Emplacement of mantle-derived magma (magmatic accretion) is often presumed or inferred to be an important cause of regional granulite facies metamorphism and crustal anatexis. The juxtaposition of mafic cumulates and regionally distributed granulite facies rocks has led some to consider the Ivrea zone (northern Italy, Southern Alps) as an important exposure that demonstrates this causal relationship. However, regional PT paths indicated by metamorphic reaction textures and PT conditions inferred from geothermobarometry indicate that the emplacement of mafic plutonic rocks (Mafic Complex) at the Ivrea zone occurred during decompression from ambient pressures at the regional thermal maximum. Field and petrographic observations, supported by PT estimates, indicate that regional retrograde decompression and emplacement of the upper parts of the Mafic Complex probably accompanied extension during the Late Carboniferous–Early Permian. A spatially restricted decompression-melting event accompanied final emplacement, depleting supracrustal rocks enclosed by an <2–3 km aureole overlying the upper Mafic Complex by 20–30% granite component. The upper Mafic Complex provided the thermal energy to reset mineral assemblages and locally overprint the regional prograde metamorphic zonation. The limited extent of the contact aureole suggests that magmatic accretion may not inexorably cause regional metamorphism and crustal anatexis.

KEY WORDS: Granulite facies; Ivrea zone; underplating; thermobarometry; migmatite; mafic complex

INTRODUCTION
The accretion of mantle-derived magma at or near the base of the crust (Wells, 1980; Bohlen, 1987) can enhance the total thermal budget of the continental lower crust, produce regional granulite facies metamorphism, and generate Y- and heavy rare earth element (HREE)-depleted granitoids (Ellis, 1987). Accretion of mafic magma and migration of partial melt will internally stratify, chemically differentiate, and deplete the continental lower crust in large ion lithophile elements (LILE). Magmatic accretion has been invoked to provide the heat and mass necessary for sustained magmatism in tectonic settings such as Phanerozoic extensional terranes (Lister et al., 1986; Gans, 1987; Fountain, 1989; Mareschal & Bergantz, 1990; Jarchow et al., 1993) and magmatic arcs (Hamilton, 1981; Kay & Kay, 1981; Bohlen & Lindsley, 1987; Hildreth & Moores, 1988).

However, as few exposed sections of lower continental crust show contiguous mafic intrusions and regionally distributed granulite facies rocks, estimates of the extent of anatexis and metamorphism accompanying magmatic accretion have relied on numerical and analog simulations. These models yield disparate results depending on whether heat transfer within the mafic intrusion and surrounding country rocks is primarily convective (Campbell & Turner, 1987; Huppert & Sparks, 1988) or conductive (Marsh, 1989; Bergantz & Dawes, 1994; Barboza & Bergantz, 1996). A comparison with field evidence is required to discriminate between the model results. In the Ivrea zone, an exposed section of mafic plutonic rocks (the Mafic Complex) has been interpreted to represent deformed cumulates of the accreted magmas that caused regional granulite facies metamorphism (Ri-valetti et al., 1975, 1980; Schmid & Wood, 1976; Sills, 1984; Pin, 1990). The continuous exposure between granulite facies crustal rocks and the mafic intrusion that...
is thought to have provided the heat during regional metamorphism has led some to consider the Ivrea zone a particularly important example of magmatic accretion (e.g. Voshage et al., 1990).

However, Barboza et al. (1999) provided field and geochemical data supporting an alternative model (Zingg et al., 1990; Schmid, 1993), in which regional granulitic facies metamorphism preceded the emplacement of the upper parts of the Mafic Complex. In this paper, we present whole-rock major- and trace-element compositional data, an inferred sequence of metamorphic reactions, and PT estimates derived from thermobarometry for supracrustal rocks proximal to the Mafic Complex. These new data suggest that the emplacement of a major part of the Mafic Complex occurred during decompression from ambient pressures at the thermal maximum during the regional granulite facies episode. Final emplacement caused anatexis and metamorphism only within a narrow (~2–3 km) aureole in the proximal supracrustal rocks. These events overprint the regional amphibolite to granulite facies prograde metamorphic zonation.

**ROCK TYPES**

**Mafic Complex**
The Mafic Complex, commonly interpreted to have been the heat source for regional metamorphism, consists of mafic plutonic rocks and mantle peridotite that form a belt spanning the length of the Ivrea zone along its western and northwestern margin. For the purpose of this study, we make an important distinction between the upper and lower parts of the Mafic Complex. Upper Val Grande and Val d’Ossola expose several hundred meters to 1 km thickness of banded pyroxene–hornblende and hornblende granofels (‘pyribolite’; Schmid, 1967). Similar rocks are exposed in upper Val Sesia and Val Strona di Omeigna (Mehnert, 1975; Zingg, 1980) and the regional baric gradient according to thermobarometry (Sills, 1984; Henk et al., 1997; Demarchi et al., 1998).

**GEOLOGICAL SETTING**

**Regional framework**
The Ivrea zone is one of three fault-bounded sections of Paleozoic basement exposed in the Southern Alps. From the originally deepest to shallowest crustal levels, the Paleozoic Southern Alpine crust comprises the Ivrea, Strona–Ceneri (or Serie dei Laghi in the Italian literature), and Val Colla zones. To the north and west, the Ivrea zone (Fig. 1) is separated from rocks of the Penninic belt by the Insubric Line, a major, NW-dipping, Neogene shear zone that juxtaposes pre-Alpine and Alpine structures and rocks (Gansser, 1968; Schmid et al., 1987). To the south and east, the Cremosina, Cossato–Mergozzo–Brissago (CMB), and the Pogallo Lines separate the Ivrea zone from plutonic rocks and amphibolite facies orthogneiss and paragneiss of the Strona–Ceneri zone (Boriani & Sacchi, 1973; Boriani et al., 1977, 1990a; Hodges & Fountain, 1994; Handy, 1987; Schmid et al., 1987).

Most regional studies have interpreted the Ivrea zone as a cross-section through attenuated lower continental crust (Fountain, 1976; Burke & Fountain, 1990). This interpretation is based on gravity and seismic studies that indicate the Ivrea zone is the surface expression of a large, high-density body (the Ivrea geophysical body) situated where the fossilized (pre-Alpine) Southern Alpine Moho comes closest to the surface (Mueller et al., 1980; Giese et al., 1982). Geophysical modeling suggests that the Ivrea body may be a NNE-striking, SE-dipping sliver of Southern Alpine crust and mantle juxtaposed against Austroalpine and Penninic crustal rocks to the north and west (Berckhemer, 1968; Giese et al., 1982; Hirn et al., 1989). This geophysical model is consistent with structural data indicating that steep tilting accompanied emplacement during Alpine collision and closure of the Tethys (Schmid et al., 1987; Nicolas et al., 1990). The mafic plutonic rocks exposed along the Insubric Line (the Mafic Complex) represent the originally deepest levels of exposed crust, and the shallowest levels are represented by supracrustal rocks (the Kinzigite Formation) juxtaposed with the Strona–Ceneri zone (Fig. 1). Although alternative interpretations have been proposed (see Boriani et al., 1990b), steep tilting of the Ivrea zone is supported by the regional increase in metamorphic grade to the NW (Schmid, 1967; Mehnert, 1975; Zingg, 1980) and the regional baric gradient according to thermobarometry (Sills, 1984; Henk et al., 1997; Demarchi et al., 1998).
thickness of 10 km. In this area, the upper Mafic Complex comprises leucodioritic to gabbroic rocks intercalated with ultramafic rocks and layers of banded granulite (Quick et al., 1994; Sinigoi et al., 1994). On the basis of field and compositional relationships, these researchers interpreted the upper Mafic Complex to have been emplaced during Late Carboniferous–Early Permian extension within the Ivrea zone. Space for the Mafic
Complex may have been partially accommodated by northward transport and ductile attenuation of the overlying Kinzigite Formation (Quick et al., 1994; Snoke et al., 1999). South of Val Sesia, lensoidal bodies of tonalite to granodiorite diatexite ≤ 200 m thick, which contain blocks and schlieren of metabasite and metapelites, separate the Mafic Complex from the overlying Kinzigite Formation (Bürgi & Klötzli, 1990; Quick et al., 1994). Phase relations and PT estimates indicate that emplacement of the Mafic Complex occurred at crustal depths from 4–5 kbar to 8–10 kbar (Demarchi et al., 1998).

Kinzigite Formation

The Kinzigite Formation comprises amphibolite-to-granulite facies metapelites and metapsammite, with subordinate metacarbonate and metabasite. Metabasite south of Val Sesia (Fig. 1) occurs as 25 cm–50 m thick, foliation-parallel lenses or layers intercalated with metapelites and minor metacarbonate. In Val d’Ossola and Val Strona di Omegna, metabasite is more common and thick layers of metabasite and metapelites alternate (Schmid, 1967; Zingg, 1980; Sills & Tarney, 1984). In amphibolite facies rocks, metapelites contain nematoblastic amphibole + plagioclase ± clinopyroxene. Orthopyroxene appears in granoblastic amphibole- and plagioclase-bearing metapelites with the transition to the granulite facies. Metacarbonate occurs in lenses and bands ≤ 40 m thick, which contain a variety of mineral assemblages (Schmid, 1967; Zingg, 1980).

