Geologic Map of the Southern Ivrea-Verbano Zone, Northwestern Italy

By James E. Quick,[1] Silvano Sinigoi,[2] Arthur W. Snoke,[3] Thomas J. Kalakay,[3] Adriano Mayer,[2] and Gabriella Peressini[2,4]

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[1]U.S. Geological Survey, Reston, VA 20192–0002.
[2]Università di Trieste, via Weiss 8, 34127 Trieste, Italia.
[3]University of Wyoming, Laramie, WY 82071–3006.
[4]Max-Planck-Institut für Chemie, J.J. Becherweg 27, 55128 Mainz, Germany.

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COVER: View of the Ponte della Gula, an ancient bridge spanning the Torrente Mastellone approximately 2 kilometers north of the village of Varallo. Diorite of Valsesia crops out beneath the bridge. Photograph by ADstudia, Silvano Ferraris, photographer, Pizza Calderini, 3–13019 Varallo Sesia (adphoto@tiscali.it)

INTRODUCTION

The intrusion of mantle-derived magma into the deep continental crust, a process commonly referred to as magmatic underplating, is thought to be important in shaping crustal composition and structure. However, most evidence for this process is indirect. High P-wave velocities and seismic-reflection profiles reveal that much of the deep continental crust is dense and strongly layered, consistent with the presence of layered mafic and ultramafic cumulates (Meissner and others, 1983; Brewer and others, 1983; Mathur, 1983; Allmendinger and others, 1987; Behrendt and others, 1990). These cumulates are thought to have crystallized from mantle-derived magmas that ponded near the base of the crust as a result of density and viscosity contrasts (Stolper and Walker, 1980; Herzberg and others, 1983; Meissner and others, 1983; Glazner and Ussler, 1989). The presence of mafic and ultramafic cumulates in the deep crust is supported by xenolith studies (for example, Griffin and O'Reilly, 1987). Furthermore, the geochemistry of volcanic rocks, mass-balance considerations, and high heat flow suggest that magmatic underplating is presently occurring beneath extending continental crust such as the Basin and Range province and the Salton trough in North America (Lachenbruch and Sass, 1985; Gans, 1987; Lachenbruch and others, 1985), as well as beneath some magmatic arcs (Hamilton, 1981; Hildreth, 1981). One of the few places in the world where magmatic underplating may be studied directly is the Ivrea-Verbano Zone, southern Alps, northwestern Italy (Mehnert, 1975). Within this zone, a 7- to 8-kilometer (km)-thick gabbro-norite-diorite composite intrusion crystallized beneath 15 to 20 km of continental crust during Permian time (Rivalenti and others, 1975, 1981, 1984; Voshage and others, 1990). Exhumed during Alpine uplift, present-day exposures reveal an impressive record of crustal anatexis, crustal assimilation, and synmagmatic deformation that accompanied the deep-crustal intrusive events.

The southern Ivrea-Verbano Zone was selected for geologic mapping because the magmatically underplated rocks are most voluminous in this area. Geologic mapping of parts of the same area was conducted previously by Artini and Melzi (1900) and Bertolani and Garuti (1970). The geologic map of the southern Ivrea-Verbano Zone presents the results of collaborative fieldwork conducted between 1991 and 2001 by the U.S. Geological Survey (USGS), the Università di Trieste, and the University of Wyoming, and was funded by the Deep Continental Studies Program of the USGS, the National Science Foundation, the Consiglio Nazionale delle Ricerche, the Ministero dell'Università e della Ricerca Scientifica e Tecnologica, and the Max-Planck-Institut für Chemie. Products of this collaboration include publications on the role of synmagmatic deformation in magmatic underplating (Quick and others, 1992, 1994); the crustal response to magmatic underplating (Snoke and others, 1999); the geochemical evolution of underplated magmas (Sinigoi and others, 1994); the role of density contrasts in magmatic underplating (Sinigoi and others, 1996); the validity of the Ivrea-Verbano Zone as a crust-mantle transition (Quick and others, 1995); and the geobarometry of the Ivrea-Verbano Zone (Demarchi and others, 1998). Also, as part of this cooperative project, an M.S. thesis on structural aspects of the Ivrea-Verbano Zone was completed at the University of Wyoming (Kalakay, 1996).

REGIONAL SETTING

The Ivrea-Verbano Zone (fig. 1) is a tectonically bounded sliver of plutonic and high-temperature, high-pressure metamorphic rocks in the southern Alps of northwestern Italy (Mehnert, 1975; Fountain, 1976). To the northwest, it is faulted against the basement of the Austro-Alpine Domain by the Insubric Line, a major suture zone that separates the European and Apulian plates (Schmid and others, 1987; Nicolas and others, 1990). To the east, it is juxtaposed against amphibolite-facies paragneiss and orthogneiss of the Strona-Ceneri Zone by two major faults identified as the Cossato-Mergozzo-Brissago (CMB) Line and the Pogallo Line (Boriani and Sacchi, 1973; Zingg, 1983; Handy, 1987; Boriani and others, 1990). Although projected into the Val Sesia area of the southern Ivrea-Verbano Zone, the CMB Line has been obscured by younger intrusions, poor exposure, and Tertiary faulting. To the southeast, the Ivrea-Verbano Zone is covered by sedimentary deposits of the Po Plain.

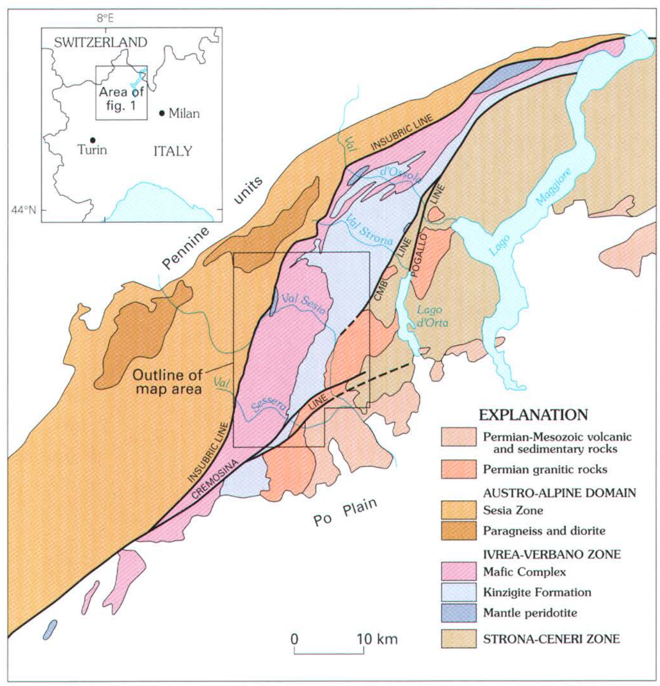


Figure 1.—Geology of the southern Alps in the vicinity of the Ivrea-Verbano Zone based on Zingg (1983) and showing tectonic boundaries, terranes, and geographic features referenced in the text. Cossato-Mergozzo-Brissago Line abbreviated as CMB Line. Ivrea-Verbano Zone contains pre-Permian peridotite, pre-Ordovician(?) paragneiss of the Kinzigite Formation, and Carboniferous(?) to Permian intrusive rocks of the Mafic Complex. Strona-Ceneri Zone contains pre-Ordovician gneiss and schist. Austro-Alpine Domain contains high-pressure, low-temperature metamorphic rocks of the Sesia Zone and thrust sheets of diorite and paragneiss. Pennine units include Pennine basement, ophiolites, and schistes lustrés.

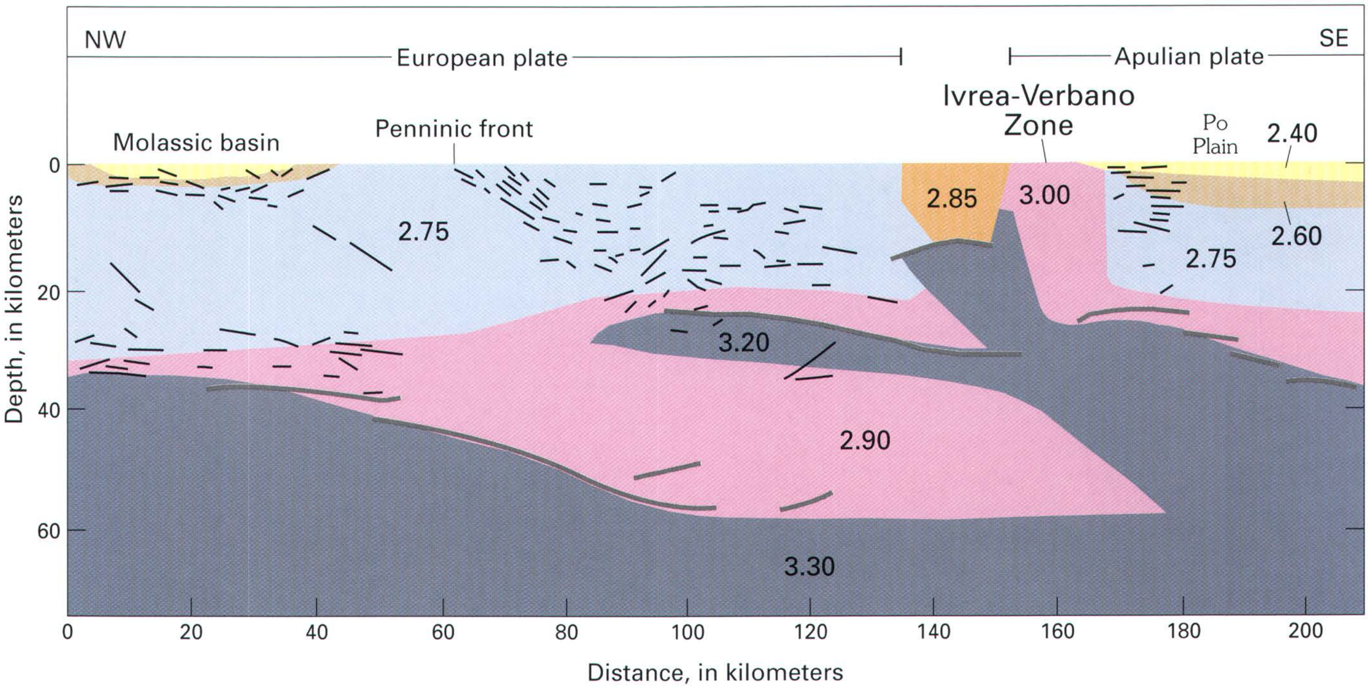


Figure 2.—Geophysical model of the ECORS-CROP traverse modified from Nicolas and others (1990) interpreting seismic and gravity data in terms of crustal-scale imbrication of the European and Apulian plates. Colored fields show rocks of uniform density (g/cm^3) in the gravity model of the traverse. Thin black lines are depth-migrated lines, and thick gray lines are wide-angle P wave reflectors.

The geophysical signature of the Ivrea-Verbano Zone (fig. 2) dips steeply to the southeast near the surface but flattens into a subhorizontal orientation at a depth of 20 to 30 km beneath the Po Plain (Berckhemer, 1968; Nicolas and others, 1990). Emplacement of the Ivrea-Verbano Zone rocks into the upper crust resulted from uplift during Mesozoic crustal thinning and subsequent lithospheric wedging related to Alpine collision (Schmid and others, 1987; Nicolas and others, 1990).

The Strona-Ceneri and Ivrea-Verbano Zones have been collectively interpreted as a more-or-less complete section through the pre-Alpine continental crust that exposes progressively deeper crustal levels toward the northwest (Mehnert, 1975; Fountain, 1976; Hamilton, 1989). According to this hypothesis, middle to upper crustal levels are exposed in the Strona-Ceneri Zone and lower crustal levels are exposed in the Ivrea-Verbano Zone. Within the latter, metamorphic grade and equilibration pressures increase from amphibolite grade and about 2 kilobars (kbar) near the CMB and Pogallo Lines to granulite grade and more than 8 kbar near the Insubic Line (Bertolani, 1968; Zingg, 1983; Henk and others, 1997). In the southern Ivrea-Verbano Zone, foliation and compositional layering are subvertical and isobars for equilibration pressures are approximately parallel to the Insubric Line (Demarchi and others, 1998). Thus, the geologic map of the southern Ivrea-Verbano Zone may be viewed as a vertical cross section through the pre-Alpine deep crust that has been tilted about 90° about a north-northeast axis.

