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Precollisional, multistage exhumation of subducted continental crust: The Sesia Zone, western Alps

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[1] The Sesia Zone within the Tertiary arc of the western Alps is a relic of the subducted part of the Adriatic continental margin along the SE border of the Tethyan ocean. The Sesia Zone comprises three basement nappes which individuated during Late Cretaceous (65–80 Ma) subduction to different depths at high-pressure (HP, blueschist, eclogite facies) conditions (peak pressures of 1.0–1.2, 1.0–1.5, and 1.5–2.0 GPa). The thrusts bounding these nappes developed where the crust was previously thinned during Jurassic rifting. Crustal-scale shear zones partly overprinted these early thrusts and exhumed coherent slices of crust containing HP rocks. Initial exhumation of the internal part of the accreted margin involved thrusting (D1) and transpressional shearing (D2) along a subvertical, E-W trending mylonitic shear zone under retrograde blueschist- to greenschist-facies conditions. This exhumation was nearly isothermal to a depth of about 25 km, where the basement nappes were juxtaposed. Subsequent exhumation of these nappes to a common depth of about 15–20 km occurred in the footwall of a greenschist-facies, top-SE extensional shear zone (D3) preserved in some of the highest mountain peaks of the Sesia Zone. New Rb-Sr mineral ages constrain D2 to have occurred at about 60–65 Ma and D3 at about 45–55 Ma. Thus top-SE extensional exhumation was broadly coeval with Eocene, SE directed subduction of the Liguro-Piemont oceanic lithosphere beneath the Adriatic margin. Slow cooling and erosional denudation of the Sesia Zone from 45 to 30 Ma occurred in the hanging wall of the Gressoney extensional shear zone (D4), which itself contributed to the exhumation of Eocene HP and ultra-HP oceanic rocks in its footwall. By 30 Ma, HP rocks of the Sesia Zone were intruded by shallow granitic plutons which were eroded and redeposited within volcanoclastic sediments. Oligo-Miocene Insubric backfolding and thrusting (D5) only exhumed northeastern parts of the Sesia Zone, where HP metamorphism is absent or was overprinted by Tertiary temperature-dominated metamorphism. Most exhumation of continental HP rocks in the Sesia Zone therefore preceded Tertiary Alpine collision and coincided with Late Cretaceous to Early Tertiary subduction of the Adriatic and Tethyan lithosphere. The transition from D2 to D3 in the Sesia Zone is interpreted to mark a change from high-stress, oblique SE directed subduction and accretion of the distal Adriatic continental margin to NW retreating, low-stress subduction of the Liguro-Piemont oceanic lithosphere. **Citation:** Babist, J., M. R. Handy, M. Konrad-Schmolke, and K. Hammerschmidt (2006), Precollisional, multistage exhumation of subducted continental crust: The Sesia Zone, western Alps, *Tectonics*, *25*, TC6008, doi:10.1029/2005TC001927.

1. Introduction

[2] The Sesia Zone in the inner arc of the western Alps (Figure 1) is the first location where eclogite-facies metamorphism of granitic rocks was identified [e.g., *Bearth*, 1959; *Compagnoni and Maffeo*, 1974] and interpreted in terms of subducted continental lithosphere [*Ernst*, 1971]. From the outset, metamorphic overprinting relationships indicated that this subduction occurred during the early stages of the Alpine evolution [e.g., *Dal Piaz et al.*, 1972; *Trümpy*, 1973; *Frey et al.*, 1974; *Hunziker*, 1974], prior to Tertiary nappe stacking and Barrovian-type overprinting metamorphism [*Compagnoni et al.*, 1977]. Yet, as in many areas affected by high-pressure (HP) and ultrahigh-pressure (UHP) metamorphism, the exhumation mechanisms for the HP rocks of the Sesia Zone have been debated to the present day. Following convention, we use the term "Alpine" to refer to events in the Alps occurring in Late Cretaceous and Tertiary time, as reviewed by *Handy and Oberhänsli* [2004].

[3] To understand the debate on exhumation, several peculiarities of the Sesia Zone must be considered: First, the Sesia Zone together with the Dent Blanche klippe and the small Pillonet klippe (Figure 1) have been interpreted by some authors to represent the remains of an Early Mesozoic passive margin of the Adriatic continent [e.g., *Lardeaux and Spalla*, 1991; *Froitzheim et al.*, 1996]. Today, the Sesia-Dent Blanche unit forms the highest tectonic unit of the western Alps; it overlies exhumed HP and UHP relics of the Liguro-Piemont ocean (Zermatt-Saas unit), the ocean-continent transition (Combin unit) and the distal part of the European continental margin (Briançonnais units). These units were affected by Tertiary, SE directed backfolding ("Rückfaltung" or "retrocharriage" in Alpine

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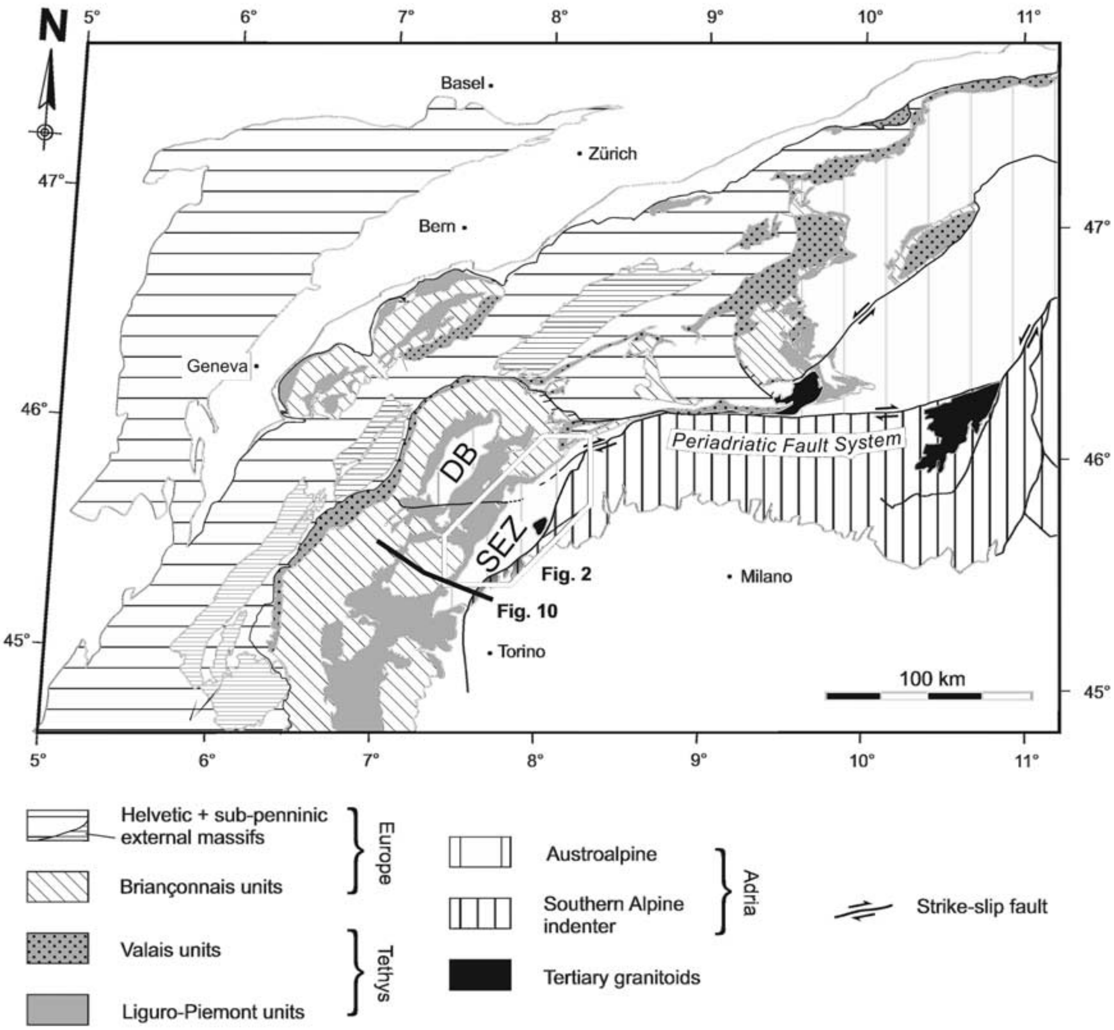


Figure 1. Tectonic map of the western and central Alps, with box enclosing area studied in this paper. The black line marks the trace of the ECORS-CROP profile in Figure 10f.

parlance [e.g., *Argand*, 1916; *Milnes et al.*, 1981]) in the retrowedge of the Alpine orogen. Second, the HP rocks of the Sesia Zone have Late Cretaceous ages, whereas those of the underying units yield Eocene ages for HP and UHP metamorphism [e.g., *Duchêne et al.*, 1997; *Rubatto et al.*, 1998, 1999]. Finally, the Sesia Zone is separated from the other units of the Adriatic passive margin (Canavese unit and Ivrea Zone, Figure 1) by greenschist-facies mylonites of the Periadriatic Fault System. These so-called Insubric mylonites accommodated late orogenic dextral strike slip and SE directed backthrusting in Oligo-Miocene time [*Schmid et al.*, 1987]. Unlike the Sesia Zone, however, none of the units south and east of the Insubric mylonite belt experienced HP Alpine metamorphism; indeed, they attained no more than anchizone to greenschist-facies Alpine metamorphism, and their pre-Alpine, amphibolite- to granulite-facies rocks escaped Alpine overprinting [*Zingg et al.*, 1990]. Thus both the northwestern and southeastern tectonic contacts of the Sesia Zone must have accommodated significant exhumation in Late Cretaceous and Early Tertiary time. We will return to this important aspect several times in the paper, but point out here that determining the age, conditions and kinematics of deformation in the Sesia Zone is key to resolving the larger issue of how large coherent blocks of subducted continental crustal are exhumed. This is the main goal of the work presented in this paper.

[4] Exhumation involves the movement of rock toward the Earth's surface [*England and Molnar*, 1990]. Not surprisingly given the complexity of the western Alps, many exhumation mechanisms have been proposed for the HP rocks of the Sesia Zone. *Ernst* [1971] reasoned that the continentally derived Sesia rocks were exhumed buoyantly due to their density contrast with surrounding mantle rocks in the early Alpine subduction zone. However, buoyant rise of subducted lithosphere only occurs to the level of isostatic equilibrium [*Platt*, 1993], so other forces must account for complete exhumation of the HP rocks to the surface. Mylonitic thrusting and folding have been held responsible

for exhumation and imbrication of thrust sheets with contrasting prograde and retrograde metamorphic evolutions in the Sesia Zone [*Gosso et al.*, 1979; *Pognante et al.*, 1980, 1987], but thrusting alone as a mechanism for exhuming the HP rocks in the Sesia Zone is unlikely in light of other structures pointing to synconvergent crustal extension [*Inger and Ramsbotham*, 1997]. Indeed, *Platt* [1986] proposed that synorogenic extension alternating with thrusting, and accompanied by erosional denudation acted together to exhume HP rocks in the retrowedge of the Alpine orogen. Yet so far, such shear zones appear to be restricted to the Liguro-Piemont ophiolitic units [e.g., *Ballèvre and Merle*, 1993; *Wheeler and Butler*, 1993] that are structurally below rather than above the Sesia Zone (Figure 1). On the basis of thermal modeling of pressure-temperature (PT) paths, *Rubie* [1984] proposed two-stage exhumation involving tectonic underplating of subducted continental lithosphere to the upper Adriatic plate followed by backthrusting localized at the Insubric mylonite belt. In this model, removal of overburden was effected by high Tertiary erosion rates. In principle, *Rubie*'s [1984] model is similar to later, numerical orogenic models of *Escher and Beaumont* [1997] in which buoyant rise of subducted HP rocks during nappe stacking is transitional to forced return flow of these rocks to the surface during backfolding and thrusting.

[5] Unfortunately, none of the models described above considers that HP rocks of the Sesia Zone were subducted to mantle depths ($\sim$60 km [e.g., *Tropper et al.*, 1999; *Zucali et al.*, 2002]) already in Late Cretaceous time, broadly coincident with the onset of oceanic subduction, and prior to mid to late Tertiary collision [e.g., *Frey et al.*, 1974; *Milnes*, 1978; *Schmid et al.*, 1996; *Weissert and Bernoulli*, 1985]. Moreover, existing models assume that the subduction and exhumation of Alpine HP rocks involved overall plane strain shortening in a NW-SE oriented vertical section, despite the fact that paleomagnetic data [*Dercourt et al.*, 1986] and paleogeographic reconstructions [*Stampfli et al.*, 1998; *Schmid et al.*, 2004] call for hundreds of kilometers of east-west dextral strike-slip motion between Briançonnais and Adria during closure of the Liguro-Piemont ocean in Late Cretaceous to Early Tertiary time. Even numerical models of Alpine-type collision belts that incorporate sophisticated assumptions on rheology, metamorphic phase transitions and erosion [*Burov et al.*, 2001; *Gerya and Stöckhert*, 2005] leave us with a decidedly ambiguous assessment of exhumation mechanisms, due partly to the simple two-dimensional (2-D) kinematic constraint and partly to the choice of thermal boundary conditions, as discussed in section 6.

[6] We therefore present new three-dimensional structural and kinematic data which, when regarded in the context of thermobarometric and geochronological data, casts an entirely different light on the early exhumation history of HP rocks in the western Alps. In section 2, we outline how basic lithological and metamorphic characteristics of the Sesia Zone can be interpreted in terms of primary thrust contacts between three basement nappes that individuated during subduction. A new type of fabric map and cross sections in section 3 reveal how these primary structures were modified and locally preserved during multistage exhumation. In section 4, new thermobarometric and geochronological constraints on exhumation indicate that although the basement units in the Sesia Zone underwent different retrograde evolutions, their transition from eclogite- to greenschist-facies conditions involved rapid, nearly isothermal decompression in Late Cretaceous to earliest Tertiary time followed by slower cooling in early to mid-Tertiary time. These field-based observations form the basis for a kinematic model of Alpine subduction and exhumation of HP rocks, presented in section 5. Contrary to existing models, our work suggests that most exhumation occurred during Late Cretaceous dextral transpression. The paper concludes in section 6 with a comparison of our field-based kinematic model with recent dynamic, physical models of exhumation.

2. Primary Lithotectonic Features in the Sesia Zone

[7] The Sesia Zone is a composite unit that can be divided into three Alpine basement nappes with contrasting pre-Mesozoic lithologies and Alpine metamorphic histories (Figure 2): (1) the Bard nappe making up the northwestern part of the Sesia Zone comprises mostly fine-grained gneiss (“gneiss minuti” in the native literature by *Gastaldi* [1871] and *Stella* [1894] as cited by *Compagnoni et al.* [1977]); (2) the Mombarone nappe in the southern and eastern parts of the Sesia Zone contains predominantly eclogitic micaschists (“micascisti eclogitici”, *Gastaldi* [1871] and *Stella* [1894] as cited by *Compagnoni et al.* [1977]) and subordinate late Paleozoic granitoid, metabasite and marble. These rocks are intruded discordantly by Oligocene granitoids (Biella and Traversella plutons in Figure 2) and, importantly, are included as boulders within Oligocene volcanoclastics exposed along the SW border of the Mombarone nappe [*Scheuring et al.*, 1974]; (3) the IIDK nappe, or “Seconda Zona Diorito Kinzigitica”, is preserved within two large synforms in the northeastern part of the Sesia Zone, and to the southwest generally follows the contact of the Bard and Mombarone nappes [*Minnigh*, 1977]. The name of this nappe belies its close lithological affinity with pre-Alpine metasediments of the Ivrea Zone (so-called kinzigites, or “Prima Zona Diorito Kinzigitica” [*Carraro et al.*, 1970; *Dal Piaz et al.*, 1971; *Isler and Zingg*, 1974; *Compagnoni et al.*, 1977]). In fact, all three of these basement units experienced prograde amphibolite- to granulite-facies Paleozoic metamorphism, but as explained in section 3, the IIDK nappe followed a somewhat different evolution which limited the degree of Alpine overprinting.