The most common lithology of supracrustal rock is high-Al migmatitic metapelite with sillimanite as the stable aluminosilicate polymorph. In lower Val Strona di Omegna, metapelites commonly contains biotite + quartz + plagioclase ± garnet ± sillimanite ± muscovite ± K-feldspar. Primary muscovite is restricted to the lowest grade rocks in lower Val Strona di Omegna and Val d’Ossola. Toward the NW, in upper Val Strona di Omegna, the modal proportions of garnet and K-feldspar increase at the expense of biotite and muscovite (Schmid & Wood, 1976; Schmid, 1978–1979). With the exception of rare occurrences in lowermost Val Strona di Omegna (Zingg, 1980) and Val d’Ossola (Peyronel Pagliani & Boriani, 1967), cordierite and hercynitic spinel are restricted to metapelites proximal to the contact with the upper Mafic Complex (Fig. 1). In agreement with Sills (1984), we observed orthopyroxene with garnet, biotite, and plagioclase in one Mg-rich and Al-poor metapelitic. Common accessory minerals include rutile, ilmenite, zircon, graphite, apatite, monazite, pyrite, chalcopyrite, and corundum.

Geochronology

Geochronology indicates a Late Carboniferous-Early Permian crystallization age for the upper Mafic Complex. Pin (1986) reported six fractions of nearly concordant zircon U–Pb data with an upper intercept of 285 ± 7/5 Ma from a diorite sample from the upper Mafic Complex. A Carboniferous–Early Permian age for emplacement is supported by Sm–Nd mineral isochrons and a whole-rock Rb–Sr age (Bürgi & Klötzli, 1990; Voshage et al., 1990). Microanalytic U/Pb ages of newly grown zircon within two minor members of the lower Mafic Complex are 299 ± 5 Ma (Vavra et al., 1999).

The imposition of the regional metamorphic zonation probably occurred within 20 My of emplacement of the upper Mafic Complex (see Hunziker & Zingg, 1980). Vavra et al. (1996, 1999) suggested regional granulite facies metamorphism occurred in a single event before 299 ± 5 Ma, on the basis of an ion microprobe study of zircons from granulite within the lower Mafic Complex and from supracrustal rocks in upper Val Strona di Omegna. A Late Carboniferous age for the regional granulite facies metamorphism is consistent with U–Pb ages on monazite (Köppel, 1974; Henk et al., 1997) and the inferred age of lead loss of 99% (granulite facies rocks) to 83% (amphibolite facies rocks) to discordant zircon populations (Köppel & Grünfelder, 1971; Köppel, 1974). These studies indicate regional metamorphism before 275 ± 2 Ma (Köppel, 1974), 292 ± 2 Ma (Henk et al., 1997), and 285 ± 10 Ma (Köppel & Grünfelder, 1971; Köppel, 1974), respectively, and may reflect cooling below 600°C (monazite) and conditions close to thermal peak (zircon). Although Henk et al. (1997) and Vavra & Schaltegger (1999) differed in their interpretation of decreasing monazite ages from 292 ± 2 Ma near the Pogallo Line to 276 ± 2 Ma near the Insubric Line, both studies support a Late Carboniferous age for the regional thermal peak of metamorphism.

RESULTS

Sample suite

To constrain the relationship between emplacement of the upper Mafic Complex and the regional metamorphic PTi history of the Ivrea zone, we collected samples for petrographic study and geochemical analyses on four traverses across the regional metamorphic zonation through the Kinzigite Formation (Fig. 1). Sample locations were selected from our mapping of the supracrustal section overlying the upper Mafic Complex south of Val Sesia and the maps of Bertolani (1959–1965) and Schmid (1967) for Val Strona di Omegna and Val d’Ossola, respectively. We compared whole-rock major- and trace-element compositions, metamorphic reaction textures,
and mineral compositions in samples collected south of Val Sesia (proximal to the upper Mafic Complex) with more distal samples collected in lower Val Strona di Omegna and Val d’Ossola. Sample locations are indicated in Figs 1 and 2.

**Whole-rock geochemistry**

*Analytical techniques*

The migmatites were separated for analysis by first sawing the sample into slabs of leucosome and melanosome. A representative split of 100–500 g of powdered melanosome sample was then taken for X-ray fluorescence (XRF) analysis. Major and trace elements were analyzed using a Philips PW 1404 XRF spectrometer. Corrections were based on a calibration using 40 international standards according to procedures described by Franzini et al. (1975) and Leoni & Saitta (1975). The detection limits are 0.01% for major elements and 10 ppm for trace elements. Analytical uncertainties are ±1% of the value for major elements, except for Mg and Na, for which they are ±5%. Analytical uncertainties of trace-element analyses are ±5–10 ppm for the concentrations at >100 ppm and ±10–20 ppm for those at 30–50 ppm.

*Results*

New whole-rock, and major- and trace-element compositions of selected metapelite melanosome samples are listed in Table 1. Locations for these selected samples are indicated in Figs 1 and 2. Whole-rock compositions projected from muscovite (Val d’Ossola samples) and K-feldspar (F. Duggia and R. Forcioula samples) are depicted in Fig. 3. AFM projections of melanosome compositions indicate that some samples are enriched in Al over Fe and Mg relative to typical high-Al metapelites. Mass balance calculations indicate a systematic depletion of SiO₂, the alkalies, and incompatible trace elements in melanosomes overlying the upper Mafic Complex relative to lower-grade metapelites in lower Val d’Ossola and Val Strona di Omegna (Barboza, 1998). Leucosome adjacent to the melanosome samples collected within Kinzigite Formation overlying the upper Mafic Complex typically are tonalitic in composition, lack mafic phases, and are depleted in K₂O relative to the inferred melt extracted from the melanomes (Barboza, 1998; Barboza et al., 1999). Banded granulite intercalated with the lower Mafic Complex and exposed in upper Val Strona di Omegna and Val d’Ossola are depleted in incompatible elements and enriched in compatible elements relative to their amphibolite facies equivalents (Sighinolfi & Gorgoni, 1978; Schnetger, 1994; Barboza et al., 1999).

**Inferred reaction sequence**

*Between upper and lower Val Strona di Omegna and Val d’Ossola*

The increase in metamorphic grade to the NW in Val Strona di Omegna and Val d’Ossola with increasing original crustal depth characterizes the regional amphibolite to granulite facies metamorphism in the Ivrea zone. We present new data and review the relevant mineralogy and regional reaction textures for comparison with rocks proximal to the upper Mafic Complex only. Interested readers are directed to other studies for a complete analysis and review (Schmid, 1967, 1993; Mehnert, 1975; Schmid & Wood, 1976; Zingg, 1980; Sills, 1984; Zingg et al., 1990; Henk et al., 1997).

On the basis of a compilation of previous petrographic work, Zingg (1980) located mineral isograds that represent prograde reactions produced during the high-grade, regional metamorphic episode (Fig. 1). These isograds include the muscovite-K-feldspar isograd in sillimanite-bearing metapelite, the first appearance of coexisting pyroxenes in metabasite, and the upper stability limit of calcite–quartz–actinolite in calcisilicate. A textural change in metapelite attends the replacement of lepidoblastic muscovite and biotite by granoblastic K-feldspar and garnet with increasing grade (Schmid, 1967; Schmid & Wood, 1976; Hunziker & Zingg, 1980). In the granulite facies rocks of upper Val Strona di Omegna and Val d’Ossola, quartz + hypersthene ± garnet granulite and calcsilicate are interlayered with granoblastic graphite + sillimanite + garnet gneiss, interpreted to be residuum from the melting of metapelite (Schmid & Wood, 1976; Sighinolfi & Gorgoni, 1978; Schmid, 1978–1979; Schnetger, 1994). Schmid & Wood (1976) attributed the progressive increase in modal garnet at the expense of biotite to the continuous reaction

\[
\text{biotite + sillimanite + quartz} \Rightarrow \text{garnet + K-feldspar + rutile + H}_2\text{O}
\]  

Two phases of muscovite growth are apparent SE of the muscovite-out isograd (Fig. 1) in lower Val Strona di Omegna and Val d’Ossola. Muscovite₂ is distinguished from muscovite, in that its long axis is oblique to the foliation defined by muscovite, biotite, and sillimanite (Fig. 4a). This relationship suggests that muscovite₂ may have appeared during retrograde cooling and hydration through the reaction

\[
\text{sillimanite + K-feldspar + H}_2\text{O} \Rightarrow \text{muscovite + quartz}
\]  

an inference supported by common textures indicating the direct replacement of sillimanite + K-feldspar by symplectitic intergrowths of muscovite + quartz.
Fig. 2. Detail sample location maps along R. Forcioula and F. Duggia (see Fig. 1). Underlined sample numbers indicate samples used for thermobarometry. Across-strike baric gradient along the R. Forcioula transect with distance from the contact between the Kinzigite Formation and the upper Mafic Complex is also shown.

Between lower Val Sesia and Val Strona di Postua
The longest sections of continuous exposure south of Val Sesia that extend across the contact between the Kinzigite Formation and the upper Mafic Complex are along Rio Forcioula in Val Strona di Postua and Fiume Duggia in Val Duggia (Figs 1 and 2). Mineral assemblages for selected samples collected along these transects are listed in Table 2. Metamorphic grade increases to the SW from the amphibolite facies in lower Val Strona di Postua to the omegna, parallel to the regional fabric toward Val
muscovite + quartz ⇌ sillimanite + K-feldspar + H₂O  

or an H₂O-conserved melting reaction such as

muscovite + plagioclase + K-feldspar + quartz ⇒ sillimanite + melt.  

