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Tectonometamorphic history of the Ivrea Zone and its relationship to the crustal evolution of the Southern Alps

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ABSTRACT

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The Ivrea Zone represents a cross-section through the lower continental crust of the Southern Alpine basement. A first long episode with regional metamorphism and associated polyphase deformation ended in the Variscan. Subsequent Late Paleozoic magmatic activity may have occurred during an early stage of crustal attenuation. Late Paleozoic and Early Mesozoic crustal thinning is accommodated by conjugate high-temperature shear zones within the granulite facies Ivrea Zone and by low-angle normal faulting within the Pogallo Ductile Fault Zone, at the base of the intermediate crust. The age and kinematics of the Pogallo Ductile Fault Zone are consistent with the occurrence of Early Mesozoic extensional basins in the Southern Alpine sediments. Exhumation and final steepening of the Ivrea Zone during the Alpine Orogeny did not substantially alter its internal structure except in the vicinity of the Insubric Line. Thus, the Ivrea Zone and the adjacent Strona–Ceneri Zone represent a good example of highly attenuated lower and intermediate continental crust.

Introduction

Exposed crustal profiles provide a valuable reference for the interpretation and refinement of geophysical profiles and models of the continental lithosphere (e.g., Fountain and Salisbury, 1981). Recent efforts to improve understanding of the lower continental crust have led to renewed interest in the Ivrea Zone, a cross-section through the Southern Alpine lower continental crust outcropping in Northwest Italy and South Switzerland. Early models of the Ivrea Zone focussed on its present lithological and regional metamorphic configuration as an analogue for deep continental crust (Giese, 1968; Mehnert, 1975; Fountain, 1976), but only recently have attempts been made to integrate the geologic history of the Ivrea Zone with the notion of evolving geophysical properties in the lower crust (Fountain, 1986). This reflects the need for understanding the tectonometamorphic history of the Ivrea Zone within the context of crustal evolution in the Southern Alps. The fact that the crust observed at the surface today is the product of a complex, possibly unique geotectonic evolution poses a fundamental problem in relating exposed crustal sections such as the Ivrea Zone to the large masses of unexposed continental crust which are imaged in deep seismic profiles of the lithosphere.

This paper considers the relationships of metamorphic, magmatic and tectonic events in the Ivrea Zone and adjacent units of the Southern Alps. Special attention is given to the age, conditions, and kinematics of shear zone activity, as well as to structures that are associated with the emplacement of the Ivrea crustal section into shallower crustal levels. The Ivrea Zone can only be used as an analogue for modelling of lower con-

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tinental crust if the exhumation process did not extensively alter its original structural, physical and chemical properties.

Geological setting

The Ivrea Zone is the surfacial expression of a large positive gravimetric and seismic anomaly (the Ivrea geophysical body) in the internal part of the Western Alpine arc (Figs. 1, 2, 3D). Modelling of this geophysical anomaly has led to the concept of an arcuate, SE-dipping sliver of Southern Alpine crust and mantle that wedges into Austroalpine and Penninic crustal rocks to the north and west (Berckhemer, 1968; Giese, 1968 and others; more recently, Kissling, 1984; Menard and Thouvenot, 1984; Wirth et al., 1987; Bayer et al., 1987, 1989, Hirn et al., 1989). The Ivrea Zone is situated where the fossilized (i.e., pre-Alpine) Southern Alpine Moho comes closest to the Earth's surface. The exposure of different levels of the Southern Alpine crust in this corner of the Alps is genetically related to this special lithospheric configuration (Fig. 3D). To the north and west the Ivrea Zone is bounded by the Insubric Line (Figs. 1 and 2).

From the originally deepest to shallowest crustal levels, the Southern Alpine crust comprises the following three Paleozoic basement units (Figs. 2 and 3): the Ivrea Zone (Zona Ivrea–Verbano or Zona Diorito–Kinzigitica in the Italian literature), the Strona–Ceneri Zone, and the Val Colla Zone. This basement is overlain by the Late Paleozoic to Tertiary sedimentary cover. All of these units are separated by faults of different ages (Pogallo, Lago Maggiore, Val Colla lines), so the section is not continuous. The originally deepest part of the Southern Alpine crustal section is cut by the NW-dipping Neogene Insubric Line. Northwest of this line, the "II. Kinzigite–Diorite" Zone (top of the Sesia Zone s.l.) and the Valpelline Series (top of the Dent Blanche Nappe) are found. Both of these Austroalpine units show similar lithologies and pre-Alpine tectonometamorphic histories to the Ivrea Zone, but were incorporated into the Central Alps and overprinted by Alpine metamorphism.

Lithologies of the Ivrea Zone and adjacent units

The Ivrea Zone contains paragneisses, mafic and ultramafic rocks and subordinate marbles, as shown in Fig. 2 and described by R. Schmid

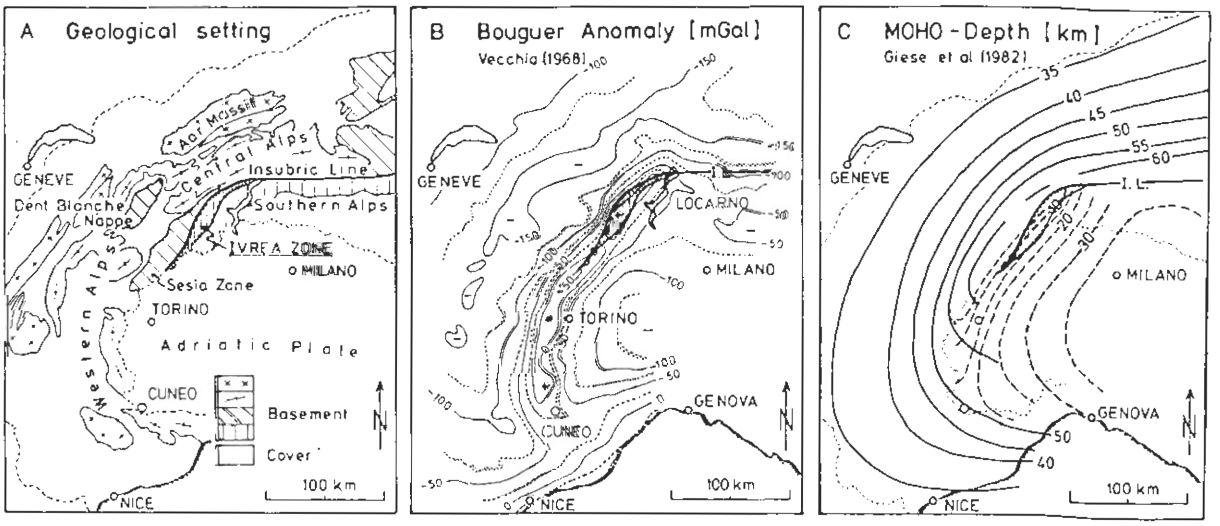


Fig. 1. (A) Location of the Ivrea Zone within the Alps. The Ivrea Zone is a piece of Paleozoic lower crust in an Alpine environment. (B) Bouguer anomaly (Vecchia, 1968). Note the abrupt end of the anomaly near Locarno and Cuneo. If the negative anomaly of the Alpine root is substracted, the effect of the Ivrea body reaches +170 mGal according to the modelling of Kissling (1984). (C) Moho-depth according to Giese et al. (1982), compare also Mueller et al. (1980). The Moho of the Southern Alps (broken contours) comes closest to the earth's surface in the Ivrea Zone region.