[8] A thin, discontinuous strand of highly deformed Mesozoic metasediments and subordinate metabasite, the Bonze unit in Figure 2, separates the Bard and Mombarone nappes along most of their mutual contact. These sediments are very similar to Late Paleozoic to Jurassic sediments of the Canavese Zone exposed along the contact of the Mombarone nappe with the Ivrea Zone. The latter sediments represent the distal part of the Adriatic passive continental margin [e.g., *Biino and Compagnoni*, 1989;

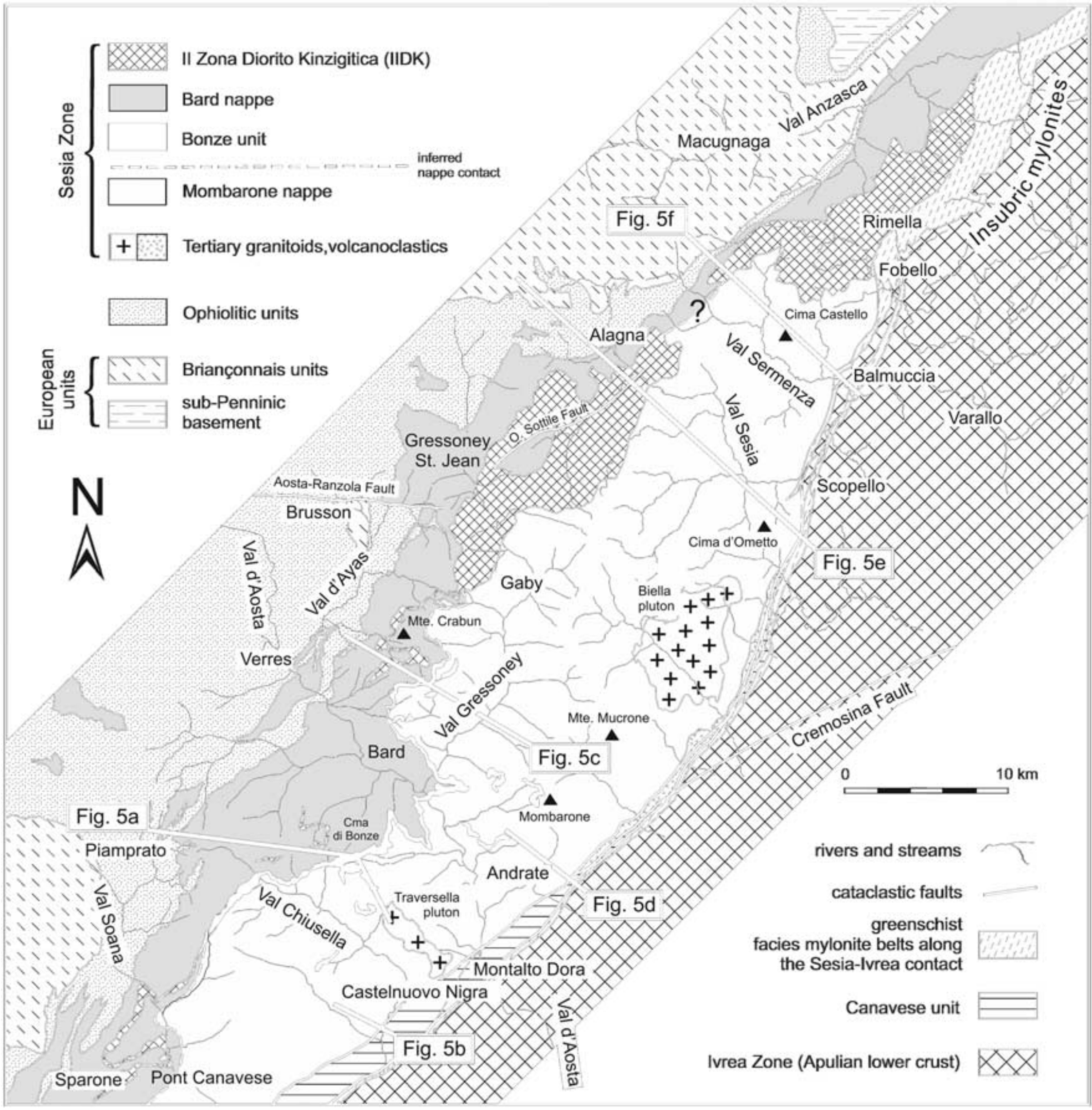


Figure 2. Lithotectonic map of the Sesia Zone and adjacent units. Map shows tectonic subdivision of the Sesia basement (Mombarone, Bard, and IIDK nappes), Mesozoic metasediments (Bonze and Canavese units), and the location of cross sections in Figure 5.

Ferrando et al., 2004]. Likewise, the Bonze unit contains lithologies (volcanoclastics, dolomitic marbles, manganiferous quartzites and calc-silicate schists; part of the GM subunit of *Pognante et al.* [1987] and *Venturini et al.* [1994, 1996]) that are diagnostic of a distal passive margin or a transitional continent-ocean setting, as discussed below). As described in section 3, both the Bonze and Canavese metasediments experienced greenschist-facies overprinting, in some places preceded by Alpine blueschist-facies metamorphism.

[9] We note that our division of the Sesia Zone differs from its classical subdivision into ''external'' gneiss-minuti, ''internal'' eclogitici micaschisti and IIDK units [e.g., *Compagnoni et al.*, 1977] in two important respects: First, we use the Mesozoic sediments of the Bonze unit, rather than the lithologies of the pre-Alpine basement as a criterion for delimiting the basement nappes. This approach is common in Alpine geology and, in the case of the Sesia Zone, is justified by thermobarometry indicating that the Bard and Mombarone nappes have different prograde metamorphic paths [*Konrad-Schmolke et al.*, 2006] during their subduction to different peak, blueschist-eclogite facies conditions (respectively, 1.0–1.5 GPa at 500–550°C and 1.5 and 2.0 GPa at 500 to 600°C [*Desmons and O'Neil*, 1978; *Koons*, 1986; *Lardeaux and Spalla*, 1991; *Tropper et al.*, 1999; *Zucali et al.*, 2002]). Pre-Alpine lithology is not a good criterion for differentiating basement units because our fabric mapping (Figure 3) revealed that the distribution of greenschist-facies assemblages in pre-Alpine lithologies (e.g., ''gneiss minuti'' in the sense of *Passchier et al.* [1981]) reflects the intensity of retrograde overprinting during exhumation, especially in the NW parts of the Sesia Zone

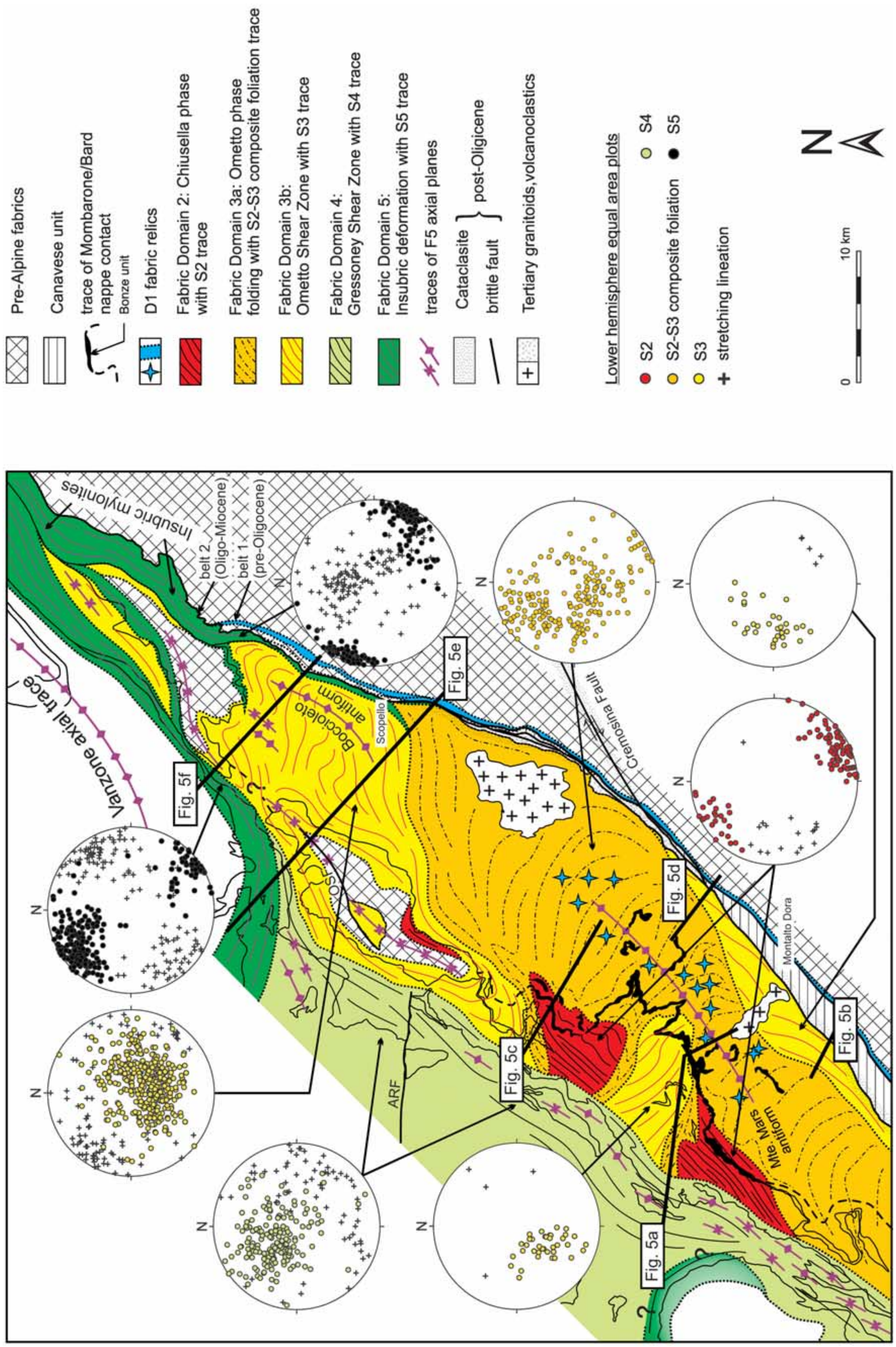

Pre-Alpine fabrics
Canavese unit
trace of Mombarone/Bard nappe contact
Bonze unit
D1 fabric relics
Fabric Domain 2: Chiusella phase with S2 trace
Fabric Domain 3a: Ometto phase folding with S2-S3 composite foliation trace
Fabric Domain 3b: Ometto Shear Zone with S3 trace
Fabric Domain 4: Gressoney Shear Zone with S4 trace
Fabric Domain 5: Insubric deformation with S5 trace
traces of F5 axial planes
Cataclasite
brittle fault
post-Oligicene
Tertiary granitoids,volcanoclastics
Lower hemisphere equal area plots
S2
S2-S3 composite foliation
S3
stretching lineation
S4
S5
0
10 km
N
Vanzone axial trace
Insubric mylonites
belt 2 (Oligo-Miocene)
belt 1 (pre-Oligocene)
Fig. 5f
Fig. 5e
Fig. 5d
Fig. 5c
Fig. 5b
Fig. 5a
Bocciolet antiform
Scopello
Cremosina Fault
Montalto Dora
Mte. Mars antiform
ARF
OSZ

Figure 3

[*Stünitz*, 1989], rather than any primary lithological differences that delimit tectonic units. To avoid further confusion, we therefore discard the historical names "gneiss minuti" and "eclogitici micascisti" in favor of the units above which define Alpine nappes and tectonic contacts formed during subduction. The second point of difference with previous approaches is that we regard the Bonze metasediments to have experienced polyphase Alpine metamorphism, rather than to be Alpine monometamorphic as previously thought (the Monometamorphic Cover Complex of *Pognante et al.* [1987] and *Venturini et al.* [1994]). We will return to this important feature when using spatial-temporal variations in polyphase Alpine metamorphism of the sediments as depth-time markers for reconstructing the evolution of the Adriatic margin during subduction and exhumation.

3. Structural and Mineralogical Records of Exhumation

3.1. Distinguishing Deformational Phases and Fabric Domains

[10] At the scale of the area in Figure 2, we distinguished five Alpine deformational phases whose extent is depicted in Figure 3 as colored fabric domains numbered from oldest (1) to youngest (5). Within a given fabric domain, the numbered (colored) deformational phase predominates, in the sense that it contains the main (i.e., visibly obvious), overprinting schistosity and/or folds. Nevertheless, older phases and mineral assemblages are preserved locally in all domains, particulary where the dominant deformation is very localized.

[11] Four types of structural and metamorphic criteria were used to distinguish these fabric domains: (1) the relative age of structures, metamorphic minerals and magmatic dikes based on crosscutting relationships; (2) structural style (e.g., congruent versus incongruent folds, discrete versus distributed shearing); (3) the orientation of structures; and (4) the kinematics of deformation related to those structures. Figure 4 shows the key observations made on different scales in each fabric domain. Where microstructural analysis revealed that metamorphic minerals grew synkinematically, structures were correlated with a metamorphic event, in some cases allowing us to estimate the thermobarometric conditions of deformation [*Konrad-Schmolke et al.*, 2006], or even to determine the age of the structures. In other cases, especially where successive deformation occurred under similar, retrograde greenschist-facies conditions (D3–D5) or had similar kinematics (D3, D4), we were restricted to using structural criteria. Taken together, the suite of criteria above allowed us to map tectonometamorphic facies across pre-Alpine lithological boundaries. It turned out that these facies define crustal-scale shear zones related to exhumation of the HP rocks. The shear zones are named below after their type locality.

[12] To simplify matters, we did not distinguish amongst pre-Alpine deformational phases, and we disregarded all but the largest brittle structures, most of which are related to the fifth Alpine (Insubric) deformational phase. It turned out that none of these structures is related to significant exhumation of HP rocks, although the impressive Miocene to Recent faulting and uplift in the Alps is partly responsible for the excellent exposure. Alpine glaciation has carved almost 3000 m of vertical relief, affording a spectacular view of the three-dimensional structure of the Sesia Zone and its adjacent units. The cross sections in Figures 5a–5f depict this crustal-scale structure, with different symbols delineating the traces of the dominant foliations in each fabric domain.