At pressures below 6 kbar, either reaction (5) or (6) in intermediate X₃₆ metapelite indicates that temperatures proximal to the contact between the upper Mafic Complex and the Kinzigite Formation exceeded >600°C (Vielzeuf & Holloway, 1988).

In lower Val Strona di Omegna, biotite is the major mineral in metapelite. However, biotite is irregularly distributed in metapelite overlying the Mafic Complex south of Val Sesia: some samples contain <5% modal biotite. The decrease in modal biotite to the SW is indicated by the appearance of abundant cordierite within metapelite proximal to the upper Mafic Complex.

The more abundant cordierite, occurs as up to 5 mm prisms that cut the foliation defined by biotite and sillimanite (Capedri et al., 1999) within migmatitic metapelite and occurs abundantly with K-feldspar porphyroblasts in the tonalite to granodiorite diatexite (Bürgi & Klotzli, 1990; Quick et al., 1994) separating the upper Mafic Complex from the overlying Kinzigite Formation south of Val Sesia. These K-poor leucosomes are restricted to the 2–3 km wide aureole that surrounds the upper Mafic Complex, coincident with the appearance of cordierite and hercynitic spinel in the melanosome assemblage. These relationships indicate that cordierite, was a peritectic phase produced during the melting reaction from which the leucosome leucosomes were derived.

The discontinuous reaction Bt + Sill = Grt + Crd is indicated by the appearance of cordierite with biotite in metapelite proximal to the upper Mafic Complex.

These K-poor leucosomes are restricted to the 2–3 km wide aureole that surrounds the upper Mafic Complex, coincident with the appearance of cordierite and hercynitic spinel in the melanosome assemblage. These relationships indicate that cordierite, was a peritectic phase produced during the melting reaction from which the leucosome leucosomes were derived.

At pressures below 6 kbar, either reaction (5) or (6) in intermediate X₃₆ metapelite indicates that temperatures proximal to the contact between the upper Mafic Complex and the Kinzigite Formation exceeded >600°C (Vielzeuf & Holloway, 1988).

In lower Val Strona di Omegna, biotite is the major mineral in metapelite. However, biotite is irregularly distributed in metapelite overlying the Mafic Complex south of Val Sesia: some samples contain <5% modal biotite. The decrease in modal biotite to the SW is indicated by the appearance of abundant cordierite within metapelite proximal to the upper Mafic Complex.

The more abundant cordierite, occurs as up to 5 mm prisms that cut the foliation defined by biotite and sillimanite (Capedri et al., 1999) within migmatitic metapelite and occurs abundantly with K-feldspar porphyroblasts in the tonalite to granodiorite diatexite (Bürgi & Klotzli, 1990; Quick et al., 1994) separating the upper Mafic Complex from the overlying Kinzigite Formation south of Val Sesia. These K-poor leucosomes are restricted to the 2–3 km wide aureole that surrounds the upper Mafic Complex, coincident with the appearance of cordierite and hercynitic spinel in the melanosome assemblage. These relationships indicate that cordierite, was a peritectic phase produced during the melting reaction from which the leucosome leucosomes were derived.

The more abundant cordierite, occurs as up to 5 mm prisms that cut the foliation defined by biotite and sillimanite (Capedri & Rivalenti, 1973). Idiomorphic cordierite, found in minor amounts, encloses fibrolite and is elongate parallel to the foliation defined by biotite, sillimanite, and the long axes of garnet augen. This observation implies that cordierite growth followed reaction (4). We infer that cordierite, may have formed from the continuous reaction

\[
garnet + sillimanite + quartz \rightarrow \text{cordierite}_1. \quad (7)
\]

Cordierite is typically heavily pinitized, but, when preserved, at least two generations are present (Capedri & Rivalenti, 1973). Cordierite, found in minor amounts, encloses fibrolite and is elongate parallel to the foliation defined by biotite, sillimanite, and the long axes of garnet augen. This observation implies that cordierite growth followed reaction (4). We infer that cordierite, may have formed from the continuous reaction

\[
garnet + sillimanite + quartz \rightarrow \text{cordierite}_1. \quad (7)
\]

The more abundant cordierite, occurs as up to 5 mm prisms that cut the foliation defined by biotite and sillimanite (Capedri & Rivalenti, 1973). Idiomorphic cordierite, is present in leucosome leucosomes (Barboza et al., 1999) within migmatitic metapelite and occurs abundantly with K-feldspar porphyroblasts in the tonalite to granodiorite diatexite (Bürgi & Klotzli, 1990; Quick et al., 1994) separating the upper Mafic Complex from the overlying Kinzigite Formation south of Val Sesia. These K-poor leucosomes are restricted to the 2–3 km wide aureole that surrounds the upper Mafic Complex, coincident with the appearance of cordierite and hercynitic spinel in the melanosome assemblage. These relationships indicate that cordierite, was a peritectic phase produced during the melting reaction from which the leucosome leucosomes were derived.
Table 1: Whole rock major- and trace-element compositions of representative metapelite melanosomes

<table>
<thead>
<tr>
<th>Sample no.:</th>
<th>-11</th>
<th>-13</th>
<th>-16</th>
<th>-21</th>
<th>-03</th>
<th>-05</th>
<th>-05</th>
<th>-02</th>
<th>-03</th>
<th>-05</th>
<th>-10</th>
<th>-01</th>
<th>-03</th>
<th>-03</th>
<th>-07A</th>
<th>-09</th>
<th>-01</th>
</tr>
</thead>
<tbody>
<tr>
<td>Location</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>RD</td>
<td>RD</td>
<td>RD</td>
<td>RD</td>
<td>RD</td>
<td>RD</td>
<td>RD</td>
<td>RD</td>
<td>RD</td>
</tr>
</tbody>
</table>

**Major oxide abundances by X-ray fluorescence analysis (wt %)**

<table>
<thead>
<tr>
<th>Oxide</th>
<th>Sample 80496</th>
<th>Sample 80496</th>
<th>Sample 80496</th>
<th>Sample 80596</th>
<th>Sample 62097</th>
<th>Sample 62097</th>
<th>Sample 70197</th>
<th>Sample 70197</th>
<th>Sample 62497</th>
<th>Sample 62497</th>
<th>Sample 61797</th>
<th>Sample 61797</th>
<th>Sample 61797</th>
<th>Sample 61797</th>
<th>Sample 61797</th>
<th>Sample 82395</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>54.72</td>
<td>47.22</td>
<td>46.49</td>
<td>42.37</td>
<td>42.33</td>
<td>38.68</td>
<td>41.78</td>
<td>52.72</td>
<td>53.75</td>
<td>59.5</td>
<td>61.61</td>
<td>55.41</td>
<td>41.78</td>
<td>58.58</td>
<td>57.96</td>
<td>54.86</td>
</tr>
<tr>
<td>TiO₂</td>
<td>1.75</td>
<td>1.82</td>
<td>1.98</td>
<td>2.59</td>
<td>2.13</td>
<td>2.31</td>
<td>1.65</td>
<td>1.65</td>
<td>1.34</td>
<td>1.31</td>
<td>1.43</td>
<td>2.31</td>
<td>1.40</td>
<td>1.40</td>
<td>1.21</td>
<td>1.62</td>
</tr>
<tr>
<td>MnO</td>
<td>0.16</td>
<td>0.21</td>
<td>0.16</td>
<td>0.18</td>
<td>0.28</td>
<td>0.21</td>
<td>0.19</td>
<td>0.11</td>
<td>0.09</td>
<td>0.09</td>
<td>0.10</td>
<td>0.13</td>
<td>0.19</td>
<td>0.09</td>
<td>0.22</td>
<td>0.08</td>
</tr>
<tr>
<td>MgO</td>
<td>4.39</td>
<td>6.19</td>
<td>4.35</td>
<td>5.24</td>
<td>6.91</td>
<td>8.00</td>
<td>4.60</td>
<td>3.6</td>
<td>3.81</td>
<td>3.15</td>
<td>4.26</td>
<td>3.24</td>
<td>4.6</td>
<td>3.12</td>
<td>2.35</td>
<td>2.75</td>
</tr>
<tr>
<td>CaO</td>
<td>2.40</td>
<td>0.91</td>
<td>0.43</td>
<td>0.54</td>
<td>2.90</td>
<td>0.59</td>
<td>1.06</td>
<td>0.37</td>
<td>0.37</td>
<td>0.37</td>
<td>1.08</td>
<td>0.4</td>
<td>1.06</td>
<td>0.31</td>
<td>0.63</td>
<td>0.39</td>
</tr>
<tr>
<td>Na₂O</td>
<td>1.31</td>
<td>0.51</td>
<td>0.2</td>
<td>0.14</td>
<td>1.77</td>
<td>0.16</td>
<td>0.66</td>
<td>0.38</td>
<td>0.23</td>
<td>0.22</td>
<td>0.68</td>
<td>0.22</td>
<td>0.66</td>
<td>0.57</td>
<td>1.48</td>
<td>1.43</td>
</tr>
<tr>
<td>K₂O</td>
<td>1.81</td>
<td>2.71</td>
<td>1.58</td>
<td>1.93</td>
<td>3.14</td>
<td>3.53</td>
<td>1.41</td>
<td>2.48</td>
<td>3.31</td>
<td>2.18</td>
<td>2.97</td>
<td>2.85</td>
<td>1.41</td>
<td>4.40</td>
<td>3.51</td>
<td>4.79</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.10</td>
<td>0.03</td>
<td>0.06</td>
<td>n.d.</td>
<td>0.04</td>
<td>0.03</td>
<td>0.07</td>
<td>0.02</td>
<td>0.05</td>
<td>0.04</td>
<td>0.21</td>
<td>0.05</td>
<td>0.07</td>
<td>0.10</td>
<td>0.14</td>
<td>0.10</td>
</tr>
<tr>
<td>Total</td>
<td>98.37</td>
<td>97.79</td>
<td>99.68</td>
<td>97.03</td>
<td>98.34</td>
<td>97.26</td>
<td>98.74</td>
<td>98.33</td>
<td>98.43</td>
<td>97.92</td>
<td>98.24</td>
<td>98.74</td>
<td>97.99</td>
<td>97.79</td>
<td>96.76</td>
<td>96.92</td>
</tr>
</tbody>
</table>

