above, is spatially related to the BBT. However, a separate lineation population exists which is sub-parallel to the foliation strike (Fig. 4). These northeast–southwest lineations occur on foliation planes that are different to, but coplanar with, the foliation planes containing the ESE–WNW lineations.

The foliation in the Petit St Bernard unit is largely subparallel with bedding, but where the foliation is discordant with the bedding, the cleavage dips at a higher angle to the east indicative of westerly vergence, possibly associated with a top-to-the-west overshear relative to stratigraphy. The shear sense on the BBT is also shown by minor fold vergence (Fig. 4) (anti-vergent limbless folds are common) and minor shear band generation, which offsets the main foliation. Kinematic indicators consistently show a top-to-the-west overshear in S-tectonites close to the BBT associated with the ESE–WNW transport. No clear shear sense was seen for the northeast–southwest (strike parallel) lineations. Foliation intensity decreases into the footwall and west vergent folds occur, with fold hinges approximately perpendicular to the lineations, that is, the fold hinges trend north–south (Fig. 4). Structurally below this the foliation gradually decreases in intensity, the interlimb angle of the folds increases and vergence direction is lost. As this

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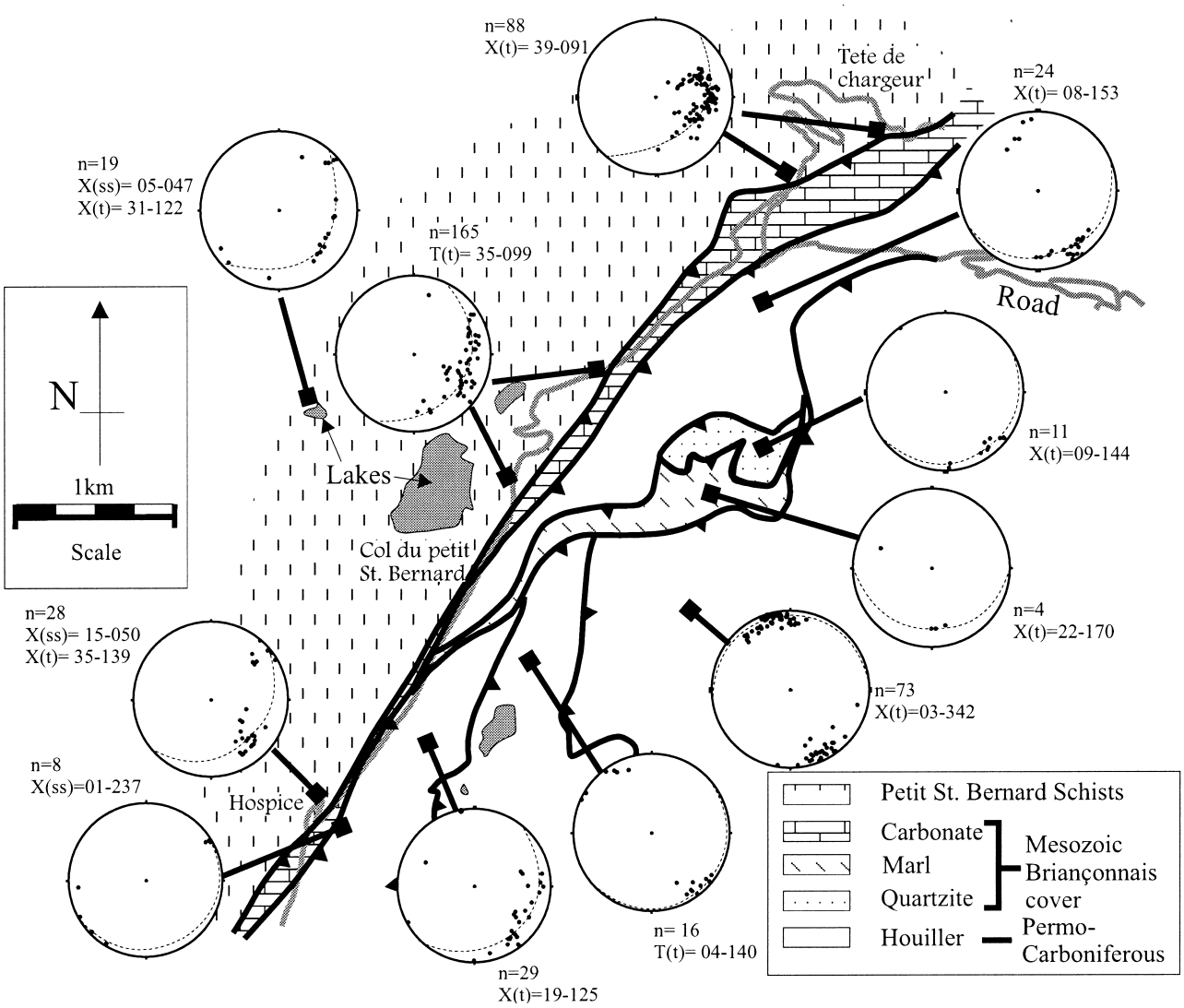


Figure 3. Geological map of the Col du Petit St Bernard region, Franco-Italian Alps. Key is inset. The BBT shear zone is predominantly in the direct footwall (western side) to the western most thrust surface. Map shows mineral elongation lineations (quartz, calcite and white mica stringers) from sub-areas within the Col du Petit St Bernard area. The data in the Basal Briançonnais Thrust hanging wall is shown for the individual thrust sheets. Representative foliation planes have been added to the stereonets. X(ss) = lineation cluster associated with strike parallel movement; X(t) = lineation cluster associated with dip parallel movement.

occurs the folds increase in wavelength from centimetre–decimetre scale to 0.5–1 m. At approximately 100 m structurally below the BBT the degree of deformation has decreased substantially, the foliation is weak, folds are open warps, and there is very limited lineation or shear band development.

The hanging wall of the BBT is composed of a series of tectonic sheets, each 50–500 m thick (Figs 3, 5). The Briançonnais basement (Houiller) sheets are composed of granitic intrusives. The Mesozoic cover sheets are formed from marl, dolomite-calc-schist and quartzite. All these sheets are separated by greenschist-facies ductile shear zones, in places decorated with cargneule. The tectonic contacts appear to duplicate the Briançonnais stratigraphy, therefore in the simplest form they represent thrust sheets. This stack of tectonic sheets overlies a carbonate sequence of variable thickness. The majority of the deformation in this zone was ductile with an intense foliation generated in places which contained suitably impure lithologies, that is, sections around Tete de Chargeur and at the Col du Petit St Bernard.

Kinematic data are provided by lineations on the foliation, defined by quartz and feldspar aggregates (Fig. 3). Lineation directions are consistent within any individual sheet, with an overall trend NNW–SSE, but vary significantly (up to 40°) between the individual tectonic sheets. This may indicate that the far field stress directions varied during this deformation phase or that strain in individual tectonic sheets was partially controlled by local as well as far field effects. Alternatively this may be recording rigid body rotations during shearing. The shear zones that separate the thrust sheets all show very well-developed S–C fabrics (as defined by Lister & Snoke (1984)) which consistently show an overshear direction top-to-the-north-

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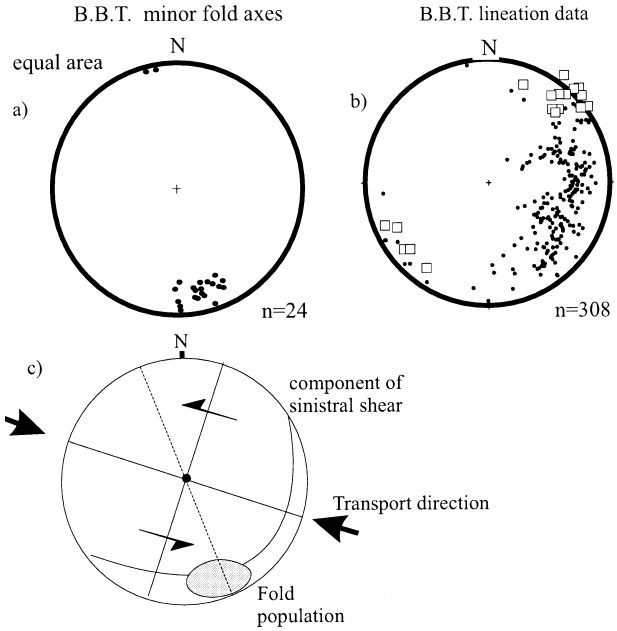


Figure 4. Stereonets of linear data from the BBT shear zone in the upper Petit St Bernard unit. (a) Fold hinges and (b) grouped mineral lineation data. Squares represent strike-parallel lineations. The fold population is counterclockwise of the expected position of formation (c).

west/north–northwest (Fig. 6). This deformation phase deforms and imbricates Mesozoic strata, so must be Alpine. The structural findings indicate that the deformation event recorded within the Briançonnais was a greenschist-facies, top-to-the-north–northwest overshear which was characterized by thin-skinned tectonics. This created ductile imbrication of cover and basement on a very local scale.