3.2. Pre-Alpine Fabrics

[13] Pre-Alpine fabric relics are ubiquitous in all basement units of the Sesia Zone [e.g., *Dal Piaz et al.*, 1972; *Williams and Compagnoni*, 1983; *Lardeaux and Spalla*, 1991] but form the dominant fabric only within the IIDK nappe (cross-hatched white areas in Figures 3 and 5a–5c). There, pegmatite dikes and migmatites truncate pre-Alpine foliations and are themselves locally overprinted by Alpine blueschist- [*Ridley*, 1989] and greenschist-facies shear zones (Figure 4b). Pre-Alpine granitoids preserve their magmatic texture, even though they underwent static recrystallization at greenschist-facies conditions (Figure 4c). The southern margin of the large IIDK body in the northeastern part of the Sesia Zone was overprinted by D5 mylonitic shearing [*Isler and Zingg*, 1974] related to a splay of the Insubric mylonite belt.

3.3. Fabrics in Domain 1: Subduction

[14] Fabric domain 1 is a composite domain that contains rocks with blueschist-to-eclogite facies, synkinematic assemblages. Up to three distinct HP assemblages have been identified in various localities, but these are only locally correlatable with structures in outcrop and thin section (e.g., Mucrone area, starred locations in Figure 3 [*Zucali et al.*, 2002]). These include relic S1 foliations preserved in the pressure shadows of mafic boudins (Figures 4d and 4e) and fold hinges, or within S2 foliation boudins [*Gosso et al.*, 1979; *Reinsch*, 1979] and garnets [*Pognante et al.*, 1987]. With the possible exception of the Bard-Mombarone contact, no large structures associated with subduction are preserved in the Sesia Zone [*Gosso*, 1977; *Gosso et al.*, 1979; *Pognante et al.*, 1980]. However, a greenschist-facies mylonite belt at the northwestern margin of the Ivrea Zone can be interpreted as a D1 shear zone, as discussed below.

Figure 3. Fabric map of the Sesia Zone and adjacent tectonic units. Criteria used to construct the map are described in the text (section 3). Orientation of main structural elements related to the different fabric domains is presented in lower hemisphere equal-area plots. Note that fabric domain boundaries crosscut the primary lithotectonic contacts shown in Figure 2. ARF, Aosta-Ranzola Fault; OSF, Ospizio-Sottile Fault.

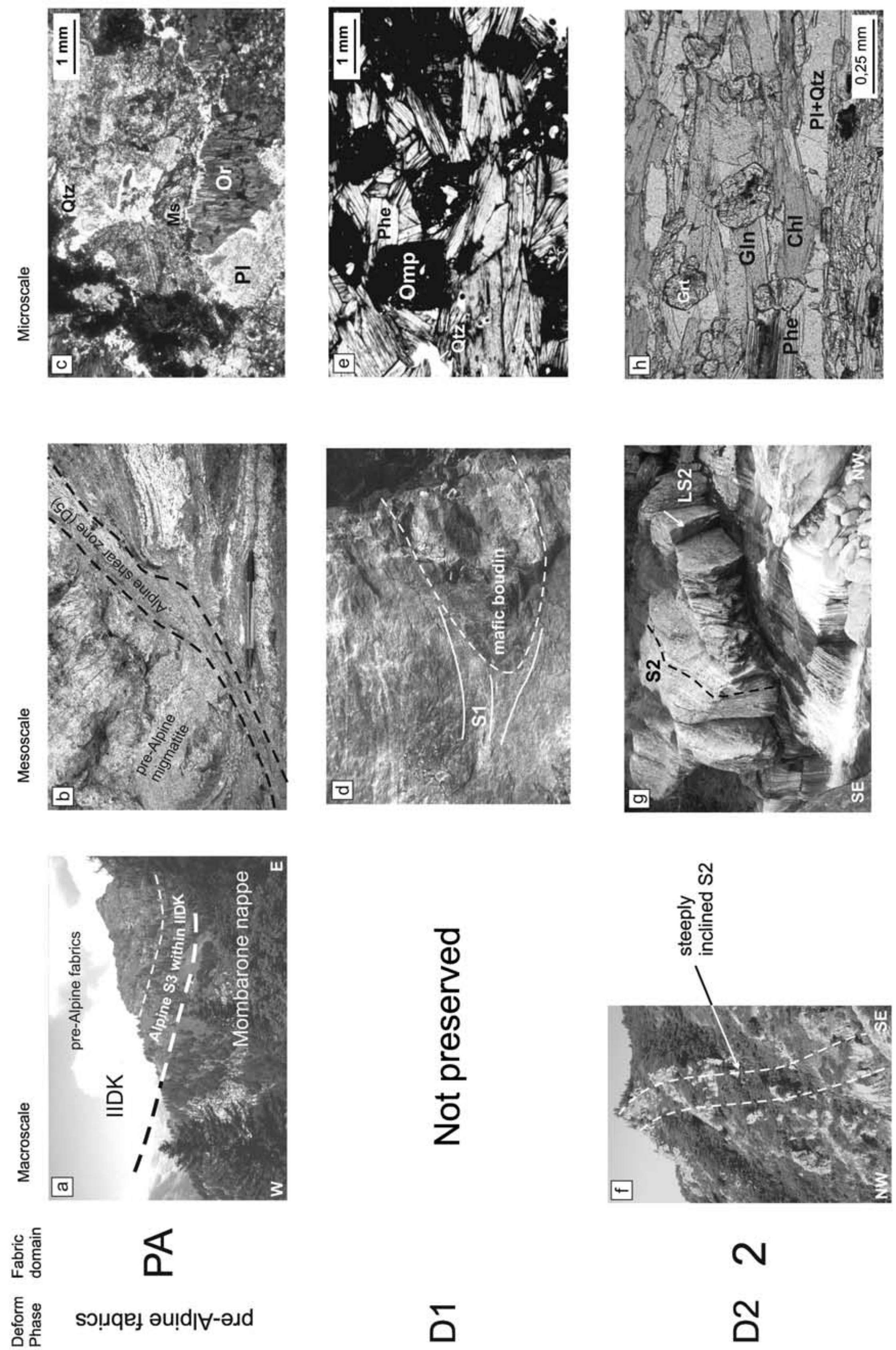

Deform Phase
Fabric domain
Macroscale
Mesoscale
Microscale
pre-Alpine fabrics
PA
a
IIDK
pre-Alpine fabrics
Alpine S3 within IIDK
Mombarone nappe
W
E
b
pre-Alpine migmatite
Alpine shear zone (D5)
c
Qtz
Ms
Or
Pl
1 mm
D1
Not preserved
d
S1
mafic boudin
e
Phe
Omp
Qtz
1 mm
D2
2
f
steeply inclined S2
NW
SE
g
S2
LS2
SE
NW
h
Grt
Gln
Chl
Phe
Pl+Qtz
0,25 mm

Figure 4

3.4. Fabric Domain 2: Chiusella Shear Zone

[15] Fabric domain 2 is clearly exposed in two tectonic windows (red areas in Figure 3, see also Figures 5b and 5c) in the footwall of fabric domains 3 and 4. These windows contain a steeply NW dipping S2 foliation (equal-area plot in Figure 3) that defines a kilometer-wide shear zone (the Chiusella shear zone) overprinting part of the boundary between the Bard, Bonze and Mombarone units. This shear zone has a fan-like foliation structure which narrows downward, as observed in the deepest, glacially eroded valleys (Val d'Aosta, Val Chiusella, Figures 5a–5c). The gentle to moderate plunge of the stretching lineation, Ls2, comprising feldspar, quartz and glaucophane on this foliation (Figures 4f and 4g) indicates a significant strike-slip component of shear during exhumation. Local shear bands indicate sinistral SE-side up motion parallel to Ls2, but outcrops with reliable kinematic indicators are rare at all scales. S2 is itself defined by synkinematic, retrograde blueschist-facies assemblages grading to greenschist-facies assemblages in the NW part of the Chiusella shear zone (Figure 4h). Toward the SE within this shear zone, low-strain domains preserve isoclinal F2 folds and even relic D1 structures with the transitional blueschist-eclogitic assemblage: garnet, omphacite, zoisite, rutile, phengitic white mica and pale blue, Fe-poor glaucophane. This D1 glaucophane is overgrown statically during D2 by dark blue Fe-rich glaucophane. The asymmetry in the synkinematic metamorphic zonation of the Chiusella shear zone suggests that this shear zone accommodated differential exhumation of the Mombarone nappe.

[16] Traces of D2 deformation are ambiguous outside of the fabric domain 2 windows mentioned above, but there is local evidence that the Chiusella shear zone extended throughout the Sesia Zone. For example, in the Val Sesia just south of the trace of the section in Figure 5e (see also Figure 3), a D2 blueschist-facies shear zone overprints the southern contact between the IIDK and the Mombarone nappes, as observed by *Ridley* [1989].

3.5. Fabric Domain 3: Ometto Shear Zone

[17] Fabric domain 3 covers most of the Sesia Zone (Figure 3) and is defined by gently to moderately dipping, retrograde greenschist-facies structures. These take the form of recumbent F3 folds (fabric domain 3a, light brown area in Figure 3) that tighten upward as the folded S2 foliation is transformed into a penetrative mylonitic S3 foliation (fabric domain 3b, yellow area in Figure 3). The change in deformational style from fold-dominated in the valleys to foliation-dominated domains at higher altitudes represents a vertical increase in D3 strain and defines a large extensional shear zone with top-east to -SE motion, here named the Ometto shear zone (Figure 4l). The folds at the base of this shear zone vary in size (amplitudes from centimeters to hundreds of meters, Figures 4i and 4j) and flatten both the D2 Chiusella shear zone and the Bard-Mombarone contact (Figures 5a, 5c, and 5d).

[18] The mylonitic foliation of the Ometto shear zone is best exposed in the highest peaks of the Sesia Zone (e.g., Monte Marzo and Cima di Bonze in Figure 5a and Mont Crabun in Figure 5c [see also *Passchier et al.*, 1981]) but also occurs at more accessible altitudes just north of Castelnuovo Nigra (Figure 5b) and Andrate (Figure 5d) and at the type locality of Cima d'Ometto (Figure 5e). Where unaffected by later deformation, the Ometto shear zone has a gently NW and SE plunging Ls3 stretching lineation (equal-area plots, Figure 3) defined by dynamically recrystallized feldspar and quartz that is often also aligned parallel to the F3 fold axes. The sense of shear parallel to Ls3 is top-east to -SE, judging from shear bands in domain 3b (Figures 4m and 4n) and the vergence of F3 folds at the transition from domains 3a to 3b. However, the lack of fold vergence in large parts of domain 3a indicates that near-coaxial, vertical shortening predominates in the footwall of the Ometto shear zone.

[19] In domain 3a, the dominant foliation is a composite of S2 and S3 (Figure 4j). It comprises phengitic white mica, albite, minor biotite and green to colorless calcic amphibole that grew at the expense of glaucophane (Figure 4k). Reequilibration during D3 was incomplete, so that relict D1 and D2 assemblages are common in this domain (stars

Figure 4. Structures and metamorphic mineral assemblages in fabric domains of the Sesia Zone. (a) Base of IIDK nappe in Val Sermenza showing overprint of pre-Alpine fabrics by Alpine deformation; (b) Alpine greenschist-facies shear zone (D5) in the IIDK nappe along its contact with the Mombarone nappe, Val Sermenza; (c) pre-Alpine magmatic textures within the IIDK nappe, streambed of Torrente Bise Rosse, St. Anna N' Rimella, Val Mastallone; (d) D1 fabric in the pressure shadow of a mafic boudin, Scalaro, lower Val d'Aosta; (e) Static growth of omphacite, phengite, and quartz along the axial plane of an F1 fold; (f) Blueschist-facies S2 foliation in the Chiusella shear zone, Tallorno, Val Chiusella; (g) Stretching lineation, Ls2, in D2 mylonite, Tallorno, Val Chiusella; (h) S2 assemblage glaucophane, garnet, plagioclase, quartz and phengite; (i) D3 folds, Sparone, Valle dell'Orco; (j) S2–S3 composite foliation parallel to F3 axial planes, Fervento, Val Sermenza; (k) Synkinematic D3 green amphibole; (l) S3 of the Ometto shear zone, M. Voghel, Val Gressoney; (m) D3 shear bands, Boccioleto, Val Sermenza; (n) D3 retrogression of glaucophane to actinolite, biotite and chlorite; (o) Base of the Bard nappe and underlying Liguro-Piemont (LP) ophiolites affected by Gressoney shear zone (D4) and by open F5 fold, eastern side Val Gressoney. Note the base of the IIDK nappe (right) affected by S3; (p) Top-SE D4 shear bands, Sesia river streambed near Alagna; (q) D4 assemblage chlorite, green biotite, actinolite, albite, and quartz; (r) S4 foliation folded by F5, near Alagna, upper Val Sesia; (s) D5 folding is associated with S3 foliation reactivated by F5 under greenschist facies conditions, San Giuseppe, Val Sermenza; (t) D5-reactivated foliation with sharp contact to adjacent S2–S3 composite foliation.

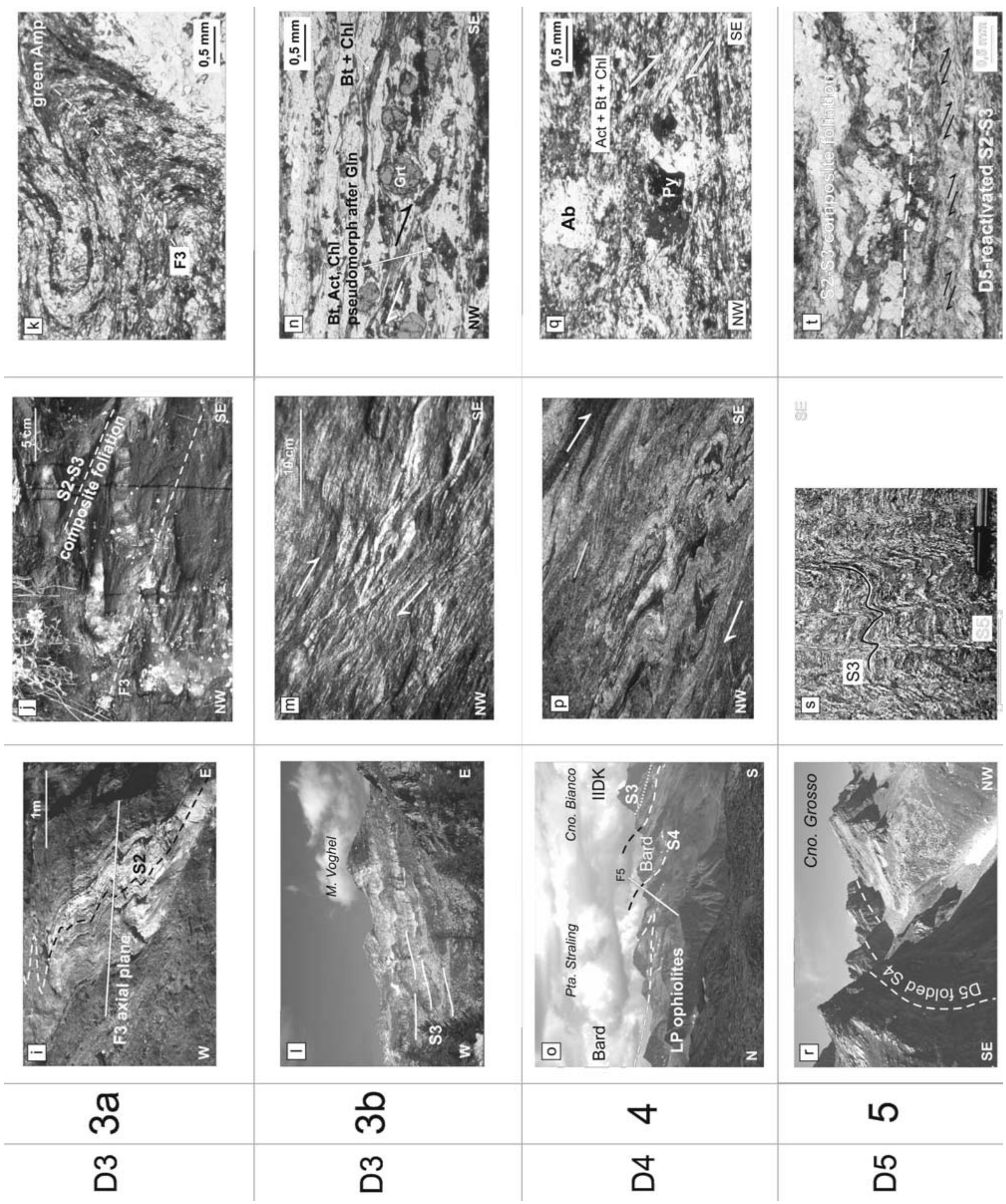


Figure 4. (continued)

in Figure 3 indicate the best preserved eclogitic relics). In contrast, S3 transposition in domain 3b is nearly complete. The same synkinematic mineral assemblage is observed as in domain 3a, but overprinting is so pervasive that eclogitic and blueschist-facies assemblages are rarely preserved (e.g., garnet in Figure 4n).