**Trace element abundances by X-ray fluorescence analysis (ppm)**

<table>
<thead>
<tr>
<th>Element</th>
<th>Sample 80496</th>
<th>Sample 80496</th>
<th>Sample 80496</th>
<th>Sample 80596</th>
<th>Sample 62097</th>
<th>Sample 62097</th>
<th>Sample 70197</th>
<th>Sample 70197</th>
<th>Sample 62497</th>
<th>Sample 62497</th>
<th>Sample 61797</th>
<th>Sample 61797</th>
<th>Sample 61797</th>
<th>Sample 61797</th>
<th>Sample 61797</th>
<th>Sample 82395</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nb</td>
<td>24</td>
<td>20</td>
<td>24</td>
<td>32</td>
<td>18</td>
<td>24</td>
<td>28</td>
<td>21</td>
<td>22</td>
<td>17</td>
<td>15</td>
<td>18</td>
<td>28</td>
<td>23</td>
<td>16</td>
<td>20</td>
</tr>
<tr>
<td>Zr</td>
<td>252</td>
<td>285</td>
<td>240</td>
<td>300</td>
<td>435</td>
<td>450</td>
<td>329</td>
<td>327</td>
<td>289</td>
<td>237</td>
<td>450</td>
<td>329</td>
<td>270</td>
<td>329</td>
<td>271</td>
<td>168</td>
</tr>
<tr>
<td>Y</td>
<td>44</td>
<td>80</td>
<td>42</td>
<td>40</td>
<td>82</td>
<td>70</td>
<td>53</td>
<td>44</td>
<td>46</td>
<td>42</td>
<td>50</td>
<td>43</td>
<td>53</td>
<td>29</td>
<td>31</td>
<td>44</td>
</tr>
<tr>
<td>Sr</td>
<td>248</td>
<td>146</td>
<td>66</td>
<td>46</td>
<td>342</td>
<td>28</td>
<td>112</td>
<td>88</td>
<td>65</td>
<td>90</td>
<td>84</td>
<td>78</td>
<td>112</td>
<td>53</td>
<td>151</td>
<td>109</td>
</tr>
<tr>
<td>Rb</td>
<td>66</td>
<td>84</td>
<td>44</td>
<td>54</td>
<td>132</td>
<td>164</td>
<td>43</td>
<td>79</td>
<td>154</td>
<td>82</td>
<td>88</td>
<td>149</td>
<td>43</td>
<td>201</td>
<td>154</td>
<td>205</td>
</tr>
<tr>
<td>Ba</td>
<td>715</td>
<td>1120</td>
<td>770</td>
<td>506</td>
<td>1045</td>
<td>1275</td>
<td>348</td>
<td>576</td>
<td>826</td>
<td>544</td>
<td>513</td>
<td>617</td>
<td>348</td>
<td>378</td>
<td>654</td>
<td>551</td>
</tr>
<tr>
<td>La</td>
<td>n.d.</td>
<td>n.d.</td>
<td>n.d.</td>
<td>n.d.</td>
<td>n.d.</td>
<td>n.d.</td>
<td>155</td>
<td>72</td>
<td>71</td>
<td>60</td>
<td>65</td>
<td>155</td>
<td>52</td>
<td>49</td>
<td>54</td>
<td>86</td>
</tr>
</tbody>
</table>

n.d., not determined. Location abbreviations: RF, Rio Forcioula; VD, Val Duggia; VDO, Val d’Ossola; VSO, Val Strona di Omegna.
Fig. 4. Electron microprobe backscatter images of (a) retrograde muscovite, nearly perpendicular to the foliation defined by muscovite, biotite, and sillimanite; sillimanite–garnet–biotite schist 82395-1, Val Strona di Omegna; (b) hercynitic spinel and ilmenite with sillimanite corona in cordierite core; sillimanite–cordierite–garnet gneiss 80596-5, Val Strona di Posuta; (c) retrograde replacement of cordierite + K-feldspar by symplectitic intergrowths of biotite + sillimanite + quartz; sillimanite–cordierite–garnet gneiss 62097-3; (d) close-up of cordierite, with reaction rim of radiating biotite + quartz + sillimanite symplectite; sillimanite–cordierite–garnet gneiss 62097-3.

Garnet grains in metapelite from lower Val Strona di Omegna and Val d’Ossola are typically idiomorphic, exhibit interior zoning, and are restricted to <250–300 μm in diameter (Fig. 4a). In contrast, garnet grains proximal to the upper Mafic Complex are typically elongate parallel to the foliation defined by biotite and sillimanite, do not exhibit interior zoning, and exceed 1100 μm in diameter. Coarse-grained garnet augen in metapelite proximal to the upper Mafic Complex commonly exhibit inclusion-rich cores surrounded by inclusion-free rims. We interpret these textural relationships to indicate at least two episodes of garnet growth, the latter of which occurred within the stability field of cordierite. Garnet growth associated with the development of peritectic cordierite implies that the reaction...
Table 2: Mineral assemblages

<table>
<thead>
<tr>
<th>Sample</th>
<th>Location</th>
<th>Qtz</th>
<th>Kf</th>
<th>Plg</th>
<th>Ga</th>
<th>Bi</th>
<th>Sil</th>
<th>Cd</th>
<th>Sp</th>
<th>Rt</th>
<th>Il</th>
<th>Pyr</th>
<th>Mn</th>
<th>Ap</th>
<th>Zr</th>
</tr>
</thead>
<tbody>
<tr>
<td>80496-6</td>
<td>RF</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>80496-11</td>
<td>RF</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>80496-13</td>
<td>RF</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>80596-3</td>
<td>RF</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>80596-5</td>
<td>RF</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>62097-5</td>
<td>RF</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>62097-3</td>
<td>RF</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>IV96-1</td>
<td>VD</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>70197-10</td>
<td>VD</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>62497-3</td>
<td>VD</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Location abbreviations as in Table 1 X indicates mineral present in sample.

biotite + sillimanite + quartz \(\Rightarrow\) garnet + cordierite, + K-feldspar + melt (8)

occurred within metapelite restricted to a 2–3 km aureole surrounding the upper Mafic Complex.

Grains of hercynitic spinel with sillimanite corona are occasionally present and are always enclosed within cordierite (Fig. 4b). This observation is consistent with the prograde reaction

biotite + sillimanite \(\Rightarrow\) cordierite, + hercynitic spinel + K-feldspar + ilmenite} + melt (9)

in locally quartz-deficient domains, or the retrograde reaction

sillimanite + melt \(\Rightarrow\) cordierite, + hercynitic spinel. (10)

In agreement with the observations of Capedri & Rivalenti (1973), we observed occasional neoblasts of fine-grained, idiomorphic, inclusion-free garnet, perhaps indicating an episode of post-tectonic garnet growth. Retrograde replacement of the assemblage cordierite, + K-feldspar is indicated by the development of symplectitic intergrowths of biotite + sillimanite + quartz (Fig. 4c and d). This observation implies retrograde hydration via reaction with a vapor phase or reaction with residual melt through the equilibrium

garnet + cordierite, + K-feldspar + H₂O/melt \(\Rightarrow\) biotite + sillimanite + quartz. (11)

Andalusite has been reported between lower Val Sesia and Val Sessera (Capedri & Rivalenti, 1973; Zingg, 1980; Demarchi et al., 1998). The reported abundance of andalusite increases to the SW, and it is common near Biella in Valle Mosso (Bertolani, 1959; Sacchi, 1962). Although Zingg (1980) interpreted andalusite to be a relict phase, Capedri & Rivalenti (1973) inferred that andalusite grew after sillimanite, on the basis of microstructural relationships. This implies a retrograde transition from the sillimanite to the andalusite stability field:

sillimanite \(\Rightarrow\) andalusite. (12)

Thermobarometry

The widespread occurrence of metapelite in the Kinzigite Formation provides favorable assemblages for thermobarometric constraints upon the regional PTi paths. We obtained pressure and temperature estimates with the GASP (3An \(\Rightarrow\) Grs + 2Als + Qtz) reaction and Fe–Mg exchange between garnet and biotite. Calculations were conducted with both MacPTAX (J. Lieberman, unpublished program, 1991) using a 1991 update of Berman’s thermodynamic database (Berman, 1988) and TWQ 2·02 (Berman, 1991). MacPTAX adopts the activity models of Berman & Aranovich (1995, 1996) for garnet and biotite, and Fuhrman & Lindsley (1988) for plagioclase. TWQ 2·02 adopts the activity models of Berman & Aranovich (1995, 1996) for garnet and biotite, and Fuhrman & Lindsley (1980) for plagioclase. MacPTAX typically returns temperatures and pressures >50–100°C and >1–2 kbar lower than TWQ 2·02.