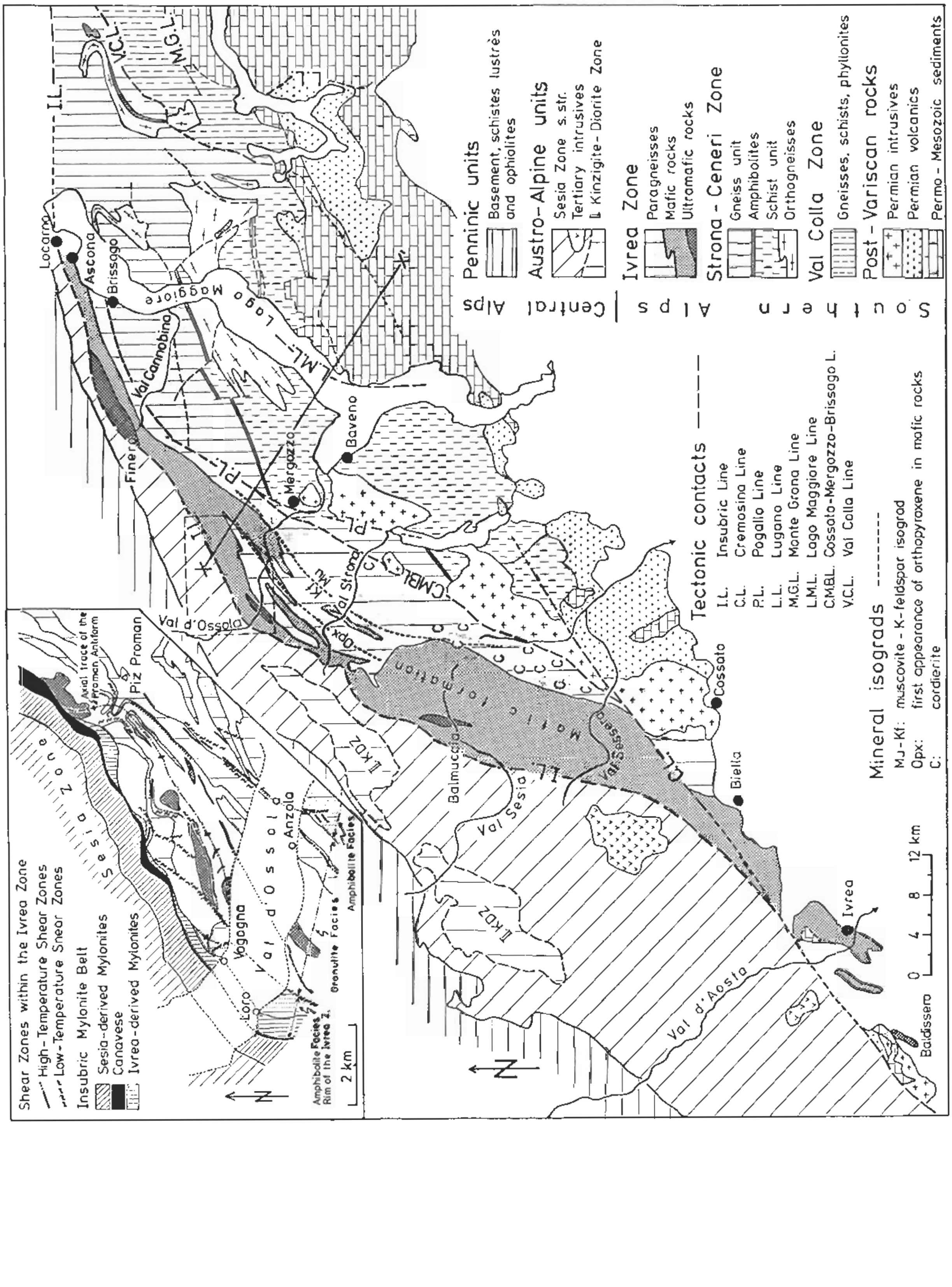


Fig. 2. Tectonic units, lithologies, tectonic lines and mineral isograds of the Ivrea and adjacent zones. The crustal section is truncated by several tectonic lines and comprises the Ivrea Zone (lower and basal intermediate crust), the Strona–Ceneri Zone and the Val Colla Zone (intermediate to shallow crust). $X - X'$ = location of profile in Fig. 3A. Inset: Detailed geological map of the northwestern rim of the Ivrea Zone in the Val d'Ossola (after R. Schmid, 1967; Brodie and Rutter, 1987; S.M. Schmid et al., 1987).

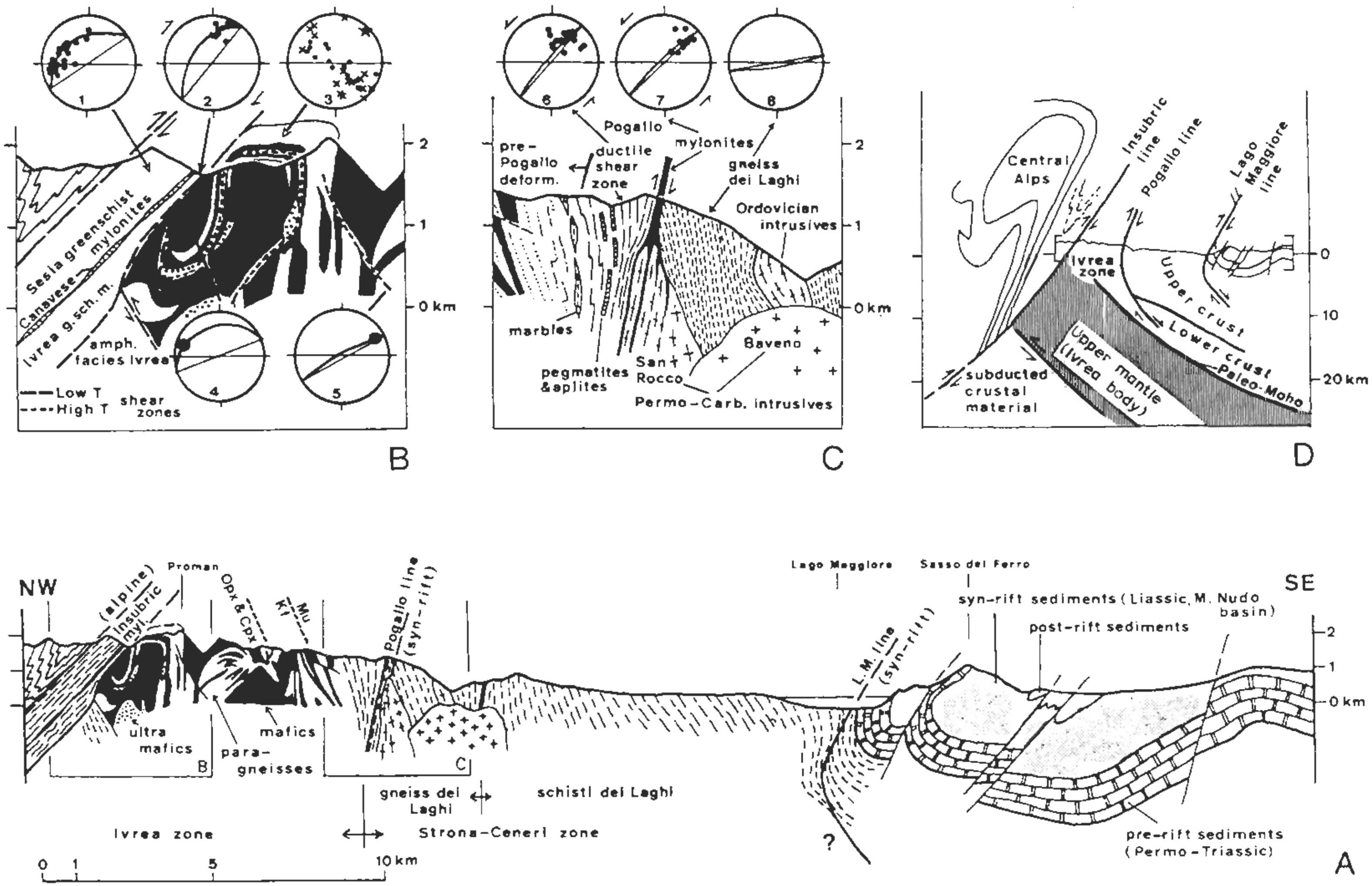


Fig. 3. (A) Geological section from the Insubric Line to the cover of the Southern Alpine basement. Location of profile indicated in Fig. 2. (b) Details of the Proman antiform and adjacent areas. The stereographic projections are lower hemisphere and indicate the average foliation and stretching lineations (dots) of the Insubric mylonites (projections 1 and 2 from S.M. Schmid et al., 1987); measurements from the Proman antiform (projection 3: dots = foliation poles; crosses = slickenside lineations; star = fold axis orientation); average orientation of foliation and stretching lineation within the high-temperature shear zones (projections 4 and 5, from Brodie and Rutter, 1987). (c) Details of the area around the Pogallo ductile fault zone. The stereographic projections are lower hemisphere and indicate the average foliation and lineation orientations (dots) within the Pogallo ductile fault zone (projection 6), the mylonites of the Pogallo Line (projection 7) and the gneiss dei Laghi (projection 8). (d) Sketch illustrating the position of the profile depicted in Fig. 3A within a larger context.

(1967), Bertolani (1968) and others. Four types of mafic rocks are distinguished: (1) mafic volcanic rocks of oceanic origin (geochemical trends: Sills and Tarney, 1984; Mazzuchelli and Siena, 1986), interlayered with paragneisses; (2) mafics forming larger bodies like the gabbroic rocks of Anzola; (3) banded and layered mafic series found in the granulite facies part of the Ivrea Zone. These three types have experienced the same regional metamorphism and deformational history as the paragneisses; (4) mafic rocks showing relict magmatic features in the Val Sesia region. The so-called Mafic Formation of the Ivrea literature is a composite of mafic rock types (2), (3) and (4) above and forms a prominent belt spanning the length of the Ivrea Zone (Fig. 2). Geochemical and petrological investigations (Rivalenti et al., 1975; Capedri et al., 1977; Shervais, 1979; Garuti et al., 1980; Rivalenti et al., 1981a, b, 1984; Sinigoi et al., 1983; Pin and Sills, 1986; Voshage et al., 1987, 1988) indicate that the different rock types forming the Mafic Formation in the Val Sesia region are deep-seated intrusions originating from the subcontinental mantle.

In the Strona–Ceneri Zone, a northwestern part rich in paragneisses of various compositions ("Gneiss dei Laghi") is separated by a series of amphibolite and amphibole-bearing gneisses from a southeastern part which is dominated by micaschists ("Scisti dei Laghi") (Boriani et al., 1977). Both parts contain large bodies of granitic gneiss (Boriani et al., 1982/1983). In addition, several Permian granite bodies occur within the Strona–Ceneri Zone. The Val Colla Zone consists of micaschists and subordinate amphibole-bearing gneisses and granitic gneisses ("Gneiss Chiari"; Reinhard, 1953, 1964; Fumasoli, 1974; Boriani and Colombo, 1979).