3.6. Fabric Domain 4: Gressoney Shear Zone

[20] Fabric domain 4 overprints the NW margin of the Bard nappe and large parts of the Liguro-Piemont unit. (light green area in Figure 3). This fabric domain corresponds to the Gressoney shear zone, first described by *Wheeler and Butler* [1993]. It is defined by a retrograde greenschist-facies S4 schistosity oriented parallel to the base of the Bard nappe (Figure 4o, equal-area plots in Figure 3) as well as by open to tight folds (amplitudes from centimeters to tens of meters) with gently SE dipping axial planes. These orientations are affected locally by D5 folds and later brittle faults, especially in the NE part of the area in Figure 3. Where S4 is undisturbed, however, it contains a stretching lineation, Ls4, that plunges gently to the east to SE (equal-area plot, Figure 3). The sense of shear parallel to Ls4 is consistently top-east to -SE (shear bands in Figures 4p and 4q), as also documented by *Wheeler and Butler* [1993]. We distinguished D4 from D3 fabrics based on refolded F3 folds (Figure 4p) as well as on the different behavior of feldspar: In D4 fabrics, feldspar displays both cataclasis and limited dynamic recrystallization, whereas in D3 fabrics it shows only dynamic recrystallization. The D4 assemblage in granitoid rocks includes dynamically recrystallized quartz, green biotite, chlorite, epidote and albite (Figure 4q). Taken together, these observations indicate that the part of the Gressoney shear zone that overprinted the Bard nappe was active at slightly lower temperature, greenschist-facies conditions than the Ometto shear zone in its hanging wall. Where the Ometto shear zone is not well developed, D4 directly overprints even older fabrics, for example in the Val d'Ayas (Figure 5c) where it deforms D2 fabrics of the Chiusella shear zone, or in the Val Sermenza (Figure 5f) where it cuts pre-Alpine fabrics of the IIDK unit.

3.7. Fabric Domain 5: Periadriatic Shear Zone

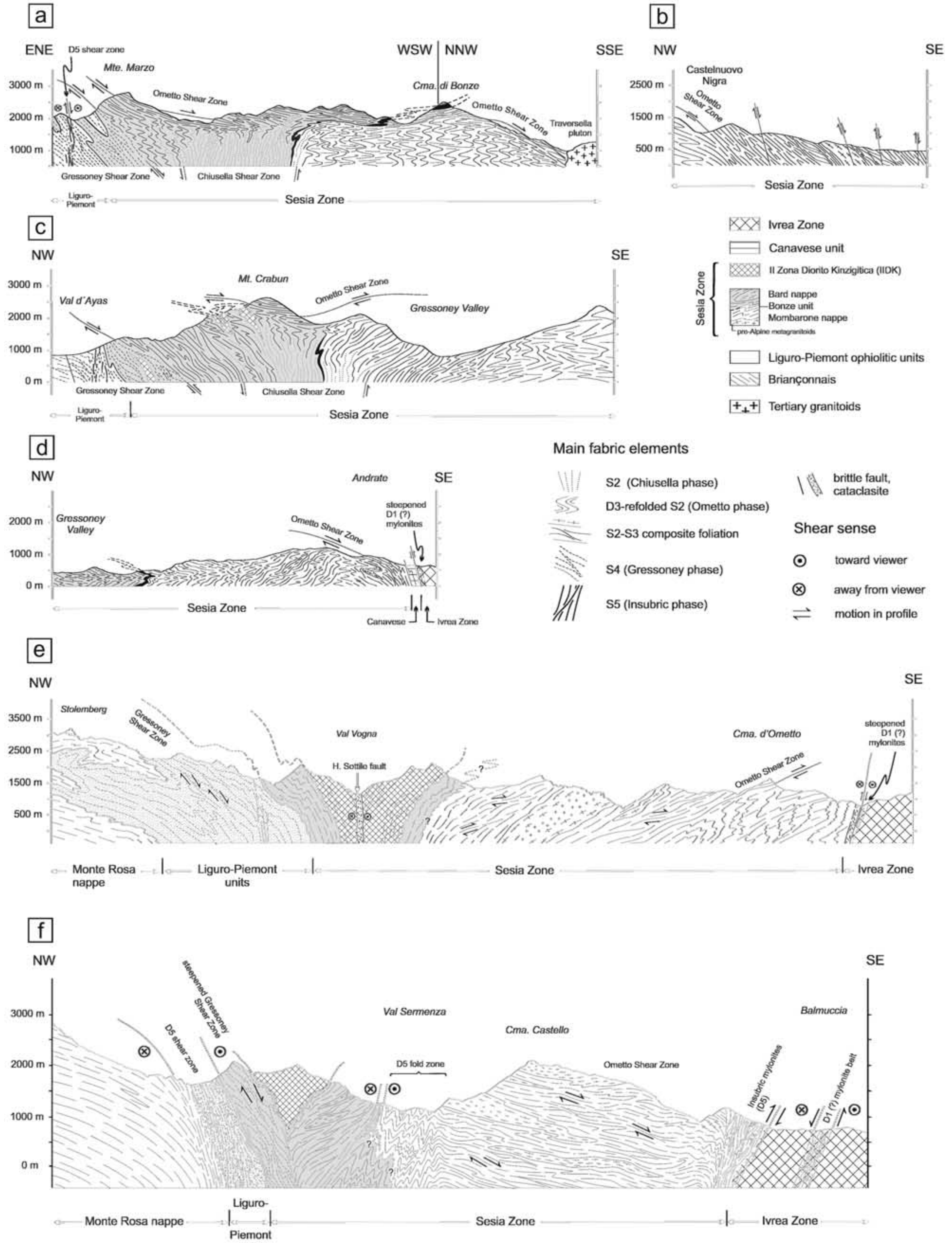
[21] D5 fabrics are characterized by steeply dipping, retrograde greenschist-facies structures that occur predominantly in the NE part of the Sesia Zone and overprint tectonic units to the north (dark green areas in Figure 3). These structures include shear zones ranging in width from tens to hundreds of meters, and upright, acylindrical folds with amplitudes from centimeters to hundreds of meters. Most prominent amongst these folds is the Vanzone antiform in the Monte Rosa nappe [e.g., *Keller et al.*, 2005], but related D5 folds in the Sesia Zone deform the S3 foliation of the Ometto Shear Zone into large, open antiforms and narrow D5 synforms. It is within the cores of such tight synforms that the two IIDK bodies are preserved (Figures 5e and 5f). The Boccioleto antiform (Figure 3) [*Steck and Hunziker*, 1994] and the Monte Mars antiform in the central part of the Mombarone nappe are broad F5 folds that arched the S3 foliation (see F5 axial traces and folded S3 traces in Figure 3). The F5 folds tighten parallel to their gently plunging axes until they merge down-plunge with greenschist-facies mylonitic shear zones. These shear zones are well developed in the NE part of the Sesia Zone, where they envelop lozenges of older fabric domains and are contiguous with the greenschist-facies, Insubric mylonite belt along the contact between the Sesia and Ivrea zones (Figure 3). *Handy et al.* [2005] interpreted these shear zones to be mylonitic splays of the Periadriatic fault system.

[22] Several overprinting relationships are key to distinguish D5 structures from earlier fabrics formed under similar greenschist facies conditions: A large F5 fold in the southern limb of the Vanzone antiform steepens domain 4 fabrics of the Gressoney shear zone at the base of the Bard nappe (Figures 4r and 5f). Earlier foliations were reactivated and transposed during F5 folding, leading to the development of S4 in the limbs of F4 folds (Figures 4s and 4t) to the exclusion of an axial plane foliation. Fold style and particularly the microstructural behavior of feldspar also help to differentiate D5 from D3 and D4 structures. F5 folds form chevrons and kinks in micaceous rocks, whereas quartz-rich rocks generally have similar folds. Feldspar clasts in a matrix of quartz and mica fracture and undergo dynamic recrystallization, suggesting that D5 folding occurred at the ductile to brittle transition of feldspar. Likewise, quartz aggregates in D5 mylonites show dynamic recrystallization (subgrain rotation und bulging recrystallization) transitional to cataclasis at the Sesia-Ivrea contact [e.g., *Schmid et al.*, 1987].

3.8. Sesia-Ivrea Contact

[23] The contact between the Sesia and Ivrea zones warrants separate consideration because it is the site of several key relations. Greenschist-facies mylonite along this contact is well known in the literature ("Scisti di Fobello e Rimella" of *Artini and Melzi* [1900], mylonite belt 2 of *Schmid et al.* [1989]) and is laterally continuous with D5 mylonitic splays in the Sesia Zone, as described above (areas colored dark green in Figure 3). This Insubric mylonite belt accommodated oblique, top-east backthrusting parallel to a steeply plunging stretching lineation (equal-area plot in Figure 3) and effected exhumation of the Sesia Zone onto the Ivrea Zone, probably in Oligo-Miocene time [*Schmid et al.*, 1989; *Zingg and Hunziker*, 1990]. This mylonite belt can be mapped as far southwest as Scopello in Val Sesia (Figure 3), where part of it appears to splay eastward into the Mombarone nappe. The other part is transitional southwestward to cataclasite that overprints the Oligocene volcanoclastic cover of the Sesia Zone (Figure 3) [*Schmid et al.*, 1989].

[24] Another greenschist-facies mylonite belt strikes parallel to the belt above but deforms the northwestern margin of the Ivrea Zone (thin blue band in Figure 3). It is cut by the Tertiary mylonite belt above and by mafic dikes of inferred Oligocene age (mylonite belt 1 of *Schmid et al.* [1989]). These relations are consistent with the prevalence of Late Cretaceous (60–75 Ma) K-Ar formational white mica ages in Canavese metasediments overprinted by this older mylonite [*Zingg and Hunziker*, 1990]. Kinematic



a
ENE
D5 shear zone
Mte. Marzo
Ometto Shear Zone
WSW
NNW
SSE
Cma. di Bonze
Ometto Shear Zone
Traversella pluton
3000 m
2000 m
1000 m
Gressoney Shear Zone
Chiusella Shear Zone
Liguro-Piemont
Sesia Zone
b
NW
SE
Castelnuovo Nigra
Ometto Shear Zone
2500 m
1500 m
500 m
Sesia Zone
c
NW
SE
Mt. Crabun
Ometto Shear Zone
Val d'Ayas
Gressoney Valley
0 m
Gressoney Shear Zone
Chiusella Shear Zone
Liguro-Piemont
Sesia Zone
Ivrea Zone
Canavese unit
Sesia Zone
II Zona Diorito Kinzigitica (IIDK)
Bard nappe
Bonze unit
Mombarone nappe
pre-Alpine metagranitoids
Liguro-Piemont ophiolitic units
Briançonnais
Tertiary granitoids
d
NW
Andrate
SE
steepened D1 (?) mylonites
Gressoney Valley
Ometto Shear Zone
Sesia Zone
Canavese
Ivrea Zone
Main fabric elements
S2 (Chiusella phase)
D3-refolded S2 (Ometto phase)
S2-S3 composite foliation
S4 (Gressoney phase)
S5 (Insubric phase)
brittle fault, cataclasite
Shear sense
toward viewer
away from viewer
motion in profile
e
NW
SE
Stolemberg
Gressoney Shear Zone
Val Vogna
H. Sottile fault
Cma. d'Ometto
Ometto Shear Zone
steepened D1 (?) mylonites
3500 m
Monte Rosa nappe
Liguro-Piemont units
Sesia Zone
Ivrea Zone
f
NW
SE
steepened Gressoney Shear Zone
D5 shear zone
Val Sermenza
Cma. Castello
D5 fold zone
Ometto Shear Zone
Balmuccia
Insubric mylonites (D5)
D1 (?) mylonite belt
Monte Rosa nappe
Liguro-Piemont units
Sesia Zone
Ivrea Zone

Figure 5

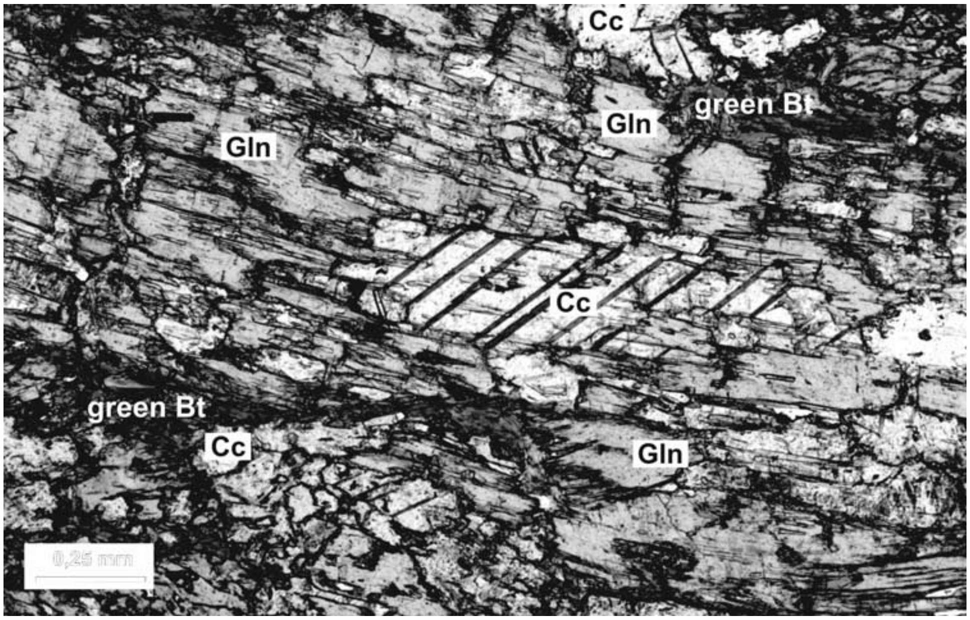


Figure 6. Na-amphibole in Mesozoic metasediment of the Canavese unit at Scopello, Val Sesia (location in Figures 2 and 3). Microprobe analyses of amphiboles available on request from the authors. Swiss topographic coordinates are 615.180/69.185.

indicators in this Late Cretaceous mylonite belt show dextral, oblique-slip motion of the Ivrea Zone parallel to a gently SW plunging stretching lineation [*Schmid et al.*, 1989]. However, paleomagnetic work on the crosscutting mafic dikes indicates that the pre-Oligocene orientation of the mylonitic foliation was gently SE dipping [*Schmid et al.*, 1989], such that mylonitization involved oblique, west directed thrusting of the Ivrea Zone. This mylonite belt can be traced from the Val Sesia (Figure 5f) to the southwest as far as the lower Val d'Aosta (Figures 5d and 5e). There, they are juxtaposed with D3 greenschist-facies mylonite along the eastern margin of the Mombarone nappe (fabric domain 3b in Figure 3). In sections 4 and 5, we argue that this older mylonite belt probably formed during D1.