GEOGRAPHY AND PHYSIOGRAPHY

The southern Ivrea-Verbano Zone is located in the foothills of the Italian Alps approximately 100 km northwest of Milan. The area is easily reached by paved roads that follow the Val Sesia, and which connect to the south with Autostrada A26 at the village of Romagnano Sesia. Villages of tens to tens of thousands of inhabitants constitute the principal population centers. Of these, only Trivero (fig. 3, on map sheet) near the southern margin of the study area is located significantly above the floor of a major valley. Isolated stone huts, which are abundant in the mountains, are mostly in decay as a consequence of decades of abandonment.

Rugged topography and poor accessibility characterize the mapped area. Steep mountains rise precipitously 1,500 meters (m) above the valley floors. Where the slopes are not cliffs, they are covered by thick deciduous vegetation (fig. 4). Paved roads connect the population centers and one ski area near the southern edge of the mapped area. A few dirt or gravel roads provide access to hydroelectric facilities, inactive mines, and some small villages and inhabited mountain huts. Most of the area mapped as the Mafic Complex is accessible only by foot; unfortunately, an extensive trail network that once connected the mountain huts in this area has been lost due to lack of maintenance and overgrowth by vegetation.

Geographic terms used in this text and on the map include words in proper Italian and in regional dialect. They translate as follows. "Oro" is regional dialect for place. "Fiume" may be read as "river," "torrente" is a significant stream, and "rio" is a small stream. "Strona" and "comba" are regional dialect for stream.

Figure 4.—View to south of the Denti di Gavala from the Balmuccia peridotite. Peaks reach 1,650 m altitude (see fig. 3, on map sheet, or geologic map for location).

"Val" translates as valley, and "cima" translates as crest. "Corno" translates as horn, and "rocca" as castle or rocky promontory. "Bec" is regional dialect for the beak of a bird. "Punta" translates as point or peak, "punte" as points or peaks, and "denti" as teeth. "Bocchetta" and "sella" identify a saddle on a ridge. "Castello" translates as castle. Following the usage of the Servizio Cartigrafico della Regione Piemonte for names of mountains, "M." is an abbreviation for "monte."

STRATIGRAPHY

The southern Ivrea-Verbano Zone, which consititutes the core of the mapped area, is traditionally subdivided into the Kinzigite Formation (Bertolani, 1954) and the Basic Formation (Bertolani, 1959). The Kinzigite Formation consists chiefly of amphibolite-facies to granulite-facies pelitic paragneiss but also includes subordinate amphibolite, impure calcite marble, and calc-silicate paragneiss. The Basic Formation is subdivided into lenses of mantle peridotite (Shervais, 1979; Rivalenti and others, 1981, 1984) and the voluminous Mafic Complex (Rivalenti and others, 1975). The Mafic Complex is a composite intrusion of gabbroic to noritic rocks with subordinate volumes of dioritic, tonalitic, charnockitic, and cumulus ultramafic rocks. In this report, we abandon the grouping of mantle peridotite with the Mafic Complex because field evidence indicates that they are genetically unrelated and were emplaced in the crust at different times.

Other important units in the mapped area are mylonitic rocks of the Insubric Line, Permian granitic and rhyolitic rocks, Ordovician(?) orthogneiss, and rocks of the Strona-Ceneri Zone. Schmid (1968) and Zingg (1983) define the Strona-Ceneri Zone to encompass all pre-Permian paragneiss, schist, amphibolite, and orthogneiss exposed east of the Ivrea-Verbano Zone and west of the Permian-Mesozoic volcanic and sedimentary cover rocks (fig. 1). This is the most widely accepted use for the Strona-Ceneri Zone, and it is adopted in this report with the exception of the orthogneiss component, which also appears to be present in the Ivrea-Verbano Zone. It should be noted, however, that Boriani and others (1977, 1995) group the same basement rocks and associated Permian intrusions under the name, Serie dei Laghi, which they subdivide into gneisses of the "Strona-Ceneri zone" and the "Scisti dei Laghi."

The geologic histories of Ivrea-Verbano and Strona-Ceneri Zones are difficult to place within the context of the geologic time scale with precision. It is impossible to assign ages to metasedimentary rocks of the Ivrea-Verbano and Strona-Ceneri Zones at the scale of a geologic period because all fossils in the metasedimentary rocks were obliterated by metamorphism and deformation and, consequently, age assignments are based solely on the isotopic geochronology of younger igneous and meta-igneous rocks. As will be discussed, the situation is further obscured in the Ivrea-Verbano Zone where credible age determinations for the Mafic Complex are close to the Carboniferous-Permian boundary. Thus, it is not possible to assign unequivocally the entire Mafic Complex to the Permian Period as some parts of the Mafic Complex may have been emplaced in Carbonifeous time. In view of these uncertainties, metamorphic and igneous rocks of the Ivrea-Verbano and Strona-Ceneri Zones are depicted on the geologic map in terms of lithologic units without definitive age assignments.

PRE-ORDOVICIAN(?) METAMORPHIC ROCKS OF (OR DERIVED FROM) THE KINZIGITE FORMATION, IVREA-VERBANO ZONE

The Kinzigite Formation consists of sedimentary and igneous rocks that have been metamorphosed at amphibolite to granulite facies (Bertolani, 1954, 1964; Bertolani and Garuti, 1970; Zingg, 1983; Sills, 1984). In the literature and in common parlance, these amphibolite-facies and granulite-facies rocks have been loosely identified as "kinzigites" and "stronalites," respectively. The most complete section and best exposure of these rocks is north of the mapped area in Val Strona (fig. 1; Bertolani, 1954, 1964, 1968; Schnetger, 1988; Henk and others, 1997). Within the southern Ivrea-Verbano Zone, the Kinzigite Formation is well exposed near the northern limit of the mapped area but is poorly exposed south of the Fiume Sesia.

The transition from amphibolite facies to granulite facies is marked by a change from lepidoblastic textures to granulitic textures, reduction in biotite abundance, increase in garnet abundance, appearance of prismatic sillimanite, and the presence of orthopyroxene in mafic meta-igneous rocks. This transition is gradual over hundreds of meters in the northern Ivrea-Verbano Zone in Val d'Ossola (fig. 1) and is marked by progressive growth of garnet and consumption of biotite with increasing metamorphic grade (Schmid and Wood, 1976). Although it is not exposed, the transition is inferred to be a ductile fault in the southern Ivrea-Verbano Zone because the transition is relatively abrupt and corresponds to the boundary between two major, northerly plunging synforms.

Amphibolite Facies

The principal amphibolite-facies unit, undifferentiated paragneiss (pg1), comprises semi-pelitic paragneiss and pelitic, biotite-rich schist with lesser amounts of white-weathering leucosome, quartzofeldspathic paragneiss, calc-silicate paragneiss, impure calcite marble, and amphibolite. The semi-pelitic paragneiss and pelitic schist contain sillimanite, garnet, and cordierite. Sillimanite is typically fibrous and difficult to recognize in hand sample. Scarce relics of kyanite testify to an early high-pressure metamorphic event that has been largely overprinted by amphibolite-grade metamorphism (Bertolani, 1959; Capedri, 1971; Boriani and Sacchi, 1973; Zingg, 1983; Demarchi and others, 1998). Subdivision of this unit is precluded in many places due to the interlayered nature of these rock types, poor exposure, and complex folding and boudinage. Where exposures permit, calc-silicate paragneiss and marble (pg2) and interlayered amphibolite and quartzofeldspathic paragneiss (pg3) are mapped as separate units.

The amphibolite-facies paragneiss (pg1) was partially melted and intensely deformed within about 100 m of the Mafic Complex (fig. 5; Quick and others, 1994). Within about 1.5 km of the Mafic Complex, tonalitic to granitic leucosomes that crystallized from anatectic melts are abundant (Snoke and others, 1999; Barboza and others, 1999).

Granulite Facies

Banded quartzofeldpathic gneiss (pg4), the principal unit in the granulite-facies paragneiss, crops out in the northern part of the map area and within the Mafic Complex where it forms thin, sheet-like inclusions termed septa. These rocks are characterized by distinctive subcentimeter- to meter-scale banding, abundant garnet, and the presence of charnockitic leucosomes (fig. 6). Within the granulite facies rocks, large bodies of mafic gneiss (mg) and marble (m) are mapped as separate units. The bulk compositions of the granulite-facies rocks are appropriate for a refractory residue from which granitic melts were removed during anatexis (Sighinolfi and Gorgoni, 1978; Schnetger, 1988).

Protoliths and Geochronology

Bulk compositions and mineral assemblages indicate that the protoliths of the Kinzigite Formation were primarily pelitic sedimentary rocks and wackes, and included impure carbonate rocks as well as mafic volcanic or intrusive rocks (Zingg, 1983; Wedepohl and others, 1989). Sills and Tarney (1984) hypothesized that the Kinzigite Formation was assembled within an accretionary prism based on the presence of amphibolites with normal-MORB (midocean ridge basalt) and enriched-MORB affinities. The presence of kyanite relics, indicating an early high-pressure, low-temperature metamorphic event, and lenses of peridotite intercalated with metapelitic rocks are consistent with this tectonic environment (Quick and others, 1995).

A pre-Ordovician age is tentatively assigned to the Kinzigite Formation based on the inference that the orthogneiss sills (ogn) cutting the Kinzigite Formation south of the village of Foresto Sesia are the same age as the Ordovician orthogneiss near the Lago Maggiore (fig. 1). However, with the exception of the most recent high-temperature metamorphic event, the age of the Kinzigite Formation is poorly constrained. This high-temperature, sillimanite-grade metamorphic event is dated by U/Pb monazite ages that range from 290 Ma to 258 Ma (Köppel, 1974; Köppel and Grünenfelder, 1979; Henk and others, 1997) and by conventional lower-intercept U/Pb zircon age data (Köppel, 1974; Köppel and Grünenfelder, 1979; Pin, 1986; Voshage and others, 1987; Teufel and others, 1989). SHRIMP (sensitive high-resolution ion microprobe) zircon U/Pb ages by Vavra and others (1996) suggest that dehydration melting in a banded paragneiss septum (pg4) in Val Sesia might have initiated as early as 296 Ma and that the rocks reached a thermal climax prior to 273 Ma. U/Pb zircon ages (Köppel, 1974; Köppel and Grünenfelder, 1979) have Proterozoic upper concordia intercepts, which are best interpreted as the ages of the sources of the sedimentary protolith. Thus, the Kinzigite Formation was derived from a protolith that contained a Proterozoic component, metamorphosed at high pressure (kyanite grade) prior to 300 Ma, and ultimately recrystallized during a Late Carboniferous to Permian high-temperature event.

PRE-ORDOVICIAN(?) METAMORPHIC ROCKS OF THE STRONA-CENERI ZONE

Outcrops of muscovite-bearing paragneiss west of the village of Borgosesia (fig. 3, on map sheet) are tentatively assigned to the Strona-Ceneri Zone. Biotite is the dominant mica in these rocks although they are distinguished from paragneiss assigned to the Kinzigite Formation by the presence of abundant muscovite. Scarce andalusite porphyroblasts are replaced by sillimanite. These rocks are divided into two units. Banded quartzo-

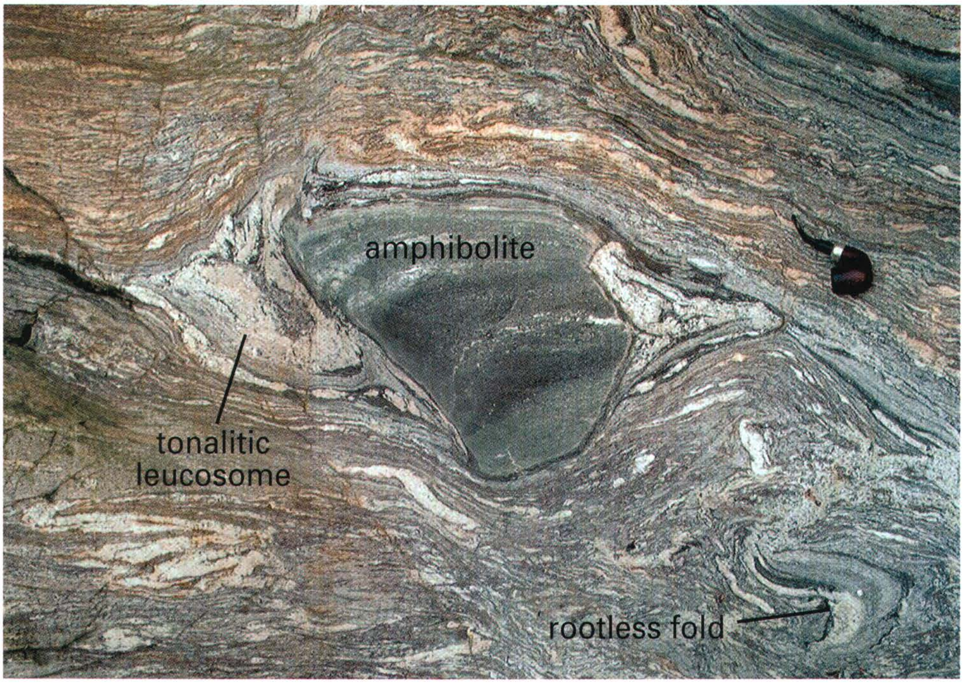


Figure 5.—Anatexis and deformation in the Kinzigite Formation (pg1) near the contact with the Mafic Complex (mc3) along the Fiume Sesia near the village of Varallo. Pelitic paragneiss contains dark-gray amphibolite boudins, white tonalitic leucosomes, and rootless folds. Note concentration of leucosomes in the pressure shadows of the amphibolite boudin. See fig. 3, on map sheet, or geologic map for location.

feldspathic paragneiss (pg5) is interpreted to be a metamorphosed, thinly to well-layered quartzose, arenitic sandstone. This unit is interfingered with interlayered schist and paragneiss (pg6) consisting of two-mica schist, banded quartzofeldspathic paragneiss, and metamafic rocks.