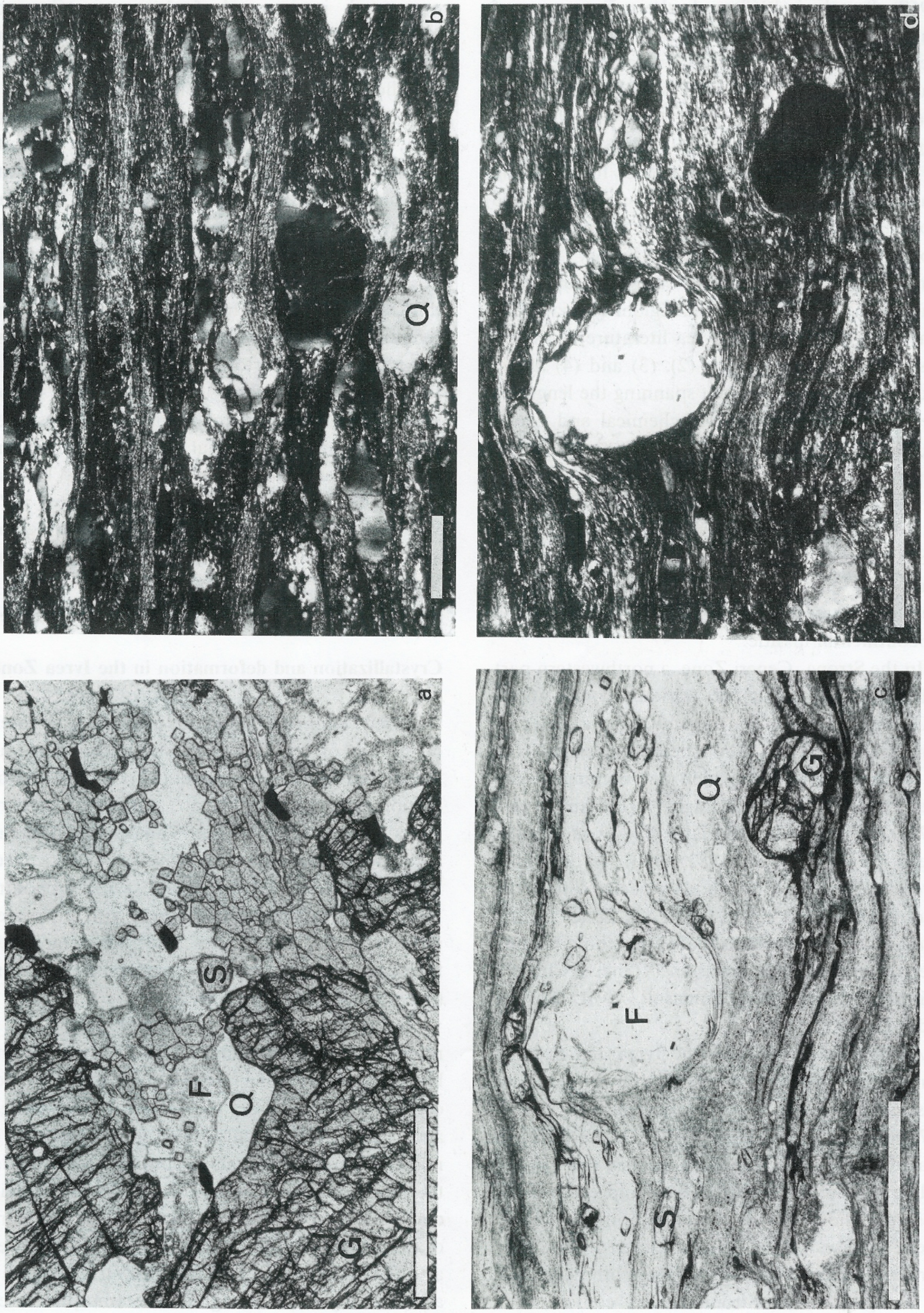
At the contact between the Ivrea and Strona–Ceneri Zones small, Late Carboniferous mafic intrusives ("appinite suite" of Boriani and Sacchi, 1973; Boriani et al., 1975, 1977) cut the foliation related to the regional metamorphism (e.g., in the Val Sesia). According to Boriani and co-workers, these intrusions caused partial melting in the adjacent gneisses of the Strona–Ceneri Zone. They attribute the concentration of these intrusive bodies along the Ivrea-Strona–Ceneri contact to a purported intracrustal fault zone (Cossato–Mergozzo–Brissago Line) that was active during the Late Paleozoic and predates the activity of the Pogallo Line (Boriani and Sacchi, 1973, 1985).

The post-Variscan Southern Alpine sedimentation began with Late Carboniferous continental deposits and Permian volcanics and volcaniclastic sediments, locally deposited in basins (Collio and Verrucano formations). Marine conditions were restored in the Middle Triassic with the deposition of dolomite and limestone series. Variations in thickness and facies indicate that localized differential subsidence started in the Late Triassic (Gaetani, 1975; Bertotti, in press). In the Lias, shallow-water platform sediments and deeper marine basinal sediments (siliceous limestones) document rapid differential subsidence and rifting at the southern continental margin of the Tethyan Ocean (Bernoulli, 1964; Kälin and Trümpy, 1977; Winterer and Bosellini, 1981). The Lago Maggiore and Lugano lines represent originally E-dipping Mesozoic normal faults at the western margin of the Mt. Nudo and Generoso basins, respectively.

Crystallization and deformation in the Ivrea Zone

Three tectonometamorphic facies with different conditions of crystallization and deformation are recognized in the Ivrea Zone: (1) polyphase deformation and amphibolite to granulite facies regional metamorphism and anatexis producing rocks with annealed microfabrics; (2) localized high-temperature shear zones with mylonitic fabrics; (3) localized shear zones with mylonites, ultramylonites, cataclasites and pseudotachylites formed under low-grade metamorphic conditions (greenschist facies and lower).

We use the term tectonometamorphic facies to refer to a set of structures and mineral assemblages formed under similar conditions. When correlating structures and mineral assemblages across the entire crustal section, it is important to realize that deformations occurring under similar thermobarometric conditions are not necessarily contemporaneous. This is because different parts of the crustal section occupied similar crustal levels at different times. In addition, the strain intensity and the degree of retrograde metamorphic over-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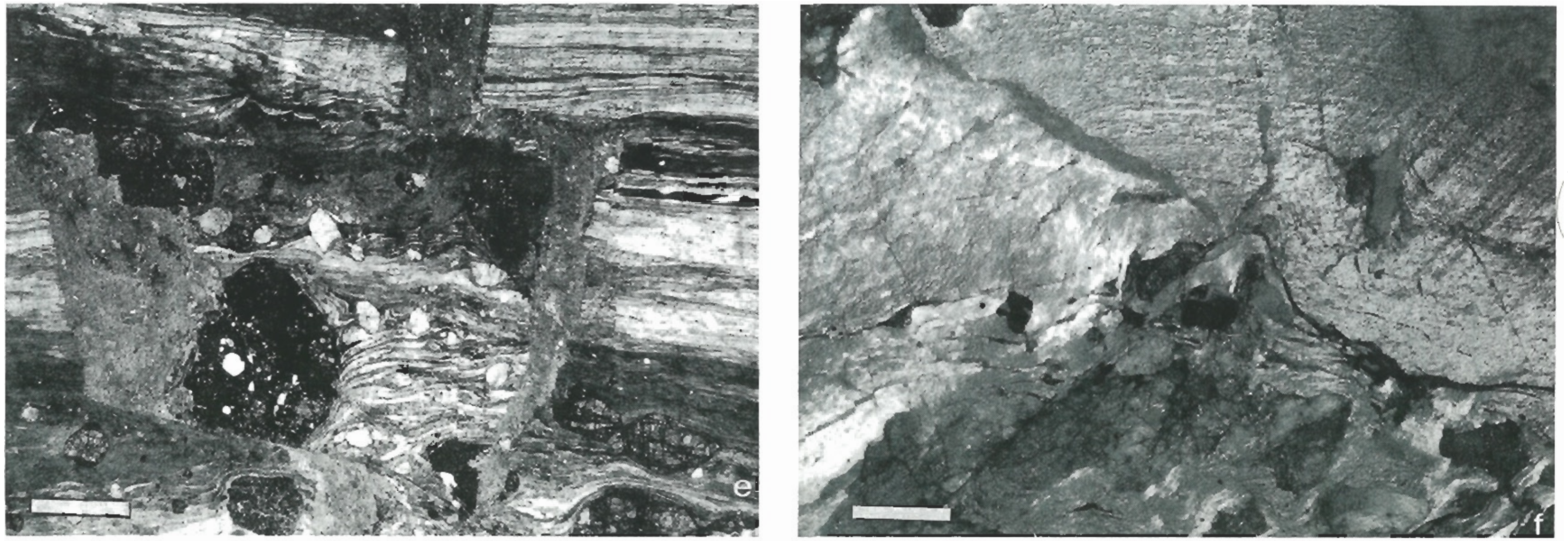

Fig. 4. Microstructures of deformed Ivrea paragneisses. Scale bar 1 mm. (a) Annealed granoblastic texture in granulite facies paragneiss (so-called stronalite) typical for the tectonometamorphic facies 1. Mineral assemblage: garnet (*G*); sillimanite (*S*); quartz (*Q*); K-feldspar and plagioclase (*F*). (b) Greenschist facies mylonite from the Insubric Line (tectonometamorphic facies 3). All minerals except quartz react to form a fine-grained matrix. Note that quartz (*Q*) forms globular grains, indicating that the fine-grained matrix underwent reaction-enhanced work softening. As a consequence of hydration and retrogression of all other minerals, quartz became relatively more flow resistant. Under anhydrous conditions quartz is the weakest mineral phase (compare Fig. 4c, d). (c, d) Details of a high-temperature mylonite (tectonometamorphic facies 2) derived from granulite facies paragneiss (c with crossed polarizers). Feldspar (*F*), garnet (*G*) and sillimanite (*S*) form porphyroclasts indicating a dextral sense of shear. The quartz-rich domains (*Q*) recrystallized to an extremely fine-grained matrix (10 μm and less). (e) Cataclastic overprint of a high-temperature mylonite, like that depicted in Figs. 4c and d, with veins of ultracataclasite (possibly pseudotachylite). Alpine deformation within the Proman antiform overprinting high-temperature mylonites of Late Paleozoic to Early Mesozoic age. (f) Box fold within a quartz-rich layer of a high-temperature mylonite that is folded around the Proman antiform. Only quartz exhibits a limited amount of intracrystalline plasticity during Alpine deformation. Note the similarity in style between this microfold and the large-scale Proman antiform depicted in Fig. 3B.

print vary strongly with rock-type. Rocks which are relatively strong at a given temperature, strain-rate and pressure (e.g., flow-resistant basic and ultrabasic rocks) are often boudinaged and tend to preserve older structures, whereas weaker rocks (e.g., quartz-rich paragneisses) usually show penetrative overprinting fabrics and younger mineral assemblages.