[25] Brittle faults along the southwestern segment of the Sesia-Ivrea border have both extensional and dextral strike-slip components (Figure 5b), which may be attributed partly to activity of the Cremosina fault (Figure 3, ''external'' and ''internal Canavese Lines'' of *Biino and Compagnoni* [1989] and *Schumacher et al.* [1997]). Although the details of the brittle deformational history are unresolved, this late faulting evidently led to pinching of the Oligocene volcanoclastic cover of the Mombarone nappe together with the Mesozoic Canavese metasediments and the Ivrea Zone [*Schmid et al.*, 1989].

[26] The Mesozoic Canavese metasediments are deformed by all of the aforementioned mylonite belts and faults [*Zingg and Hunziker*, 1990]. In most locations, including the type area in the lower Val d'Aosta (Montalto Dora in Figure 3), these metasediments are affected only by greenschist-facies metamorphism and deformation [*Zingg et al.*, 1976; *Biino and Compagnoni*, 1989]. However, we discovered that they also contain blue, Na-rich amphibole (Figure 6) at Scopello in Val Sesia (Figure 2). Although blueschist-facies metamorphism in the Canavese sediments is so far limited to this single locality, it indicates that the Sesia-Ivrea contact was the southern limit of Alpine subduction. This is consistent with the occurrence of prograde blueschist-facies mylonites along the southeastern margin of the Sesia Zone, SW of the area in Figure 2 [*Pognante*, 1989]. Hence pre-Oligocene mylonitic deformation along the Sesia-Ivrea contact must have involved exhumation and juxtaposition of subducted and nonsubducted fragments of the Canavese sediments. In section 4, we discuss why this early exhumation can be ascribed to late D2, dextral strike-slip movement along the mylonite belt in the Ivrea Zone.

[27] To summarize section 3, fabric mapping reveals crustal-scale shear zones that developed under retrograde metamorphic conditions and overprint all subduction-related structures, including the thrust contact between the Bard and Mombarone basement nappes. These shear zones are responsible for exhuming the HP rocks in the Sesia Zone, beginning with oblique-slip D2 deformation under retrograde eclogite- to blueschist- to greenschist-facies conditions, and followed by D3 top-east to -SE extensional shearing under retrograde greenschist-facies conditions. D4

Figure 5. Structural cross sections of the Sesia Zone and adjacent tectonic units. (a) Val Chiusella section; (b) Castelnuovo section; (c) and (d) Val d'Aosta sections; (e) Val Sesia section; (f) Val Sermenza section. Symbols indicate fabric domain and kinematics as depicted in the legend. Location of cross sections is shown in Figures 2 and 3.

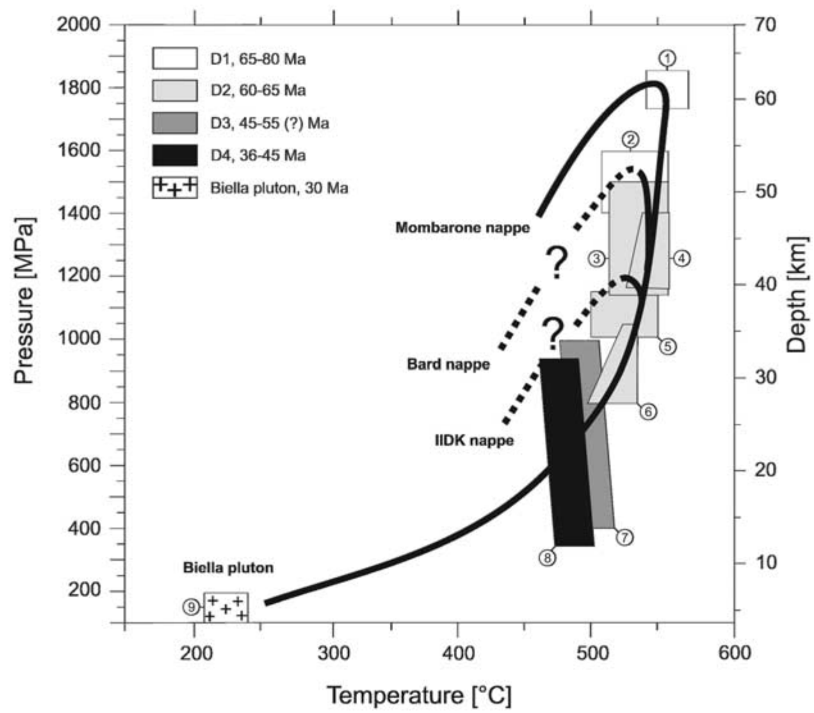


Figure 7. Pressure versus temperature diagram showing the thermobarometric evolution of the Mombarone, Bard, and IIDK nappes in the vicinity of the cross section in Figure 5d. PT curve for the Mombarone nappe is from thermodynamic modeling of *Konrad-Schmolke et al.* [2006]; curves for the Bard and IIDK nappes are constructed from constraints within the boxes. Dashed parts of curves are speculative. Numbered boxes show PT constraints obtained from the following petrological work: (1) peak conditions in the Mombarone nappe [*Lardeaux et al.*, 1982]; (2) peak conditions in the Bard nappe [*Lardeaux and Spalla*, 1991]; (3) peak to late D1 conditions in the Bonze unit [*Venturini*, 1995]; (4) early D2 conditions from thermodynamic forward modeling of glaucophane zonation patterns [*Konrad-Schmolke et al*., 2005]; (5) thermobarometric multiequilibrium calculation (TWQ 2.02 after *Berman* [1988]) of the syn-D2 assemblage albite, glaucophane, garnet, chlorite, quartz; (6) late D2 conditions from thermodynamic forward modeling of glaucophane-calcic amphibole zonation patterns [*Konrad-Schmolke et al*., 2005]); (7) conditions from syn-D3 calcic amphiboles obtained from amphibole-feldspar geothermometer of *Holland and Blundy* [1994] in the work by *Konrad-Schmolke et al*. [2005]; (8) conditions from D4 calcic amphiboles obtained from amphibole-feldspar geothermometer of *Holland and Blundy* [1994] in the work by *Konrad-Schmolke et al*. [2005]; (9) emplacement depth of Biella pluton [*Rosenberg*, 2004]. Shading of boxes indicates the Alpine fabric domain. Age ranges for the Alpine fabric domains are based on radiometric, stratigraphic and paleobotanical data discussed in the text. Depth axis calculated with an assumed density of 2.8 g/cm^3.

extensional shearing exhumed HP rocks in the footwall of the Sesia Zone, whereas D5 Insubric deformation effected differential exhumation only of the northeastern part of the Sesia Zone. The Sesia-Ivrea contact was a site of repeated Alpine movement, beginning with subduction and recurring during all of the exhumational phases described above.

4. Pressure-Temperature-Time Evolution of HP Rocks in the Sesia Zone

[28] Figure 7 shows the pressure-temperature-time (PTt) paths of the Mombarone, Bard and IIDK nappes in the vicinity of the cross section in Figure 5d, i.e., away from areas with pervasive D5 deformation. Age ranges for the Alpine fabric domains are based on radiometric, stratigraphic and paleobotanical data discussed and cited below. We point out that the nappes' PT evolutions were staggered in time, based on considerations of structural and metamorphic overprinting relationships. Unfortunately, constraints on the absolute ages are poor pending further geochronological work.

[29] The age of prograde blueschist- to eclogite-facies metamorphism (D1) is constrained only by high-retentivity ages in the Mombarone nappe that range from 65 Ma (U-Pb SHRIMP zircon ages on leucocratic layers in mafic eclogite [*Rubatto et al.*, 1999]) to 69 Ma (Hf-Lu ages on coexisting garnet and phengite [*Duchêne et al.*, 1997]). These ages fall within a broad range of 60–90 Ma Rb-Sr and $^{40}Ar/^{39}Ar$

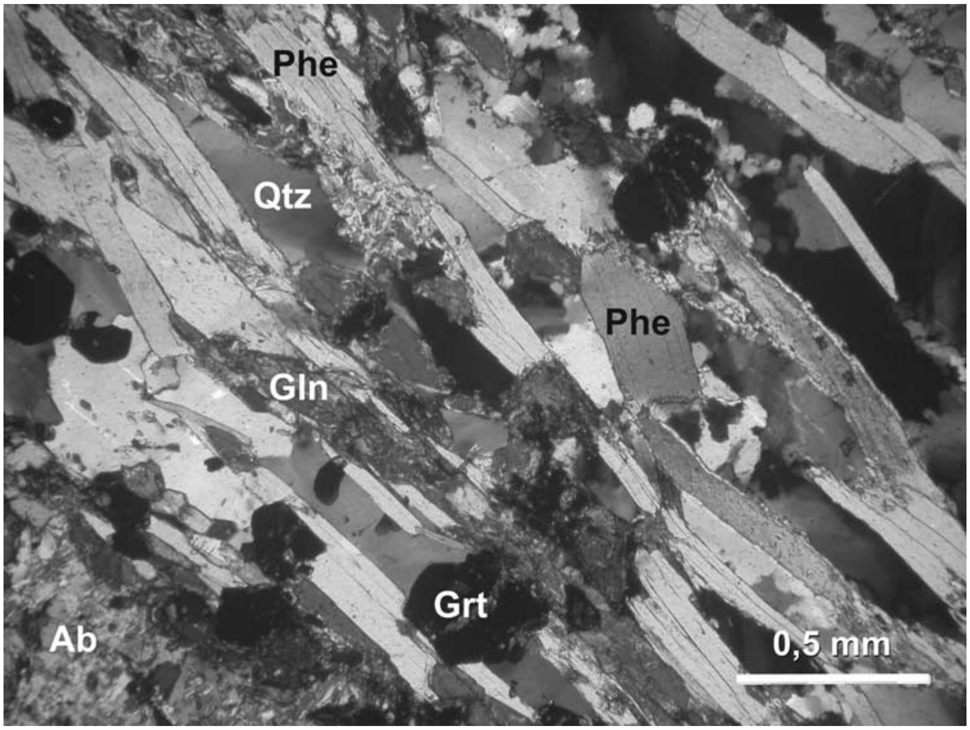


Figure 8. Microstructural relationship between zoned D2 glaucophanes and phengites. Rb-Sr phengite-feldspar isochron yields a younger age limit for D2 Chiusella shearing of ~64 Ma (see Appendix A). Sample MK52, location, Val Chiusella; H, 5043395; R, 396392 (referred to Gauss-Boaga net).

phengite ages [*Desmons et al.*, 1999; *Handy and Oberhänsli*, 2004, and references therein]. Eclogite-facies metamorphism in the Bard nappe has not been dated yet. The fact that peak pressures in the Mombarone nappe exceed those in the Bard nappe, plus that both units experienced D2 shearing at somewhat lower pressures (Figure 7) indicates that the Mombarone nappe was exhumed relative to the Bard nappe before or during the early stages of D2 Chiusella shearing. Certainly, thrusting along the nappe contact marked by the Bonze unit (Figure 2) preceded greenschist-facies shearing of both basement nappes by the end of D2.

[30] Synkinematic, coarse-grained phengite and albite making up S2 (Figure 8) yield a 64 Ma Rb-Sr phengite-feldspar minimum age of D2 deformation (see Appendix A, Figure A1). This age falls near the younger limit of the 60–90 Ma Rb-Sr phengite ages cited above, all from the Mombarone nappe, spanning the time for D1, as well as for the D2 transition from eclogite-facies to retrograde blueschist- and greenschist-facies conditions.

[31] The near-isothermal decompression path for D2 in the Mombarone nappe (Figure 7) combined with the synkinematic, blueschist-to-greenschist-facies field gradient across the Chiusella shear zone suggest that Late Cretaceous D2 exhumation was rapid, possibly occurring within only a few million years of the baric peak.

[32] Peak pressures in the IIDK nappe (1000–1200 MPa) are less than those attained in the Mombarone and Bard nappes (Figure 7), but within the same range of peak values obtained for D2 Chiusella shearing (1050 MPa [*Konrad-Schmolke*, 2006]). This corroborates previous work indicating that the blueschist facies metamorphism is prograde in the IIDK nappe, but retrograde with respect to the eclogite-facies assemblages in the Mombarone nappe [*Ridley*, 1989]. Thus the IIDK nappe was juxtaposed with the other two basement units under blueschist-facies conditions [*Dal Piaz et al.*, 1971; *Pognante et al.*, 1987; *Ridley*, 1989] during D2 Chiusella shearing, but before D3, which affected all three basement units.

[33] D3 Ometto extensional shearing involved pronounced cooling relative to decompression (Figure 7) in all three basement units. It obviously occurred after D2, but its maximum age is otherwise unconstrained. A younger age limit for D3 is provided by an undeformed trachy-andesitic dike that cuts the S2–S3 composite foliation in fabric domain 3a of the Mombarone nappe (Figure 9). A Rb-Sr feldspar-hornblende isochron defines an age of 44 Ma, which is interpreted as the minimum age of dike intrusion and therefore also of D3 (Appendix A, Figure A2). This youngest D3 age is consistent with overprinting relations along the northwestern margin of the Bard nappe indicating that D3 must be older than the onset of D4 Gressoney shearing at about 45 Ma (Rb-Sr phengite ages of *Reddy et al.* [2003]).