The stratigraphic assignment of these rocks is consistent with an apparent 3 km of right-lateral offset across the Agnona fault, which, if restored, would place them in the Strona-Ceneri Zone where similar schist and paragneiss have been described (Boriani and Sacchi, 1973; Boriani and others, 1977). U/Pb zircon and monazite ages from the Strona-Ceneri Zone indicate a significant metamorphic event at about 450 Ma (Köppel and Grünenfelder, 1971; Köppel, 1974; Ragettli and others, 1994).

IGNEOUS ROCKS OF ORDOVICIAN(?) AGE

Orthogneiss (ogn) crops out in Val Sesia near the village of Quarona and forms sheet-like bodies in the Kinzigite Formation south of the village of Foresto Sesia. The orthogneiss is granodioritic in composition, consisting of quartz, potassium feldspar, plagioclase, biotite, and minor muscovite. It is easily recognized by gneissosity with augen structures. Similar orthogneiss intrudes the paragneiss of the Strona-Ceneri Zone east of the

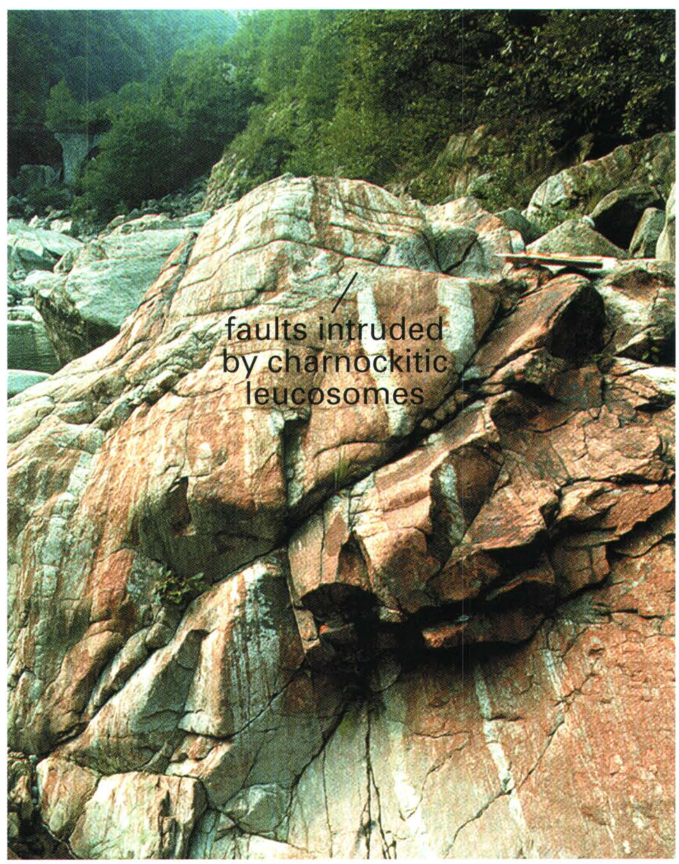


Figure 6.—View to west of septum of banded quartzofeldspathic gneiss (pg4) in Val Sessera. White- and rusty-red-weathering bands are cut by small faults, which dip approximately 45° to the left (south) and which are intruded by charnockitic leucosome. Field case on rock is approximately 30 cm wide. See fig. 3, on map sheet, or geologic map for location.

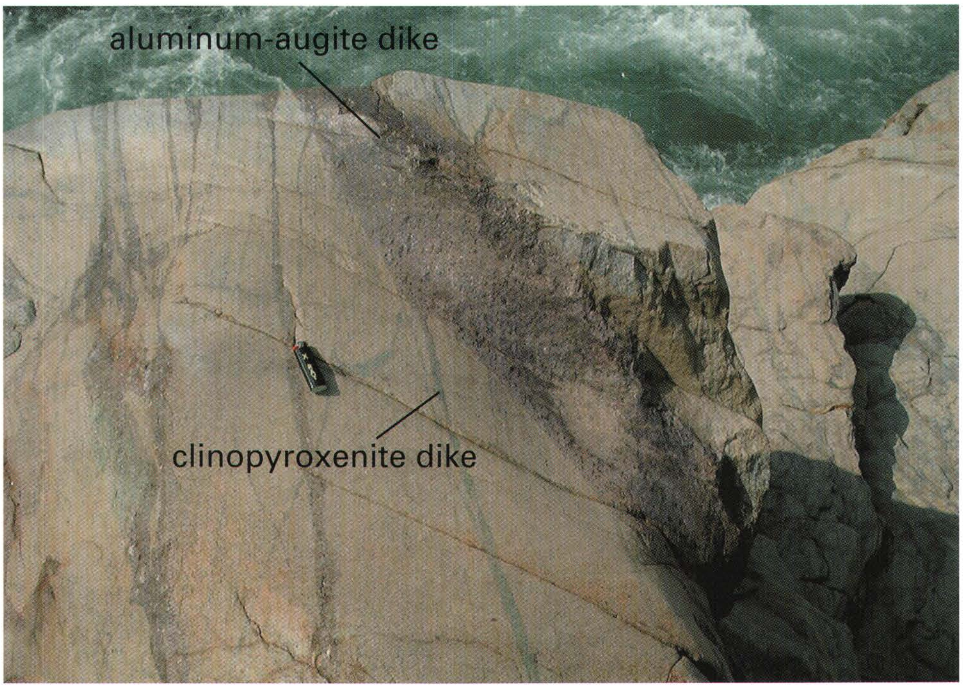


Figure 7.—Outcrop of mantle peridotite (p) along the north bank of the Fiume Sesia approximately 1.1 km east of the village of Balmuccia. Foliation in peridotite and subparallel clinopyroxenite dikes are cut by lavender-weathering, aluminium-augite dikes. See fig. 3, on map sheet, or geologic map for location.

mapped area in the vicinity of the Lago Maggiore (fig. 1) (Boriani and others, 1982–83). That occurrence has calc-alkaline geochemistry (Boriani and others, 1995), U/Pb zircon ages of about 450 Ma (Köppel and Grünenfelder, 1971; Köppel, 1974), muscovite K/Ar ages that range from 325 Ma to 311 Ma, and biotite K/Ar ages that range from 310 Ma to 234 Ma (Boriani and others, 1995). Boriani and others (1995) interpreted these data to indicate that the orthogneiss originated as subduction-related, calc-alkaline intrusions that invaded the metasedimentary rocks of the Strona-Ceneri Zone in Ordovician time and that were subsequently deformed and metamorphosed along with the host rocks during the Hercynian event.

PRE-PERMIAN MANTLE ROCKS

Pre-Permian mantle rocks in the Ivrea-Verbano Zone comprise small lenses of peridotite that are interpreted to have been derived from the mantle based on the presence of a pervasive spinel foliation, a preferred orientation of olivine consistent with high-temperature flow, and whole-rock elemental and isotopic compositions (Shervais, 1979; Rivalenti and others, 1981; Boudier and others, 1984; Voshage and others, 1990; Shervais and Mukasa, 1991). The dominant peridotite lithology in the study area is lherzolite; less abundant lithologies are harzburgite, dunite, and dikes and bands of green-weathering clinopyroxenite and lavender-weathering aluminum-augite dikes (fig. 7).

The largest lense of mantle peridotite, the famous Balmuccia peridotite, is well exposed along the banks of the Fiume Sesia approximately 1 km east of the village of Balmuccia (fig. 3, on map sheet). About 0.75 km west of the village of Meula, the Balmuccia peridotite is repeated by a northeast-striking fault. Similar rocks also crop out in Comba di Valmala approximately 2 km east of the village of Scopa.

The crustal emplacement of these rocks is constrained only to have been prior to the intrusion of the Mafic Complex. The Sm/Nd systematics and geochemistry of the peridotite suggest that it was affected by multiple magmatic events while resident in the mantle (Shervais and Mukasa, 1991).

CARBONIFEROUS(?) TO PERMIAN INTRUSIVE ROCKS OF THE MAFIC COMPLEX, IVREA-VERBANO ZONE

Rivalenti and others (1975, 1981, 1984) first described the Mafic Complex in the southern Ivrea-Verbano Zone as a layered, stratiform intrusion that was emplaced into the lower crust or at the crust-mantle boundary in a tectonically active environment. The Mafic Complex in this region is composed mainly of gabbroic and noritic rocks with subordinate ultramafic, intermediate, and silicic plutonic rocks. To facilitate description of the complex and to underscore its similarity to layered, stratiform intrusions, Rivalenti and others (1984) introduced a stratigraphy based on traverses in Val Sesia and Val Mastallone between the Balmuccia peridotite and the eastern contact of the Mafic Complex. This stratigraphy, while valid in these valleys, does not incorporate significant rock types south of Val Sesia nor does it lend itself to the description of the lithologic variation that exists along the structural grain of the Mafic Complex. Thus, it is not used in the accompanying geologic map, which presents the geology in terms of informal units representing mappable rock types and associations of rock types.

Many of the rocks in the Mafic Complex could be properly described with terminology appropriate for mafic granulites

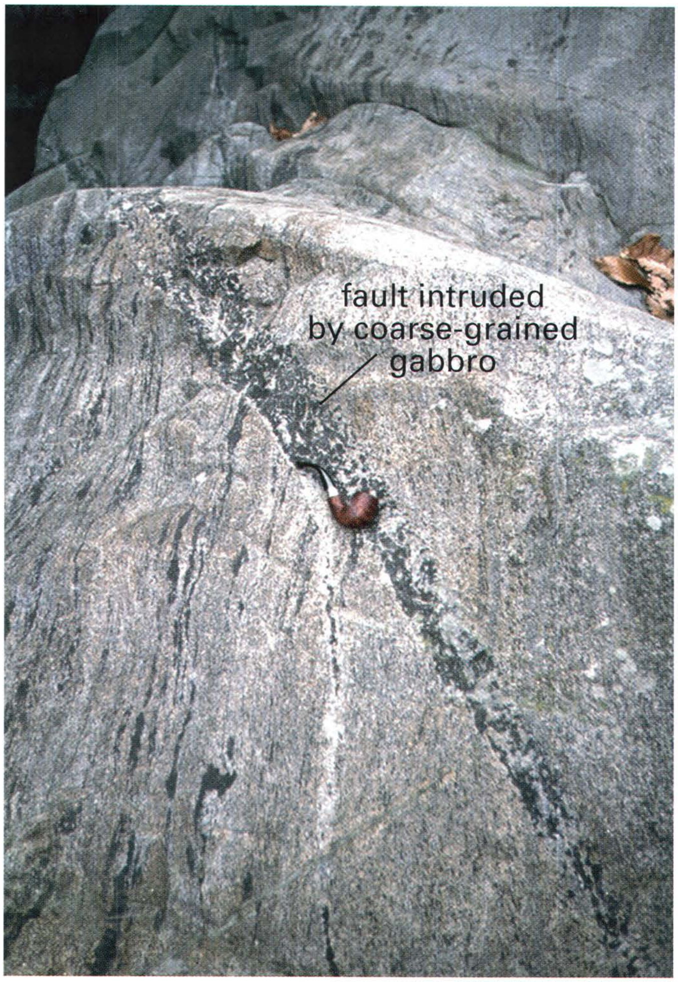


Figure 8.—Amphibole gabbro (mc1) in the channel of the Torrente Sessera showing foliation and banding cut by a stretching fault, which is intruded by coarse-grained gabbro. See fig. 3, on map sheet, or geologic map for location.

because igneous textures and structures, which are well preserved within 2 to 3 km of the eastern contact, are progressively overprinted by granoblastic textures and gneissic banding toward the west. With this caveat, we describe these rocks with igneous names to emphasize their plutonic origin and because the granoblastic textures and gneissic structure are interpreted to be the result of late-stage synmagmatic deformation and recrystallization of cumulus rocks (Quick and others, 1994).