Regional metamorphism and anatexis

The metamorphic grade increases from the mid-amphibolite to the granulite facies towards the northwest (i.e., perpendicular to the general strike of the compositional banding of the Ivrea Zone as shown in Fig. 2). The progressive changes of modes, textures and mineral compositions in the central part of the zone are described by Peyronel Pagliani and Boriani (1967), R. Schmid (1967), Bertolani (1968), R. Schmid and Wood (1976), Hunziker and Zingg (1980), Sills (1984). Well equilibrated, annealed textures with large post-kinematic minerals (e.g., garnet) are common features (Fig. 4a). Prograde reactions can only rarely be derived from textural observations. Zingg (1980) mapped several mineral isograds in the Ivrea Zone, including the replacement of muscovite by K-feldspar in sillimanite-bearing paragneisses, the first appearance of clinopyroxene and orthopyroxene in mafic rocks and the upper stability limit of the assemblage calcite–quartz–actinolite in impure marbles. The pressure and temperature increase towards the northwest, perpendicular to the strike of the zone. The common occurrence of cordierite in the southwestern part of the Ivrea Zone suggests an additional minor pressure gradient parallel to the strike of the zone (i.e., a pressure decrease towards the southwest).

For the amphibolite to granulite facies in the Ivrea Zone a temperature range between 550° and 940°C is obtained from geothermometry and 8–9 kbar represent the best pressure estimate in the granulite facies (R. Schmid and Wood, 1976; Garuti et al., 1978/1979; Hunziker and Zingg, 1980; Zingg, 1983; Sills, 1984; R. Schmid et al., 1988).

In the upper amphibolite facies part of the Ivrea Zone, two phases of leucosome are locally recognized. The older leucosome is isoclinally folded, whereas the younger leucosome occurs in the axial plane of more open folds. In addition, the bulk composition of paragneisses becomes progressively depleted in granitophile elements with increasing metamorphic grade, suggesting a large-scale removal of granitic melts (R. Schmid, 1972, 1978/1979; Sighinolfi and Gorgoni, 1978). Thus, the paragneisses can be considered as restites, a view that is supported by Nd-isotope ratios that are more primitive in the amphibolite facies paragneisses than in the granulite facies paragneisses (Voshage et al., 1987).

Regional metamorphism and associated deformation are the dominant features in the Ivrea Zone (R. Schmid, 1967; Steck and Tièche, 1976; Kruhl and Voll, 1976; Handy, 1986, 1987). However, the mineral assemblages in parts of the Mafic Formation (type 4 mafic rocks, see above) in the Val Sesia region do not match the mineral isograd pattern (Zingg, 1980) and its rocks are only moderately deformed compared with adjacent paragneiss series. Thus, the intrusion of type 4 mafic rocks postdates the peak of the regional metamorphism.

High-temperature shear zones

High-temperature mylonitic shear zones have been observed in many parts of the Ivrea Zone (Walter, 1950; R. Schmid, 1967; Boriani, 1971; Kruhl and Voll, 1976; Steck and Tièche, 1976; Brodie, 1980, 1981; Brodie and Rutter, 1987; Rutter and Brodie, 1988) and new occurrences are found in most areas that are reinvestigated. Brodie and Rutter (1987) have mapped these shear zones in detail in the Val d'Ossola (see inset Fig. 2). The width of most high-temperature shear zones ranges from a few centimeters to several meters; they are commonly very narrow and form anastomozing networks in the mafic and ultramafic rocks, but are generally broader and more diffuse in the paragneisses. After the removal of the effects of later low-grade faulting and folding (e.g., the Proman Antiform; see below), the shear zones accommodate extension parallel to the compositional banding of the series (Brodie and Rutter, 1987). The proximity of many shear zones to the amphibolite-granulite facies transition in the Val d'Ossola suggests that the contact between the amphibolite and granulite facies terrains in this part of the Ivrea Zone is tectonic (R. Schmid and Wood, 1976; Brodie and Rutter, 1987).

In these high-temperature mylonites, the minerals that formed during regional metamorphism are found as porphyroclasts and relics in the fine-grained mylonitic matrix, so mylonitization postdates the regional metamorphism (Fig. 4c, d). Moderately sheared mafic rocks often contain undeformed kelyphitic rims around garnets (Walter, 1950; R. Schmid, 1967; Schenk, 1981; Brodie and Rutter, 1985). These rims consist of orthopyroxene, plagioclase, hercynite and opaques that are undeformed. Thus, the high-temperature conditions indicated by the kelyphitic assemblages outlasted the shearing.

The mylonitic matrix, which is often annealed, consists of fine-grained recrystallized quartz, K-feldspar, plagioclase and biotite in the sheared paragneisses; plagioclase, amphibole and clinopyroxene in the sheared mafic rocks, and olivine, amphibole, clinopyroxene, orthopyroxene and phlogopite in the sheared peridotites. These matrix minerals sometimes have compositions equal to those of the larger porphyroclasts formed during regional metamorphism. In other cases, slight differences in composition between matrix minerals and porphyroclasts are reported that indicate both retrograde (Kruhl and Voll, 1978/1979) and prograde (Brodie, 1981; Brodie and Rutter, 1987) conditions during shearing compared to the regional metamorphism.

A striking microstructural feature of some high-grade mylonitized paragneisses is the small grain-size of dynamically recrystallized quartz (Fig. 4c, d). Such fine grain-sizes indicate high flow stresses and are typically found in greenschist facies mylonites. However, the granulite facies assemblages in these high-temperature mylonites are largely unaltered. The coexistence of small, dynamically recrystallized quartz grains and unaltered high-grade minerals probably reflects anhydrous conditions prior to and during the shearing. This may explain the discrepancy in composition amongst recrystallized grains in the mylonitic matrix. If the rocks were dry during shearing, equilibrium amongst the newly formed mineral phases may not have been attained and their compositions may not be very reliable indicators of the syntectonic metamorphic conditions.

Syntectonic temperatures in the mylonites are further constrained with microstructural observations of the dominant deformation mechanisms. Dynamic recrystallization of minerals such as feldspar, amphibole, clinopyroxene and olivine in sheared mafics and ultramafics indicates that these minerals deformed by power-law creep. The minimum temperatures for power-law creep at geological strain-rates is estimated to be about 500°C for feldspar and at least 600–650°C in the case of olivine (Kirby, 1983).

In summary, the exact metamorphic conditions of shearing are not yet established. The apparently conflicting compositions of the matrix minerals may be explained by one or a combination of the following factors: (1) the compositions of the newly formed matrix minerals do not reflect equilibrium because of the dry conditions; (2) repeated activity of this high-temperature shearing over a long time interval and under changing metamorphic conditions; (3) high thermal gradients due to the juxaposition of hot, deep crustal rocks against cooler, shallower rocks during extension, the increased heat-flow in extensional environments and shear heating (Brodie and Rutter, 1987).

The transition from high- to low-temperature deformation: the Pogallo Ductile Fault Zone

The retrograde transition from high-temperature (> 500°C) to low temperature (< 300 °C) deformation in the Ivrea Zone is observed within the Pogallo Ductile Fault Zone (PDFZ). A complete description and discussion of this crustal-scale shear zone is given in Handy (1987). Pogallo shearing primarily affected paragneisses within the southern margin of the Ivrea Zone and the northernmost rim of the Strona–Ceneri Zone. Deformation occurred under retrograde amphibolite facies conditions in the bulk of the PDFZ and greenschist facies conditions at the Pogallo Line, forming part of the southeastern border of the PDFZ (see figs. 1 and 2 in Handy, 1987). The systematic kinematic and microstructural development across and along strike of the PDFZ indicates that amphibolite and greenschist facies deformation was contemporaneous and involved oblique sinistral uplift of the Ivrea Zone with respect to the Strona–Ceneri Zone. These findings are interpreted as resulting from a continuous extension under retrograde metamorphic conditions during the Early Mesozoic (see p. 186). Northeast of the Val Cannobina (Fig. 2) metamorphic conditions in the PDFZ reach a higher grade, and the fault zone becomes broader and very diffuse, while southwest of the Val d'Ossola the available evidence suggests that the PDFZ becomes a discordant, brittle dislocation within the Strona–Ceneri Zone. According to Boriani and Sacchi (1973), the sinistral movement along this fault is about 10 km based on the displacement of an amphibolite gneiss band within the Strona–Ceneri Zone. The Cossato–Mergozzo–Brissago Line (Boriani and Sacchi, 1973, 1985) along the contact of the Ivrea and Strona–Ceneri Zones (to the extent that it is observed) appears to pre-date activity of the PDFZ.

Low-temperature deformation

Deformation under greenschist facies and lower conditions is found along major lineaments at the northwestern (Insubric Line) and southeastern (aformentioned Pogallo Line) borders of the Ivrea Zone. However, the age and tectonic setting of these two occurrences are completely different.

In the greenschist facies shear zones the paragneisses are hydrated and retrogressed to quartz–phengitic muscovite–chlorite–clinozoisite–albite mylonites (Fig. 4b). The amphibolites are transformed to albite–actinolite–chlorite–clinozoisite–sphene mylonites and the peridotites are serpentinized. Cataclasites and pseudotachylites are more frequently observed in the mafic rocks, which are more resistant to mylonitization and associated hydration reactions than the paragneisses.