The lineation data show a disparity in movement direction between the basement–cover stack and the BBT shear zone (Figs 3, 6). The BBT movement direction was top-to-the-west–northwest compared to north–northwest for the previously imbricated hanging wall. In the structurally lower hanging wall thrust sheets closest to the BBT, the shearing on the BBT has reworked and reoriented the early northwest fabric generating both WNW–ESE and northeast–southwest fabrics (Figs 3, 6). On a map scale the timing of the deformation within the BBT hanging wall and the BBT shear zone is again clear. The hanging wall thrust sheets are truncated by the BBT (Fig. 6), with the BBT cutting up hanging wall tectonic stratigraphy towards the southwest. Thus the BBT shear zone developed in response to westerly overshear seen in the PSB schists which occurred after the north–northwest directed stacking of the Briançonnais Domain. The present BBT must therefore have propagated from within the Briançonnais units, cutting through and carrying the original basal décollement.

From field data it is clear that the units record a two stage deformation history. Initially there was a greenschist-facies north–northwest directed overshear, preserved within the Briançonnais units. This early stack was then truncated and carried passively on the greenschist-

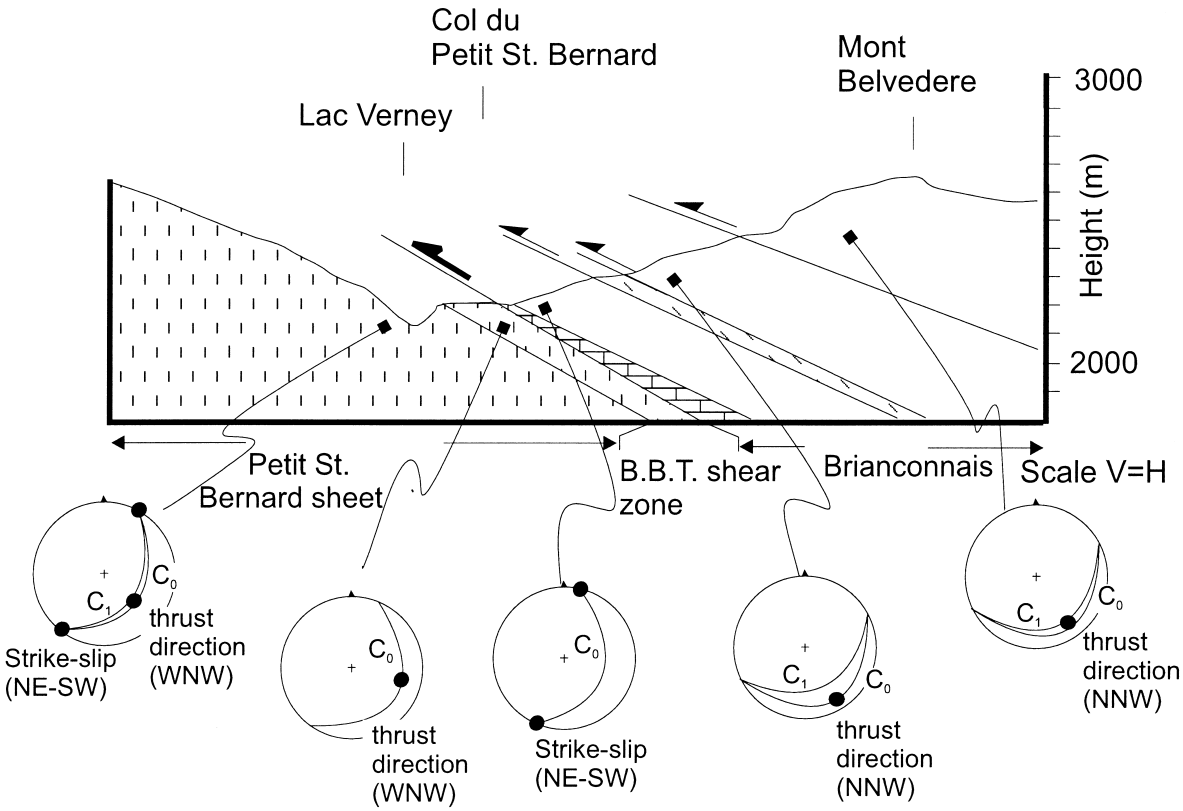


Figure 5. Cross-section through the Col du Petit St Bernard area. Section is drawn along the transport direction of the Basal Briançonnais Thrust. Stereonets show representative mineral lineations, primary and secondary fabrics. Shear direction as indicated by minor structures within the fault zones are shown.

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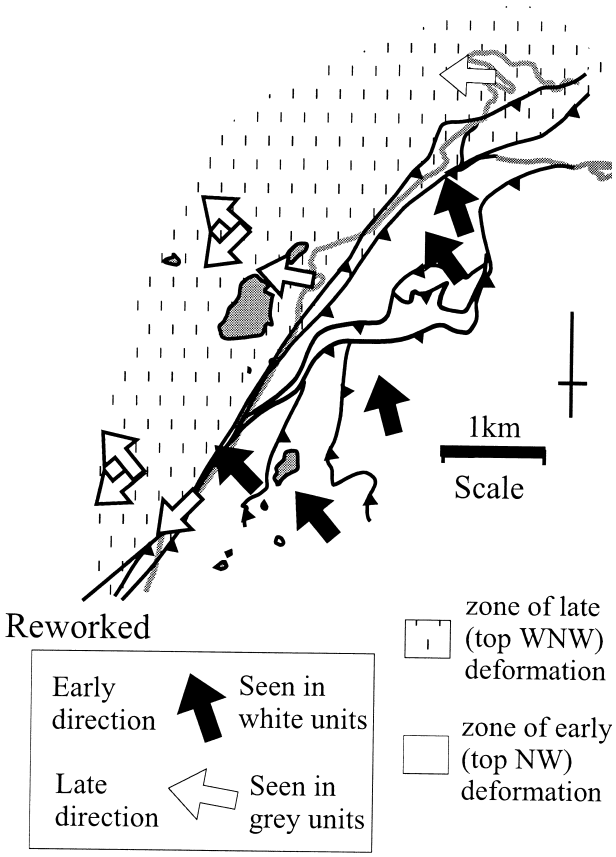


Figure 6. An interpretative map of the Col du Petit St Bernard area showing the variation in tectonic transport of the different thrust sheets as indicated by the mineral lineations. The transport directions of the first event (recorded in the Briançonnais basement units) are shown as black arrows. The transport directions of the later shearing, associated with the Basal Briançonnais Thrust, are shown as white arrows.

facies Basal Briançonnais Thrust over the foreland Valais basin sediments to the west–northwest, imposing ductile shear fabrics on the underlying Petit St Bernard sediments.

3.a. Geochronology and metamorphism

Deformation can be dated by isotopic analysis of minerals which underwent dynamic recystallization below their closure temperature during shear fabric formation (Freeman *et al.* 1997). In this study the most useful system is Rb–Sr in white mica because the rock was deformed below, and has remained below, the closure temperature of approximately 550 °C (Cliff, 1993). The implications of the white mica isotopic analyses depend upon:

(1) the proportion of different mica populations;

(2) the deformation and recrystallization mechanism of the white micas;

(3) the temperature of mica formation; and

(4) the temperature evolution of the micas since formation.

These factors can be assessed by detailed study of mineral chemistry and textures.

3.a.1. Metamorphic and crystallization conditions

Within the basement rocks in the basement–cover stack (top-to-the-north–northwest) the mineralogical variation is limited. The main assemblage is quartz–feldspar (albite)–phengite ± calcite ± chlorite ± biotite ± epidote. The assemblage quartz–feldspar–phengite appears to relate to an early fabric-forming event, overprinted to varying degrees along discrete shear planes by finer-grained biotites, chlorites and fine-grained white mica. The fine-grained intergrowths of chlorite and biotite formed along the shear planes are typical of lower greenschist-facies conditions.

Following the formation of the deformation fabric a limited static recrystallization created very fine-grained, randomly orientated clots of white mica characteristic of low temperature fluid infiltration.

By contrast with the basement units, the Briançonnais cover units can only have been deformed and metamorphosed in the Alpine event, therefore representing the best sources of information about the early north–northwest-directed Alpine event because they lack any earlier metamorphic or igneous histories. Two of the cover sheets are mono-mineralic (a quartzite and a marble) and therefore of limited use, while the third cover sheet, directly above the BBT, is formed from a micaceous carbonate that is selectively reworked by the BBT (Fig. 3).

The mineralogy of the Mesozoic carbonate schist unit is calcite + quartz + albite + chlorite + white mica (paragonite + phengite), placing an upper temperature limit of 400–550 °C due to the breakdown of paragonite + calcite + quartz to albite + clinozoisite + CO_2 + H_2O at these temperatures (Chatterjee, 1972). The lack of garnet, amphibole and chloritoid also indicates temperatures less than 480 °C at pressures of ~5 kbar (Will *et al.* 1990). A pressure estimate can be inferred from the lack of jadeite, giving an upper pressure limit of ~10 kbar. The quartz-rich bands and the calcite domains both show crystal-plastic deformation structures, indicating temperatures in excess of ~300 °C. The formation of the north–northwest directed shear fabric in the Briançonnais material was therefore approximately 300–450 °C below ~10 kbar.