Amphibole Gabbro

Amphibole gabbro (mc1) constitutes the westernmost exposures of the Mafic Complex from the southwestern corner of the mapped area to the Fiume Sesia. To the west, these exposures are bounded by the Insubric Line. This unit is identified by its distinctive granoblastic texture, high color index (40–50), and a mafic mineralogy dominated by black-weathering amphibole. The rock can be massive or banded at the centimeter to meter scale (fig. 8). Foliation and lineation are weakly defined by oriented trails of amphibole grains. In Valmala, the rocks are locally garnetiferous. The intrusive origin of the amphibole gabbro is demonstrated by sparse inclusions of banded quartzofeldspathic paragneiss (pg4) derived from the Kinzigite Formation. Assimilation of crustal material also is indicated by initial $^{87}Sr/^{86}Sr$ ratios in the amphibole gabbro that range from 0.7068 to 0.7086 (Sinigoi and others, 1991; Sinigoi and others, 1996).

Gabbro and Norite

Gabbro and norite (mc2) comprise a heterogeneous assemblage that is difficult to subdivide in the rugged terrain of the Ivrea-Verbano Zone. These rocks are distinguished from the amphibole gabbro (mc1) by having a significantly greater abundance of pyroxene. A foliation and weak lineation are defined by pyroxene and amphibole. Centimeter-scale to meter-scale banding (fig. 9) defined by abrupt changes in color index is most clearly developed in the western third of the unit.

In Val Sesia and Val Mastallone, the dominant rock type is a two-pyroxene, amphibole-bearing gabbro. Here, the unit comprises the "Main Gabbro and the Basal, Intermediate, and Upper Zones" of Rivalenti and others (1984). In these rocks, black amphibole locally forms distinctive large (up to 3 cm in diameter) oikocrysts. Less abundant rock types in this unit are layers of anorthosite and amphibole gabbro; scarce olivine-bearing gabbro crops out in the Val Sesia, Val Mastallone, and Val Strona di Postua. The gabbro and norite contain large garnets (0.5–1 cm in diameter) in the western half of the unit where these rocks are in contact with or along strike with paragneiss septa (pg4) or charnockitic rocks (ck). In the western Val Sesia, garnet and pyroxene-spinel coronas between plagioclase and olivine indicate recrystallization during slow cooling at high pressure. Biotite is a locally minor mineral in the eastern half of the unit where the gabbro and norite are within about 100 m of charnockitic rocks (ck), the diorite of Val Sesia (mc3), and the biotite granodiorite of Val Strona di Postua (mc4).

The gabbro and norite are in contact with the amphibole gabbro (mc1) from Val Sesia to the southern edge of the mapped area. Where the principal contact is not faulted, it is conformable, devoid of crosscutting relations, and interlayered over a few tens of meters. However, south of Cima della Mora (fig. 3, on map sheet) in the southwestern corner of the mapped area, norite dikes and sills of mc2 cut the foliation of the amphibole gabbro (mc1), demonstrating that the gabbro and norite are a younger intrusive phase of the Mafic Complex.

The gabbro and norite (mc2) intrude the Kinzigite Formation and the mantle peridotite (p). Septa of quartzofeldspathic paragneiss (pg4), which were derived from the Kinzigite Formation, are abundant within about 2 km of the western contact of the gabbro and norite (mc2). These septa are 0.1 to 100 m thick. The chemical and isotopic composition of gabbro and norite in contact with the septa indicates increased levels of assimilation of high $^{87}Sr/^{86}Sr$ material from the septa (Sinigoi and others, 1996). Gabbro cuts the foliation and internal structure of the Balmuccia peridotite.

Pegmatitic Clinopyroxenite

Pegmatitic clinopyroxenite (pc) rims and cuts the foliation of the peridotite bodies (p) (fig. 10). The clinopyroxenite is locally

Figure 9.—Banded gabbro (mc2) in small tributary to the Fiume Sesia approximately 800 m south-southwest of the village of Isola. Banding is cut by a small stretching fault, which dips steeply to the right and which is intruded by leucogabbro. See fig. 3, on map sheet, or geologic map for location.

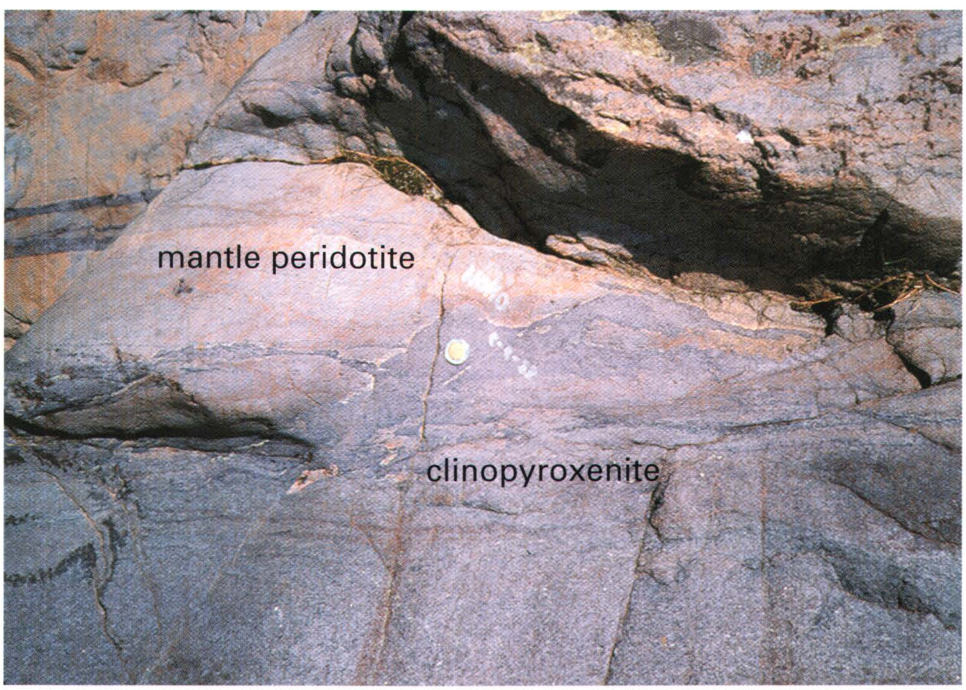


Figure 10.—Contact between dark-gray-weathering clinopyroxenite (pc) and orange-weathering mantle peridotite (p) bearing the poorly discernible epitaph in white characters of "Moho" courtesy of a previous geologic field trip. Mantle peridotite cut by small dikes of clinopyroxenite. Coin is a 500 Italian lire, which has a diameter of 2.5 cm. Outcrop is located on the south bank of the Fiume Sesia approximately 900 m southwest of the village of Isola. See fig. 3, on map sheet, or geologic map for location.

feldspathic; as plagioclase abundance increases, it grades into gabbro away from the peridotite. This rimming relation is reminiscent of reaction rims that form on mantle inclusions in calc-alkaline plutons.

Cumulus Ultramafic Rocks

Cumulus ultramafic rocks (uc) crop out in a variety of settings and comprise dunite, harzburgite, and websterite. A cumulus origin is indicated by the mineralogy of these rocks, which is dominated by olivine, orthopyroxene, and clinopyroxene, and by relics of heteroadcumulus texture. Amphibole is present in some bodies. Local concentrations of copper-rich and nickel-rich sulfides were mined prior to the end of World War II at Valmaggia (fig. 3, on map sheet) and Castello di Gavala, where surface exposures of ultramafic rocks no longer exist, and at Bec Galline, Doccio, Isola, and Sella Bassa. Bodies that are near or within the undivided paragneiss (pg1) of the Kinzigite Formation contain local concentrations of amphibole and phlogopite. Accessory minerals include iron-titanium oxides, zircon, apatite, sphene, and calcite.

Cumulus ultramafic rocks cut the Kinzigite Formation, forming large discordant intrusions near Cima di Rondo at the northern edge of the mapped area and dikes at Val Duggia northwest of Quarona. Intrusion of the Val Duggia dikes must have occurred relatively early in the evolution of the Mafic Complex because mappable xenoliths from the Val Duggia dikes are included in the easternmost Mafic Complex. Isolated bodies of cumulus ultramafic rock within the gabbro and norite (mc2) may also be fragments of an early intrusive phase that were reincorporated into the growing Mafic Complex; these are located at Bec Galline southwest of Varallo, on Monte Luvot, and in the upper Val Strona di Postua.

Cumulus ultramafic rocks form a large sheet-like body in the gabbro and norite (mc2) near Cima del Cavallo and numerous thin, sill-like bodies within the westernmost kilometer of the gabbro and norite (mc2). The sill-like bodies are typically less than 10 m thick and may be traced along strike for up to 2 km, yet are concordant to the foliation and banding in the host gabbro and norite (mc2) and to nearby septa of banded quartzofeldspathic paragneiss (pg4). One explanation for the scarcity of crosscutting relations is that, following their intrusion, these sill-like bodies were transposed into parallel with other structural elements during large-scale, pervasive deformation of the Mafic Complex (Quick and others, 1994).

Cumulus ultramafic rocks form numerous small, concordant sill-like bodies, and discordant intrusions located northwest of Denti di Gavala, south of Cima della Mora, and at Rocca d'Argimonia. The discordant intrusions at Rocca d'Argimonia and Cima della Mora are associated with norite, which suggests a consanguineous relation.

These field relations suggest that the cumulus ultramafic rocks may have been emplaced at various times during the evolution of the Mafic Complex. Garuti and others (2001) report zircon Pb/Pb evaporation ages of 288±3 Ma and 287±3 Ma for the bodies at Bec Galline and Fei di Doccio, respectively.

Charnockitic Rocks

Leucocratic granitic rocks ranging in composition from tonalite to granodiorite are mapped as charnockitic rocks (ck) based on the presence of iron-rich orthopyroxene. Charnockitic rocks are mapped at Monte Barone (fig. 3, on map sheet), Monte Luvot, Castello di Gavala, and west of Bec d'Ovaga. These bodies are transverse to the trend of the Mafic Complex but are concordant with its internal structural grain. The charnockitic rocks at Monte Luvot and Castello di Gavala are associated with diorite (mc3), biotite granodiorite (mc4), and septa of banded quartzofeldspathic paragneiss (pg4). Based on an association of charnockitic leucosomes with granulite-facies, banded quartzofeldspathic paragneiss (pg4), and on the major- and trace-element geochemistry of the charnockitic rocks in the Ivrea-Verbano Zone, Sinigoi and others (1994) concluded that the charnockitic rocks crystallized from anatectic melts derived by partial melting of the banded quartzofeldspathic paragneiss (pg4) of the Kinzigite Formation.

Diorite of Val Sesia

Biotite-rich plutonic rocks that crop out in Val Sesia and Val Mastallone in the vicinity of the village of Varallo are mapped as diorite of Val Sesia (mc3). This unit corresponds to the "Diorites" of Rivalenti and others (1984). Included are diorite, the dominant rock type, as well as quartz diorite and monzogabbro. Pyroxene and amphibole are present in some rocks, but biotite is invariably the most abundant mafic mineral. In most places, the contact between the diorite and the gabbro and norite (mc2) is gradational. Biotite decreases in abundance relative to pyroxene and amphibole toward the gabbro and norite; the contact is placed at the point where biotite constitutes less than 50 percent of the mafic minerals. Near the contact with the Kinzigite Formation, garnet is abundant (fig. 11), and quartz and potassium feldspar are locally present. Locally, an igneous foliation is weakly defined by orientation of biotite.

A cumulate origin is indicated by a hypidiomorphic-granular texture, positive europium anomalies characteristic of the accumulation of plagioclase, and barium enrichment resulting from accumulation of biotite (Sinigoi and others, 1994). Intrusion into the Kinzigite Formation is demonstrated by the presence of submeter-sized paragneiss xenoliths within about 100 m of the contact and larger septa of paragneiss (pg4) that are found up to 1 km from the contact. Also, the oxygen isotopic composition of garnet in the diorite indicates that these grains are in part composed of xenocrysts from the Kinzigite Formation (Clemens-Knott, 1992).