[34] D4 Gressoney extensional shearing in the Liguro-Piemont units lasted until about 36 Ma [*Reddy et al.*, 1999] but probably stopped much earlier in the overlying Bard nappe (circa 40–45 Ma) based on the progressive decrease

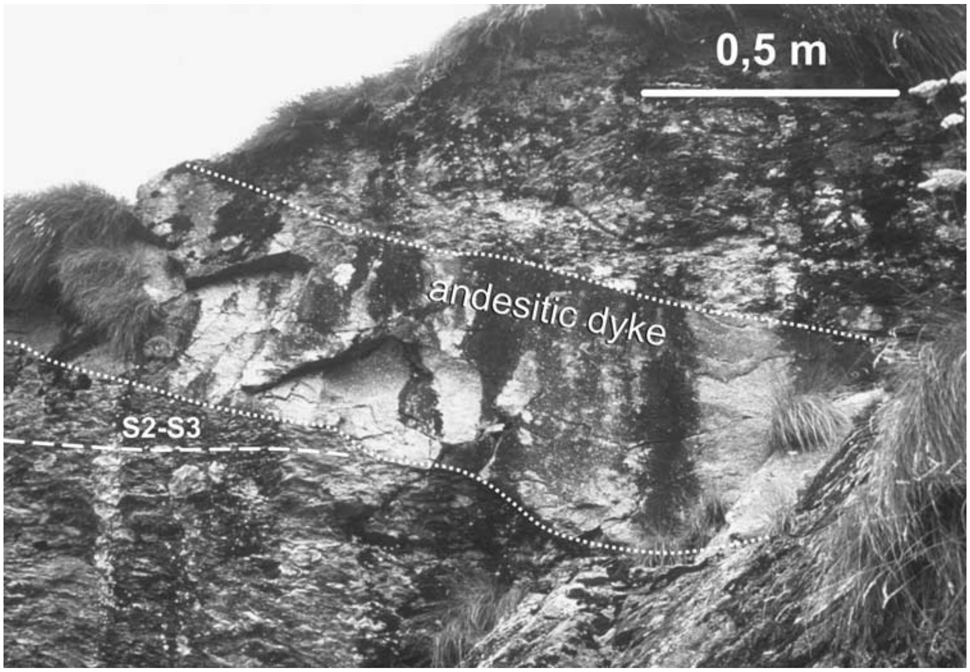


Figure 9. Undeformed 44 Ma trachyandesitic dike truncates S2–S3 composite schistosity in fabric domain 3a of the Mombarone nappe. This dike places a younger age limit on activity of the Ometto shear zone. Location is Val Chiusella; Rb-Sr age data are discussed in Appendix A.

in Rb-Sr phengite ages down into the Tertiary nappe pile [*Reddy et al.*, 1999]. The Gressoney shear zone affected the base of the Sesia Zone, so it cannot have exhumed the Sesia HP rocks in its hanging wall. However, it is potentially responsible for exhumation of Early Tertiary HP rocks in its footwall, as discussed in section 5. From 45 Ma until the end of D4 shearing at 36 Ma, the Bard and Mombarone nappes experienced only modest decompression (Figure 7). During this time and until 30 Ma, both nappes cooled from about 440° [*Konrad-Schmolke*, 2006] to below 225°C (Figure 7), as documented by Rb-Sr mica cooling ages [*Hunziker et al.*, 1992; *Reddy et al.*, 1999] and zircon fission track ages [*Hurford and Hunziker*, 1985; *Hurford et al.*, 1991; *Bistacchi et al.*, 2001]. This prolonged cooling may be responsible for the partial annealing of D3 and older fabrics in the Sesia Zone. The Oligocene Biella and Traversella plutons are shallow intrusions (~5 km [*Rosenberg*, 2004]), and the part of the Mombarone nappe adjacent to the Ivrea Zone is overlain by 33 Ma volcanoclastic sediments containing boulders of Sesia eclogite [*Scheuring et al.*, 1974]. Thus the HP rocks of the central Sesia Zone had reached the surface before the onset of D5 deformation.

[35] D5 mylonitic backthrusting along the Sesia-Ivrea contact exhumed the northeastern part of the Sesia Zone by some 10 km in the interval of 30–19 Ma [*Handy et al.*, 2005, and references therein]. Zircon and apatite fission track ages in this area are only about 27 and 9 Ma, respectively [*Hurford et al.*, 1991]. This contrasts with the lack of ductile D5 deformation and exhumation of the Sesia Zone southwest of the section in Figure 5e, where zircon and apatite yield ages are 32–39 Ma [*Hurford et al.*, 1991] and 24–29 Ma [*Bistacchi et al.*, 2001]. Thus post-Insubric exhumation of the northeastern part of the Sesia Zone probably occurred less than 9 Myr ago and amounted to no more than a few kilometers. This may be related to post-D5 brittle faulting observed throughout the area (e.g., Aosta-Ranzola and Ospizio-Sottile faults, Figures 2 and 3 [*Diamond and Wiedenbeck*, 1986; *Hurford et al.*, 1991; *Bistacchi et al.*, 2001]).

[36] To conclude this section, the HP rocks of the Sesia Zone were exhumed mostly in Late Cretaceous time, during D2 Chiusella transpressional shearing, and to a lesser extent in Early Tertiary time in the footwall of the D3 Ometto extensional shear zone. The HP rocks cooled slowly from 45 to 30 Ma, when they were exposed to erosion at the surface. Since then, brittle deformation had little effect on D2 and D3 structures, except in the northeastern part of the Sesia Zone where all tectonic units remained buried at about 10 km depth until their exhumation due to middle Tertiary D5 backfolding and backthrusting and post-Oligocene brittle faulting.

5. A Kinematic Model for Exhuming HP Rocks

5.1. Original Orientation of Subduction and Exhumation Structures

[37] Any kinematic reconstruction must account for the modifying effects of post-Early Tertiary deformation on the orientation of D1, D2 and D3 structures responsible for subducting and exhuming the HP rocks in the Sesia Zone.

Fortunately, most of this late deformation is restricted to the southeastern margin of the Mombarone nappe, south of the Biella pluton (Figure 2) where brittle faults at the confluence of the Sesia-Ivrea contact and the Cremosina fault overprint mylonites described above (Figure 3). There, paleomagnetic data of *Lanza* [1977, 1984] from the Oligocene volcanoclastic cover of the Mombarone nappe indicate that the adjacent Sesia basement rocks underwent a post-Oligocene clockwise rotation of some 40° looking NNE about a horizontal axis. This rotation diminishes to 0° toward the axial trace of the Monte Mars antiform within the Sesia Zone, suggesting that at least some of the 40° rotation can be attributed to folding in the southeastern limb of this F5 antiform (Figures 3, 5c, and 5d), as already suspected by *Lanza* [1977] and *Gosso et al.* [1979]. Other post-Oligocene brittle faults incurred minor displacement (2000–3000 m of north block down extensional displacement on the Aosta-Ranzola normal fault (Figures 2 and 3 [*Diamond and Wiedenbeck*, 1986; *Hurford et al.*, 1991; *Bistacchi and Massironi*, 2000; *Bistacchi et al.*, 2001]) and are located away from areas containing HP rocks (e.g., Ospizio-Sottile Fault, Figure 5e [*Bistacchi and Massironi*, 2000]). The kinematic model discussed below therefore applies to that part of the Sesia Zone which contains HP rocks and is unaffected by D5 faulting and differential exhumation.

[38] In comparison to the Sesia Zone, the Ivrea Zone experienced much more Tertiary block rotation [*Schmid et al.*, 1989; *Handy et al.*, 1999]. This entailed a post-Oligocene clockwise rotation of 60–85° looking NNE about a NNE-SSW trending, subhorizontal rotational axis. As discussed above, this rotation implies that the Late Cretaceous, greenschist-facies mylonite belt at the northwestern rim of the Ivrea Zone (blue in Figure 3) was originally gently ESE dipping.

5.2. From Rifting to Subduction and Early Exhumation

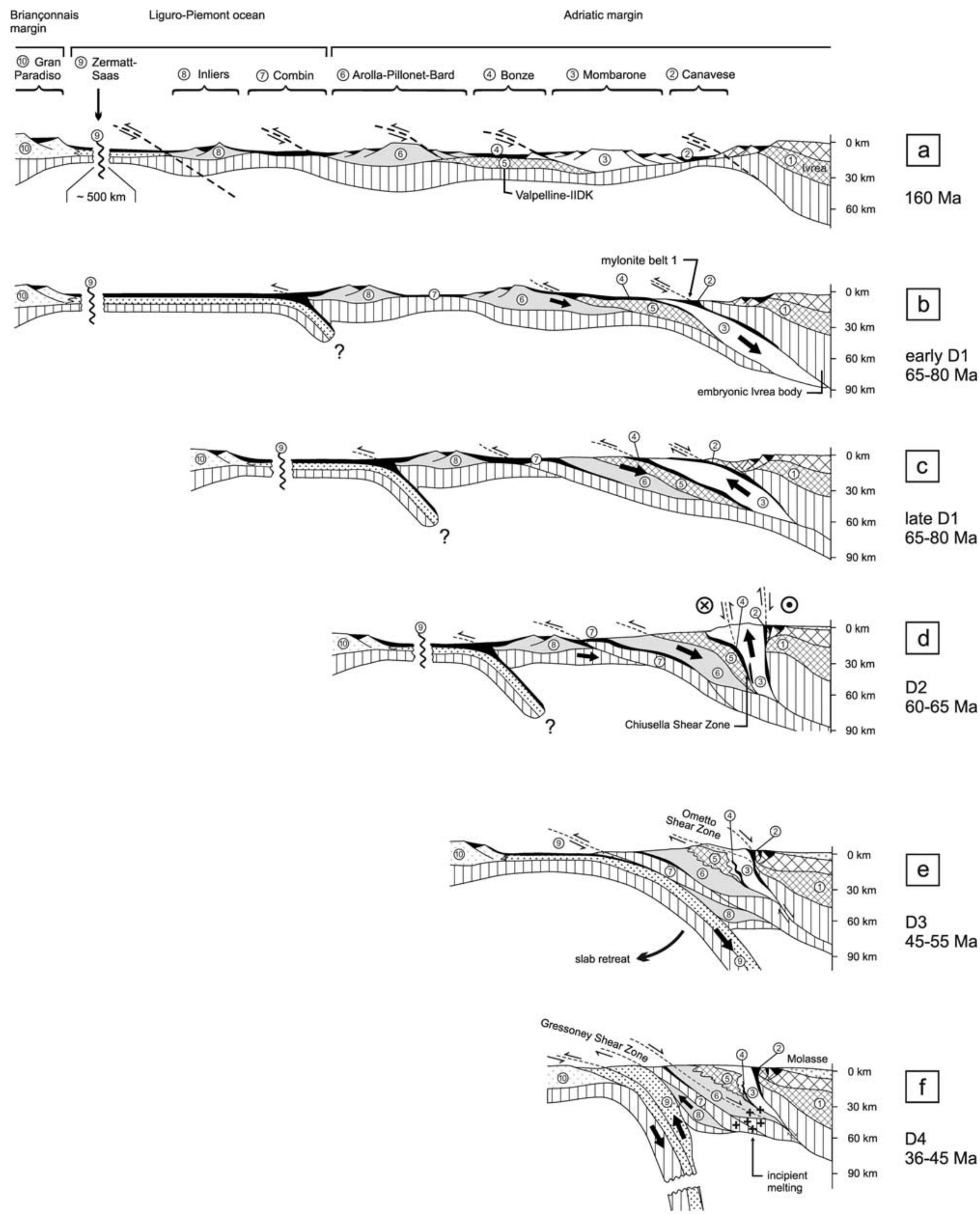
[39] The individualization of basement nappes along narrow zones containing Early Mesozoic metasediments in the Sesia Zone reflects the important effect of rift tectonics on later subduction and exhumation in the western Alps. The cross sections in Figure 10 show the evolution of the Adriatic continental margin since the end of rifting and subsidence in Jurassic time. We emphasize the importance of strike-slip motions perpendicular to these sections for the exhumation history, as discussed below.

[40] The structurally highest Austroalpine units in the western Alps are interpreted to have formed extensional allochthons at the distal part of this margin [*Dal Piaz et al.*, 2001] just prior to oblique-sinistral opening of the Liguro-Piemont ocean in Middle Jurassic time (units 3, 6, and 8 in Figure 10a [e.g., *Froitzheim et al.*, 1996]). We positioned the Canavese and Bonze metasediments (units 2 and 4 in Figure 10a) in accordance with their present position, respectively, at the junction of the Ivrea crustal section and the Mombarone nappe, and along the Mombarone-Bard nappe contact. However, we hasten to add that all of these units may have occupied originally different positions in map view, i.e., out of the plane of the sections in Figure 10a, due an unknown amount of strike-slip motion during D1 and D2. The occurrence of metabasites in the Bonze unit as well as the preservation of fragments of lower crust and fault rocks in the Canavese metasediments [*Ferrando et al.*, 2004] suggest that these units were the sites of Early Jurassic extensional faulting at or near the continent-ocean transition. In the case of the Bonze unit, this extensional faulting may have emplaced the Mesozoic sediments directly onto thinned, pre-Mesozoic lower continental crust represented by Paleozoic metabasite (intrusion age of 350 Ma [*Rubatto et al.*, 1999]).

[41] Incipient accretion and possibly also oceanic subduction documented by Late Cretaceous flysch preserved in klippe in the western Alpine foreland (Figure 1) [e.g., *Matter et al.*, 1980] are interpreted to have coincided broadly with early D1 thrusting along the transitional oceanic-distal continental margin, including future units of the Sesia Zone (Figure 10b). Much of this thrusting is inferred to have localized at the necks of the extensional allochthons, leading to the individuation of the basement nappes (units 3, 5, 6, and 8 in Figure 10) separated by Mesozoic metasediments, as seen today. The southern border of this broad zone of accretion and subduction is marked by prograde blueschist-facies assemblages in the Mesozoic Canavese sediments along the Sesia-Ivrea contact (Figure 6), and possibly also by the Late Cretaceous, greenschist-facies mylonite belt at the rim of the Ivrea Zone (marked blue in Figure 3) that accommodated WSW directed thrusting of the Ivrea Zone. The latter belt may represent a higher structural level within the subduction-accretion complex and was later juxtaposed with blueschist-facies mylonites along the Sesia-Ivrea contact [e.g., *Pognante*, 1989] during D2 and D3 exhumation.

[42] Three nappes developed within the accretionary Austroalpine margin, as shown in Figure 10b. The Mombarone nappe formed the internal part of this margin (unit 3 in Figure 10). At the external base of this margin, the Bard nappe and part of the subducting Dent Blanche unit are depicted as a continuous thrust sheet (unit 6 in Figure 10) due to their close lithological affinity (gneiss minuti of the Bard nappe correlated with the Arolla Series of the lower part of the Dent Blanche klippe [*Compagnoni et al.*, 1977]). Likewise, the similarity of lower crustal rocks in the upper part of the Dent Blanche klippe (the Valpelline Series) to the pre-Alpine IIDK lithologies [*Compagnoni et al.*, 1977] betrays a common origin within the former passive margin, possibly beneath the Bonze metasediments (unit 5 in Figure 10a). Note that our derivation of the IIDK-Valpelline nappe from between the Arolla-Pillonet-Bard and Mombarone units contrasts with earlier interpretations in which the IIDK nappe is derived directly from the Ivrea Zone [e.g., *Compagnoni et al.*, 1977]. Today, the basal thrust of the Bard and Dent Blanche units, as well as the overlying thrust within the Dent Blanche klippe and at the base of the IIDK nappe are no longer continuously traceable at the surface due to Tertiary backfolding and erosion (Figures 2 and 10g).