We estimated PT conditions for nine samples collected on traverses along the Rio Forcioula (6) and Fiume Duggia (3) (Fig. 2). To estimate ‘peak’ conditions, we compared garnet core compositions free of local zoning near inclusions (determined by X-ray mapping and microprobe analyses) with plagioclase cores and neighboring matrix biotite not in contact with garnet. Mineral compositions and garnet and plagioclase zoning patterns
Table 3: Representative* matrix biotite analyses

<table>
<thead>
<tr>
<th>Sample no.:</th>
<th>80496</th>
<th>80496</th>
<th>80596</th>
<th>80596</th>
<th>62097</th>
<th>62097</th>
<th>IV96</th>
<th>70197</th>
<th>62497</th>
</tr>
</thead>
<tbody>
<tr>
<td>Location</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>VD</td>
<td>VD</td>
<td>VD</td>
</tr>
<tr>
<td>SiO₂</td>
<td>36.78</td>
<td>36.66</td>
<td>36.91</td>
<td>36.76</td>
<td>35.62</td>
<td>36.52</td>
<td>35.34</td>
<td>36.28</td>
<td>36.45</td>
</tr>
<tr>
<td>TiO₂</td>
<td>5.56</td>
<td>5.04</td>
<td>3.94</td>
<td>5.23</td>
<td>4.63</td>
<td>4.28</td>
<td>3.78</td>
<td>4.84</td>
<td>4.45</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>16.57</td>
<td>16.31</td>
<td>16.59</td>
<td>17.21</td>
<td>16.84</td>
<td>17.66</td>
<td>18.55</td>
<td>17.59</td>
<td>18.11</td>
</tr>
<tr>
<td>FeO⁷</td>
<td>13.60</td>
<td>13.08</td>
<td>12.91</td>
<td>14.08</td>
<td>14.15</td>
<td>16.06</td>
<td>19.11</td>
<td>15.39</td>
<td>14.00</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.09</td>
<td>0.05</td>
<td>0.14</td>
<td>0.09</td>
<td>0.40</td>
<td>0.11</td>
<td>0.13</td>
<td>0.10</td>
<td>0.16</td>
</tr>
<tr>
<td>Total</td>
<td>94.22</td>
<td>94.4</td>
<td>94.3</td>
<td>95.56</td>
<td>95.01</td>
<td>96.2</td>
<td>95.96</td>
<td>95.75</td>
<td>95.37</td>
</tr>
<tr>
<td>Si</td>
<td>5.51</td>
<td>5.49</td>
<td>5.51</td>
<td>5.45</td>
<td>5.36</td>
<td>5.42</td>
<td>5.33</td>
<td>5.41</td>
<td>5.41</td>
</tr>
<tr>
<td>Ti</td>
<td>0.63</td>
<td>0.57</td>
<td>0.44</td>
<td>0.58</td>
<td>0.52</td>
<td>0.48</td>
<td>0.43</td>
<td>0.54</td>
<td>0.50</td>
</tr>
<tr>
<td>Al</td>
<td>2.93</td>
<td>2.88</td>
<td>2.92</td>
<td>3.01</td>
<td>2.99</td>
<td>3.09</td>
<td>3.30</td>
<td>3.09</td>
<td>3.17</td>
</tr>
<tr>
<td>Al⁴⁺</td>
<td>2.49</td>
<td>2.51</td>
<td>2.49</td>
<td>2.56</td>
<td>2.64</td>
<td>2.58</td>
<td>2.67</td>
<td>2.60</td>
<td>2.59</td>
</tr>
<tr>
<td>Al⁶⁺</td>
<td>0.44</td>
<td>0.36</td>
<td>0.43</td>
<td>0.45</td>
<td>0.34</td>
<td>0.51</td>
<td>0.63</td>
<td>0.49</td>
<td>0.58</td>
</tr>
<tr>
<td>Fe</td>
<td>1.57</td>
<td>1.51</td>
<td>1.48</td>
<td>1.60</td>
<td>1.64</td>
<td>1.83</td>
<td>2.22</td>
<td>1.76</td>
<td>1.60</td>
</tr>
<tr>
<td>Mg</td>
<td>2.71</td>
<td>3.00</td>
<td>3.19</td>
<td>2.76</td>
<td>3.04</td>
<td>2.65</td>
<td>2.21</td>
<td>2.61</td>
<td>2.74</td>
</tr>
<tr>
<td>Na</td>
<td>0.03</td>
<td>0.01</td>
<td>0.04</td>
<td>0.02</td>
<td>n.d.</td>
<td>0.03</td>
<td>0.04</td>
<td>0.03</td>
<td>0.05</td>
</tr>
<tr>
<td>K</td>
<td>1.81</td>
<td>1.88</td>
<td>1.80</td>
<td>1.83</td>
<td>1.88</td>
<td>1.82</td>
<td>1.78</td>
<td>1.86</td>
<td>1.86</td>
</tr>
<tr>
<td>Xᵣ⁺</td>
<td>0.29</td>
<td>0.28</td>
<td>0.27</td>
<td>0.30</td>
<td>0.30</td>
<td>0.34</td>
<td>0.40</td>
<td>0.33</td>
<td>0.30</td>
</tr>
<tr>
<td>X₅⁺</td>
<td>0.50</td>
<td>0.56</td>
<td>0.57</td>
<td>0.51</td>
<td>0.55</td>
<td>0.48</td>
<td>0.48</td>
<td>0.48</td>
<td>0.51</td>
</tr>
<tr>
<td>Xₛ⁻</td>
<td>0.12</td>
<td>0.10</td>
<td>0.08</td>
<td>0.11</td>
<td>0.09</td>
<td>0.09</td>
<td>0.08</td>
<td>0.10</td>
<td>0.09</td>
</tr>
<tr>
<td>X₆⁻</td>
<td>0.09</td>
<td>0.07</td>
<td>0.08</td>
<td>0.08</td>
<td>0.06</td>
<td>0.09</td>
<td>0.11</td>
<td>0.09</td>
<td>0.11</td>
</tr>
</tbody>
</table>

*All mineral analyses obtained using the JEOL 733 electron microprobe at the University of Washington. The term ‘representative’ indicates that data are actual analyses (not averages) and are characteristic of the sample analyzed.

†Mole fraction $Z = Xᵣ⁺(Xₛ⁻ + X₅⁺ + X₆⁻ + X₄⁺)$. 

were obtained by X-ray mapping and electron microprobe analyses. Guided by the X-ray maps and microprobe transects, we analyzed 3–5 points per mineral and 4–6 inferred near-peak equilibrium mineral combinations (garnet–biotite–plagioclase) per sample. The compositional variability of minerals within a sample results in a range of calculated pressures and temperatures. Sample locations are provided in Fig. 2, and representative mineral analyses are listed in Tables 3–5. In accordance with Guidotti & Dyars’ (1991) empirical model, we assumed that ferric iron accounted for 8% of FeO⁷ in biotite. We assumed FeO⁷ = FeO in garnet.

The results of the thermobarometric calculations are depicted in Fig. 5 and tabulated in Table 6. Although estimated pressures (∼3–6 kbar) and temperatures (650–750°C) are consistent with the mineral assemblages, these conditions probably do not correspond to a local thermal maximum. As a result of retrograde Fe–Mg exchange and the fact that exchange and net transfer equilibria tend to have different closure temperatures (Frost & Chacko, 1989), our calculation of ‘peak’ conditions should be regarded as a minimum estimate of the PT conditions during a local thermal maximum.

High temperatures are also indicated by a change of garnet zoning patterns with proximity to the upper Mafic Complex. South of Val Sesia, the predominant garnets are medium- to coarse-grained augen, elongate parallel to the foliation defined by sillimanite and biotite. At least three phases of garnet growth are evident. Garnet augen consist of inclusion-rich cores, surrounded by inclusion-free rims. Late, fine-grained, inclusion-free, idiomorphic garnet cuts the foliation defined by biotite, sillimanite, and the long axes of garnet augen. Garnet grains from lower Val Strona di Omegna and Val d’Ossola samples are weakly zoned in grossular and, to a lesser extent, spessartine contents with the most extreme zonation being within ∼5–10 µm of their rims. In contrast, garnet grains in samples collected south of Val Sesia are virtually...
Table 4: Representative garnet core compositions.