Magma mingling is evidenced by the presence of mafic enclaves (fig. 12). These are well exposed in Val Mastallone and consist of decimeter-sized, rounded to subrounded blebs of very fine grained, nonfoliated gabbro. Such structures are thought to form as pillow-like intrusions of mafic magma injected into cooler, more silicic melt or crystal mush (for example, Dorais and others, 1990). The isotopic, major- and trace-element composition of the mafic enclaves is consistent with crystallization from melt in equilibrium with the cumulus minerals of the gabbro and norite in Val Sesia (Sinigoi and others, 1994). West of the confluence of the Fiume Sesia and the Torrente Mastallone, enclaves appear to be flattened, indicating compaction; they are also less distinct from the host diorite, indicating that they have been modified by reaction with and assimilation by the host crystal mush.

Biotite Granodiorite of Val Strona di Postua

Biotite granodiorite (mc4) is well exposed in Val Strona di Postua and crops out near the eastern contact of the Mafic Complex south of the Fiume Sesia. These rocks also form pro-

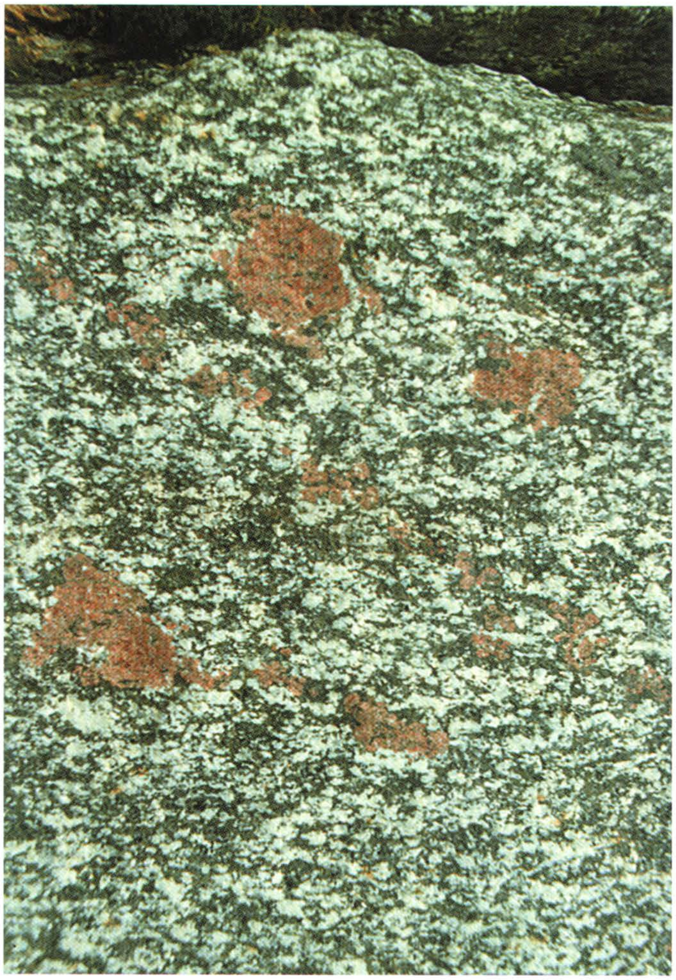


Figure 11.—Garnet-bearing diorite (mc3) of the Mafic Complex near the contact with the Kinzigite Formation (pg1) in Val Sesia. See fig. 3, on map sheet, or geologic map for location.

jections into the Mafic Complex south of Monte Luvot that are concordant with its internal structural grain.

The most striking feature of this unit is the great abundance of very fine grained, subangular to subrounded mafic inclusions (fig. 13) and less abundant inclusions of quartzofeldspathic paragneiss derived from the Kinzigite Formation. These rocks are further distinguished from the diorite of Val Sesia by a significantly lower color index, ubiquitous presence of quartz, and local porphyritic texture. The biotite granodiorite (mc4) grades into the gabbro and norite (mc2); this gradation is characterized by decreasing biotite and quartz content, increasing color index, and decreasing abundance of inclusions. The mafic inclusions are interpreted to be fragments derived from fine-grained gabbro dikes and sills that intruded the Kinzigite Formation and that were subsequently incorporated by the Mafic Complex.

Leucotonalite of Val Strona di Postua

Leucotonalite of Val Strona di Postua (t) forms a thin layer between the Mafic Complex and the Kinzigite Formation south of the Fiume Sesia. The unit comprises a heterogeneous suite of leucocratic granitic rocks that is mainly tonalite but that includes granodiorite and granite. These rocks contain abundant inclusions of paragneiss (pg1) derived from the Kinzigite Formation. They are distinguished from the inclusion-rich granodiorite of Val Strona di Postua (mc4) by having a distinctly lower color index (<3) and a greater abundance of quartz. They are distinguished from the charnockitic rocks of the Mafic Complex (ck) by the presence of biotite and muscovite, although the local presence of iron-rich orthopyroxene suggests that some of the rocks have charnockitic affinity.

The leucotonalite grades into agmatitic paragneiss in the wall rocks as the ratio of paragneiss inclusions to leucotonalite increases. The incompatible-element composition of the leucotonalite is consistent with crystallization from anatectic melts derived from the amphibolite-facies paragneiss (pg1) of the Kinzigite Formation (Sinigoi and others, 1994).

Dioritic and Gabbroic Dikes

Dioritic and gabbroic dikes (dg) intrude the Kinzigite Formation north of the village of Coggiola. Included in this unit are biotite diorite, gabbro, and ultramafic cumulus rock, all of which are inferred to have crystallized from melts related to the Mafic Complex. The dikes are up to 100 m thick. Leucosomes of the Kinzigite Formation are abundant and mutually intrusive with the mafic dikes, suggesting that heat introduced by the dikes increased melting of the Kinzigite Formation.

Age of the Mafic Complex

SHRIMP zircon U/Pb ages for the Mafic Complex were determined for four samples collected during the course of this investigation. Sample RA616 was collected from the amphibole gabbro (mc1) in Val Sessera. RA234 and J649 were collected from the gabbro and norite (mc2) south of Val Sessera and in Val Duggia, respectively. Sample VS612 was collected from the diorite of Val Sesia (mc3) southwest of the village of Varallo and close to the sample locality from which Pin (1986) obtained a conventional U/Pb zircon age of 285+7-5 Ma.

Samples were processed by conventional mineral separation techniques at the Max-Planck-Institut für Chemie in Mainz, Germany. A representative fraction of the zircon population was handpicked for each sample and mounted in epoxy. To identify internal structures and to target areas for analysis, each zircon was examined and photographed in transmitted light, carbon-coated, and photographed with a JEOL JSM 5600 scanning electron microscope in cathodoluminescence mode at Stanford University, Palo Alto, California. Carbon coats were subsequently removed and mounts were gold-coated for isotopic analysis with the USGS/Stanford SHRIMP-RG at Stanford University. Isotopic analyses were performed following the methods described by Compston and others (1984) and Williams and Claesson (1987). A primary oxygen ion beam, operating at 7 to 11 nA, produced 20- to 30-µm diameter, 1-µm deep, flat-bottomed analysis pits. Prior to each analysis, the beam was rastered for 90 seconds to remove surface contamination. For each spot, peaks of $^{90}Zr_2O$, ^{204}Pb, ^{206}Pb, ^{207}Pb, ^{208}Pb, ^{238}U, $^{232}Th^{16}O_2$, and $^{238}U^{16}O$ were scanned. After every third analysis, zircon standard R33 (age 419 Ma; Joseph Wooden, USGS, written commun., 2002) was analyzed to check reproducibility of the data and to correct elemental fractionation. Data reduc-

Figure 12.—Dark-gray mafic enclaves in light-gray diorite (mc3). Enclaves up to 30 cm in diameter. Outcrop located on the east bank of the Torrente Mastallone approximately 1 km north of the village of Varallo. See fig. 3, on map sheet, or geologic map for location.

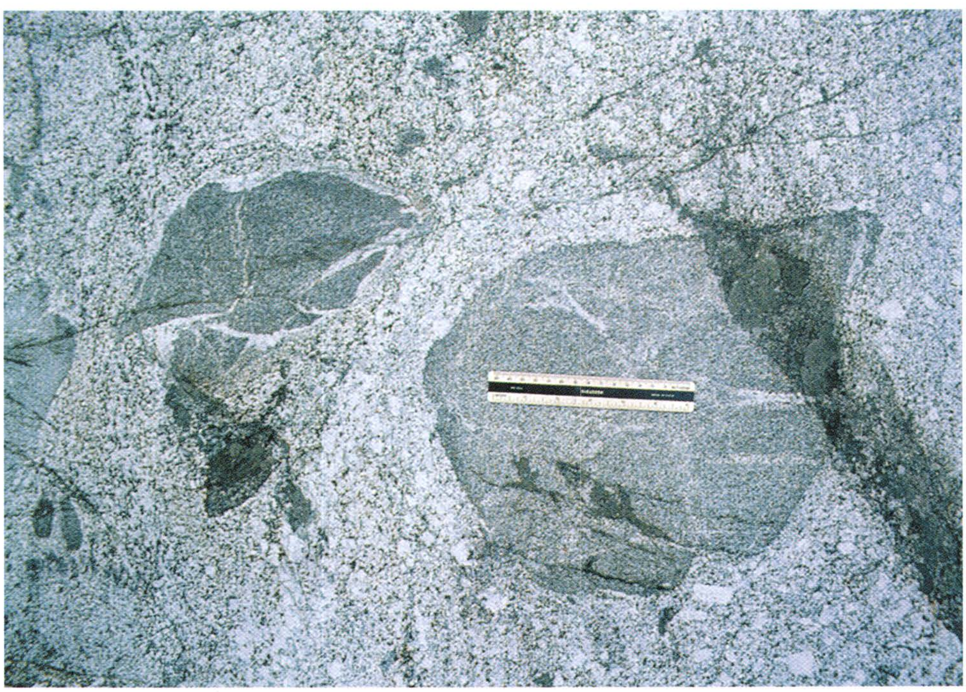


Figure 13.—Mafic inclusions in porphyritic biotite granodiorite (mc4). Scale is 15 cm long. See fig. 3, on map sheet, or geologic map for location.

tion was made with the Isoplot/Ex and Squid programs (Ludwig, 1999, 2001) using decay constants recommended by the International Union of Geological Sciences (IUGS) (Steiger and Jäger, 1977) and performing a common lead correction using the measured $^{207}Pb/^{206}Pb$ ratio. Following isotopic analysis, cathodoluminescence imaging was performed on a HITACHI S450 scanning electron microscope at the Max-Planck-Insitut für Chemie to verify positions of the analysis pits relative to the internal structures of the grain.

The physical characteristics of the zircon separates indicate fundamental differences between RA616 and the other samples. Zircons from RA616 display a variety of colors, including milky white, light yellow, and pink. Most of these zircons were relatively equant, rounded, and 100 to 200 µm in diameter. Cathodoluminescence images reveal a considerable variability in zoning with some grains showing well-defined, high-frequency zonation and others appearing devoid of zoning. In contrast, samples J649 and RA234 yielded elongate prismatic zircons terminated by pyramids; the pyramids in RA234 appeared slightly rounded. Individual zircons were up to 100 µm wide and 200 to 500 µm long. Zircons from RA234 and the most elongate zircons from J649 were pink. Zircons from sample VS612 were also elongate, well-terminated, and pink; individual zircons were 20 to 150 µm wide and 140 to 700 µm long. Cathodoluminescence images of zircons from RA234, J649, and VS612 reveal coarse zonation with a wide range in brightness between zones.

Analytical results are reported in table 1 and figure 14. In the zircons of samples J649, VS612, and RA234, measured uranium abundance ranged from 8 to 422 ppm and Th/U ratios ranged from 0.36 to 1.11. In contrast, a much wider range in uranium abundance (3 to 1,227 ppm) and Th/U ratio (0.16 to 716) was found in zircons from sample RA616. For each sample, with the exception of RA616, $^{206}Pb/^{238}U$ ages of individual analyses agree within analytical error, consistent with the presence of a single age population. Thus, the zircon ages from samples J649, VS612, and RA234 are interpreted to date a single episode of magmatic zircon growth during the crystallization of the Mafic Complex. In contrast, the individual SHRIMP zircon U/Pb ages for sample RA616 do not define a single age population within analytical error, suggesting the presence of multiple ages of zircons. This interpretation is supported by the variations in morphology, color, internal structure, and Th/U ratios exhibited by the analyzed grains.