Within the BBT shear zone (top-to-the-west–northwest event) the mineral assemblage defining the deformation fabrics is the same as that of the north–northwest shear fabric observed in the Briançonnais carbonate schist unit. The same *P–T* limits therefore apply, that is, 300–450 °C at less than ~10 kbars.

All the evidence suggests assemblages and fabrics formed below the Sr closure temperature in white mica. Micas which have not undergone recrystallization during subsequent deformation will have preserved their isotopic signatures, while micas crystallizing during the deformation events should yield the age of fabric formation.

Mineral characteristics preserved from a number of growth periods during a rock's evolution, as commonly seen in white mica fabrics (Dempster, 1992), are important influences on the isotopic record. Such mineral populations are most easily characterized chemically by

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electron-optic and microbeam techniques which allow chemical compositions to be related to textural data and isotopic analyses of bulk separates.

The white mica fabric within the Houiller units (Briançonnais basement) is complex in detail. Three mica populations can be distinguished via their atomic number contrast and morphology (Fig. 7). The earliest population is a large (~200–300 microns) light-coloured (high mean atomic number) population. This is overgrown by a darker (lower mean atomic number) large white mica population (again ~200–300 microns). These white micas have since been deformed under a brittle regime. The third mica population displays a decussate texture, therefore this mica population crystallized post-shearing.

Major element analyses of the micas yields some very interesting results (Table 1). Both the large early micas show extremely high Ti contents. As the Briançonnais basement samples all contain coexisting rutile, suggesting that the rock was saturated with respect to Ti during crystallization, the values indicate temperatures in excess of possible metamorphic phengite (in excess of ~650 °C) using the data of Guidotti (1978) (Fig. 8). This indicates that the two large early mica populations retain an original igneous chemical signature. Isotopic analyses of these micas should therefore yield cooling ages, not deformation ages.

The third, fine-grained, metamorphic phengite population shows a Ti content which relates to upper greenschist-facies conditions, much higher than that predicted by the co-existing fine-grained biotite and chlorite intergrowths. The data of Guidotti (1978) may yield temperatures 50–100 °C higher than reality due to crystal chemical variations.

Since the units record no history of exceeding the closure temperature of strontium in white mica following magmatic crystallization, Rb–Sr dating of the igneous micas should yield information relating to the time of crystallization and cooling of the igneous protolith. Dating of the decussate micas should yield information about the timing of the post-deformation, metamorphic event.

Within the Briançonnais carbonate schist unit (top-to-the-north–northwest) microprobe (Table 1) and backscattered SEM analyses indicate that coexisting phengite and paragonite were initially stable with each other (Fig. 9). A later mica crystallization period generated small micas (~10 microns). These micas occur as minor overgrowths on the original population. The early large crystals have a relatively high Si value of approximately 3.40–3.45 pfu (per formula unit), whereas the later overgrowths show lower Si contents, and higher alkali contents (Fig. 10). This pattern can be interpreted as recording either a drop in pressure or an increase in temperature (Massonne & Schreyer, 1987).

The morphological data combined with the structural data indicate that the early mica population must have formed during some period of enhanced diffusion which was prior to the BBT shear event, but during the Alpine

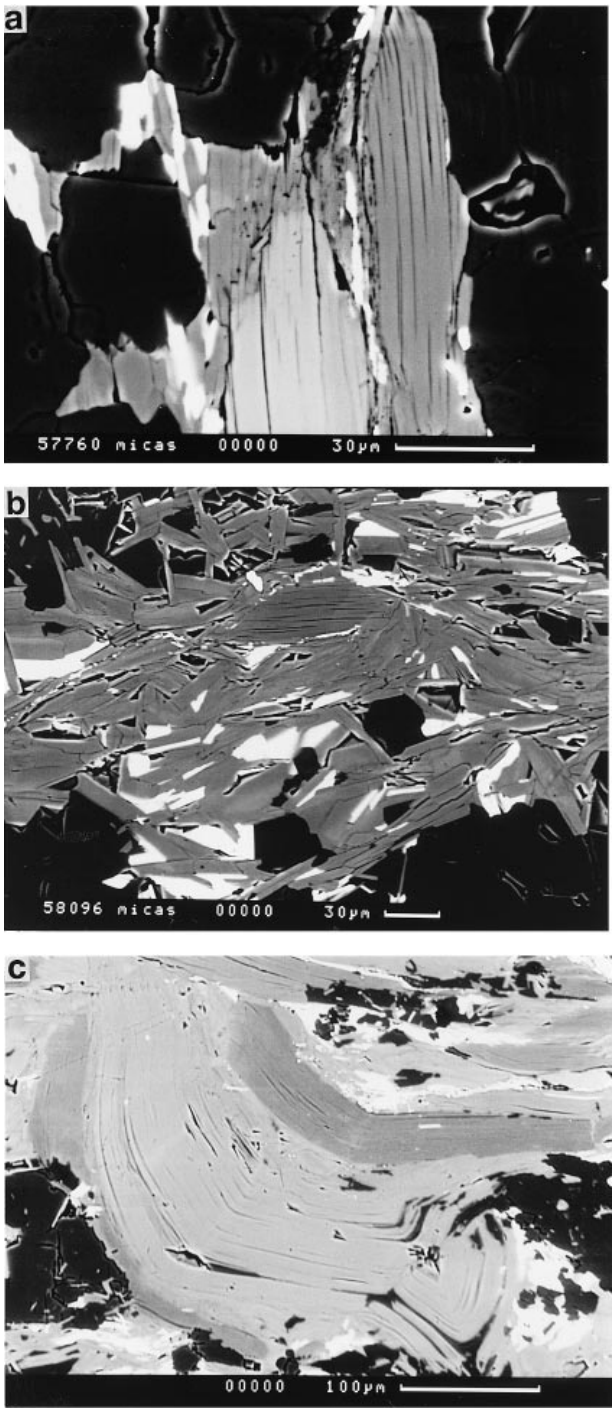


Figure 7. Backscattered electron images of large early phengites seen in the Briançonnais basement units (Houiller). (a) Original light-coloured (high mean atomic number) phengites are overgrown by later darker phengites (lower mean atomic number). (b) Dark phengites are subsequently overgrown by decussate fine-grained white micas. (c) The two early mica populations are deformed whereas the fine-grained population is unaffected.

event. This was most likely the north–northwest directed overshearing. Dating of the early high Si, low Na/Na+K mica population is therefore likely to yield the timing of the early north–northwest directed overshearing event.

The phengites from within the BBT shear zone display very uniform compositions (Fig. 11, Table 1). The white

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Table 1. White mica probe analyses of phengites from the Col du Petit St Bernard