<table>
<thead>
<tr>
<th>Sample no.:</th>
<th>RF</th>
<th>RF</th>
<th>RF</th>
<th>RF</th>
<th>RF</th>
<th>RF</th>
<th>RF</th>
<th>IV96</th>
<th>70197</th>
<th>62497</th>
</tr>
</thead>
<tbody>
<tr>
<td>Location</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>VD</td>
<td>VD</td>
<td>VD</td>
</tr>
<tr>
<td>SiO₂</td>
<td>38.13</td>
<td>38.44</td>
<td>38.46</td>
<td>38.72</td>
<td>38.16</td>
<td>38.78</td>
<td>37.84</td>
<td>37.96</td>
<td>38.19</td>
<td>38.78</td>
</tr>
<tr>
<td>TiO₂</td>
<td>&lt;d.l.</td>
<td>0.02</td>
<td>0.04</td>
<td>&lt;d.l.</td>
<td>0.02</td>
<td>0.02</td>
<td>&lt;d.l.</td>
<td>&lt;d.l.</td>
<td>&lt;d.l.</td>
<td>&lt;d.l.</td>
</tr>
<tr>
<td>FeO⁺</td>
<td>30.26</td>
<td>29.68</td>
<td>31.22</td>
<td>30.26</td>
<td>31.75</td>
<td>33.37</td>
<td>34.83</td>
<td>32.34</td>
<td>31.53</td>
<td>32.58</td>
</tr>
<tr>
<td>MnO</td>
<td>0.49</td>
<td>0.72</td>
<td>0.62</td>
<td>0.65</td>
<td>0.55</td>
<td>0.46</td>
<td>0.87</td>
<td>0.44</td>
<td>0.74</td>
<td>0.57</td>
</tr>
<tr>
<td>MgO</td>
<td>7.66</td>
<td>7.50</td>
<td>7.05</td>
<td>7.48</td>
<td>6.89</td>
<td>5.31</td>
<td>4.29</td>
<td>6.37</td>
<td>6.64</td>
<td>6.69</td>
</tr>
<tr>
<td>CaO</td>
<td>1.12</td>
<td>1.52</td>
<td>1.22</td>
<td>1.67</td>
<td>1.29</td>
<td>1.26</td>
<td>1.15</td>
<td>1.12</td>
<td>1.33</td>
<td>1.32</td>
</tr>
<tr>
<td>Total</td>
<td>99.19</td>
<td>99.54</td>
<td>99.87</td>
<td>99.21</td>
<td>100.38</td>
<td>99.43</td>
<td>100.4</td>
<td>100.12</td>
<td>100.14</td>
<td>100.38</td>
</tr>
<tr>
<td>Si</td>
<td>6.00</td>
<td>6.01</td>
<td>6.03</td>
<td>6.10</td>
<td>5.97</td>
<td>6.17</td>
<td>6.01</td>
<td>5.97</td>
<td>5.99</td>
<td>6.09</td>
</tr>
<tr>
<td>Ti</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>Al</td>
<td>3.99</td>
<td>3.99</td>
<td>3.93</td>
<td>3.79</td>
<td>4.01</td>
<td>3.79</td>
<td>4.01</td>
<td>4.05</td>
<td>4.01</td>
<td>3.78</td>
</tr>
<tr>
<td>Fe</td>
<td>3.98</td>
<td>3.88</td>
<td>4.10</td>
<td>3.99</td>
<td>4.15</td>
<td>4.44</td>
<td>4.63</td>
<td>4.25</td>
<td>4.13</td>
<td>4.28</td>
</tr>
<tr>
<td>Mn</td>
<td>0.07</td>
<td>0.10</td>
<td>0.08</td>
<td>0.09</td>
<td>0.07</td>
<td>0.06</td>
<td>0.12</td>
<td>0.06</td>
<td>0.10</td>
<td>0.08</td>
</tr>
<tr>
<td>Mg</td>
<td>1.79</td>
<td>1.75</td>
<td>1.65</td>
<td>1.76</td>
<td>1.61</td>
<td>1.26</td>
<td>1.02</td>
<td>1.49</td>
<td>1.55</td>
<td>1.57</td>
</tr>
<tr>
<td>Ca</td>
<td>0.19</td>
<td>0.26</td>
<td>0.21</td>
<td>0.28</td>
<td>0.22</td>
<td>0.21</td>
<td>0.20</td>
<td>0.19</td>
<td>0.22</td>
<td>0.22</td>
</tr>
<tr>
<td>Xₐm⁺</td>
<td>0.66</td>
<td>0.65</td>
<td>0.68</td>
<td>0.65</td>
<td>0.69</td>
<td>0.74</td>
<td>0.78</td>
<td>0.71</td>
<td>0.69</td>
<td>0.70</td>
</tr>
<tr>
<td>Xₘp⁺</td>
<td>0.01</td>
<td>0.02</td>
<td>0.01</td>
<td>0.01</td>
<td>0.01</td>
<td>0.02</td>
<td>0.01</td>
<td>0.01</td>
<td>0.02</td>
<td>0.01</td>
</tr>
<tr>
<td>Xₚpyr</td>
<td>0.30</td>
<td>0.29</td>
<td>0.27</td>
<td>0.29</td>
<td>0.27</td>
<td>0.21</td>
<td>0.17</td>
<td>0.25</td>
<td>0.26</td>
<td>0.26</td>
</tr>
<tr>
<td>Xₚgrs</td>
<td>0.03</td>
<td>0.04</td>
<td>0.03</td>
<td>0.05</td>
<td>0.04</td>
<td>0.04</td>
<td>0.03</td>
<td>0.03</td>
<td>0.04</td>
<td>0.04</td>
</tr>
</tbody>
</table>

<d.l., below detection limit.  
* Mole fraction $Z = X_alm + X_m + X_p + X_grs$.

unzoned (Barboza, 1998). The absence of compositional zoning within garnet grains from samples overlying the upper Mafic Complex probably indicates that homogenization at temperatures exceeding 600–650°C was followed by rapid cooling (Yardley, 1977; Whitney & Dilek, 1998).

Pressure–temperature history

**Petrogenetic grid**

Our regional PT path is based on the sequence of metamorphic reactions deduced from microstructural interpretations with additional constraints provided by PT estimates derived from geothermobarometry. The petrogenetic grid we used (Fig. 6a) includes the beginning of melting in the system Qz–Ab–An–Or–H₂O (Johannes, 1984) and relevant subsolidus and melting reactions for intermediate Xₐm pelites (Vielzeuf & Holloway, 1988) in the system KFMASH. The bulk composition of typical metapelite in the Ivrea zone approximates the intermediate Xₐm metapelite for which the grid applies (Fig. 3). For clarity, single lines represent the divariant fields for the equilibria in Fig. 6. The omission of Fe₂O₃, TiO₂, Na₂O, and CaO from the KFMASH model system precludes the consideration of reactions involving Fe–Ti oxides and plagioclase. The inclusion of an albite component shifts melting curves to >40°C lower temperatures (Vielzeuf & Holloway, 1988); an anorthite component reduces this effect (Thompson & Tracy, 1979; Carrington & Harley, 1995).

**Summary of inferred PTt path**

The thick shaded line in Fig. 6b illustrates PTt path ‘A’, for rocks proximal to the upper Mafic Complex south of Val Sesia. PTt path ‘B’ is for rocks distal to the upper Mafic Complex in Val Strona di Omegna. Both paths were inferred for rocks at the approximate structural level of the contact between the upper Mafic Complex and the Kinzigite Formation south of Val Sesia. Arrows indicate constraints on the PTt path for which we interpret microstructural evidence that the indicated reaction boundaries were crossed during metamorphism. PT and age constraints obtained from geothermobarometry (Fig. 5) and geochronology are also shown. Reported occurrences of relict kyanite and staurolite (Bertolani, 1959; Boriani & Sacchi, 1973; Zingg, 1980)
Table 5: Representative matrix plagioclase analyses.

<table>
<thead>
<tr>
<th>Sample no.:</th>
<th>80496</th>
<th>80496</th>
<th>80496</th>
<th>80596</th>
<th>80596</th>
<th>62097</th>
<th>62097</th>
<th>62497</th>
<th>IV96</th>
<th>70197</th>
<th>62497</th>
</tr>
</thead>
<tbody>
<tr>
<td>Location</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>RF</td>
<td>VD</td>
<td>VD</td>
<td>VD</td>
<td>VD</td>
</tr>
</tbody>
</table>

\[
\begin{array}{cccccccccccc}
\text{SiO}_2 & 57.64 & 55.51 & 58.97 & 59.72 & 58.37 & 58.92 & 60.41 & 57.76 & 56.80 & 58.57 \\
\text{Al}_2\text{O}_3 & 26.04 & 27.79 & 25.93 & 25.36 & 25.68 & 26.16 & 25.15 & 25.49 & 27.44 & 25.54 \\
\text{CaO} & 7.60 & 8.58 & 7.81 & 7.25 & 8.42 & 7.54 & 6.08 & 6.88 & 7.10 & 7.41 \\
\text{Na}_2\text{O} & 0.21 & 0.14 & 0.14 & 0.26 & 0.09 & 0.19 & 0.12 & 0.24 & 0.15 & 0.14 \\
\text{K}_2\text{O} & 0.01 & 0.01 & 0.01 & 0.01 & 0.01 & 0.01 & 0.01 & 0.01 & 0.01 & 0.01 \\
\text{FeO} & n.d. & <d.l. & 0.02 & <d.l. & n.d. & n.d. & 0.08 & n.d. & 0.08 & 0.04 \\
\text{Si} & 2.61 & 2.53 & 2.64 & 2.67 & 2.62 & 2.63 & 2.55 & 2.64 & 2.55 & 2.64 \\
\text{Al} & 1.39 & 1.49 & 1.37 & 1.33 & 1.36 & 1.34 & 1.45 & 1.37 & 1.45 & 1.36 \\
\text{Ca} & 0.37 & 0.42 & 0.37 & 0.35 & 0.40 & 0.36 & 0.44 & 0.33 & 0.44 & 0.36 \\
\text{Na} & 0.65 & 0.57 & 0.58 & 0.62 & 0.62 & 0.63 & 0.54 & 0.68 & 0.54 & 0.63 \\
\text{K} & 0.01 & 0.01 & 0.01 & 0.02 & 0.01 & 0.01 & 0.01 & 0.01 & 0.01 & 0.01 \\
\text{Fe} & n.d. & 0.00 & 0.00 & 0.00 & n.d. & n.d. & 0.00 & n.d. & 0.00 & 0.00 \\
\text{X}^a & 0.36 & 0.42 & 0.39 & 0.35 & 0.39 & 0.36 & 0.45 & 0.32 & 0.45 & 0.36 \\
\text{X}^b & 0.63 & 0.57 & 0.60 & 0.63 & 0.60 & 0.63 & 0.55 & 0.67 & 0.55 & 0.63 \\
\text{X}^c & 0.01 & 0.01 & 0.01 & 0.02 & 0.01 & 0.01 & 0.01 & 0.01 & 0.01 & 0.01 \\
\end{array}
\]

\*Mole fraction \( Z = \text{X}^a/(\text{X}^a + \text{X}^b + \text{X}^c) \).