Insubric Line

The northwestern limit of the Ivrea Zone comprises the greenschist facies mylonite belt of the Insubric Line (see inset of Fig. 2), the so-called "Scisti di Fobello et Rimella" (e.g., Sacchi, 1977). This belt consists of mylonitized protoliths from three units (i.e., gneisses of the Austroalpine Sesia Zone, Permo-Mesozoic Canavese sediments and Ivrea rocks; Ahrendt, 1980; S.M. Schmid et al., 1987). The Ivrea-derived mylonites show horizontal to moderately E- and NE-plunging stretching lineations with a dextral sense of shear (Fig. 3B). They can be related to the Neogene transpressive tectonics during the generally westward motion of the Adriatic plate with respect to the European plate. In the Sesia-derived mylonites, down-dip lineations (Fig. 3B) are associated with a sense of shear that indicates uplift and backthrusting of the Central Alps over the Southern Alps (Fig. 3D). Local overprinting relations within the Ivrea-derived mylonites indicate that the down-dip lineations are older than the horizontal lineations (S.M. Schmid et al., 1987).

Alpine folding within the northern Ivrea Zone

A huge box-type antiform is observed at the northwestern margin of the Ivrea Zone in the Val d'Ossola section (inset Fig. 2 and Fig. 3A). R. Schmid (1967) observed that the high-temperature mylonites predate the formation of the antiform since they are folded around its hinge. Figure 3B summarizes structural data which support an Alpine age of formation as postulated by S.M. Schmid et al. (1987) and Brodie and Rutter (1987).

The folding involved foliation-parallel flexural slip. Slickensides which formed by fibrous growth of greenschist facies or lower grade minerals are found to define a great circle perpendicular to the Proman fold axis (Fig. 3B). R. Schmid (1967) and Brodie and Rutter (1987) noted that the lineations associated with high-temperature mylonites lie approximately on a small circle around the Proman fold axis. Their average orientation and sense of shear are different on both limbs of the fold (Fig. 3B). Internal strain is associated with cataclasites and pseudotachylites (Fig. 4e). Only quartz in the paragneisses exhibits a limited amount of intracrystalline plasticity (Fig. 4f). The deformation under very low metamorphic grade, the kinematics of conjugate brittle shear zones, the overall geometry of the Proman Fold and its proximity to the Insubric Line suggest that it formed during NW–

TABLE 1

Rb–Sr whole-rock isochrons from the Ivrea and Strona–Ceneri zones

	AGE (Ma)	NUMBER OF SAMPLES, SAMPLE SIZE, ROCK TYPE	REFERENCES
IVREA ZONE	478 ± 20	9, 30-50 kg, amphibolite and granulite facies paragneisses	1
	362 ± 21	6, 20-30 kg, charnockitic granulites (b) within the Mafic formation	2
	338 ± 41 (a)	7, 1-4 cm thick slabs of 3-5 kg, sheared granulite facies paragneisses	3
	329 ± 1	3, 20-30 kg, diorites from the rim of the Mafic formation	2
	293 ± 13	7, 20-40 kg, phlogopite-bearing ultramafics	4
	274 ± 17	4, 20-30 kg, charnockites from the migmatitic contact zone between the Mafic formation and the paragneisses	2
STRONA-CENERI ZONE	539 ± 55 (a)	5, amphibolite facies basement rocks	5
	473 ± 29	4, 30-50 kg, amphibolite facies basement rocks	1
	466 ± 5	13, orthogneisses	6
	276 ± 5	6, 30-50kg, granites of the postmetamorphic Baveno suite and 1 quartz-diorite	1
	283 ± 14	4, postmetamorphic granites, Montorfano	7

(a) age recalculated with the new constant, see Hunziker and Zingg (1980)
(b) in the sense of Bertolani & Garuti (1970)

References: 1) Hunziker and Zingg (1980); 2) Bürgi (1987); 3) Graeser and Hunziker (1968); 4) Voshage et al. (1987); 5) Hamet and Albarède (1973); 6) Boriani et al. (1982/83); 7) Boriani et al. (1987).

SE shortening related to the Neogene Insubric backthrusting event. However, the subsequent Insubric strike-slip movements truncated the curved axial trace of this antiform both to the northeast and southeast (see inset of Fig. 2).

The formation of the Proman Antiform as a compressional flexural slip structure would not have been possible if the main lithological boundaries and earlier foliations had been steep prior to Alpine compression. Hence, we infer an Alpine age of steepening of the Ivrea crustal cross-section. Due to the very minor internal strain associated with flexural slip folding, the pre-Alpine structures remained undisturbed in the Ivrea Zone.

The present structure of the Ivrea section can be viewed as part of a huge Alpine antiform. Its southeastern limb corresponds to the steepened section between the Proman Antiform and the Lago Maggiore (Fig. 3A). The synformal flexure connecting the steepened basement section with the sediments (which are approximately in their original horizontal position in the southeast) is not observable (discussion in Handy, 1987). Thus, this synform and the interpretation of the Lago Maggiore Line as a rotated former low-angle extensional fault remain conjectural. Recent work proposes a similar scenario of steepening of the Lugano/Monte Grona Line (Fig. 2), another originally E-dipping extensional fault further to the east (Bertotti, in press). Most of the northwestern limb of this huge antiformal structure has been sheared off during the movements along the Insubric Line. Several occurrences of pre-Alpine amphibolite facies Ivrea rocks were found in the vicinity of the Insubric Line (i.e., to the northwest of the granulite facies region; S.M. Schmid et al., 1987), and one of these amphibolite facies rims is shown in the inset of Fig. 2. The location of these amphibolite facies rocks and Canavese sediments (i.e., the Permo-Mesozoic sedimentary cover of the Southern Alpine basement) support the idea of the Ivrea section as part of a huge antiformal structure.

Radiometric age determinations

Figure 5 gives a synoptic view of the radiometric ages from the Ivrea and Strona–Ceneri zones. The whole-rock isochrons and the U–Pb ages are summarized in Tables 1 and 2. The ages vaguely reflect "Caledonian", "Variscan" and Permian events. There are also numerous Mesozoic mica

TABLE 2

U–Pb age determinations from the Ivrea and Strona–Ceneri zones. The old data were kindly recalculated with the new constants by V. Köppel.

	MINERAL, ROCK TYPE,	SAMPLE No.	AGES: $^{206}Pb/^{238}U$; $^{207}Pb/^{235}U$	REFERENCES
IVREA ZONE	zircon, paragneisses	STRO2	discordant	1
		STRO3	discordant	1
		ANZ1	discordant, upper Intersection: 1900 Ma	1
			lower Intersection: 285 Ma	
	monazite, paragneisses	STRO2	273; 274 Ma	1
		ANZ1	272; 272 Ma	1
		near Candoglia	295 - 310 Ma	2
		Valle Strona	280; 280 Ma	3
	zircon, dioritic rim of the Mafic Fm.		285+7/-5Ma	4
STRONA - CENERI ZONE	zircon, paragneisses	MAL2	discordant }	5
		CEN1	discordant } upper Intersection:	5
		CAS1	discordant } 1900- 2500 Ma	5
		2FM5	discordant } lower Intersection:	5
		FM5	discordant } 450 - 500 Ma	5
		2FM7	discordant }	5
	zircon, orthogneisses	MAL1	441; 459 Ma (150-75 μm)	5
			429; 436 Ma (< 42 μm)	
		FM17	440; 450 Ma	5
		FM7	481; 487 Ma	5
		FM12	450; 452 Ma	5
		STRO1	discordant	1
	monazite, orthogneiss	STRO1	287; 287 Ma (> 125 μm) NW-Border of the	1
			294; 297 Ma (< 125 μm) Strona-Ceneri Zone	
	monazite, paragneisses	CAS1	438; 438 Ma (total fraction)	5
			448; 452 Ma (> 75 μm)	
			453; 453 Ma (< 42 μm)	
		A.Ruspessa	440 - 450 Ma	2
		Mergozzo	320 - 330 Ma	2

References: 1) Köppel (1974); 2) Köppel and Grünenfelder (1978/79); 3) Teufel and Schärer (1989); 4) Pin (1986); 5) Köppel and Grünenfelder (1971).

TABLE 3

Summary of the evolution of the Ivrea Zone

	AGE	EVENT	FABRIC	
CRUSTAL GENESIS AND OROGENY		high-pressure/low-temperature event: age and extent unknown		1
	Paleozoic	polyphase deformation amphibolite to granulite facies metamorphism, anatexis	annealed, medium to coarse grained	
		Late Paleozoic magmatic activity in all crustal levels		
OVERALL COOLING AND CRUSTAL ATTENUATION	Permian - Triassic	high-temperature shearing	mylonitic locally annealed	2
	Triassic - Lias	retrograde amphibolite and greenschist facies shearing (Pogallo Shear Zone)	mylonitic locally cataclastic	
ALPINE TRANSPRESSIVE TECTONICS	Cretaceous - Tertiary	density inversion at the base of the Ivrea Geophysical Body		3
		steepening of the Ivrea crustal section localized folding and thrusting	cataclastic	
		NW-rim of the Ivrea Zone: greenschist facies shearing along the Insubric Line	mylonitic	
			TECTONOMETAMORPHIC FACIES	

Rb–Sr and K–Ar ages. Pin and Vielzeuf (1983), Boriani (1987) and Boriani et al. (1988) discuss the possibility that these may represent mixed ages between Variscan and Alpine thermal events. However, most of the Ivrea biotites are not altered and generally yield concordant Rb–Sr and K–Ar ages, a feature not observed for mixed ages (Fig. 6A; see also Dempster, 1986). The minor microscopic and submicroscopic alteration reported by Bürgi (1987) rejuvenated the K–Ar ages by less than 8% compared with the Rb–Sr ages. Several biotite determinations from the Finero area yield substantially lower K–Ar ages (120–150 Ma; Klötzli, 1988). At least some of these lower ages are related to alteration of the biotites, evidenced by chlorite layers and low potassium content. In conclusion, most of the biotite ages from the Ivrea Zone that cluster around 190 Ma (Fig. 6B) must