Figure 14 presents concordia diagrams and $^{206}Pb/^{238}U$ ages for the analysed samples. Error ellipses in the concordia diagrams and error bars on the $^{206}Pb/^{238}U$ ages are shown at the 1σ level. Computed $^{206}Pb/^{238}U$ weighted average ages are reported at the 2σ level. For samples J649, VS612, and RA234, these $^{206}Pb/^{238}U$ weighted average ages are 285±4 Ma, 287±3 Ma, and 287±2, Ma, respectively. These ages agree within uncertainty with each other and with a conventional U/Pb zircon age of 285+7-5 Ma for the diorite of Val Sesia (mc3) (Pin, 1986). They also agree within uncertainty with zircon Pb/Pb evaporation ages of 288±3 Ma and 287±3 Ma for the cumulus ultramafic rocks at Bec d'Ovaga and Fei di Doccio (Garuti and others, 2001), and with a zircon Pb/Pb evaporation age of 292±4 Ma on a diorite at Bec d'Ovaga (Garuti and others, 2001). Considered collectively, these data suggest that zircon in the rocks in the eastern third of the Mafic Complex crystallized circa 285 Ma to 290 Ma.

For RA616, insufficient data are available for an accurate identification of multiple events but it is possible that this sample contains zircons ranging in age from greater than 295 Ma to 270 Ma or less (fig. 14). A similar range of SHRIMP zircon U/Pb ages was obtained by Vavra and others (1996) from a paragneiss septum (pg4) near the village of Isola. Vavra and others (1996) concluded that the thermal event attending intrusion of the Mafic Complex may have begun as early as 296 Ma and reached a peak by 273 Ma.

Additional age data for the Mafic Complex include a concordant U/Pb age of 250 Ma for a single zircon crystal from the Mafic Complex near the Balmuccia peridotite (Wright and Shervais, 1980). The meaning of this age is unclear in light of the more recent age determinations, but it is unlikely that it dates the formation of the bulk of the Mafic Complex. Also, Voshage and others (1987) report a garnet-plagioclase-whole rock Sm/Nd isochron for gabbro from Val Mastallone of 248±8 Ma, and Mayer and others (2000) report Sm/Nd mineral isochrons for rocks from Val Sesia and Val Sessera that range from 274±11 Ma to 244±4 Ma. However, these Sm/Nd ages do not date the intrusion of the Mafic Complex but instead date the closure of the Sm/Nd system during subsolidus cooling of the rocks.

IGNEOUS ROCKS OF PERMIAN AGE

Biotite granodiorite (bgd) crops out on both sides of the Fiume Sesia between Roccapietra (fig. 3, on map sheet) and Borgosesia. Inclusions of fine-grained, subangular mafic rock are locally abundant; inclusions of biotite schist and quartzofeldspathic paragneiss are less abundant. In places, the rock displays a magmatic flow foliation defined by orientation of biotite grains and by biotite-rich schlieren that drape inclusions. Inclusion-rich, flow-foliated rocks grade into inclusion-free, nonfoliated rocks. The biotite granodiorite intrudes orthogneiss (ogn) north of Quarona.

Undivided granitic rocks (gru) form numerous small intrusions in the Kinzigite Formation and a continuous but heterogeneous body that extends from Quarona to southeast of Coggiola. These consist of mainly leucotonalite and lesser amounts of leucogranodiorite and leucogranite. There is significant variability in the abundance of biotite, muscovite, and garnet. The leucotonalite is similar in mineralogy and outcrop appearance to the leucotonalite of Val Strona di Postua (t). The undivided granitic rocks are distinguished from the biotite granodiorite (bgd) by having a lower color index (<5). They are in gradational contact with the biotite granodiorite (bgd) south of Agnona. The undivided granitic rocks intrude and contain xenoliths of paragneiss from the Kinzigite Formation (pg1), schist and quartzofeldspathic paragneiss of the Strona-Ceneri Zone (pg5, pg6), and orthogneiss (ogn). West of Quarona, the undivided granitic rocks are intimately associated with the orthogneiss (ogn), a relation that might be interpreted to indicate that some of these granitic rocks crystallized from anatectic melts derived from the orthogneiss. The biotite granodiorite (bgd) and undivided granitic rocks (gru) are tentatively assigned to the Permian, post-metamorphic granitic suite of Baveno-Monte Orfano (Zingg, 1983).

Table 1.—SHRIMP U/Th/Pb data for zircon from rocks of the Mafic Complex, Ivrea-Verbano Zone, Italy.

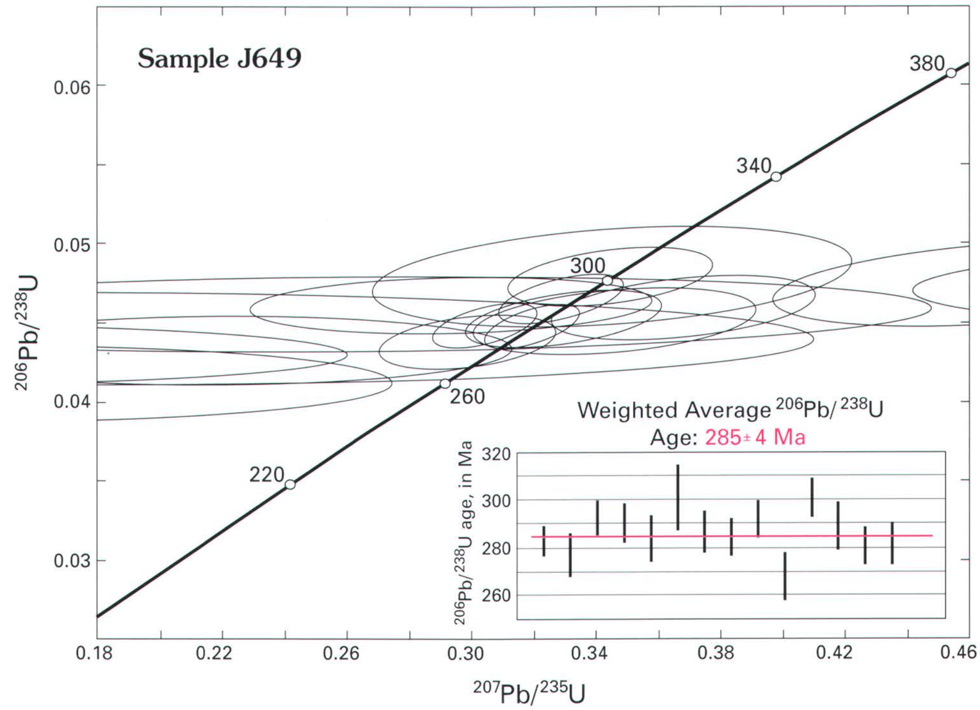
[Analyses are numbered in order of acquisition for each sample. All zircons analyzed during a 5-day period in July 2002]

Sample no.	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{206}Pb$	Percent common ^{206}Pb[a]	U, in ppm	Th/U	$^{206}Pb/^{238}U$ age, in Ma	Error, in Ma[b]	$^{238}U^{c}/^{206}Pb$	Error, in percent[b]
J649.1	—	0.050	0.00	50	0.56	277	7	22.84	2.5
J649.2	—	0.052	0.00	234	0.62	282	6	22.36	2.1
J649.3	—	0.056	0.00	21	0.51	282	9	22.22	3.1
J649.4	6,800	0.052	0.27	336	0.63	283	6	22.32	2.0
J649.5	531	0.065	3.40	22	0.42	277	9	23.15	3.2
J649.7	—	0.051	0.00	118	1.04	293	6	21.52	2.2
J649.8	545	0.069	3.31	52	0.51	290	8	21.98	3.5
J649.10	375	0.069	4.81	19	0.44	284	9	22.81	4.4
J649.11	—	0.052	0.00	8	0.47	301	13	20.89	4.4
J649.12	—	0.058	0.00	32	0.36	287	8	21.78	2.9
J649.14	—	0.053	0.00	41	0.51	285	7	22.13	2.6
J649.15	1,370	0.057	1.31	55	0.46	292	7	21.70	2.6
J649.17	364	0.067	4.96	14	0.41	268	9	24.31	3.8
J649.18	—	0.052	0.00	37	0.46	301	8	20.90	2.6
J649.19	—	0.072	0.00	10	0.38	291	11	21.11	3.8
J649.21	363	0.062	4.97	38	0.45	281	7	23.32	3.1
J649.22	—	0.070	0.00	14	0.39	284	10	21.04	4.0
VS612.1	—	0.056	0.51	49	0.37	294	8	21.28	2.9
VS612.2	—	0.059	0.85	30	0.48	280	9	22.32	3.2
VS612.3	—	0.054	0.20	104	1.11	290	8	21.66	2.7
VS612.4	464	0.061	1.08	27	0.52	289	9	22.40	4.0
VS612.5	—	0.054	0.26	29	0.62	281	9	22.41	3.1
VS612.6	—	0.062	1.22	34	0.41	293	9	21.22	3.0
VS612.7	—	0.054	0.28	56	0.64	282	8	22.31	2.8
VS612.8	364	0.085	4.12	68	0.70	275	9	23.14	3.5
VS612.9	—	0.053	0.08	74	0.39	295	8	21.36	2.8
VS612.10	396	0.052	0.00	18	0.37	279	10	21.62	4.5
VS612.11	—	0.050	0.00	24	0.45	296	10	21.39	3.2
VS612.12	841	0.057	0.61	23	0.42	286	9	22.43	3.4
VS612.13	—	0.053	0.07	32	0.44	291	9	21.63	3.1
VS612.14	—	0.048	0.00	16	0.57	292	10	21.72	3.4
VS612.15	—	0.053	0.09	70	0.70	297	8	21.20	2.7
VS612.16	5,030	0.054	0.31	112	0.80	282	7	22.36	2.7
RA616.1	—	0.125	9.20	3	0.19	279	20	22.61	1.49
RA616.2	—	0.051	0.00	673	0.25	287	4	22.00	0.33
RA616.3	—	0.054	0.74	27	0.96	297	8	21.32	0.57
RA616.4	4,310	0.052	0.00	274	1.97	268	4	23.34	0.38
RA616.5	12,100	0.057	1.91	51	0.28	273	6	23.39	0.54
RA616.6	6,460	0.052	0.25	670	0.23	284	4	22.27	0.33
RA616.7	1,500	0.054	0.00	866	1.62	267	4	23.60	0.34
RA616.8	2,590	0.059	1.78	294	1.90	260	5	24.50	0.43
RA616.9	2,180	0.055	0.38	501	1.54	268	4	23.52	0.35
RA616.10	14,100	0.057	1.30	228	1.47	274	4	23.18	0.37
RA616.11	—	0.053	0.13	1227	0.16	279	4	22.65	0.32
RA616.12	—	0.053	0.18	135	2.04	275	5	22.95	0.40
RA616.13	1,350	0.051	0.19	747	0.30	284	4	22.25	0.33
RA616.14	1,350	0.060	1.34	282	716	284	4	22.27	0.36
RA616.15	8,620	0.053	0.21	283	315	274	4	23.03	0.36
RA616.16	6,360	0.054	0.28	449	589	297	4	21.24	0.32
RA616.17	6,090	0.054	0.30	1,115	266	264	4	23.95	0.34
RA234.1	841	0.069	2.14	241	0.91	276	4	22.88	1.8
RA234.2	1,020	0.064	1.54	133	0.42	283	5	22.30	1.9
RA234.3	1,380	0.063	1.38	181	0.68	292	5	21.54	1.9
RA234.4	2,620	0.056	0.50	303	1.09	287	4	22.04	1.7
RA234.5	4,670	0.056	0.53	299	0.83	287	4	21.90	1.7
RA234.6	1,840	0.057	0.63	390	0.89	281	5	22.39	1.8
RA234.7	2,150	0.062	1.23	422	0.94	283	4	22.20	1.7
RA234.8	2,980	0.058	0.69	318	0.86	295	4	21.40	1.7
RA234.9	374	0.056	0.50	400	0.83	290	4	21.74	1.7
RA234.10	3,160	0.092	4.94	189	0.72	280	5	22.47	2.1
RA234.11	788	0.056	0.42	415	0.77	293	4	21.54	1.7
RA234.12	1,430	0.074	2.71	78	0.54	293	6	21.43	2.1
RA234.13	842	0.060	0.95	250	0.82	290	4	21.86	1.8

[a] Common Pb computed on the basis of measured $^{207}Pb/^{206}Pb$ ratio.

[b] 1σ error.

[c] ^{207}Pb corrected.