[43] The Mombarone nappe is interpreted to have reached its maximum depth of subduction during early D1, somewhat before maximum subduction of the Bard nappe during



Briançonnais margin
Liguro-Piemont ocean
Adriatic margin
⑩ Gran Paradiso
⑨ Zermatt-Saas
⑧ Inliers
⑦ Combin
⑥ Arolla-Pillonet-Bard
④ Bonze
③ Mombarone
② Canavese
~ 500 km
Ivrea
Valpelline-IIDK
0 km
30 km
60 km
90 km
a
160 Ma
mylonite belt 1
embryonic Ivrea body
b
early D1
65-80 Ma
c
late D1
65-80 Ma
Chiusella Shear Zone
d
D2
60-65 Ma
Ometto Shear Zone
slab retreat
e
D3
45-55 Ma
Gressoney Shear Zone
Molasse
incipient melting
f
D4
36-45 Ma

Figure 10

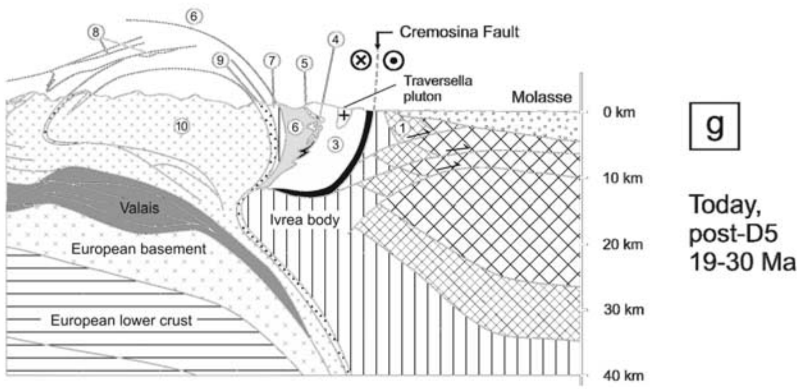


Figure 10. (continued)

late D1 (Figure 10c). This NW propagating, in-sequence subduction of the distal Adriatic margin is considered likelier than out-of-sequence subduction, which would require normal fault motion along the Bard-Mombarone nappe contact during D1. No such normal fault is observed. Initial exhumation of the Mombarone nappe from greater depth is inferred to have involved D1 thrusting at its base in order to account for partial overprinting of the Mombarone-Bard nappe contact by the D2 Chiusella shear zone. Subduction and initial exhumation during D1 progressed from SE to NW, i.e., from right to left in Figures 10b and 10c. Accordingly, exhumation of the Mombarone nappe occurred above the downgoing Bonze, II DK and Bard units, and directly beneath a rigid mantle wedge ("embryonic Ivrea body") that formed when the distal continental margin was subducted beneath cold lithospheric mantle that had already been extensionally exhumed during Early Mesozoic rifting [*Schmid et al.*, 1987]. This may explain the local imbrication of small slivers of serpentinized ultramafic rock within eclogitic schists of the Mombarone nappe, as mapped by *Ferraris and Compagnoni* [2003] and *Pognante et al.* [1980].

5.3. Transpressional Exhumation

[44] D2 shearing is responsible for transpressional exhumation of the HP rocks in earliest Tertiary time (Figure 10d). The retrograde blueschist- to greenschist-facies zonation across the steep D2 Chiusella Shear Zone (Figures 5a and 5c) is consistent with rapid, nearly isothermal east-side-up exhumation of the Mombarone nappe with respect to the Bard nappe. However, subhorizontal to moderately plunging Ls2 stretching lineations (Figure 3) suggest strike-slip or oblique-slip motion. Taken together, these structural and metamorphic features indicate an overall transpressional setting, as depicted in Figure 10d. As argued in section 5.2, D2 transpressional shearing also brought the IIDK nappe to its greatest depth, where it was juxtaposed with the Bard and Mombarone nappes.

[45] For kinematic reasons, D2 transpression must have involved another major shear zone along the Sesia-Ivrea contact to accommodate west-side-up exhumation of the Mombarone nappe with respect to the Alpine unmetamorphosed rocks of the Ivrea Zone (Figure 10d). This shear zone remains a kinematic necessity in need of verification in the field, as no Late Cretaceous mylonite belt with kinematics appropriate for exhuming the Sesia Zone has been found to date. Possibly, such a mylonite belt was obliterated by D3 shearing.

[46] Regarded in a plate tectonic context, overall dextral transpression during D1 and D2 is consistent with paleomagnetic and paleogeographical data calling for eastward displacement of the Ibero-Briançonnais peninsula with respect to both Europe and Adria from Late Cretaceous to Early Tertiary time [e.g., *Dewey et al.*, 1989; *Stampfli et al.*, 1998]. Estimates of this displacement vary with the plate tectonic reconstruction, but most authors agree that a minimum displacement of some 500 kilometers is necessary

Figure 10. Model for the evolution of the Adriatic continental margin and Tethyan ocean since Early Mesozoic time: (a) Adriatic passive margin with location of future thrust planes during subduction; (b) incipient subduction of the Adriatic margin, early D1 thrusting and HP metamorphism of the Mombarone nappe (unit 3); (c) subduction of the Bard and IIDK nappes (units 5 and 6) in external parts of the Adriatic margin during late D1 thrusting and initial exhumation of the Mombarone nappe; (d) D2 transpressional exhumation of the Adriatic margin involving thrusting and strike-slip shearing during high-stress oblique subduction of LP oceanic lithosphere; (e) D3 extensional exhumation of Adriatic margin in the upper plate of a low-stress, retreating subduction zone; subduction and HP-UHP metamorphism of LP oceanic lithosphere; (f) slow cooling and erosional denudation of the Adriatic margin during rapid D4 exhumation of HP-UHP rocks of the LP oceanic lithosphere, subduction of European (Briançonnais) and Valais oceanic lithosphere, incipient magmatism; (g) present lithospheric structure along ECORS-CROP section (Figure 1) showing D5 doming, Insubric shearing, and Oligocene plutons. Note change in scale of Figure 10g with respect to Figures 10a–10f.

to allow for subduction of the Liguro-Piemontese ocean [e.g., *Schmid et al.*, 2004]. We suspect that only a modest proportion of this amount was taken up by D2 shear zones in the accreted continental crust represented by the Sesia Zone; most was accommodated by oblique subduction along a Late Cretaceous to Early Tertiary trench at the southern limit of the Liguro-Piemont ocean, as documented by Late Cretaceous flysch in thrust nappes of the Préalpes [e.g., *Schmid et al.*, 2004, Figure 1].

5.4. Precollisional Extensional Exhumation

[47] Early Tertiary D3 deformation exhumed the Sesia Zone in the footwall of the Ometto extensional shear zone (Figure 10e), bringing the HP rocks to depths of about 15–20 km within the orogenic wedge. The southern limit of this extensional shear zone is the southeastern margin of the Mombarone nappe along the Sesia-Ivrea contact (Figure 3). The younger age limit of Ometto shearing (45 Ma) falls within the 44–50 Ma age range for HP and UHP metamorphism in the underlying Liguro-Piemontese units and Austroalpine inliers [*Dal Piaz et al.*, 2001; *Handy and Oberhänsli*, 2004, and references therein]. This supports the idea that D3 extensional shearing in the Sesia Zone was coeval with subduction of oceanic lithosphere in its footwall, as depicted in Figure 10e. Thus the entire Sesia Zone containing partly exhumed HP rocks from different depths of the accreted basement wedge was extruded toward the foreland and juxtaposed in its roof with rigid pre-Alpine rocks of the Ivrea Zone. We note that *Inger and Ramsbotham* [1997] also proposed coeval extension and thrusting as a means of exhuming the HP rocks in the Mombarone nappe (containing their "eclogitici micaschisti unit") but that they instead related D3 extensional shearing to purported blueschist-facies thrusting at the contact with the Bard nappe (containing their "gneiss minuti unit"). As shown in section 5.3, blueschist-facies mylonites at this contact [*Avigad*, 1996] are actually related to D2 Chiusella shearing.

[48] D4 represents a continuation of top-SE extensional shearing to beneath the Sesia Zone (Figure 10f) within the Liguro-Piemont units (Gressoney shear zone of *Wheeler and Butler* [1993]). It effected extensional exhumation of HP and UHP rocks in the oceanic Zermatt-Saas units [*Reddy et al.*, 2003; *Bucher et al.*, 2003]. In the hanging wall of this extensional system, the Sesia Zone experienced only slow cooling and minor exhumation (Figure 7). We attribute most of this exhumation to erosional denudation, as documented by the influx of detritus derived from the Sesia Zone in Oligo-Miocene molasse sediments of the southern Alpine Piedmont basin [*Carrapa et al.*, 2003; *Carrapa and Wijbrans*, 2003].

[49] The decrease in age of extension into the units below the Sesia Zone is paralleled by a decrease in the age of HP and UHP metamorphism, from 65–80 Ma in the Sesia Zone to 38–50 Ma in the Liguro-Piemont [*Lapen et al.*, 2003, and references therein] and Monte Rosa units [*Bucher*, 2003]. These trends point to northwestward migration of subduction toward the European continental margin, as also inferred from palinspastic reconstructions of the Penninic cover nappes in the Préalpes [*Trümpy*, 1980], and the volume and migration of flysch sedimentation in the Alps [e.g., *Royden and Burchfiel*, 1989; *Schmid et al.*, 1997] (Figure 1). Migrating D3 and D4 extension above a retreating, Early Tertiary subduction zone may have induced cooling of the upper plate, until possible slab drop-off in Oligocene time triggered final uplift and exhumation [*von Blanckenburg and Davies*, 1995], ultimately exposing the HP rocks of the Sesia Zone to erosion (Figure 10f).

5.5. Collisional Exhumation

[50] D5 backfolding of the entire Tertiary nappe stack during the late stages of collision modified the former subducted Adriatic margin, as shown in Figure 10g for the ECORP-CROP transect (Figure 1). D5 exhumation of the Sesia Zone in this section was minor, but as pointed out above, this exhumation increased toward the northeast along the strike of the western Alpine arc. This D5 exhumation gradient probably reflects an eastward increase in the amount of Oligo-Miocene north-south shortening combined with wedging of the Adriatic lower crust beneath the retrowedge of the Alpine orogen, as discussed by *Schmid and Kissling* [2000].

6. Discussion of Exhumation Mechanisms

[51] The kinematic model presented above departs from existing kinematic and physical models of exhumation in several respects. First, we have shown that the Sesia Zone is a composite unit whose basement nappes followed different subduction and exhumation paths. Second, the exhumation of the HP rocks within these nappes is multistage during Late Cretaceous and Early Tertiary time. Third, most exhumation occurred in a transpressional, precollisional setting that involved significant dextral strike-slip motion. These aspects have major implications for assessing the mechanisms of exhuming HP rocks.

[52] From the start, certain previously proposed mechanisms can be ruled out for the Sesia Zone. Exhumation of the Sesia Zone by corner flow within a low-viscosity accretionary channel bounded above by the rigid Ivrea upper mantle body [*Schmid et al.*, 1987] is incompatible with the preservation of coherent basement units within the Sesia Zone. There is no evidence in the exhumed Sesia rocks for the kind of laminar deformation required for corner flow [*Cowan and Silling*, 1978] or channel flow [*Cloos*, 1982] of a weak matrix containing HP rocks. This renders tectonic overpressure [*Mancktelow*, 1995] a rather unlikely explanation for the HP metamorphism in the Sesia Zone, a notion that has been previously doubted on mechanical grounds as well [e.g., *Burov et al.*, 2001].

[53] A period of plate-scale extensional exhumation between subduction and collisional episodes [e.g., *Ballèvre and Merle*, 1993] can be excluded in light of evidence that the convergence of Europe and Adria was uninterrupted in Late Cretaceous to Early Tertiary time [*Schmid et al.*, 1996; *Schmid and Kissling*, 2000]. Arguments against older models of exhumation in the Sesia Zone center on dated or questionable assumptions regarding the age, conditions and

kinematics of exhumation, as reviewed in the introduction. We therefore focus the remainder of this discussion on possible mechanisms for D2 transpressional exhumation, and D3 and D4 extensional exhumation.

[54] A recurring theme in discussions of exhumation is the relative roles played by buoyancy, isostasy, lithospheric rheology, and tectonic and erosional denudation. In the case of the HP rocks of the Sesia Zone, the importance of these factors changed both in time and space. The Mombarone nappe was subducted beneath a cold, dense upper mantle wedge (Figures 10b–10d), leading to significant density contrasts with its overlying rocks. Thus buoyancy is a viable force for initial exhumation of the HP rocks during late D1 and D2 transpression. Indeed, buoyancy forces may have driven exhumation up to about 20–30 km depth, especially if the rising Mombarone rocks heated and hydrated the surrounding rocks, thus reducing their viscous shear strength [*Konrad-Schmolke*, 2006]. This makes detachment and return-block motion of coherent pieces of continental crust at the top of the downgoing slab [*Chemenda et al.*, 1995] a potentially attractive mechanism for the initial exhumation of the Mombarone nappe. This mechanism can also explain the lack of erosional relics of HP continental rocks in Late Cretaceous sediments of the Southern Alps, because exhumation balanced by coeval subduction of an equivalent volume of intervening crust would preclude formation of significant topographic relief and deposition of erosional products in the hinterland (Figure 10c).

[55] Buoyancy-driven exhumation obviously did not work where the nappes containing the HP rocks were too small, where they were surrounded by rocks of similar density, or where shear resistance to their rise was too great. This pertains to the IIDK nappe and probably also to the Bard nappe, because these two units occupied the core of the subducted and accreted continental crust within the active margin (Figures 10c and 10d). Their exhumation may have involved tectonic wedging during oblique slip on steep shear surfaces in a downward tapering continental accretion zone (Figure 10d). This tapered geometry is favored at depth by the combination of confining pressure and horizontal maximum force acting during dextral transpression. Indeed, exhumed HP rocks are ubiquitous at restraining bends in large strike-slip systems [*Mann and Gordon*, 1996]. A close relationship of exhumed HP rocks to NNW trending strike-slip faults was noted in the western Alps [*Ricou and Siddans*, 1986; *Pfiffner*, 1992], but in contrast to our study, these faults were attributed to north to NW directed Tertiary motion of Adria before the ages of the faults and the HP metamorphism in the western Alps were known. The NNW trend of these purported strike-slip faults is clearly incompatible with the E-W trend of the Late Cretaceous Chiusella shear zone documented above.

[56] The switch from D2 transpression to D3 extension in the Sesia Zone was associated with an increase in cooling during decompression. This is important because cooling during D3 extensional exhumation rules out two other exhumational processes that are associated with heating. The first is thickening and heating following buoyancy-driven exhumation to the base of the crust within the upper plate. This has been proposed for HP rocks of the Variscan orogen [*O'Brien and Vrana*, 1995] and UHP rocks of the Western Gneiss Region in Norway [*Walsh and Hacker*, 2004] which experienced flattening at increasing temperatures (amphibolite-facies conditions) during their emplacement within the crust. The second is break-off of the downgoing oceanic slab, which, acting on its own, would have led to the buoyant rise and heating of the overlying, exhuming crust [*Davies and von Blanckenburg*, 1995; *Ernst et al.*, 1997]. In the Alps, the putative break-off of the Liguro-Piemont oceanic lithosphere slab occurred at about 30–35 Ma [*von Blanckenburg and Davies*, 1995], at least 10 Myr after most exhumation of the Sesia HP rocks.

[57] Therefore the mechanism that best fits the D3 thermal and kinematic history of the Sesia HP rocks involves D3 extension in the upper plate, possibly as the trench retreated to the NW under the weight of the descending Liguro-Piemont slab (Figures 10e). Heating of the extending crust in the upper plate was prevented by the accretion of cold slices of transitional (Combin unit) and oceanic (Zermatt-Saas unit) lithosphere in its footwall. These slices effectively ‘‘refrigerated’’ the upper plate and maintained a low geotherm. A similar scenario of retreating subduction was also proposed for Late Cretaceous extensional exhumation of the accreted Adriatic margin in eastern Alps [*Handy*, 1996; *Froitzheim et al.*, 1997] and, before that, for Miocene extension of the Pannonian basin behind the retreating Carpathian orogen [e.g., *Royden and Burchfiel*, 1989]. Extensional exhumation of HP rocks during thrusting and subduction has also been propounded for Eocene HP and UHP rocks in the Liguro-Piemont and Briançonnais units [e.g., *Bucher et al.*, 2003; *Wheeler et al.*, 2001]. As in the Sesia Zone, these rocks experienced decompression from blueschist- to greenschist-facies conditions [*Lapen et al.*, 2003; *Keller et al.*, 2005], precluding heating during exhumation and supporting the hinge-rollback model of subduction in Figure 10f.