The presence of peritectic cordierite, [reaction (8)] requires that temperatures exceeding 650°C were maintained proximal to the upper Mafic Complex during decompression from the regional thermal maximum. Decompression at lower temperatures leads to the breakdown of biotite producing cordierite in the presence of a vapor phase rather than a melt (Fig. 6a). Temperatures exceeding 650°C at moderate pressures are consistent also with \( PT \) estimates derived from geothermobarometry (Fig. 5) and with the absence of compositional zoning within garnet in proximal metapelite (Barboza, 1998). Finally, high-grade conditions for rocks proximal to the upper Mafic Complex are consistent with mass-balance calculations that indicate proximal metapelites are depleted by \( >20–30\% \) of a granite component relative to muscovite stability, constraining the conditions along this portion of the prograde path to \( P < 8 \) kbar and \( T < 700 \) °C. Additionally, we speculate that the common occurrence of fibrolite on biotite grains indicates that the staurolite-out reaction boundary was also crossed within the stability field of sillimanite, limiting this portion of the prograde path to \( P < 7 \) kbar and \( T < 700 \) °C. Together, these observations indicate that moderate pressures prevailed for much of the prograde \( PT \) evolution of the Ivrea zone. Henk et al. (1997) inferred that rocks in lower Val Strona di Omegna reached final equilibration from 4 to 6 kbar and from 650 to 700°C (marked ‘I’ in Fig. 6b).

The presence of peritectic cordierite, [reaction (8)] requires that temperatures exceeding 650°C were maintained proximal to the upper Mafic Complex during decompression from the regional thermal maximum. Decompression at lower temperatures leads to the breakdown of biotite producing cordierite in the presence of a vapor phase rather than a melt (Fig. 6a). Temperatures exceeding 650°C at moderate pressures are consistent also with \( PT \) estimates derived from geothermobarometry (Fig. 5) and with the absence of compositional zoning within garnet in proximal metapelite (Barboza, 1998). Finally, high-grade conditions for rocks proximal to the upper Mafic Complex are consistent with mass-balance calculations that indicate proximal metapelites are depleted by \( \sim 20–30\% \) of a granite component relative to muscovite stability, constraining the conditions along this portion of the prograde path to \( P < 8 \) kbar and \( T < 700 \) °C. Additionally, we speculate that the common occurrence of fibrolite on biotite grains indicates that the staurolite-out reaction boundary was also crossed within the stability field of sillimanite, limiting this portion of the prograde path to \( P < 7 \) kbar and \( T < 700 \) °C. Together, these observations indicate that the upper Mafic Complex provided the heat to reset local mineral assemblages during decompression from the baric gradient at the thermal maximum. These observations indicate that the upper Mafic Complex provided the heat to reset local mineral assemblages during decompression from the baric gradient at the thermal maximum. Mineral assemblages in rocks more distal to the upper Mafic Complex accompanied decompression from the regional baric gradient at the thermal maximum. These observations indicate that the upper Mafic Complex provided the heat to reset local mineral assemblages during decompression from the baric gradient at the thermal maximum. Mineral assemblages in rocks more distal to the upper Mafic Complex are incompletely reset and preserve evidence of the pre-existing regional metamorphic zonation.

The predominant microstructural feature of retrograde metamorphism that is preserved in metapelite proximal to the upper Mafic Complex is the replacement of cordierite and K-feldspar by symplectitic intergrowths of biotite + sillimanite + quartz [reaction (10)]. The d\( P/\)dT slope (Fig. 6a) of this equilibrium is shallow (Vielzeuf...
Fig. 5. (a) Shaded polygons depict the results of our thermobarometric calculations (TWQ 2·02) for metapelite overlying the upper Mafic Complex. Dimensions of the polygons determined by the compositional range of each sample. The results of the TWQ 2·02 calculations correlate with the results of Demarchi et al. (1998). The results of their study are summarized by the unfilled polygons that enclose the intersections of GASP and garnet–cordierite equilibria with \( P_{\text{H2O}} = P_{\text{total}} \) (a) and \( P_{\text{H2O}} = 0 \) (b) calculated for metapelite overlying the upper Mafic Complex (Demarchi et al., 1998). Insets (b) and (c) are example calculations for two samples illustrating the difference between the results of the two algorithms. Calculated \( PT \) positions of thermobarometric equilibria from individual garnet–biotite and garnet–plagioclase pairs bound the shaded regions. Continuous lines indicate MacPTAX calculations; dashed lines indicate TWQ 2·02 calculations. Complete results of MacPTAX and TWQ 2·02 calculations are given in Table 6.

DISCUSSION AND CONCLUSIONS

Timing of regional and contact metamorphism

Previous interpretation

Although Zingg et al. (1990), Schmid (1993), and Barboza et al. (1999) took an alternative view, most regional studies have presumed or inferred that final emplacement of the upper Mafic Complex caused regional metamorphism (e.g., Schmid & Wood, 1976; Hunziker & Zingg, 1980; Sills, 1984; Birgi & Klozzi, 1990; Schnetger, 1994; Sinigoi et al., 1996). This interpretation is based on geochronology, an inferred temperature gradient across the southeastern margin of the upper Mafic Complex (Schmid & Wood, 1976), and a textural gradation from hypidiomorphic–granular to granoblastic with increasing original depth within the upper Mafic Complex (Rivalenti et al., 1980; Quick et al., 1994). Rivalenti et al. (1980) and Pin (1990) suggested that this textural gradation is consistent with cooling subsequent to synmetamorphic intrusion.

However, some \( PT \) estimates (Sills, 1984) do not indicate a regional temperature gradient with proximity to the upper Mafic Complex. Alternative interpretations for the origin of the textural gradation within the upper Mafic Complex (Zingg et al., 1990; Quick et al., 1994)
Table 6: Calculated metamorphic conditions

<table>
<thead>
<tr>
<th>Sample</th>
<th>Location</th>
<th>MacPTAX</th>
<th>TWG 2-02</th>
</tr>
</thead>
<tbody>
<tr>
<td>80496-06</td>
<td>RF</td>
<td>600–650</td>
<td>700–750</td>
</tr>
<tr>
<td>80496-11</td>
<td>RF</td>
<td>600–650</td>
<td>675–725</td>
</tr>
<tr>
<td>80496-13</td>
<td>RF</td>
<td>550–625</td>
<td>650–700</td>
</tr>
<tr>
<td>80596-03</td>
<td>RF</td>
<td>600–650</td>
<td>650–725</td>
</tr>
<tr>
<td>80596-05</td>
<td>RF</td>
<td>600–650</td>
<td>650–700</td>
</tr>
<tr>
<td>62097-05</td>
<td>RF</td>
<td>650–750</td>
<td>650–800</td>
</tr>
<tr>
<td>62097-03</td>
<td>RF</td>
<td>575–625</td>
<td>625–800</td>
</tr>
<tr>
<td>IV96-01</td>
<td>VD</td>
<td>600–700</td>
<td>675–700</td>
</tr>
<tr>
<td>70197-10</td>
<td>VD</td>
<td>600–775</td>
<td>700–825</td>
</tr>
<tr>
<td>62497-03</td>
<td>VD</td>
<td>565–625</td>
<td>650–750</td>
</tr>
</tbody>
</table>

Location abbreviations as in Table 1.

Tectonic implications

Figure 7 depicts a schematic illustration of our new model for metamorphism and magmatism at the Ivrea zone. The presence of relic kyanite and staurolite may suggest that an early phase of crustal thickening accompanied prograde metamorphism (Fig. 7a). Although not the only interpretation of the early prograde metamorphic evolution of the Ivrea zone (see Zingg, 1983; Pin, 1990; Handy & Zingg, 1991), such a prograde contractional event is consistent with evidence of an early phase of high-pressure metamorphism in the Strona Ceneri zone (Boriani & Villa, 1997).

The ubiquitous presence of metapelite with sillimanite in the matrix and as inclusions within garnet indicates that, subsequent to the early contractional phase of metamorphism, much of the prograde PT path occurred within the stability field of sillimanite. The limits of both muscovite and staurolite stability were probably exceeded within the stability field of sillimanite, further indicating that moderate pressures prevailed along much of the prograde path (<7 kbar). At the regional thermal maximum, PT conditions for rocks in lower Val Strona di Omegna were 40–60°C/km and >650–700°C (Henk et al., 1997), implying a quasi-steady-state thermal gradient of ~40–60°C/km.