SiO_2	TiO_2	Al_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	F	Total	O=F	Total	Si	Ti	Al	Fe^2	Mg	Na	Ca	K	OH	F	Total	Fe#
57760 – Briançonnais basement																								
47.8	0.69	32.26	1.62	1.56	0	0.76	10.19	4.45	0	99.38	0	99.4	6.39	0.07	5.08	0.18	0.31	0.20	0	1.74	3.9	0	17.97	36.8 (L.L)
46.9	0.58	33.66	1.34	1.11	0.09	1.41	8.66	4.41	0	98.33	0.1	98.3	6.29	0.06	5.32	0.15	0.22	0.37	0	1.48	3.9	0	17.92	41.4 (L.D.)
48.0	0.22	33.77	1.5	0.96	0.04	0.55	10.05	4.4	0	99.79	0.1	99.7	6.37	0.02	5.28	0.17	0.19	0.14	0	1.70	3.9	0	17.89	47.8 (F. dec)
57566 – Briançonnais basement																								
47.6	0.93	32.76	2.15	1.42	0	0.4	10.41	4.49	0	100.0	0	100.0	6.33	0.09	5.14	0.24	0.2	0.10	0	1.77	4.0	0	17.95	46.2 (L.L)
47.1	0.51	35.57	0.95	0.62	0	1.15	8.92	4.42	0	99.6	0.1	99.5	6.23	0.05	5.54	0.11	0.12	0.29	0	1.50	3.9	0	17.85	47.1 (L.D.)
47.8	0.13	3.08	1.67	1.36	0	0.24	10.27	4.47	0	99.02	0	99.0	6.40	0.01	5.22	0.19	0.27	0.06	0	1.75	4.0	0	17.89	40. 8 (F. dec)
C17 phengites – Briançonnais cover unit																								
52.7	0.15	27.24	0.52	4.2	0.34	0.26	10.4	4.25	0.68	100.8	0.29	100.5	6.91	0.02	4.21	0.05	0.82	0.06	0.04	1.74	3.7	0.28	17.87	7.6
50.2	0.31	30.08	0.19	3.25	0.46	0.45	10.54	4.22	0.69	100.4	0.29	100.1	6.62	0.03	4.68	0.02	0.64	0.06	0.12	1.77	3.7	0.29	17.95	3.9
51.6	0.23	28.93	0.28	3.79	0.43	0.31	10.16	4.32	0.55	100.6	0.23	100.4	6.77	0.02	4.47	0.03	0.74	0.06	0.08	1.70	3.8	0.23	17.86	4
52.1	0.18	28.51	0.47	3.85	0.44	0.2	10.42	4.34	0.52	101	0.22	100.7	6.81	0.02	4.40	0.05	0.75	0.06	0.05	1.74	3.8	0.22	17.87	6. 6
C17 paragonite – Briançonnais cover unit																								
47.7	0.04	37.66	0.11	0.73	0.44	6.28	1.24	4.55	0.19	98.95	0.08	98.86	6.15	0.00	5.73	0.01	0.14	1.57	0.06	0.20	3.92	0.08	17.86	8.1
C7 phengites – BBT shear zone																								
53.0	0.08	32.83	0.81	2.68	0.05	0.26	7.2	4.35	0.79	102.1	0.33	101.8	6.71	0.01	4.90	0.09	0.51	0.06	0.01	1.16	3.7	0.32	17.44	14.5
51.0	0.12	33.58	0.63	2.13	0.16	0.25	7.09	4.43	0.44	99.83	0.19	99.64	6.59	0.01	5.12	0.0	0.41	0.06	0.02	1.17	3.8	0.18	17.45	14.4
53.3	0.09	31.72	0.75	2.78	0.07	0.14	7.24	4.47	0.48	101.1	0.2	100.9	6.80	0.01	4.77	0.08	0.53	0.04	0.01	1.18	3.8	0.19	17.41	14.4
52.7	0.13	31.86	0.69	2.73	0.05	0.16	7.26	4.43	0.5	100.5	0.21	100.2	6.76	0.01	4.82	0.07	0.52	0.04	0.01	1.19	3.8	0.20	17.42	12.4
C8 phengites – BBT shear zone																								
52.1	0.13	33.23	0.65	2.29	0.38	0.18	6.87	4.46	0.48	100.8	0.2	100.6	6.66	0.01	5.01	0.07	0.44	0.04	0.05	1.12	3.8	0.19	17.40	13.7
51.2	0.09	33.5	0.68	2.11	0.11	0.14	7.41	4.35	0.62	100.2	0.26	99.97	6.61	0.01	5.10	0.07	0.41	0.04	0.02	1.22	3.7	0.25	17.46	15.3
51.5	0.06	33.9	0.5	2.29	0.45	0.28	6.98	4.47	0.46	101	0.19	100.8	6.58	0.01	5.11	0.05	0.44	0.07	0.06	1.14	3.8	0.19	17.46	12.8
52.4	0.12	33.7	0.47	2.29	0.2	0.16	7.32	4.37	0.75	101.7	0.32	101.4	6.65	0.01	5.04	0.05	0.43	0.04	0.03	1.19	3.7	0.30	17.43	10.4

The table shows analyses from the Briançonnais basement units (57760 and 57566), the Briançonnais cover unit (C17), and the Petit St Bernard unit (C7 and C8). For the Briançonnais basement micas L.L. – large light (high mean atomic number); L.D. – large dark (lower mean atomic number); F. dec – Fine decussate.

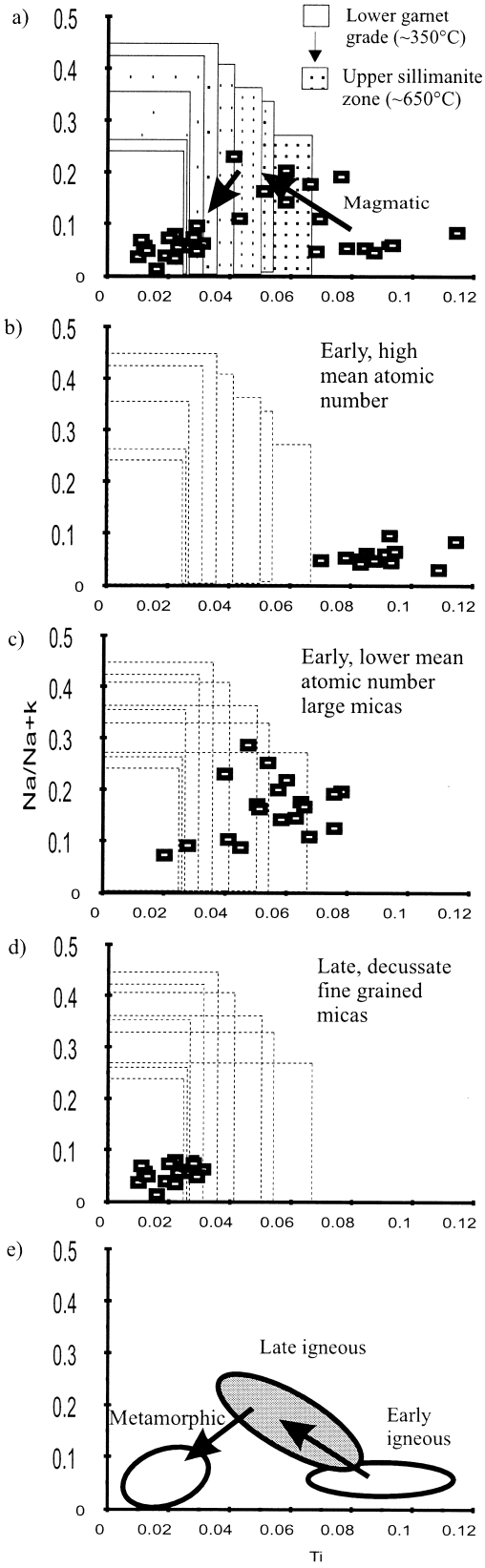


Figure 8. Alkali vs. Ti plots of the phengites contained within the Houiller basement units. The graphs indicate very high formation conditions (> 650 °C) for both the large early phengite populations seen. The late, fine, decussate phengites indicate temperatures of equilibration of approximately mid-greenschist-facies conditions. (a) All micas analysed; (b) light, large early phengites; (c) early dark micas; (d) decussate micas; (e) interpretation of mica evolution.

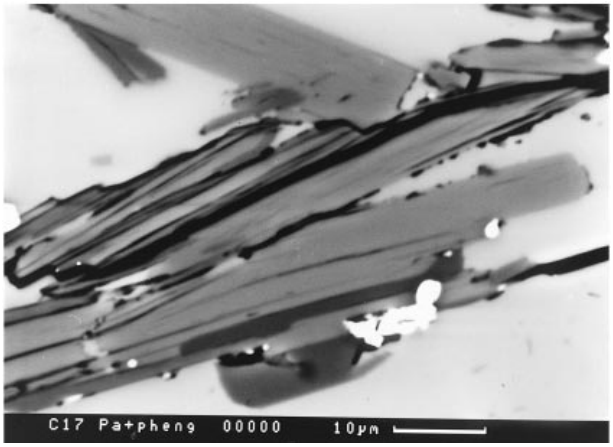


Figure 9. Backscattered electron image of white micas sitting in calcite. The white micas are composed of coexisting phengite (mid-grey) and paragonite (black). Late small-scale overgrowths of white mica on the larger primary mica rims occur.

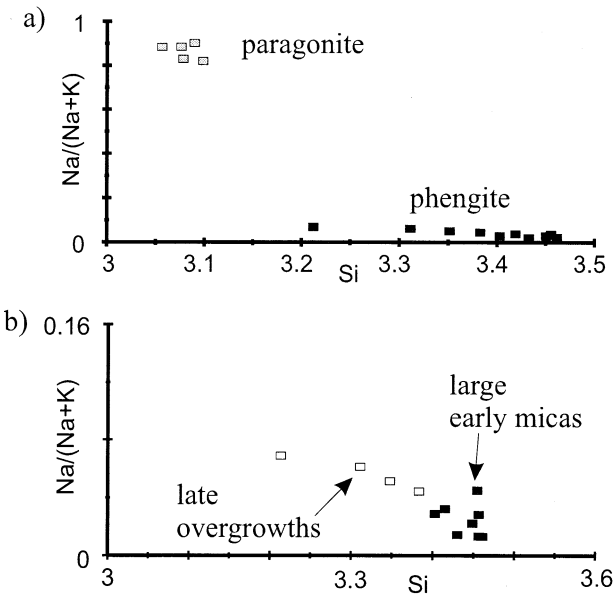


Figure 10. (a, b) White mica solid solution plots of co-existing phengite and paragonite of the sample shown in Figure 9. The sample is from the Mesozoic carbonate schist unit (see Fig. 3). (a) The plot shows the range of compositions of phengite (black squares) and paragonite (grey squares). (b) This plot shows the relationship between the phengites morphologies and compositions. The large foliated early micas are shown as black squares and the later fine overgrowths are shown as open squares.

micas are comprised of coexisting paragonite and phengite in several samples, and just phengite in others. All the samples show very tight compositional groupings, indicating equilibrium was approached during one single crystallization event (Fig. 12). These micas define the mylonitic fabric related to the BBT. No other mica population exists within these samples. These micas most probably formed during the shearing associated with the BBT, thus Rb–Sr analyses of these micas will yield the time since the BBT shearing ceased to crystallize or recrystallize white micas.