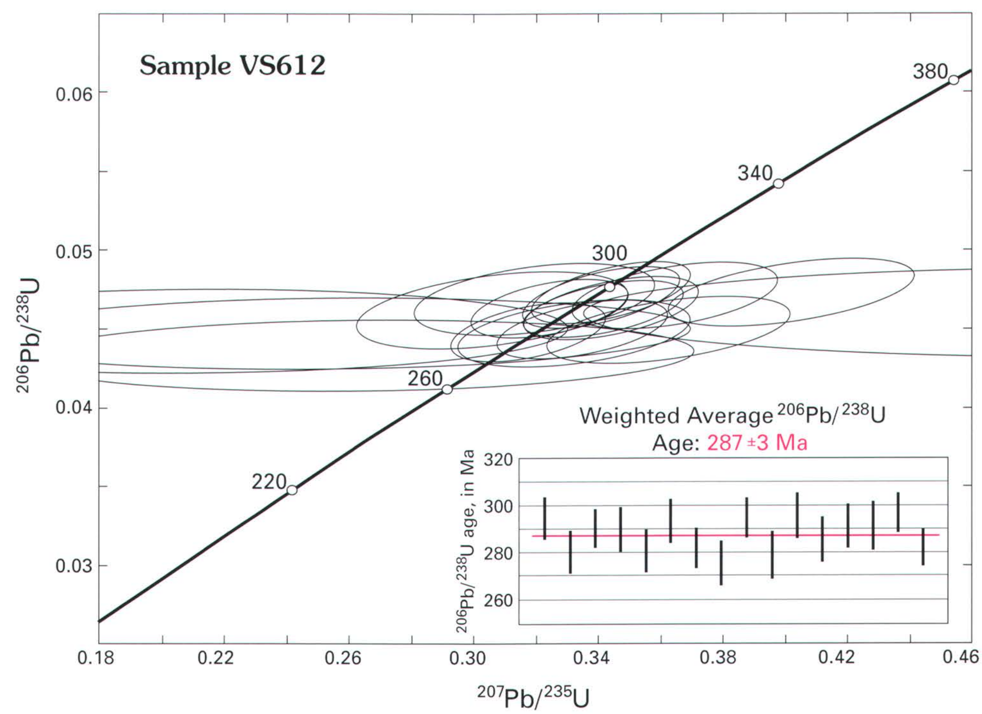
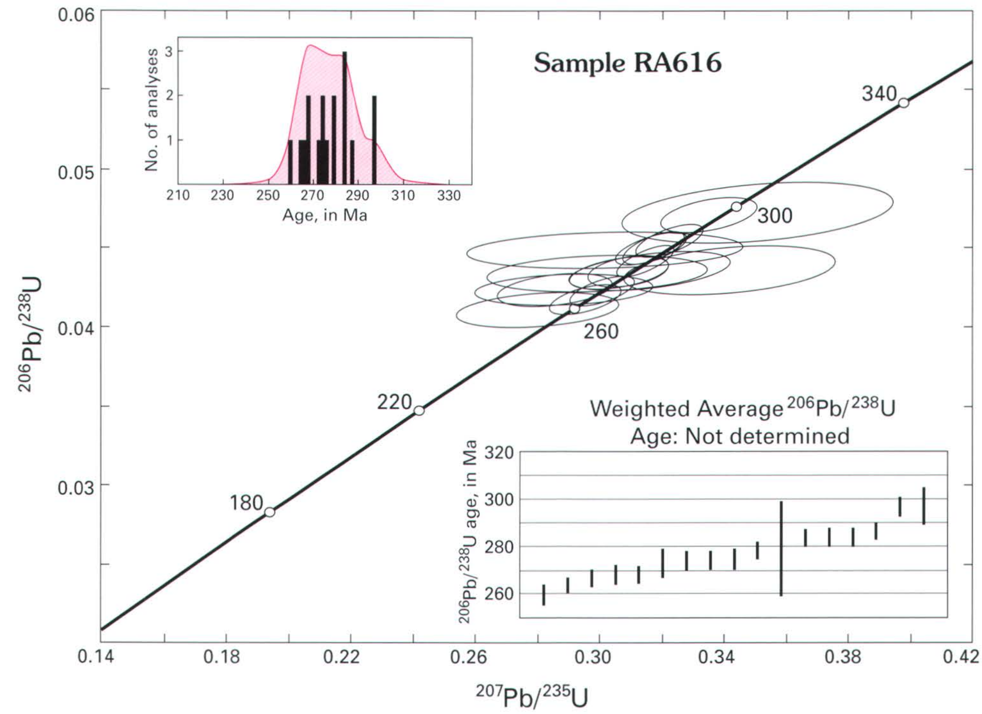


Figure 14.—U/Pb concordia and $^{206}Pb/^{238}U$ weighted average plots of SHRIMP data for the Mafic Complex of the Ivrea-Verbano Zone. Concordia plotted with 1σ error ellipses. Individual $^{206}Pb/^{238}U$ ages shown with 1σ error bars. Order of presentation of individual $^{206}Pb/^{238}U$ ages corresponds to order in table 1 for all samples with the exception of RA616, which presents individual $^{206}Pb/^{238}U$ ages in order of increasing age to emphasize absence of a coherent age group. Calculated $^{206}Pb/^{238}U$ weighted average ages reported with 2σ errors. Histogram of $^{206}Pb/^{238}U$ ages for RA616 shows distribution could suggest discrete ages at about 295, 287, and 268 Ma, although additional data are required. See text for discussion of data reduction and analysis.



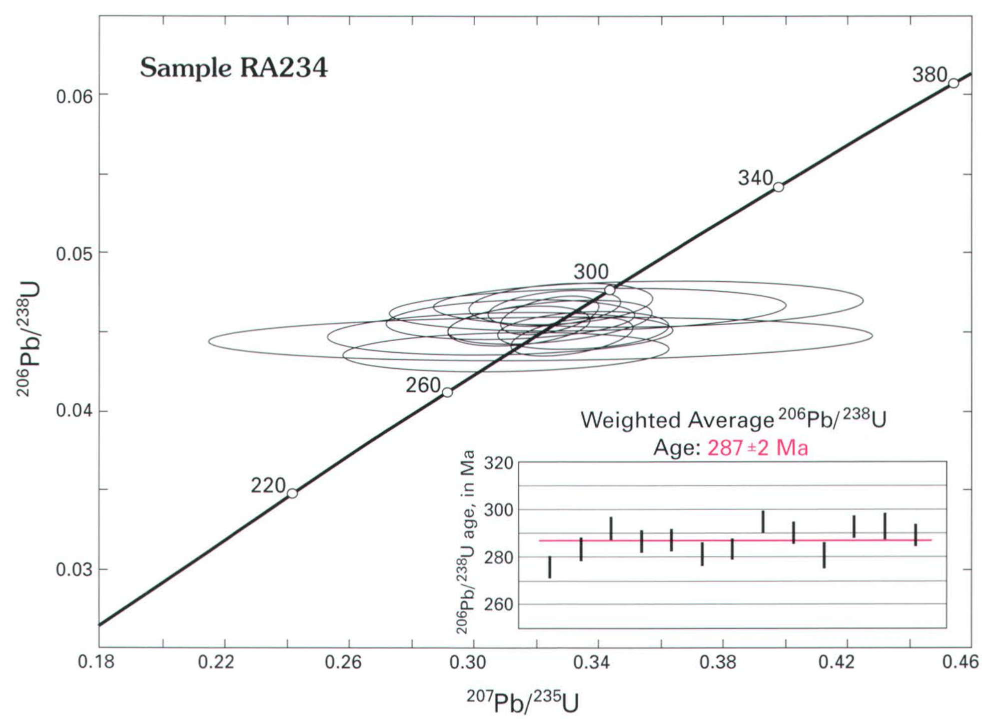


Figure 14.—Continued.

Rhyolitic rocks (r) crop out south of the Cremosina Line in the southeastern corner of the map area. The rocks include ignimbrite with abundant fiamme, volcanic breccia, and massive porphyry, which may be the hypabyssal equivalent of the volcanic rocks.

ROCKS OF THE INSUBRIC LINE

Serpentinite

Massive to foliated serpentine-rich rock containing veins and boudins of quartz and carbonate is mapped as serpentinite (s) within the belt of mylonitic rocks of the Insubric Line (il) about 1 km northwest of the village of Meula. It is unknown if serpentinization occurred during incorporation of these rocks into the Insubric Line or if it was an older event.

Mylonitic Rocks

Mylonitic rocks (il) are mapped where synkinematic recrystallization under greenschist-facies conditions has overprinted older textures within the 1-km-thick Insubric Line. In the vicinity of Val Sesia and Val Sessera, this belt includes mylonite, phyllonite, and schist formed by plastic deformation and recrystallization of rocks derived from the Ivrea-Verbano and the Sesia Zones (fig. 1), and small, discontinuous imbricates of Permian to Mesozoic metasedimentary rocks (Zingg and Hunziker, 1990).

CENOZOIC SURFICIAL DEPOSITS

Moderately consolidated deposits of conglomerate, sandstone, and shale with minor carbonate are mapped as undivided Tertiary to Quaternary(?) sedimentary deposits (QTu) in the southeastern part of the map area. These are distinguished from other surficial deposits by greater consolidation and moderate dips of up to 20°.

Poorly consolidated to unconsolidated materials ranging in grain size from boulders to clay and deposited up to 30 to 40 m above the currently active flood plains are mapped as terrace deposits (Qt). Detritus deposited on slopes by debris flows, avalanches, streams, and frost action is mapped as colluvium (Qc). Boulders, cobbles, gravel, sand, and silt underlying present flood plains are mapped as alluvium (Qa). Small, unmapped relic deposits of glacial erratics derived from sources west of the Insubric Line occur at elevations of up to 1,200 m above sea level.

STRUCTURAL GEOLOGY

STRUCTURES PREDATING THE MAFIC COMPLEX

Folding in the Kinzigite Formation and Strona-Ceneri Zone

The rocks of the Kinzigite Formation are deformed by tight to isoclinal, steeply plunging, mesoscopic to kilometer-scale folds with northeast-plunging hinge lines. Truncation of these structures by the Mafic Complex demonstrates their pre-Permian age. Zingg (1983) reports that, in the high-grade rocks of Val d'Ossola (fig. 1), two phases of folding occurred prior to granulite-facies metamorphism.

Paragneiss of the Strona-Ceneri Zone also is deformed by tight mesoscopic to kilometer-scale folds. These folds are cut by the undivided granitic rocks (gru), indicating a probable pre-Permian age for the deformation.

Massa del Turlo Fault

A major north-northeast-striking, ductile shear zone, herein named the Massa del Turlo fault, is inferred at the contact between the amphibolite-facies (pg1) and granulite-facies (pg4) rocks of the Kinzigite Formation north of Val Mastallone. Although this structure is not exposed, a fault is suggested by the juxtaposition of two, kilometer-scale, north-facing synforms and local discordance in attitudes of foliation. The Massa del Turlo fault projects toward the Forno fault in Val Strona north of the mapped area (fig. 1; Brodie and others, 1992). To the south, the diorite of Val Sesia (mc3) appears to be undeformed by the Massa del Turlo fault and lies athwart its projection, indicating that the Massa del Turlo fault must be an older structure.

STRUCTURES RELATED TO THE EMPLACEMENT OF THE MAFIC COMPLEX

Original Orientation

Thermobarometric studies (Henk and others, 1997; Demarchi and others, 1998) suggest that the present surface of exposure in the central and southern Ivrea-Verbano Zone presents a cross section through the pre-Alpine deep crust that has been tilted approximately 90° about a north-northeast axis. Thus, the pre-Alpine orientation of the southern Ivrea-Verbano Zone may be approximated by viewing the map from the northwest so that the trace of the Insubric Line appears horizontal. In this orientation, the roof of the Mafic Complex corresponds to its contact with the Kinzigite Formation between Val Sesia and Val Sessera. North of Val Sesia, the contact curves downward to deeper crustal levels. The amphibole gabbro (mc1) and mantle peridotites (p) are at the deepest levels of the complex.

Kinzigite Formation

Snoke and others (1999) describe a 1- to 1.5-km-thick zone of high-temperature shearing and increased partial melting in the Kinzigite Formation adjacent to the roof of the Mafic Complex, south of Varallo. The zone is characterized by concentrated ductile deformation, penetrative foliation subparallel to the intrusive contact, and northeast-plunging sillimanite lineation. In the Kinzigite Formation, evidence of non-coaxial strain and transposition, as shown by rootless isoclinal folds, boudinage, and agmatite, increases toward the contact with the Mafic Complex. Leucosomes are abundant as layer-parallel concentrations and also fill pressure shadows of boudins, tension gashes, and small stretching faults. Thus, it appears that partial melting attended shearing in the Kinzigite Formation. Melt migration upward from high-strain to low-strain regions is suggested by leucosome concentrations, which range from approximately 20 percent near the Mafic Complex to as much as 60 percent 1.5 km from the contact. The peak metamorphic grade was upper-amphibolite facies based on the absence of orthopyroxene in the mafic gneisses and the absence of muscovite except as a retrograde mineral. Snoke and others (1999) interpret this zone of deformation and melting to be a high-temperature stretching fault across which overlying paragneiss deformed synchronously with the emplacement and synmagmatic deformation of the Mafic Complex.