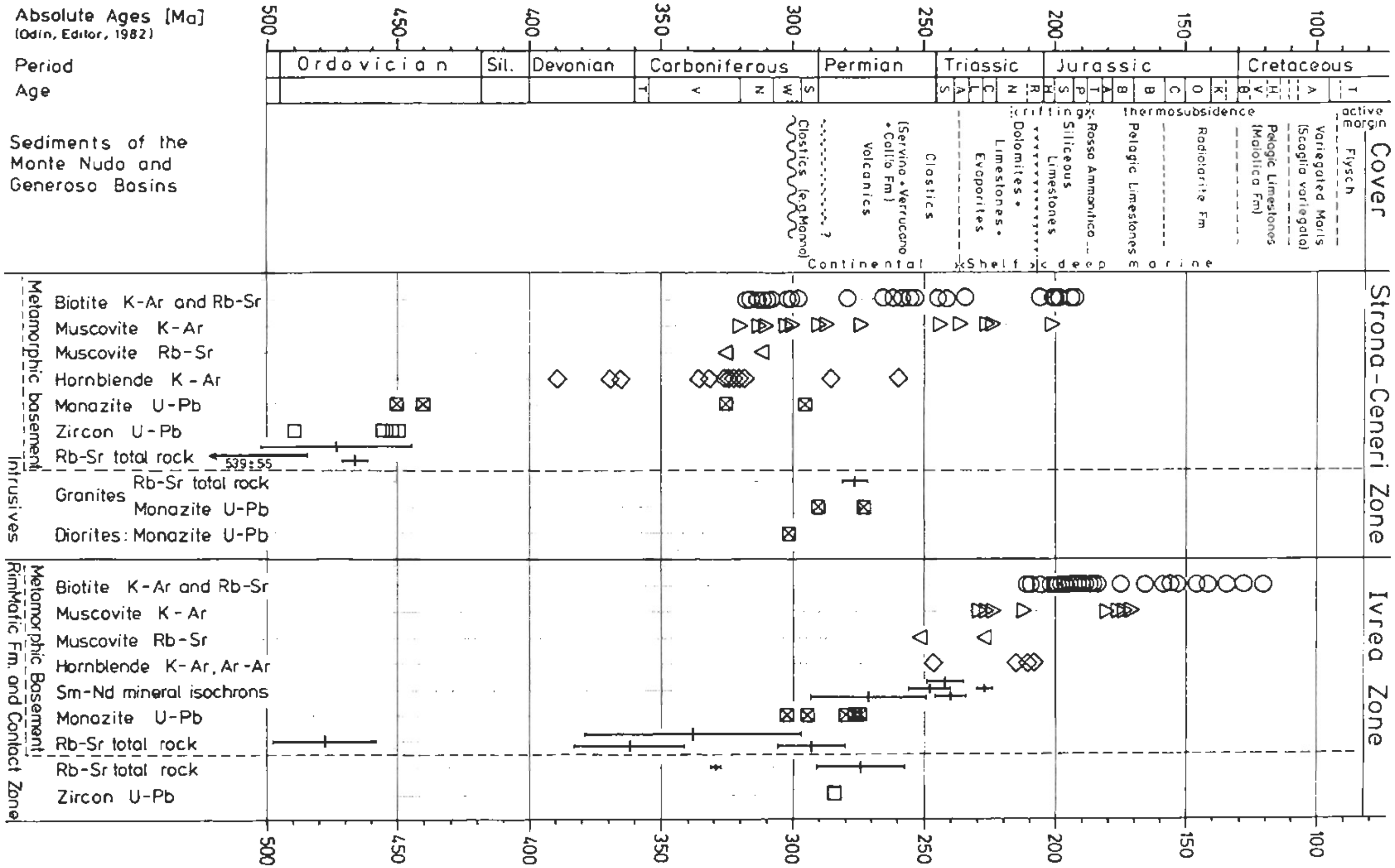

Fig. 5. Synoptic view of the radiometric age determinations from the Ivrea and Strona–Ceneri zones. Time scale after Odin (1982). Data from Jäger et al. (1967), Graeser and Hunziker (1968), McDowell and Schmid (1968), McDowell (1970), Köppel and Grünenfelder (1971, 1978/1979), Hamet and Albarède (1973), Hunziker (1974), Köppel (1974), Hunziker and Zingg (1980), Boriani et al. (1982/1983), Polvé (1983), Pin (1986), Bürgi (1987), Voshage et al. (1987), Klötzli (1988), and Brodie et al. (1989).

be considered as real cooling ages and give the time of cooling to below about 300°C. The biotite ages tend to become younger towards the Insubric Line (i.e., towards the deeper part of the crustal section).

Tertiary ages were obtained only by fission-track dating (Fig. 7). The apatite ages cluster around 12 Ma (except Anzola with 20 Ma) both in the Ivrea and Strona–Ceneri Zones and the three zircon ages scatter between 31 and 63 Ma (Wagner and Reimer, 1972; Hurford, 1986; Klötzli, 1988; Hurford et al., 1989).

Figure 5 also illustrates the post-Variscan thermal history of the different Southern Alpine crustal segments. The classical Variscan angular unconformity is Carboniferous in age (about 305 Ma) and is marked by trangressive Westphal B/C clastic sediments (Venzo and Maglia, 1947; Jongmans, 1960). At that time, high temperatures prevailed in the Ivrea Zone (> 600°C in the granulite facies part).

Radiometrically determined time-temperature paths are shown in Fig. 7 for the amphibolite facies rocks of the Ivrea and Strona–Ceneri zones in the Ossola. This diagram indicates overall post-Variscan cooling starting in the Permian. However, Permo-Carboniferous magmatic activity and Early Mesozoic rifting suggest a more complicated thermal evolution that involved temporary temperature increases (Laubscher and Bernoulli, 1982).

Dating and discussion of the tectonometamorphic facies

Despite the dominance of the regional metamorphism (tectonometamorphic facies 1), traces of the earlier evolution are recognized. The upper intercepts of the zircon discordia lie between 1900 and 2500 Ma (Köppel and Grünenfelder, 1971; Köppel, 1974) and point to Early Proterozoic source rocks for the paragneisses. An Early Paleozoic accretionary wedge setting was proposed by Sills and Tarney (1984) to account for the intimate association of these continentally derived sediments with oceanic crust (type 1 mafics).

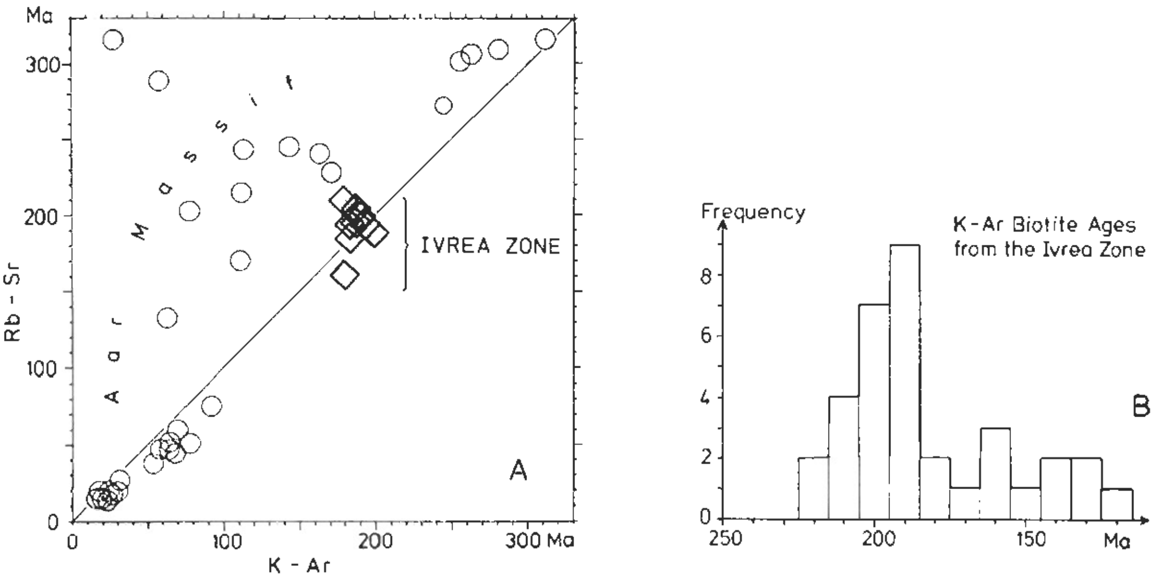

Fig. 6. (A) K–Ar versus Rb–Sr biotite ages from the Ivrea Zone and, for comparison, from the Aar massif (Variscan external massif with variable Alpine overprint). Data from Hunziker (1974), Bürgi (1987), Dempster (1986) and Schaltegger (1986). Nearly concordant K–Ar and Rb–Sr biotite ages are found in the Aar massif only where Variscan structures and mineralogy are largely preserved or where the Alpine overprint is very strong or even complete. The samples with discordant biotite ages show partial recrystallization and alteration of the Variscan mineral assemblages (Voll, 1976), the growth of new minerals, and the incipient development of a new foliation. Most of the Ivrea biotites are nearly concordant and do not show such a partial overprint (see text for exceptions). (B) Frequency distribution of the K–Ar biotite ages of the Ivrea Zone, age data from Hunziker (1974), Bürgi (1987), Klötzli (1988).