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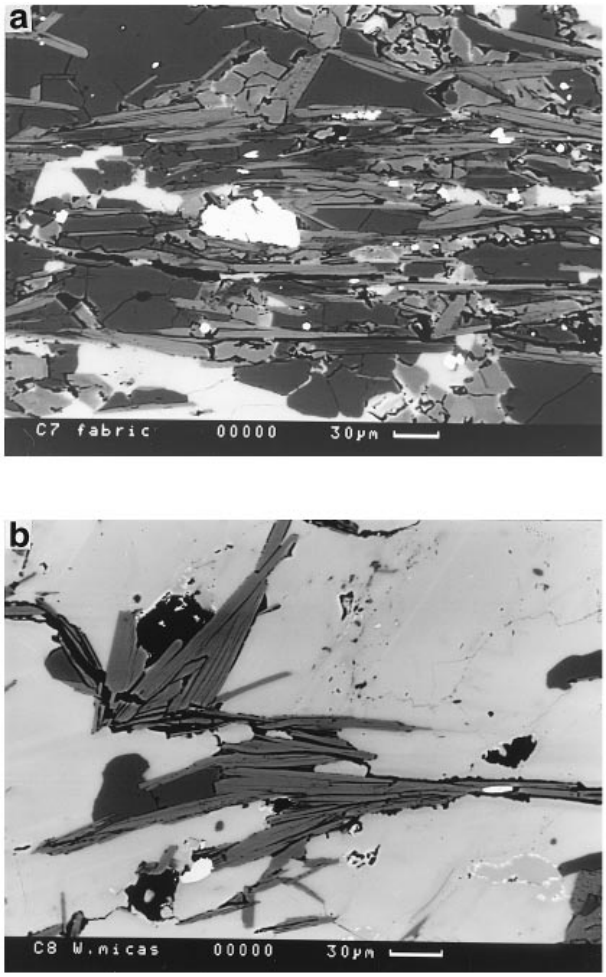


Figure 11. (a, b) Backscattered electron images of samples from within the Basal Briançonnais Thrust shear zone. The white micas show very uniform compositions. The white micas define the foliation within these high strain rocks.

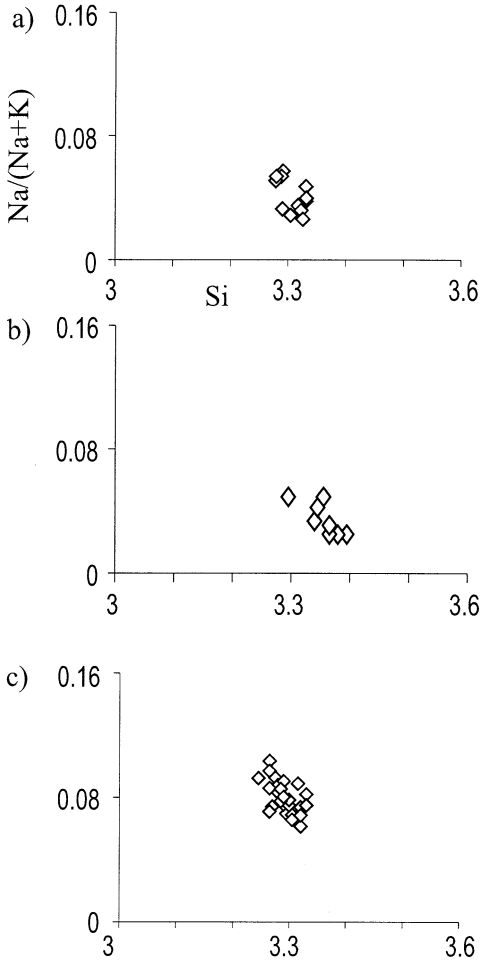


Figure 12. (a, b, c) White mica solid solution plots for the phengites taken from within the Basal Briançonnais shear zone. All samples show uniform compositions with tight clustering on the diagrams. All the samples also show very similar silica contents, indicative of formation at similar *P–T* conditions.

3.a.2. Rb–Sr white mica dating

Ages were obtained by analysis of coexisting phases defining 'two-point isochrons'. In addition to the white mica separate, a mineral was selected that most probably had the same Sr isotopic composition at the time of mica crystallization, that is, one that would have equilibrated with the same isotopic reservoir. In the quartzo-feldspathic basement units the feldspars are the major Sr reservoir, while for the carbonates the calcite is the major Sr reservoir. The feldspar and the calcite are likely therefore to have controlled the grain boundary fluid Sr composition which in turn would have controlled the Sr composition of the crystallizing micas. The ages were calculated using the Isoplot program of Ludwig (1990). With a knowledge of the different mica populations and the origin of these populations, the data yielded by isotopic analyses of the micas can be interpreted in terms of different tectonic events.

The dating technique used here is described fully in Freeman *et al.* (1997) as is the experimental procedure and the age determination method.

3.a.3. Geochronological results

Ten samples were analysed, the results from which are shown in Table 2 and Figure 13. The four samples which were formed from the micas relating exclusively to the BBT shear zone activity (that is, Si ~ 3.3) produce young ages, between 27 and 32 Ma. This range is interpreted as representing the age that white mica ceased to recrystallize within the BBT shear zone. Large errors in sample C15 result from the presence of paragonite in the white mica sample, but this has no significant effect on the age produced; it simply increases the analytical error by diluting the phengite analysis.

The likelihood of sustained deformation on the BBT is high because it is a major regional structure, thus the spread of ages seen in the geochronological results for the BBT and associated footwall may be assigned to continued growth and recrystallization of white mica during the lifetime of the structure (32–27 Ma).

The sample from the cover unit within the basement–cover stack contains the mica population relating to the initial north–northwest directed compression (that is,

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Table 2. Rb–Sr white mica analyses from the Col du Petit St Bernard

Sample	Rb (ppm)	Sr (ppm)	Rb/Sr ratio (ppm)	$^{87}Sr/^{86}Sr$	$^{87}Rb/^{86}Sr$	Age ± 2σ
BBT (WNW related)						
C7 mica	74.1983	59.0140	1.2573	0.70967	3.64	32.9 ± 0.97
C7 calcite	14.4091	806.712	0.0179	0.70799	0.05	
C8 mica	199.291	338.528	0.5887	0.70868	1.71	30.9 ± 1.9
C8 calcite	8.05133	1278.14	0.0063	0.70795	0.02	
C15 mica	64.6301	219.911	0.2939	0.70832	0.85	27.1 ± 3.9
C15 calcite	14.4682	1482.57	0.0098	0.70800	0.03	
C22 mica	171.844	56.706	3.0300	0.71198	8.77	31.1 ± 0.52
C22 calcite	6.61433	1215.35	0.0054	0.70811	0.02	
NNW related						
C17 mica	272.247	15.208	17.9016	0.74135	51.97	45.3 ± 0.54
C17 calcite	9.89719	512.00	0.0193	0.70795	0.06	
Igneous related						
C25 mica (60-90)	268.664	122.415	2.1947	0.74276	6.37	320 ± 5.1
C25 feldspar	44.6246	83.463	0.5347	0.72081	1.55	
C25 mica (120-150)	186.510	87.713	2.1264	0.73981	6.17	288 ± 4.7
C25 feldspar	44.6250	83.463	0.5347	0.72081	1.55	
C9 mica (60-90)	99.4622	56.733	1.7532	0.73643	5.09	7100 ± 6900
C9 feldspar	72.2627	41.992	1.721	0.72503	4.99	
C24 mica (60-90)	273.434	117.231	2.3324	0.74439	1.83	322 ± 5.4
C24 feldspar	46.5092	73.459	0.6331	0.72176	6.77	
C1 mica (60-90)	276.481	139.664	1.9796	0.74214	5.75	187.7 ± 2.5
C1 feldspar	26.5944	192.936	0.1378	0.72787	0.40	
C1 mica (120-150)	268.728	130.309	2.0622	0.74321	5.99	193 ± 2.5
C1 feldspar	26.5944	192.936	0.1378	0.72787	0.40	
58096 mica (60-90)	209.804	101.612	2.0648	0.73913	5.99	324 ± 7.1
58096 feldspar	41.1935	47.605	0.8653	0.72307	2.51	

Where (60–90) or (120–150) is shown this refers to mica size (mesh spacing per inch).

Si ~ 3.45). This yielded an age of 45 Ma. This single age is interpreted as representing the time that white mica ceased to recrystallize during the north–northwest overshear event.

The igneous micas extracted from the basement in the basement–cover stack produced ages of ~320 Ma. A single age between 320 Ma and 45 Ma in sample C1 probably represents contamination of the igneous micas with the later metamorphic micas, producing a mixed age. For most of the basement samples very fine-grained mica separates which would have a much greater proportion of the late metamorphic micas were analysed to assess the likely contamination effect of these micas. For all these duplicate analyses the age produced was either within error or less than the large mica separate comprised dominantly of the igneous micas (Table 2). The shift towards a lower age was what would have been predicted, but the shift was generally insignificant in relation to the age difference between the igneous (~320 Ma) and the metamorphic micas (< 45 Ma). The consistency of the igneous mica ages (~320 Ma) in the basement is surprising considering the metamorphic and deformation history of the unit, and suggests that the Sr isotopic system remained undisturbed through subduction and then Alpine collision as indicated by the metamorphic studies.