These temperatures are elevated with respect to an average crustal geotherm and may require elevated heat flow through a combination of magmatic accretion, aqueous fluid flow, and/or lithospheric extension (De Yoreo et al., 1991). Stable isotopes, however, indicate limited fluid infiltration (Baker, 1988) and the metamorphism we associate with the emplacement of the upper Mafic Complex overprints the regional metamorphic zonation. We suggest that crustal attenuation through tectonic extension (Fig. 7b and c) could explain the moderate pressures along the prograde PT path and the elevated geotherm at the regional thermal maximum. Passive upwelling of the asthenosphere accompanied the extension leading to a steepening of the geotherm, thereby enhancing the lower-crustal heat budget (Furlong & Londe, 1986; De Yoreo et al., 1991) and imposing granulite facies conditions at relatively shallow crustal levels. Lachenbruch & Sass (1978) inferred a geotherm of >40°C/km for the Battle Mountain High in the Basin and Range Province, a thermal gradient sufficiently steep to explain the PT conditions at the regional thermal maximum at the Ivrea zone (Fig. 5).

Decompression subsequent to the regional thermal maximum is documented by the appearance of a lower-pressure assemblage (cordierite, hercynitic spinel) within depleted metapelite proximal to the upper Mafic Complex. The absence of these phases and the presence of retrograde muscovite require that cooling accompanied...
Fig. 6. (a) Partial petrogenetic grid for intermediate $X_{Mg}$ metapelites. Single lines represent divariant fields for clarity. Reaction boundaries adapted after Jones & Brown (1990) and references therein. Reaction $\text{Chl} + \text{Ms(ss)} \Rightarrow \text{St} + \text{Bt} + V$ from Xu et al. (1994). (b) Proposed $PTt$ path. The two paths represent the $PTt$ paths of rocks proximal (path A) and distal (path B) to the upper Mafic Complex. (I), $PT$ conditions and age constraints at the regional thermal maximum; (II), thermal maximum for rocks proximal to the upper Mafic Complex discussed in text; (III), cooling age below $\sim 300°C$ based on K–Ar biotite ages (Bürgi & Klotzli, 1990). Alternative interpretations of the prograde metamorphic evolution reflect the range of uncertainty in the dashed portion of the prograde $PTt$ path.

decompression in rocks distal to the upper Mafic Complex. Emplacement of the upper Mafic Complex occurred during retrograde decompression from the regional thermal maximum (Fig. 7c) and provided the thermal energy to reset proximal mineral assemblages. Mineral assemblages in distal rocks were incompletely reset and preserve evidence of higher pressures at the regional thermal maximum (Fig. 5).
We attribute the decompression from the regional baric gradient at the regional thermal maximum to crustal thinning by continued tectonic extension. This interpretation is consistent with other observations in the Ivrea zone. Henk et al. (1997) inferred a baric gradient of \( \approx 0.41 \) kbar/km in Val Strona di Omegna. Although uncertainty in baric gradients calculated from mineral equilibria is large, this estimate exceeds the lithostatic gradient of \( \approx 0.3 \) kbar/km (Burke & Fountain, 1990), suggesting crustal attenuation by a factor of \( \approx 1.4 \) subsequent to the closure temperature of the net transfer reactions from which pressure was estimated. Regional extension is consistent with Brodie & Rutter’s (1987) inference that 2 km of E-W-directed attenuation within the lowermost 5 km of the Ivrea zone occurred along mylonitic shear zones. Oriented symplectic intergrowths of orthopyroxene, plagioclase, and spinel in metapelites from upper Val Strona di Omegna support an episode of regional, near-isothermal decompression (Brodie, 1995). Handy (1987) estimated 8–10 km of E-W-directed extension along normal faults near the boundary between the Ivrea and Strona–Ceneri zones. Quick et al. (1994) interpreted the arcuate structure of relict magmatic foliation in the upper Mafic Complex as implying emplacement into a zone of active extension, an inference consistent with age estimates for the onset of tectonic extension along the regional retrograde \( PT_t \) path (Brodie et al., 1989). Finally, Snoke et al. (1999) reported evidence supporting progressive, non-coaxial, ductile deformation within the carapace of supracrustal rocks overlying the upper Mafic Complex, an observation consistent with ductile attenuation through extension during emplacement. Although compositional variability and the thin (\( \approx 3 \) km) carapace of Kinzigite Formation overlying the upper Mafic Complex south of Val Sesia preclude an equilibria estimate, this estimate exceeds the lithostatic gradient of \( \approx 0.3 \) kbar/km (Burke & Fountain, 1990), direct calculation of the baric gradient in Val Strona di Postua, significant crustal attenuation can be accommodated by our results (Fig. 2).

We conclude that extension characterized the Ivrea zone from the Late Carboniferous to the Early Permian. The peak thermal episode of regional granulite facies metamorphism resulted from elevated heat flow during extension and was perhaps enhanced by magmatic accretion below the current level of exposure (Henk et al., 1997). Remnants of the prograde and near-peak magmatism that accompanied extension and regional granulite facies metamorphism may be preserved in parts of the lower Mafic Complex (Vavra et al., 1999). Emplacement of the upper Mafic Complex accompanied decompression from ambient pressures at the regional peak of metamorphism during continued extension. The heat released during crystallization of the upper Mafic Complex provided the thermal energy for contact metamorphism up to the granulite facies in proximal crustal
rocks. The occurrence of cordierite, hercynitic spinel, and coarse-grained garnet augen preserves evidence of low-$P$, high-$T$ contact metamorphism.

Crustal attenuation after the regional thermal maximum led to decompression melting of metapelitic overlying the upper Mafic Complex and its depletion by up to $\sim 30\%$ granite component (Barboza, 1998). The depleted accumulations of cumulate and peritectic mineral phases that document this melting event are probably represented by the abundant leucotonalitic leucosomes intercalated with metapelite within the upper Mafic Complex contact aureole (Barboza, 1998; Barboza et al., 1999; Snoke et al., 1999). Foliation-parallel melt movement and melt segregation to higher crustal levels may have been directed by penetrative deformation within the carapace of supracrustal rocks overlying the upper Mafic Complex (Snoke et al., 1999). Weakly deformed granitic intrusive bodies in the eastern Ivrea zone may represent the mobilized melt extracted from the contact aureole (Snoke et al., 1999). The cessation of extension and high-grade conditions in rocks proximal to the upper Mafic Complex may have coincided with its final emplacement (Demarchi et al., 1998) and was followed by near-isobaric cooling (Fig. 7d) until Tertiary orogenic activity led to emplacement of the Ivrea zone within the upper crust.

Magmafic accretion models

Our field, petrographic, and geochemical evidence implies that anatexis and metamorphism caused by emplacement of the upper Mafic Complex was confined to a $\sim 3$ km aureole in supracrustal rocks that overlie the intrusion. Models that rely upon rapid cooling and convective heat transfer during magmatic accretion (Wells, 1980; Campbell & Turner, 1987) overpredict the extent of crustal heating and anatexis we observe in the Ivrea zone. Huppert & Sparks (1988) calculated that emplacement of basaltic magma at $1200^\circ$C into crust at 500–830$^\circ$C would generate a volume of silicic magma roughly 2:0–0:6 times that of the basalt. This amount exceeds the anatexis we have documented within the upper Mafic Complex contact aureole. Even simulations based on a conductive model for heat transfer (Barboza & Bergantz, 1996) overestimate the extent of the contact aureole and degree of partial melting in $<0:1$ My after the initiation of magmatism.

The discrepancy between the model results and the field observations may be reduced if the thickness of the contact aureole was reduced by incorporation of residual rocks within the underlying upper Mafic Complex (Voshage et al., 1990; Sinigoi et al., 1996) and attenuation through non-coaxial ductile deformation along the roof of the intrusion (Snoke et al., 1999). According to this view, the contact aureole surrounding the upper Mafic Complex represents the thermal equilibration of crustal rocks with heat released during the final stages of emplacement of the upper Mafic Complex. Large-scale regional metamorphism resulting from emplacement, however, is not apparent, demonstrating that the intrusion of large volumes of mafic magma within the lower continental crust may not inexorably imply regional-scale metamorphism and anatexis of crustal rocks. Reference to studies that predict regional granulite facies metamorphism and extensive crustal anatexis are a necessary consequence of magmatic accretion (Wells, 1980; Campbell & Turner, 1987; Wickham, 1987; Huppert & Sparks, 1988) should be tempered by an awareness that such effects are not apparent in the Ivrea zone, often regarded as an important example of such a process.

ACKNOWLEDGEMENTS

Dr. J. Quick, Professor S. Sinigoi, Professor M. Brown, Dr. L. Burlini, and Professor A. Boriani are thanked for fruitful discussions and logistical support. Informal reviews by Professor B. W. Evans and Dr. Quick, and formal reviews by Professors A. W. Snoke, B. R. Frost, R. Schmid, and B. Hacker substantially improved the manuscript. The detailed editorial assistance of Dr. Sorena Sorensen is also greatly appreciated. Partial funding for this work was provided by National Science Foundation grant EAR-9508291, and by a Royalty Research Fund grant from the University of Washington.

REFERENCES

Berman, R. G. (1988). Internally-consistent thermodynamical data for minerals in the system $\text{Na}_2\text{O}–\text{K}_2\text{O}–\text{CaO}–\text{MgO}–\text{FeO}–\text{Fe}_2\text{O}_3–\text{Al}_2\text{O}_3–\text{SiO}_2–\text{TiO}_2–\text{H}_2\text{O}–\text{CO}_2$. *Journal of Petrology* 29, 445–522.