[58] Sophisticated numerical codes have allowed modelers to differentiate the competing effects of rheology, metamorphic phase transitions and erosion on the PTt paths of subducted crust during Alpine-type orogenesis. *Burov et al.* [2001] proposed that subducted and metamorphosed crust at depths of about 100 km can return to the surface by a combination of buoyant rise along a crustal channel at the top of the downgoing slab, and corner flow within an orogenic wedge at depths less than 40 km. Buoyant rise within the crustal channel is facilitated by incomplete metamorphic equilibration at eclogite-facies conditions. *Gerya and Stöckhert* [2005] went further in predicting that subducted fragments of continental crust can rise buoyantly through a convectively overturning mantle wedge beneath the upper plate margin. It is very doubtful that such buoyant-convective exhumation actually occurred in the western Alps, especially given abundant structural and radiometric evidence from the Sesia and Ivrea zones that the upper plate lithosphere of the Alpine subduction system was cool (<200–300°C) and therefore rigid already in Mid-Jurassic time, well before the onset of subduction [e.g., *Zingg et al.*, 1990].

[59] Though these models are fascinating for the animated pictures they provide, their only testable features are the synthetic PT paths, the densities and the lithosphere-scale structures produced by the end of the numerical experiments. Even after taking into account problems with scaling, one finds that the model PT curves and structures differ significantly from the natural features in Figures 5 and 7. The usefulness of such models as simulations and controlled experiments is obviously limited by the degree of testability and by the choice of boundary conditions.

7. Conclusions

[60] Our work has shown that the exhumation of continental HP rocks in the Sesia Zone preceded Tertiary Alpine collision and coincided with oblique subduction of the Adriatic and Tethyan lithospheres in Late Cretaceous to Early Tertiary time. The three basement nappes making up the Sesia Zone (the Bard, Mombarone and IIDK nappes) are exhumed slivers of subducted and tectonically eroded, Adriatic continental margin. The HP rocks in these nappes experienced quite different PTt paths as subduction and exhumation proceeded from internal to external parts of the accreted continental margin, i.e., from SE to NW across the Sesia Zone.

[61] The Bard and Mombarone nappes individuated during Late Cretaceous subduction (65–80 Ma) and the thrust separating them appears to have nucleated where the continental crust was previously thinned during Jurassic rifting. The Mesozoic cover sediments of this passive margin, the Bonze and Canavese units, presently occupy narrow zones between the Bard and Mombarone nappes and the Ivrea Zone. Both sedimentary units contain blueschist- and greenschist-facies assemblages, indicating that they were also subducted and exhumed.

[62] Kilometer-scale shear zones partly overprinted the nappe contacts and exhumed coherent slices of crust containing the HP rocks. Most early exhumation involved thrusting (D1) at the base of the Mombarone nappe and transpressional shearing (D2) along a steep, E-W trending mylonitic shear zone (the Chiusella shear zone). This kilometer-wide shear zone accommodated both dip-slip and strike-slip motion under retrograde eclogite- to blueschist- to greenschist-facies conditions. Exhumation of the Mombarone nappe was nearly isothermal (500–550°C) and rapid. D2 transpressional shearing also juxtaposed the IIDK nappe with the Bard and Mombarone nappes at about 40 km before exhuming all three nappes to a common depth of about 20 km. New Rb-Sr mineral ages constrain D2 to have occurred at about 60–65 Ma.

[63] Continued exhumation of the amalgamated basement nappes to 10 km or less occurred in the footwall of a subhorizontal, greenschist-facies extensional shear zone (D3, the Ometto shear zone) that accommodated top-SE motion. This shear zone is best exposed in the highest mountain peaks of the Sesia Zone, but in lower structural levels along valley walls is manifested by recumbent folds that flatten older foliations. D3 extensional exhumation involved cooling to 400° and was broadly coeval with Early Tertiary, SE directed subduction of oceanic Liguro-Piemont units beneath the Sesia Zone. A new Rb-Sr age on a crosscutting dike combined with existing mineral ages constrains D3 to have occurred at about 45–55 Ma.

[64] From 45 to 36 Ma the Sesia Zone underwent slow cooling and denudation in the hanging wall of the Gressoney extensional shear zone (D4 [*Wheeler and Butler*, 1993; *Ballèvre and Merle*, 1993]). This greenschist-facies shear zone accommodated top-SE motion, mostly within the oceanic Liguro-Piemont units, but also along the northwestern margin of the Sesia Zone (Bard nappe). D4 extensional shearing was broadly coeval with HP and UHP metamorphism in the Zermatt-Saas and Monte Rosa units beneath the Sesia Zone, as proposed in previous work [e.g., *Reddy et al.*, 2003].

[65] At 30 Ma, the HP rocks of the Mombarone unit were intruded by the shallow Biella and Traversella plutons and incorporated as boulders within volcanoclastic sediments. Oligo-Miocene Insubric backfolding and backthrusting under greenschist facies conditions (D5) only exhumed northeastern parts of the Sesia Zone where Alpine HP metamorphism is absent. The southeastern rim of the Sesia Zone and its border with the Ivrea Zone preserve evidence of repeated movement, including blueschist- and greenschist-facies mylonite belts that are locally overprinting by cataclasites.

[66] Regarded from a plate tectonic perspective, the transition from D2 to D3 deformation in the Sesia Zone coincided with a first-order change in the nature of Tethyan subduction, from high-stress transpressional subduction and thickening of the Adriatic continental margin to low-stress subduction of the Liguro-Piemont oceanic lithosphere. This change had a profound impact on the mechanisms of exhumation within the thickened Adriatic crust. Whereas initial exhumation from depths within the mantle was probably buoyancy-driven, transpression during subduction favored tectonic wedging and forced upward extrusion within the crust. At the onset of low-stress subduction in Early Tertiary time, the main driving mechanism of exhumation switched to a combination of extension and erosion of the thickened margin above the NW retreating hinge of the subducting Liguro-Piemont oceanic lithosphere.

[67] Finally, we emphasize the value, indeed the necessity, of mapping fabrics across large tracts of deeply eroded crust. Only structures and metamorphic assemblages that are related in time and correlatable with radiometric ages can be interpreted in a regional context. This is time consuming but will form the basis for the next generation of tectonic models.

Appendix A

A1. Sample Description

[68] Sample MK52 is a white-mica bearing sodic-amphibole garnet gneiss from a steeply NNW dipping S2 shear zone in Val Chiusella (sample coordinates: H 5043395; R 396392 in a Gauss-Boaga net). Albite, sodic amphibole, coarse-grained phengite and garnet ($Prp_{5.6}$, Alm_{64}, Grs_{30}) document blueschist facies conditions. Sodic amphibole is zoned, with

Table A1. Rb and Sr Concentration and Sr Isotope Composition of the Samples MK52 and RH1 and Their Minerals[a]

Description	Rb, ppm	Sr, ppm	$^{87}Rb/^{86}Sr$	$\pm 2\sigma$	$^{87}Sr/^{86}Sr$	$\pm \sigma_m$
Sample MK 52						
Feldspar	2.464	14.53	0.4908	4	0.714677	18
White mica	117.2	73.10	4.635	42	0.718452	15
Sample RH 1						
Andesitic dike (WR)	72.12	882.2	0.2365	21	0.708022	52
WR – cc[b]	32.41	557.6	0.1682	15	0.707907	13
Calcite[b]	1.9527	442.7	0.01276	12	0.708523	17
Plagioclase	200.5	849.0	0.6832	61	0.708092	17
Amphibole	1.578	603.5	0.007565	68	0.707668	16
Epidote	2885	1.988	0.001993	18	0.707748	18

[a]Errors in element concentrations are smaller than 0.3%.
[b]$C_2H_3O_2Cl$-leachate.

barroisitic amphibole at the rim. The modal content of barrosite in the sample is estimated to be less than 1%. Coarse-grained phengite and Na-amphibole are oriented parallel to S2 (Figure 8). Fe and Mg contents increase at the rims of the amphiboles. A second generation of fine-grained, muscovitic mica in patchy aggregates without any preferred orientation forms rims on phengite, indicating a greenschist-facies overprint. The Rb and Sr concentrations and isotopic compositions are given in Table A1.

[69] Sample RH1 is from the fine-grained andesitic dike pictured in Figure 9 (sample coordinates H 5043434; R 0395278 in a Gauss-Boaga net, altitude 1395 m above sea level). Ca-amphibole and plagioclase in this sample show magmatic features and contain no traces of deformation. Magmatic epidote and plagioclase with oscillatory zonation form 1–5 mm long phenocrysts and are interpreted to represent a quenched melt. Epidote has magmatically resorbed grain boundaries. Smaller plagioclase laths with polysynthetic twins occur interstitially. Magmatic Ca-amphibole is also found in the matrix. Accessory minerals are titanomagnetite and, very rarely, pyrite. Secondary calcite is attributed to hydrothermal alteration that fills tiny vein as well as forming amoeboid aggregates within the matrix.

A2. Sample Preparation

[70] Both samples (MK52 and RH1) were crushed and sieved. Minerals from the sieved fractions were separated by conventional techniques (magnetic separation, heavy liquids, and hand picking). The minerals were dissolved in a mixture (4:1) of hydrofluoric and nitric acid over five days. After drying, the samples were diluted in hydrochloric acid (2.5 M) and divided into two aliquots, one for measuring the isotopic composition and the other for determining the concentration of Sr and Rb by adding spike solutions of both elements. After separating Sr from Rb with ion chromatographic columns, the isotopic composition and the concentration were determined in a mass spectrometer (Finnigan MAT 261). During the analysis, the $^{87}Sr/^{86}Sr$ of a standard (NBS 987) was measured at 0.71026 ± 4 (2σ; n = 12). The standard error of the concentration was estimated to be smaller than 0.4%. Errors of the ratios are given in Table A1.

[71] Because of alteration of sample RH1, the specimen was leached with ultrapure chloro-acetic acid to dissolve calcite. The insoluble residual (WR-cc) was treated by the same procedure as the other rock constituents. The results are given in Table A1.

A3. Results and Interpretation

[72] Coexisting albite and phengite from sample MK52 yield an age of 63.6 ± 0.8 Ma (Figure A1). We interpret this value as a minimum age for the formation of S2. This is a

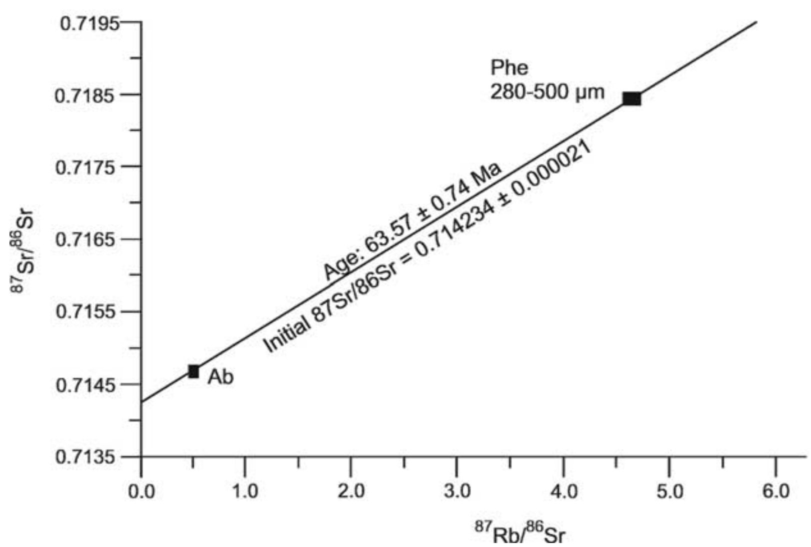


Figure A1. Rb-Sr isochron diagram of a white-mica-bearing sodic-amphibole garnet gneiss from a steeply NNW dipping D2 shear zone in Val Chiusella (sample MK 52). Ab, albite; Phe, phengite. The slope of the line connecting albite and phengite defines an age of 63.57 ± 0.74 Ma with an initial Sr ratio of 0.714234 ± 0.000021.

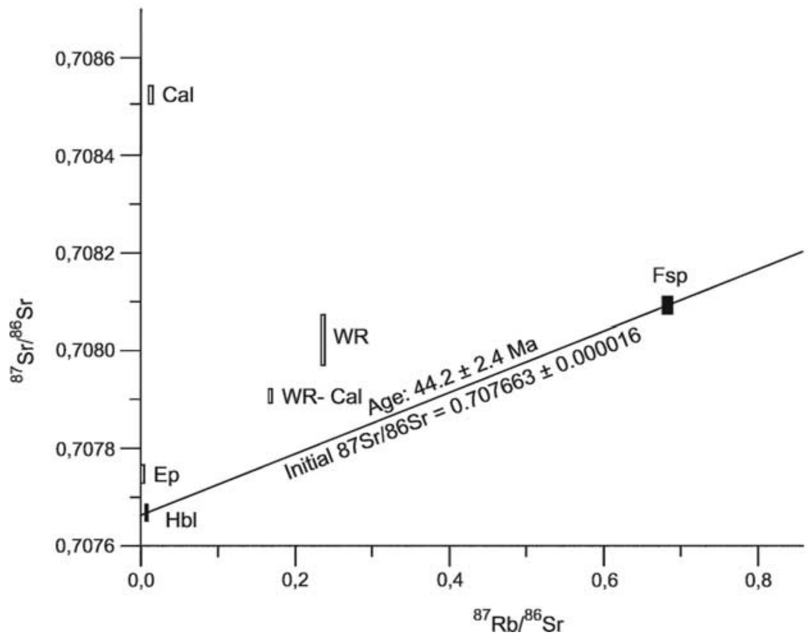


Figure A2. Rb-Sr isochron diagram of a trachyandesitic dike (sample RH 1). Plag, plagioclase; Amph, amphibole; Ep, epidote; Cc, leachate of chloro-acetic acid; WR, whole rock; WR–Cc, whole rock without the chloro-acetic acid leachate. The slope of the line between plagioclase and amphibole defines an age of 44.2 ± 2.4 Ma with an initial Sr ratio of 0.707663 ± 16.

minimum age of D2 deformation because, despite our efforts to separate the grain size fraction carefully, we cannot preclude the possibility that some fine-grained mica remained in the coarse-grained phengite fraction to lower its age value.

[73] All minerals and the whole rock of sample RH1 have $^{87}Sr/^{86}Sr$ ratios between 0.7077 and 0.7080 (Figure A2). However, calcite has a very high $^{87}Sr/^{86}Sr$ ratio (0.7085) despite its very low Rb content (1.95 ppm), indicating that this radionuclide did not contribute to the high ratio. The source of this high Sr ratio is a probably a secondary hydrothermal fluid, derived from continental crust. WR – cc has a lower $^{87}Sr/^{86}Sr$ ratio than the whole rock sample, in agreement with the leached out calcite. An age of 44.2 ± 2.4 Ma was calculated from the slope of the line connecting plagioclase and amphibole (Figure A2). These minerals are probably less affected by alteration. The fact that magmatic epidote does not lie on the line defined by plagioclase and amphibole indicates that at least part of its Sr is hydrothermally influenced. Therefore we cannot exclude the possibility that plagioclase and amphibole are partly altered and we therefore interpret the calculated age to be a minimum age for the emplacement of the dike.

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