4. Discussion

The two-stage compression pattern seen at the Col du Petit St Bernard has been recognized previously in the Briançonnais Domain, principally by Platt *et al.* (1989). In this and previous studies early NNW–SSE lineations were found to be overprinted by a very extensive east–west compression event. Figure 14 shows a compilation

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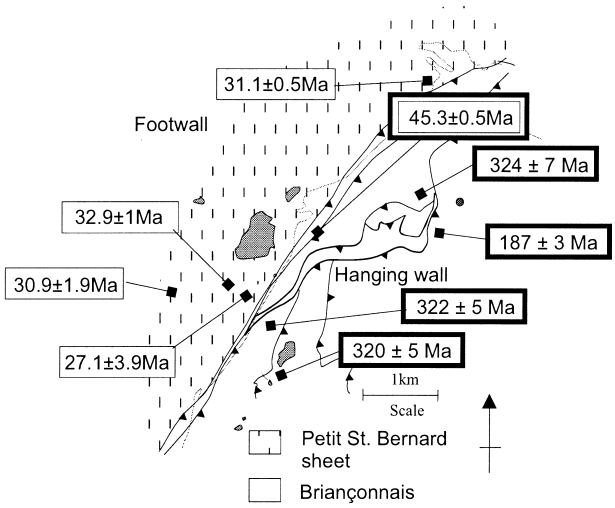


Figure 13. Rb–Sr phengite geochronological data from the Col du Petit St Bernard area. The white boxes with thin outlines are the results from the phengites which relate to shearing on the Basal Briançonnais Thrust; the double box is the result from the phengites which relate to the early north–northwest directed shearing; and the white boxes with thick rims are the results from the phengites which indicated an igneous origin.

of stereonets with pressure estimates from the deformation events that occurred in and around the central Briançonnais of the Western Alps. In previous studies the NNW–SSE lineations were found to be formed in part from rotated and recrystallized glaucophane (Fig. 14b, c), suggesting relatively high pressures for the formation of this fabric (9–12 kbar: Platt & Lister, 1985). The early north–northwest lineations seen in the Briançonnais units in the Col du Petit St Bernard area (Fig. 14a) appear to be analogous with the glaucophane lineations seen elsewhere in the Briançonnais Domain.

Pressure estimates for the later WNW–ESE forethrusting and backthrusting events elsewhere in the Briançonnais (Fig. 14d–h) are 4–8 kbar (Platt & Lister, 1985); this is within the bracket established for the main fabric forming event on the BBT at the Col du Petit St Bernard (< 10 kbar), and therefore it suggests that the BBT structure was probably formed in response to the same deformation forces that created the 'late' forethrusting and backthrusting seen elsewhere in the Briançonnais Domain.

It has been established that the eastern margin of the Briançonnais Domain was a major top-to-the-east–southeast backthrust accommodating at least 20 km of shortening (Dal Piaz, 1928; Butler & Freeman, 1996). This backthrust was dated using Rb–Sr in recrystallized white micas at 34 ± 1 Ma (Freeman *et al.* 1997). It has been suggested that the BBT may represent the basal décollement from which this backthrust propagated (Butler & Freeman, 1996). Lineations within the BBT shear zone and the Entrelor backthrust (Fig. 14 d,e) show a remarkable similarity in transport direction. From the kinematic studies of the BBT (that is, top-to-the-west–northwest) the hypothesis of Butler & Freeman (1996) that the backthrust system on the eastern margin of the Briançonnais links onto the BBT appears to be valid.

If the BBT does represent the basal décollement of the backthrust on the eastern margin of the Briançonnais Domain, then it may be possible to work out the rate of propagation of the deformation loci on the BBT if the Briançonnais can be restored to the time when the eastern backthrust (Entrelor/Nomenon backthrust) was active. Unfortunately the lack of both marker horizons and an accurate pre-deformation template makes this difficult.

Since the publication of the ECORS CROP Alps 1 seismic line (Nicolas *et al.* 1990) it has been suggested that the BBT and the Frontal Pennine Thrust (FPT) are out of sequence structures which cut through previously stacked European continental crust (Mugnier, Loubat & Cannic, 1993). This is because both of these structures appear to truncate major units supposedly juxtaposed in the Alpine orogeny. The inference from this is that the compressional event relating to full continental collision had already occurred in this region and that the FPT and the BBT were break-back structures (Butler, 1987). This interpretation is only valid if the units that the BBT and the FPT cut can be shown to have been juxtaposed during the Alpine continental collision, and not juxtaposed prior to the collision event. This model makes testable predictions about the age of other structures. The model states that structures within the foreland should be older than the FPT and BBT and structures in the hinterland should be significantly older.

The structural evidence suggests that the BBT represents a westerly propagating basal décollement for the forethrusts and backthrusts within the Briançonnais Domain. Dating of the most easterly, and structurally lowest structure (the Entrelor/Nomenon shear zone) has shown that the deformation was earlier in the hinterland (34 Ma, Entrelor/Nomenon) than in the foreland (31 Ma, BBT). Crystallization ages on the Diableret thrust, which is foreland of both the FPT and the BBT, records an Ar–Ar age of ~15 Ma (Crespo-Blanc *et al.* 1995). Structures further towards the foreland have been shown by reconstructing the syn-orogenic surface and thermal structure to have been active between 12–6 Ma (Butler, 1992b). The BBT is clearly an 'in sequence' structure in relation to the Alpine continent–continent compressional event (Fig. 15).

The BBT and FPT do cross-cut earlier juxtaposed contacts (Antoine, 1971) and slice through previously deformed units, but this earlier deformation can be ascribed to early movement within the subduction zone (Piedmont over Briançonnais) or closure of the sedimentary basins (folds within the Valais sediments).

The structural pattern observed in the Col du Petit St Bernard of an early north–northwest directed compression followed by a west–northwest directed compressional event appears to fit with the pattern observed throughout the central Briançonnais Domain by previous workers (e.g. Choukroune *et al.* 1986; Platt *et al.* 1989). The compressional event that duplicated stratigraphy in a north–northwest direction took place at ~45 Ma in this

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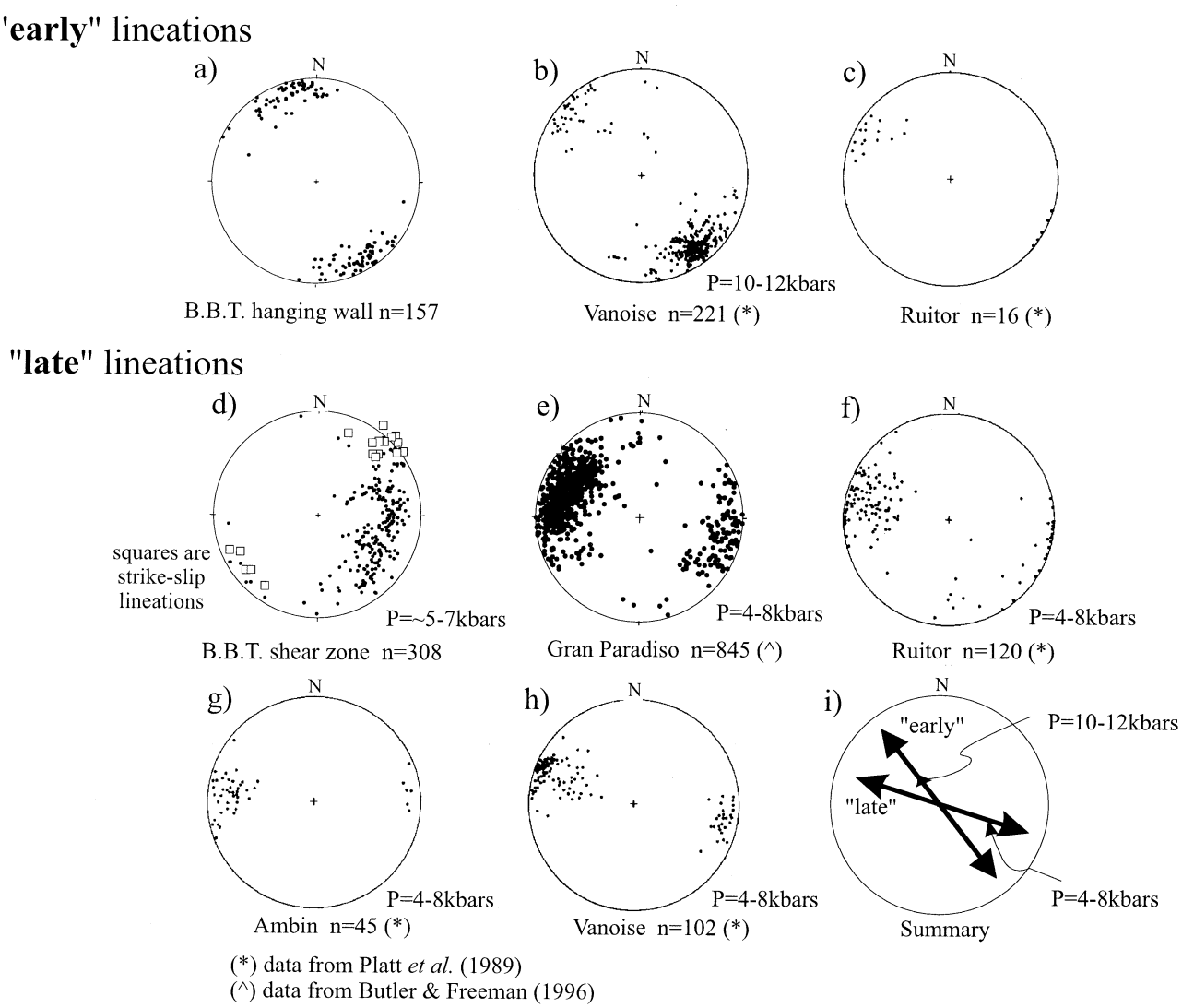