Subduction tectonics during the early evolution of the Ivrea and Strona–Ceneri rocks is also suggested by rare occurrences of high-P/low-T relics like eclogitic amphibolites (Boriani and Peyronel Pagliani, 1968; Buletti, 1983) and kelyphitic peridotites (Lensch and Rost, 1972).

Regional metamorphism

When discussing the age of the regional metamorphism, we must be aware that regional metamorphism is associated with several processes such as polyphase deformation, dehydration, anatexis and melt mobilization, annealing and mineral growth. The isotopic systems used for dating will react in different ways to these processes and record different parts of the thermal history. In addition, these processes may not be synchronous in the different crustal levels considered.

Based on different interpretations of the Rb–Sr whole-rock and U–Pb data in Tables 1 and 2 (see also Fig. 5), Ordovician (Caledonian; Hunziker and Zingg, 1980), Carboniferous (Variscan; e.g., Boriani et al., 1982/1983) or Early Permian (Late Variscan; Pin, 1986; Bürgi, 1987; Teufel and Schärer, 1989) ages have been proposed for the peak of the regional metamorphism.

The Ordovician age proposed by Hunziker and Zingg (1980) for the peak of the regional metamorphism is based on a 478 Ma Rb–Sr whole-rock isochron (20–40 kg samples) from Ivrea paragneisses and on zircon and monazite ages from the Strona–Ceneri Zone. Discordant paragneiss zircon suites have lower concordia intercepts between 450 and 500 Ma (i.e., in the same range as the concordant monazites; Köppel and Grünenfelder, 1971, 1978/1979; Köppel, 1974). The paragneiss zircons from the Ivrea Zone with their lower intersection of the concordia at about 280 Ma (Köppel, 1974), however, do not reflect an Ordovician event. The interpretation of the Ordovician Rb–Sr whole-rock isochron as the age of the peak of the

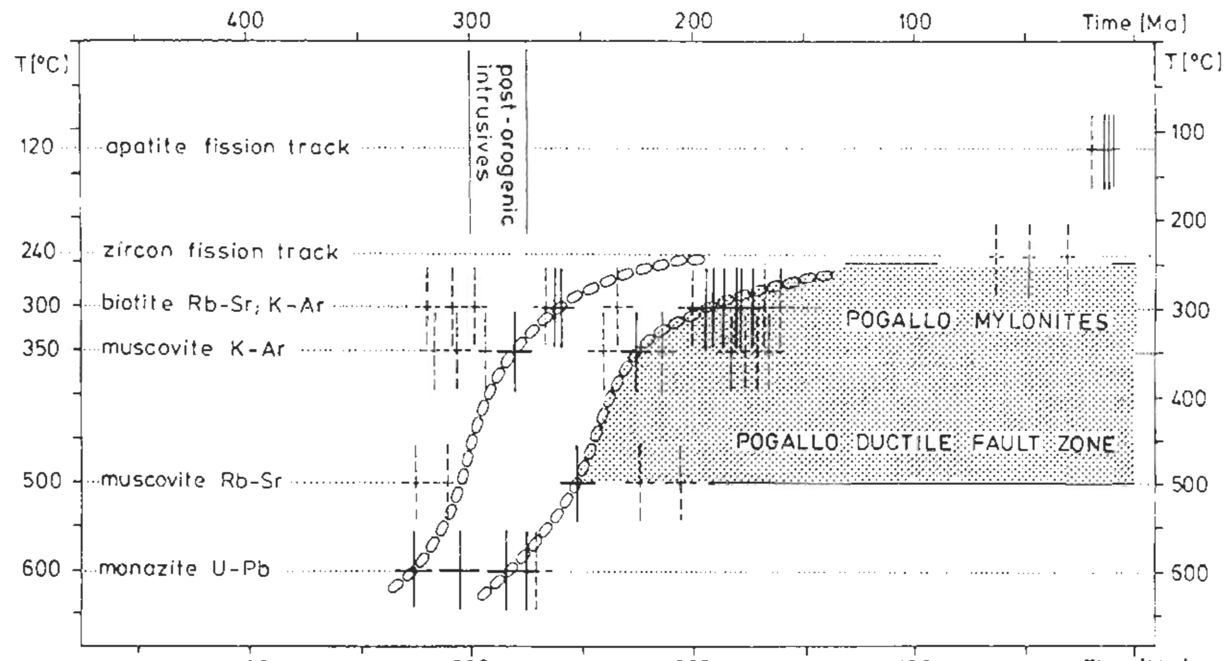

Fig. 7. Time–temperature curves for the amphibolite facies part of the Ivrea Zone and the adjacent part of the Strona–Ceneri Zone. Heavy symbols: age data from the Val d'Ossola, broken symbols: age data from the region east of the Val d'Ossola. The temperature conditions for the Pogallo deformation were estimated from the synkinematic mineral assemblages of the shear zones (see Handy, 1987). Data from Jäger et al. (1967), McDowell and Schmid (1968), McDowell (1970), Wagner and Reimer (1972), Hunziker (1974), Köppel (1974), Köppel and Grünenfelder (1978/1979), Boriani et al. (1982/1983), Hurford (1986), Klötzli (1988), Hurford et al. (1989).

regional metamorphism in the Ivrea Zone is based on the assumption that anatexis and "degranitization" of the paragneisses (R. Schmid, 1978/1979) caused a large-scale homogenization of Sr-isotopes.

Boriani and coworkers (e.g., Boriani et al., 1982/1983, 1985) interpret the Ordovician and Variscan ages as follows: Ordovician granites (orthogneisses of the Strona–Ceneri Zone dated at 466 Ma) intruded low-grade sedimentary series causing contact metamorphism and granitization of these rocks and a partial reset of the paragneiss zircon ages. The whole complex was then metamorphosed during the Variscan orogeny (around 320 Ma).

It is tempting to assume an Ordovician intermediate-grade metamorphism overprinted by a Variscan high-grade metamorphism. However, the concordant Ordovician orthogneiss zircon and the monazite ages from the Strona–Ceneri Zone do not support such an interpretation.

The quartz–dioritic rim of the Mafic Formation yields a U–Pb zircon age of 286 Ma (Pin, 1986). Bürgi (1987) obtain Rb–Sr whole-rock isochron of 274 ± 17 Ma on anatectic paragneisses from the contact zone of the Mafic Formation. The authors conclude that the regional metamorphism is Early Permian. This disregards the fact that the mineral isograds cannot be traced across the rim of the Mafic Formation (type 4 mafics), indicating a late intrusion with respect to the regional metamorphism. These Late Carboniferous ages seem more relevant to the intrusion age of that particular part of the Mafic Formation than to the regional metamorphism. This intrusion occurred in a deep crustal level where high temperatures prevailed. In shallower crustal levels (Strona–Ceneri Zone and contact with the Ivrea Zone) dioritic and granitic intrusions of 290–270 Ma clearly postdate regional metamorphism and cut the regional foliation (lower Val Sesia, Montorfano, Baveno).

All authors agree on high-temperature metamorphic conditions during the Variscan but none of the proposed interpretations give a satisfactory explanation of all available data, as discussed by Zingg (1983) and Boriani et al. (1985).

Dating the shear zones

There are basically three ways of dating shear zones within basement rocks: (1) the dating of the synkinematically formed minerals; (2) the comparison of the thermal history of rocks on either side of the movement zone; and (3) the dating of intrusive rocks that cross-cut tectonites or are themselves deformed by shear zones. A further constraint is obtained by correlating the synkinematic metamorphic grade in the shear zone with radiometrically derived time-temperature curves (Fig. 7). The first method is limited by the characteristics of the mineral systems, since the dating is successful, only if a formational age is obtained rather than a mixing or a cooling age. The second method is restricted to movement zones with a major vertical component, since the differential cooling of the two crustal blocks is compared.

High-temperature shear zones

Brodie et al. (1989) obtained a minimum age of 247 Ma with the Ar–Ar method on hornblende of a relatively undeformed gabbroic rock adjacent to a high-temperature shear zone near Anzola in Val d'Ossola. The fine-grained, newly formed hornblende in the mylonitic matrix of the high-temperature shear zone yields ages of 215 and 210 Ma. According to these authors, all these ages record cooling. The blocking temperature is interpreted to be grain-size dependent and lower for the fine-grained mylonitic matrix. Thus, the shearing is probably older than Late Triassic. From the correlation of the synkinematic metamorphic conditions with the time vs. temperature curve, they infer mylonitization to have initiated prior to 280 Ma (i.e., prior to Early Permian). This age corresponds roughly to the Late Variscan magmatic activity.

Since the metamorphic grade during shearing is debatable (see above) and the geothermal gradient at the time of shearing is unknown, the correlation with time–temperature curves of the type shown in Fig. 7 only broadly constrains the shearing to have occurred between the Late Carboniferous and Triassic. Thus three geological interpretations of the significance of the high-temperature shear zones are possible:

(1) High-temperature shearing is associated with incipient Tethyan crustal attenuation during the Late Paleozoic (i.e., prior to 280 Ma according to Brodie et al., 1989) that affected the lower crust and lithospheric mantle without substantial subsidence and formation of marine basins in the upper crust. The high-temperature shear zones thus predate the Pogallo Ductile Fault Zone. Models of non-uniform stretching (Royden, 1986) could account for such a situation.

(2) High-temperature shearing was contemporaneous with Pogallo ductile faulting. In this case, a high geothermal gradient must be assumed during shearing. This interpretation implies a much more rapid cooling history than that obtained simply by correlating syntectonic mineral assemblages in the shear zones with the radiometrically determined cooling path. Accordingly, such rapid extensional uplift and cooling may have occurred either during the Late Paleozoic or during Early Mesozoic rifting.