Figure 14. Compilation of mineral elongation lineation data from within and on the margins of the Briançonnais Domain. Pressure estimates for the mineral lineation formation are shown. The data show that an early southeast–northwest event is evident in many places within the Briançonnais Zone. Following this event there was a highly pervasive ESE–WNW event. Data from this study, Butler & Freeman (1996) and Platt *et al.* (1989).

area, although further analyses are required to confirm this. In the regional context this area is in a relatively external part of the Briançonnais Domain and is therefore likely to be one of the most westerly thrust stacks of the early compressional northwest directed event still exposed. If the deformation style was a foreland propagating piggyback thrust sequence, the basement–cover thrust stack at the Col du Petit St Bernard is one of the youngest thrust stacks of this early event preserved. The more internal NNW–SSE in-sequence structures should therefore be older than 45 Ma.

The end of the major compressive NNW–SSE directed event seen in the present exposure levels in the central Briançonnais was apparently recorded by the 45 Ma mica fabric age. The 34 Ma Entrelor event (Freeman *et al.* 1997) apparently dates the onset of the ESE–WNW directed deformation. Does this 10 million year age gap in deformation reflect a true tectonic hiatus? One possibility is that compressional tectonics continued during this period, but the whole of the Briançonnais Domain was being carried passively on a basal décollement whose location is unknown. The décollement relating to this stack must have been truncated by the present BBT, implying that the early basal thrust relating to the ~45 Ma compressional event must exist at depth internally (towards the ESE) from the present BBT and underlies this structure. Such a structure would be an obvious candidate to accommodate subduction-related strain between 45 and 34 Ma.

After the cessation of compressional tectonics in this area from ~45 Ma to 34 Ma, the compression was restarted when the relict high pressure metamorphic zone of the internal European massifs, Piedmont and Austroalpine system indented into the Briançonnais Domain at approximately 34 Ma (Butler & Freeman, 1996; Freeman *et al.* 1997). From then on the deformation propagated through the Briançonnais Domain from the east at 34 Ma to the west at 31 Ma.

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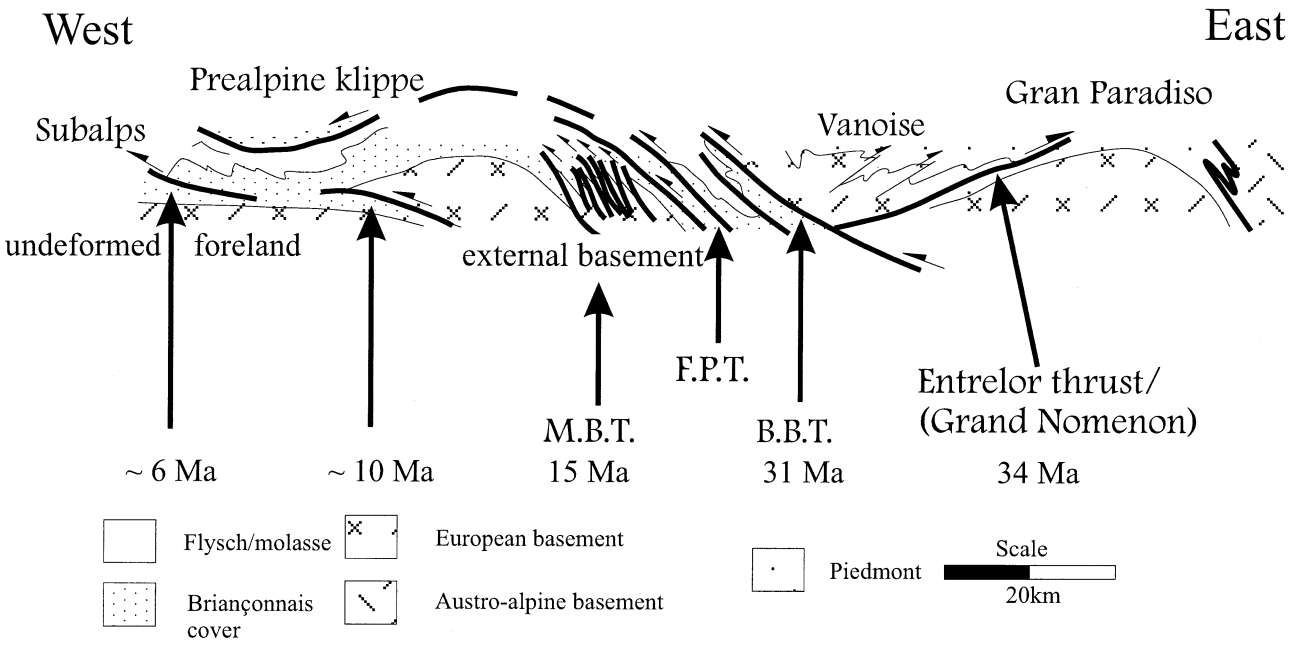


Figure 15. Ages of east–west overshear deformation structures in the western Alps. The section shows the migration of the deformation loci from the hinterland at 34 Ma, through the Briançonnais Domain at ~31 Ma, then out into the foreland between 31 and 15 Ma, then out into the sub-Alps after 15 Ma. The section shows that both the internal and the external zones acted as a bulk foreland propagating sequence. The deformation ages are from the following sources; BBT – this study; Entrelor thrust – Freeman *et al.* (1997); Diableret thrust – Crespo-Blanc (1995); Sub-Alps – Butler (1992). The section is after Butler (1986). B.B.T. – Basal Briançonnais Thrust; M.B.T. – Mount Blanc Thrust; F.P.T. – Frontal Pennine Thrust.

The sequence of tectonic events established from this study is very similar to those associated with the hanging wall of the Basal Briançonnais Thrust in the Swiss Central Alps (Marquer *et al.* 1994), although in the Central Alps the temperatures and pressures were somewhat higher. If the Central Alps are recording the same tectonic history at greater depths, there may have been a major lateral ramp in the BBT climbing up through tectono-stratigraphic levels towards the southwest, or the central Alpine Briançonnais may have been buried to deeper levels under the Austro-alpine thrust. In any case, the structural data suggest that prior to ~30 Ma the Briançonnais Domain of the Western and Central Alps enjoyed a similar tectonic history.

Our data may help to constrain the transition between subduction, which produced the Eocene high- and ultra-high-pressure metamorphism in the Alps, and true continental collision that imbricated the European foreland. Fission track dating within the Briançonnais and surrounding zones shows that overall there is a significant decrease in zircon fission track (f.t.) ages from east to west (Hurford, Hunziker & Stöckhert, 1991; Seward & Manktelow, 1994; Fig. 16). A zircon f.t. age of 30 Ma for the Gran Paradiso (Hurford, Hunziker & Stöckhert, 1991) implies that while the Gran Paradiso was being carried passively in the hanging wall of the active BBT it underwent cooling from approximately 400 °C at 34 Ma (Freeman *et al.* 1997) to 225 °C at 30 Ma, presumably related to extensive exhumation. Hurford, Hunziker & Stöckhert (1991) also report a zircon f.t. age of 26 Ma, and an apatite age of 12 Ma from Courmayeur (Fig. 16). Courmayeur is further up transport direction and lower within the hanging wall of the BBT than the Gran Paradiso Massif, therefore it appears that extensive denudation of all the BBT hanging wall was occurring during the thrusting on the BBT. This period of rapid exhumation also fits with fission track data from the Sesia Lanzo zone (higher within the BBT hanging wall), which shows a drop from ~225 °C to 100 °C between 33 Ma to 26 Ma.

A time–temperature path of the Sesia, Gran Paradiso, Courmayeur and the Col du Petit St Bernard shows that units in the BBT's immediate hanging wall all cooled at a rate of around 40 °C per million years during the activity of the BBT (Fig. 17). The more easterly and higher levels of the hanging wall (e.g. the Sesia-Lanzo zone), although undergoing an increased rate of exhumation during the activity of the BBT (see Hurford, Flisch & Jäger, 1989), did not undergo such a great amount of cooling compared to more westerly areas lower within the sheet.

Previous work suggests that there was an increase in the exhumation rate since ~35 Ma (e.g. Hurford, Flisch & Jäger, 1989). If the rate of exhumation is assumed to be primarily controlled by the surface topography, then an increased rate of exhumation implies an increased topographic height. To form such an increased topography either the lateral forces which are compressing the mountain chain need to change, or the buoyancy of the mountain chain needs to alter. Overthrusting of Adria has been occurring since the Cretaceous; plate convergence models do not suggest a great change in plate convergence since ~35 Ma (Dercourt *et al.* 1986; Dewey & Helman,

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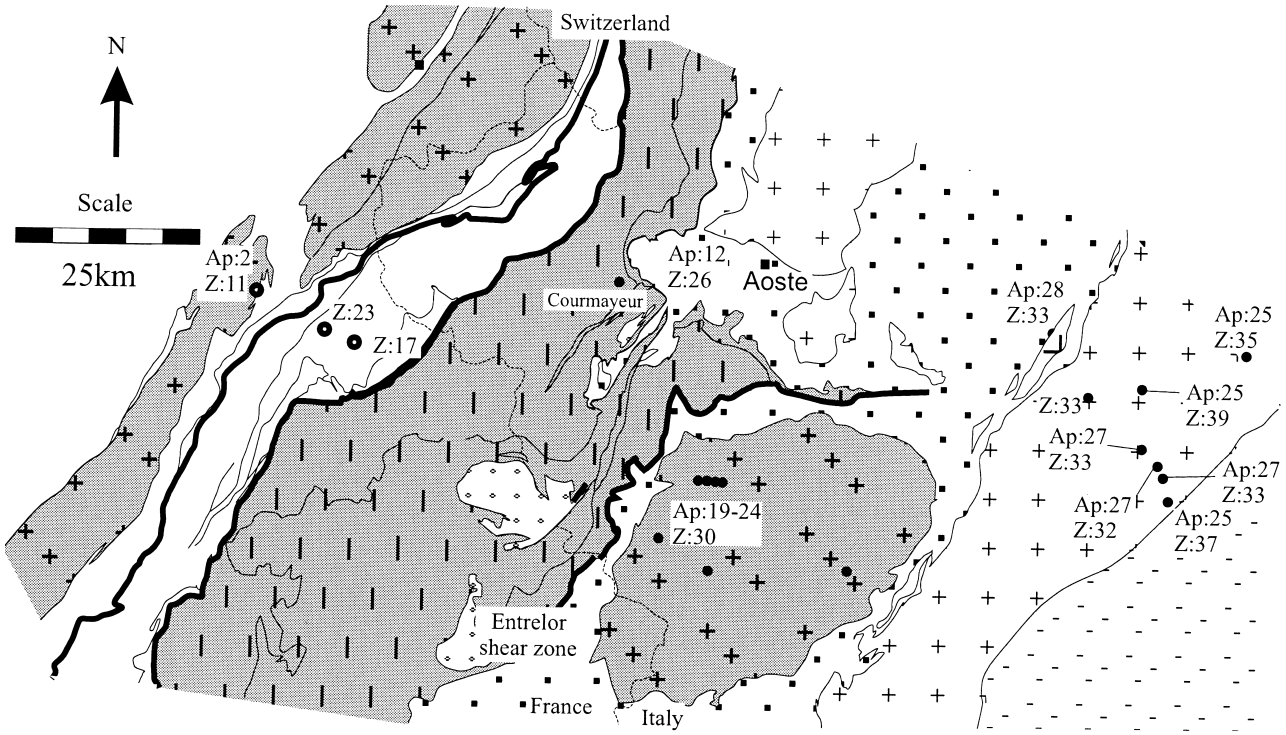


Figure 16. Fission track data of zircon and apatite from the internal western Alps. The data show an overall decrease in annealing ages from the west towards the east. The data is from Hurford, Flisch & Jäger (1989) and Seward & Mancktelow (1994).

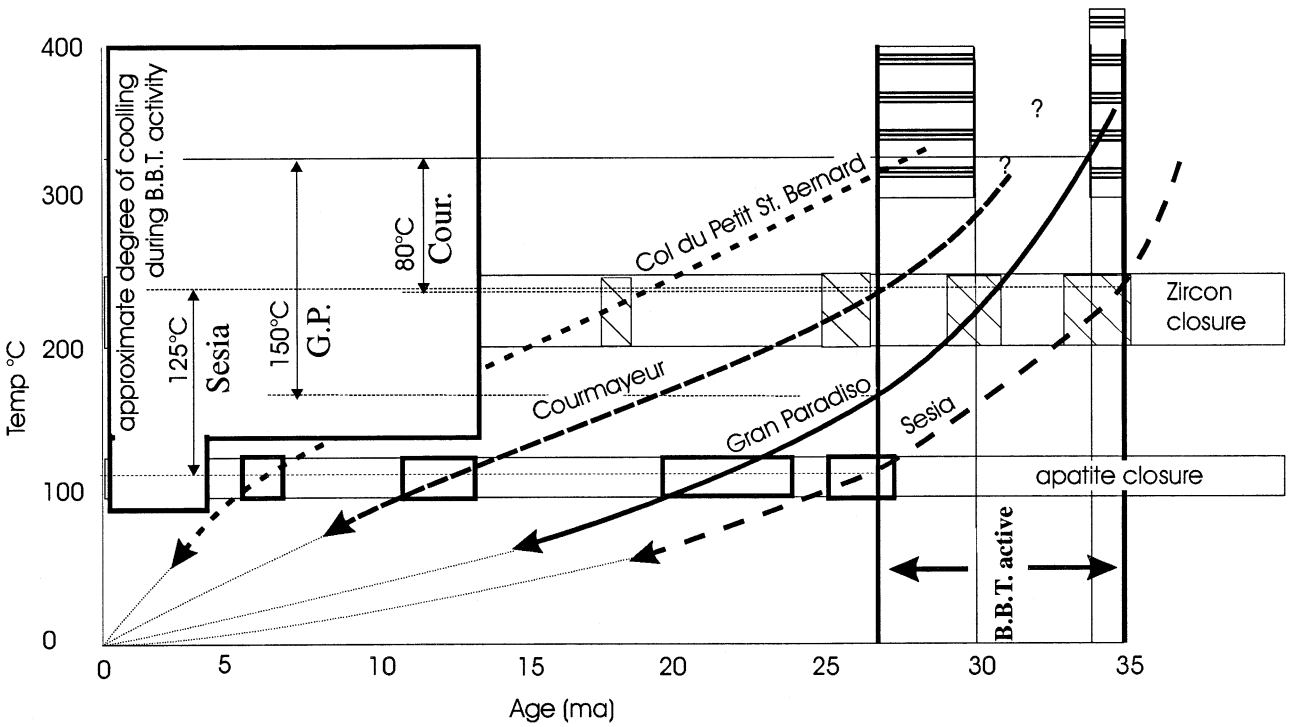


Figure 17. Time–temperature paths for various localities within the internal western Alps. The data show a rapid decrease in temperature of the hanging wall of the Basal Briançonnais Thrust while the Basal Briançonnais Thrust was propagating through the Briançonnais Zone. The data shows that during the propagation of the Basal Briançonnais Thrust the hanging wall reduced in temperature by ~ 100 °C. Data is from this study, Freeman *et al.* (1997), Seward & Mancktelow (1994) and Hurford, Flisch & Jäger (1989)

1988). This suggests that the change in exhumation rate was probably related to a change in the buoyancy of the crust. The most effective way for this to happen is crustal thickening. Our model predicts a substantial thickening of the crust in the Briançonnais Zone from ~34 Ma in the east and ~31 Ma in the west, consistent with the increased rates of denudation at this time. Moreover, the orogen-wide switch to west–northwest-directed structures may

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reflect the first true mechanical interaction between European and Adrian lithospheres, and therefore the onset of continent–continent collision.

5. Conclusions

The Briançonnais Domain records several major orogen-forming events. Two temporally separate compressive events took place. The first records a present day transport direction of top-to-the-north–northwest. The second event is recorded by forethrusts, backthrusts, and backfolds, with an overall overshear direction towards the west–northwest. The initial compressive phase is tentatively dated at 45 Ma, and is conjectured to be in response to subduction and docking of the Briançonnais Domain into the subduction complex under the Adria margin. The second event started in the east of the Briançonnais Domain at 34 Ma, and propagated westward through the Briançonnais Domain to the present western edge at approximately 31 Ma. This second event is believed to be in response to the front of the Adria plate mechanically interacting with the Briançonnais Domain (Butler & Freeman, 1996). Coeval with this compressional phase was rapid uplift and exhumation of the Basal Briançonnais hanging wall. The age of 34 Ma may therefore time the onset of true continent–continent collision within the Alpine orogen.

The structure which initially formed the basal décollement for the Briançonnais Domain separating it from the Petit St Bernard and Valais zones to the west (the 45 Ma compressional event) cannot be the present BBT. The original structure must lie below the present BBT buried towards the east.

Mica phases can be very resistant to isotopic resetting within the greenschist facies even during significant deformation, but different mica generations can be related via their chemistry to different igneous, metamorphic and deformation events and so isotopically dated.

If the separate mica populations can be related from the microscale to the macroscale this technique can place constraints on tectonic models that previously had only a geometric basis. In this example we have shown that within the Western Alps, deformation probably occurred as a foreland-propagating locus of contractional structures.

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