(3) Although the high-temperature shear zones and the Pogallo Ductile Fault Zone presently have the same structural orientation and kinematics, they belong to two independent tectonic settings. Initial high-temperature shearing and uplift of the Ivrea Zone are related to transtension during Permian strike-slip tectonics and magmatism (Handy and Zingg, in press), whereas subsequent high- to low-temperature shearing (Pogallo ductile faulting) at high geothermal gradients occurs during Late Triassic to Early Jurassic Tethyan rifting. The latter interpretation best explains the observed syntectonic mineral assemblages in the shear zones, the radiometric ages, and the oblique orientation of the Permian basins with respect to the Triassic–Early Jurassic rift basins.

Pogallo Ductile Fault Zone

The Pogallo Ductile Fault Zone formed under a large range of metamorphic conditions. This indicates either a considerable time span for the movements or high-temperature gradients both across and along strike of the fault zone. Thus, the Ivrea crustal block probably does not have a uniform cooling history. Greenschist facies mylonitization of Permian and Permo-Carboniferous intrusions places an upper age limit on the activity of the Pogallo Line. Comparison of the Strona–Ceneri and Ivrea cooling curves (Fig. 7) gives only a crude estimate of the timing of the movement. Therefore, the age was further constrained by correlating the synkinematic metamorphic grade derived from the paragenesis of the mylonitic matrix with the radiometrically derived cooling curves. According to these curves, Pogallo ductile faulting may have initiated as early as ca. 240 Ma (lower amphibolite facies conditions), but certainly ended no later than ca. 160 Ma (greenschist facies conditions). Evidence from extensional basins in the Southern Alps indicates that rapid differential subsidence stopped in the Early Jurassic. The ductile faulting is inferred to be rapid with respect to the rate of thermal equilibration across the shear zone since both amphibolite and greenschist facies deformational fabrics in the Pogallo Ductile Fault Zone show little or no evidence of post-tectonic annealing or grain-growth (Handy, 1987). Hence, a Late Triassic to Early–Mid Jurassic age is inferred for Pogallo extensional movement.

If the Pogallo structures are rotated around a horizontal axis parallel to the general strike of the compositional banding and mineral isograds of the Ivrea Zone (i.e., approximately around the axis of the Proman Antiform), a SW–NE striking low-angle normal fault with gently ENE-plunging stretching lineations is obtained (Handy,1986, 1987; S.M. Schmid et al., 1987). The coincidence in the age and orientation of the Pogallo Ductile Fault Zone with the Late Triassic to Early Jurassic Southern Alpine sedimentary basins that document the opening of the Tethys support the interpretation of the Pogallo Ductile Fault Zone as a deep crustal extensional fault (Hodges and Fountain, 1984 and the authors above).

Insubric Line and steepening of the Ivrea crustal section

A Late Oligocene to Early Miocene uplift and backthrust of the Central Alps along the Insubric Line was established from a radiometric age profile across the Insubric Line near Locarno (Hurford, 1986). The southern limit of the uplifted Central Alpine block comprises the Sesia-derived mylonites with the down-dip lineations (see above). Late Oligocene and Early Miocene K–Ar ages (range: 19–26 Ma) were also obtained on white mica from the mylonitized Canavese sediments between the Val d'Ossola and Locarno (Zingg et al., 1976; Zingg and Hunziker, in press). The locally observed overprinting relationships among the mylonitic structures indicate that the Ivrea-derived mylonites with subhorizontal stretching lineations are somewhat younger than the Sesia-derived mylonites (S.M. Schmid et al., 1987). However, all mylonites show approximately the same greenschist facies synkinematic metamorphic grade, suggesting that the Ivrea-derived mylonites are not substantially younger than Late Oligocene to Early Miocene.

In addition to the arguments discussed above, the rotation of Oligocene magnetic paleopoles (S.M. Schmid et al., 1989) independently suggests that the formation of the large antiforms (Proman, Ascona) and the steepening of the planar structures of the Ivrea Zone are coeval with Neogene backthrusting.

The Ivrea Geophysical Body

The Ivrea Geophysical Body (i.e., a sliver of mantle and lower crust in upper crustal levels overriding material of lower density) is cut by the Neogene Insubric Line (Fig. 3D). Thus the emplacement age of this body cannot be determined from direct structural observations but must be inferred from regional considerations. According to Laubscher (e.g., 1971, 1984), the Ivrea Body formed during the Neogene westward motion of the Adriatic plate. S.M. Schmid et al. (1987) relate the density inversion to the Cretaceous subduction of continental crust, which is documented by the high-pressure, low-temperature assemblages of the adjacent Sesia Zone (e.g., Dal Piaz et al., 1972; Compagnoni et al., 1977).

Geophysical implications of the evolution of the Ivrea Zone

Does the tectonometamorphic evolution of the Ivrea Zone render it a good example or a general model of the in-situ lower continental crust? Vai and Cocozza (1986) express the view that the Ivrea Zone is the highly metamorphosed axial zone of the Variscan Orogen. Accordingly, the process of exhumation and emplacement was intimately related to the Variscan Orogeny. It is evident that exhumation during the high-grade deformation associated with this orogenic cycle in the Southern Alpine basement would have substantially altered the original lower crustal configuration. A consequence of this interpretation is that the presently exposed Ivrea Zone would have very little in common with in-situ lower crust.

The evidence indicates to us, however, that the lower crustal Ivrea rocks were consolidated during the Paleozoic, subsequently attenuated, cooled, and uplifted during the Late Paleozoic and Early Mesozoic, and finally emplaced as a relatively rigid body during the Alpine Orogeny. This long and rather complicated evolution is summarized in Table 3. Significantly, the final emplacement of the Ivrea crustal section into shallow crustal levels did not substantially modify the original lower crustal features.

The presently observed compositional and regional metamorphic zonation of the Ivrea crustal section was established by the end of the Paleozoic. Therefore, the close relationship between seismological properties, bulk composition and regional metamorphic grade in the Ivrea Zone (Fountain, 1976) suggests that the zone had acquired its basic seismological signature by this time. The major change in seismic velocities and densities across the amphibolite to granulite facies transition may correspond to the Paleozoic Conrad discontinuity (Fountain, 1976) while peridotite bodies at the base of the section (Fig. 2) might represent bright spots near the top of the Paleozoic Moho discontinuity.

Extensional detachment faulting during Late Paleozoic and Early Mesozoic times is likely to have enhanced seismic reflectivity of the lower crust, especially in the vicinity of the Pogallo Line and the anastomozing high-temperature shear zones. This enhancement would be primarily due to anisotropy effects associated with ductile shearing and retrograde syntectonic mica growth (Handy, 1987). Extension may also have enhanced seismic reflectivity by accentuating pre-existing lithological and metamorphic discontinuities within the crustal section. The thickness of the Ivrea section decreased from the Late Paleozoic value of ca. 25–30 km (pressure estimates of 8–9 kbar) to only 10–20 km by the end of the Early Mesozoic rifting (Handy, 1987).

Alpine deformation of the Ivrea Zone was very localized, so that most of the Ivrea section retained its original deep crustal features while being folded into its present steep orientation near the surface. The steepening of the crustal section within upper crustal levels is closely linked to the final emplacement of the Ivrea Geophysical Body into its present fan-like configuration within the Western Alpine arc during the Neogene (Figs. 1 and 3D). This final emplacement postdates an unusual interplay of pre-Neogene geotectonic circumstances, the most important of which are: (1) the inherited Paleozoic crustal edifice; (2) the Mesozoic and Tertiary thermal regimes; and (3) the kinematic situations during Early Mesozoic extension and Cretaceous compression. Early Mesozoic rifting is inferred to have involved oblique extension, based on the reconstructed original orientation of the Pogallo Ductile Fault Zone (Handy, 1987) and on the sedimentological evidence in dismembered Southern Penninic (i.e., Tethyan) ophiolites (Weissert and Bernoulli, 1985). Thus, rifting probably effected sloping of the Southern Alpine Moho into a SE-dipping orientation (fig. 8 of S.M. Schmid et al., 1987), analogous to that presently observed at the base of the Armorican continental margin (e.g., Le Pichon and Barbier, 1987). Important for the kinematics of emplacement is that this southeast dip was oblique to the subsequent E–W directed Cretaceous Eoalpine compression. At this time, thinned Austroalpine continental crust (Sesia Zone) was subducted beneath both upper mantle and Southern Alpine continental crust (Ivrea and Strona–Ceneri zones). This gave rise to the density inversion that presently characterizes the base of the Ivrea Geophysical Body (fig. 9 in S.M. Schmid et al., 1987). Subsequent Oligo-Miocene dextral transpression (i.e., the Insubric event) induced a steepening of the Ivrea Geophysical Body to its present attitude (Fig. 3D and fig. 10 in S.M. Schmid et al., 1987). The steeply dipping Ivrea Zone is only a surficial manifestation of this impressive Alpine structure that resulted from the obliquity of the Alpine movement vector with respect to the pre-existing Mesozoic structures.

What emerges from the discussion above is that the Ivrea Zone is a composite of different magmatic, metamorphic and tectonic features. When we propose the Ivrea section as an example of attenuated deep continental crust, we refer to the configuration acquired at the end of the Mesozoic. The Ivrea section differs from in-situ lower crust in that it has been locally folded and mylonitized during emplacement into the upper crust. The main effects of the Alpine emplacement history are the steepening of the crustal section and the formation of a density inversion below the Ivrea Body (Fig. 3D). These modifying effects have to be removed if the Ivrea Zone is to be used for geophysical modelling. However, Alpine tectonics did not substantially alter the original internal configuration. We conclude that the Ivrea Zone represents an excellent example of attenuated lower continental crust.

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