Mafic Complex

Within the Mafic Complex, layering, foliation, and mappable units define an arcuate structure that is focused on the village of Varallo (Quick and others, 1992, 1994). Paragneiss septa (pg4) and concordant charnockitic (ck), dioritic (mc3), and granodioritic (mc4) bodies are traceable for kilometers around this structure without a major break although they are increasingly attenuated with depth in the complex. Isobars in the complex parallel its roof and cut across the arcuate structure (Demarchi and others, 1998), indicating that the structure was established before equilibration pressures were locked in, and is, therefore, a pre-Alpine feature. The pre-Alpine orientation of the gross structure of the complex would have been a northeast-trending trough that was roofed and flanked on the northern side by the Kinzigite Formation.

Within the Mafic Complex, layer-parallel stretching is evidenced by boudinage, stretching faults, and mineral lineation. Banding is locally deformed by tight to isoclinal folds with axial surfaces concordant with the regional foliation and hinge lines subparallel to the stretching lineation. Locally, undeformed, poikilitic amphibole has grown across the foliation indicating interstitial melt was present when the foliation formed. Consistent with this interpretation, stretching faults are healed by undeformed veins of leucogabbro, which crystallized from segregations of late-stage interstitial melts (fig. 9) and late-stage melts segregated into undeformed patches that crosscut foliation and fill tension gashes and pressure shadows at the ends of boudins. Analogous relations involving banding and charnockitic leucosomes are found in the paragneiss septa (pg4). The scarcity of crosscutting throughout much of the complex can be explained by nearly complete transposition of intrusive contacts.

STRUCTURES POSTDATING THE MAFIC COMPLEX

Insubric Line

The Insubric Line consists of a belt of greenschist facies mylonite, phyllonite, and cataclasite up to 1 km thick that juxtaposes the European plate to the northwest against the Apulian plate to the southeast (Schmid, 1968; Zingg, 1983; Nicolas and others, 1990). Within the mapped area, these rocks strike north-northeast, dip generally 30° to 60° to the west-northwest, and overlie the Mafic Complex. For the purposes of the present study, the boundary of this structure was placed at the easternmost occurrence of mylonitic rocks. North of Bocchetta della Boscarola (fig. 3, on map sheet), slivers of phyllonite, serpentinite, and mylonitized gabbro with recognizable relic texture are present within more strongly deformed rocks. South of Bocchetta della Boscarola, a large lens of undivided paragneiss of the Kinzigite Formation (pg1) is incorporated into the Insubric Line.

Zingg and Hunziker (1990) have analyzed and dated the movement of the Insubric Line between Val Sesia and Val Sessera. Along this segment, they recognize a pre-Oligocene greenschist-facies mylonite with moderately north-dipping lineations and a dextral sense of shear. This is overprinted in the vicinity of Val Sessera by post-Oligocene cataclastic deformation.

Cremosina Line and Agnona Fault

The Cremosina Line is a regionally significant, northeast-striking, steeply dipping fault that constitutes the southeastern boundary of the Ivrea-Verbano Zone within the mapped area. Regional compilations show it joining the Insubric Line to the southwest and having an apparent dextral slip of about 12 km (Boriani and Sacchi, 1973). Fault gouge of the Cremosina Line is well exposed near the ridge crests northwest of the village of Crevacuore (fig. 3, on map sheet).

The Agnona fault is a subvertical structure that is subparallel to and about 2 km north of the Cremosina Line. It cuts the Mafic Complex and displaces the Kinzigite Formation and undivided granitic rocks approximately 3 km in a dextral sense. The Agnona fault is well exposed in a small stream southeast of the village of Trivero where a sheared and altered gabbro is imbricated with steeply dipping cataclasite and ultra-cataclasite in a zone up to 100 m thick. Tertiary movement on the Cremosina Line and Agnona fault is indicated by the cataclastic style of deformation.

Minor Faults of Permian(?) to Alpine Age

East of the Insubric Line and north of the Agnona fault, rocks of the mapped area are cut by numerous north- to northeast-striking, steeply dipping faults. Most of these faults are right lateral and have offset mapped units less than 500 m. In many places, these faults are clearly brittle structures characterized by recessive-weathering, cataclastic zones less than 1 to 10 m thick. Relics of resistant-weathering mylonite (fig. 15) indicate that at least some of these faults had an older history of ductile deformation. Considered collectively, most of the minor faults have appropriate orientation, displacement, and style of deformation to be riedel shears (Sylvester, 1988) related to motion on the Insubric Line. Analogy to the Insubric Line suggests that the mylonites could have been products of early Tertiary or Late Cretaceous Alpine deformation. However, an older age cannot be definitely ruled out because Permian-Triassic high-temperature, ductile faults related to crustal extension have been described in the northern Ivrea-Verbano Zone (Brodie and Rutter, 1987; Brodie and others, 1989; Handy and Zingg, 1991).

EMPLACEMENT OF THE MAFIC COMPLEX

Widespread, synmagmatic extension in the Mafic Complex and Kinzigite Formation is consistent with evidence that many faults in the northern Ivrea-Verbano Zone are deep crustal expressions of Permian to early Mesozoic extensional tectonics (Hodges and Fountain, 1984; Handy, 1987; Brodie and Rutter, 1987; Brodie and others, 1989; Rutter and others, 1993). It is likely that the Mafic Complex began to form as mantle-derived melts invaded the lower crust at the onset of extension and ponded at their horizon of neutral buoyancy (Sinigoi and others, 1995). This horizon may have corresponded to the granulite- to amphibolite-facies transition in the Kinzigite Formation. The strength of the Kinzigite Formation was probably reduced by heating and anatexis and the internal structure of the Mafic Complex indicates that it behaved as a thick zone of ductile cumulates, the strength of which was probably reduced sig-

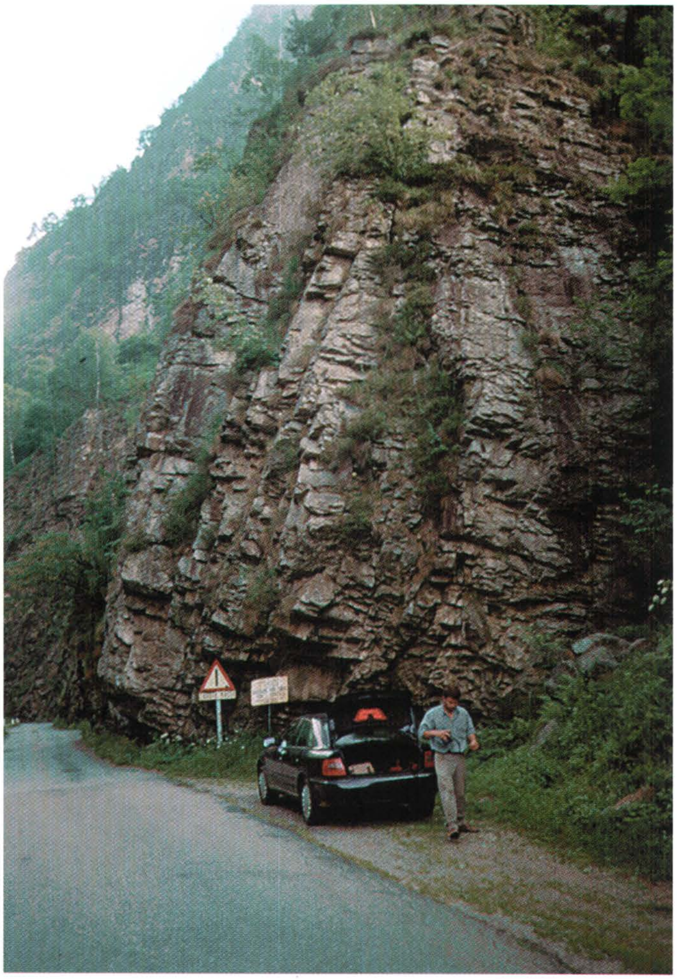


Figure 15.—Resistant-weathering, high-temperature mylonite below the Corno di Gula in Val Mastallone. Entire outcrop is mylonite with a foliation defined by plastically deformed minerals that parallels the obvious, steeply dipping banding. See fig. 3, on map sheet, or geologic map for location.

nificantly by the presence of small amounts of interstitial melt. Hypersolidus conditions were probably maintained for a long time by the insulating effect of the overlying continental crust, and the Mafic Complex and anatectic roof rocks may have acted as a zone of weakness that facilitated extension and lateral movement of the overlying crust. Well-studied analogs for the arcuate structure and synmagmatic deformation of the Mafic Complex are found in ophiolitic gabbro (Nicolas, 1989, 1992). Numerical modeling (Quick and Denlinger, 1992a,b, 1993; Phipps Morgan and Chen, 1993; Henstock and others, 1993) demonstrates that these characteristics can be produced by large-scale necking of a thick section of partially molten cumulates beneath a small (<1 km wide) magma chamber as the crust moves away from an axis of spreading. Application of this model to the Mafic Complex is provided by Quick and others (1992, 1994).

EMPLACEMENT OF THE MANTLE PERIDOTITES

The Ivrea-Verbano Zone has been repeatedly cited as an exposed crust-mantle transition (Mehnert, 1975; Fountain, 1976; Fountain and Salisbury, 1981; Hamilton, 1989). In this context, the eastern contact of the Balmuccia peridotite (fig. 10) has been described in the literature and on innumerable field excursions as an exposed relic of the subcontinental, petrologic Moho. Popular models have interpreted the Balmuccia body to be either a peridotite diapir that pushed into preexisting lower crust represented by the Mafic Complex (Shervais, 1979; Boudier and others, 1984) or the mantle basement over which the Mafic Complex crystallized during underplating at the crust-mantle interface (for example, Rivalenti and others, 1984; Hamilton, 1989; Voshage and others, 1990; Quick and others, 1992).

Quick and others (1995) demonstrated that the peridotite bodies were tectonically incorporated into the Kinzigite Formation as lensoidal bodies prior to the intrusion of the Mafic Complex. Sharp, magmatic contacts with pegmatitic clinopyroxenite are present on both the eastern and western sides of the Balmuccia peridotite. Thus, the Mafic Complex magmatically incorporated the Balmuccia peridotite and is not present west of the Balmuccia peridotite as a consequence of tectonic repetition. The eastern contact of the Balmuccia peridotite dips from subvertical to about 50° to the west. The western contact dips 30° to 40° to the west (cross section *A–A′*). West of Meula, the bottom of the body (fig. 16) is exposed in a waterfall. These orientations, coupled with the outcrop pattern, suggest that the shape of the Balmuccia peridotite is lensoidal or boudin-like. Similar relations are present in the Valmala peridotite, which crops out in two valleys and is encased by a sheath of clinopyroxenite (cross section *B–B′*). Here, the contact between the peridotite and the Mafic Complex is clearly convex and closed upward, consistent with the top of a lens.

It appears that the lensoidal shape of these peridotites was acquired prior to incorporation into the Mafic Complex. Paragneiss septa (pg4) derived from the Kinzigite Formation are present to the east and the west of the Balmuccia and Valmala peridotites. Septa and peridotite are intruded by gabbroic rocks of the Mafic Complex. Foliation, banding, paragneiss septa (pg4), and ultramafic cumulus layers (uc) drape the peridotite bodies in a manner suggesting ductile flow around boudins. These relations indicate that Balmuccia and other peridotite lenses were intercalated with the Kinzigite Formation prior to intrusion of the Mafic Complex and that, following incorporation into the Mafic Complex, they behaved as relatively competent bodies. Quick and others (1995) suggested that intercalation of the mantle peridotite within the Kinzigite Formation may have occurred within the accretionary prism of a subduction zone. Alternatively, the peridotites may have been tectonically derived from the subcontinental mantle and introduced into the Kinzigite Formation when it was at deep crustal levels. In either case, reference to the Ivrea-Verbano Zone as a crust-mantle transition must be tempered by the facts that a relic petrologic Moho is not exposed at the Balmuccia peridotite, and that present exposures were exhumed from unknown distance above the pre-Alpine crust-mantle boundary.

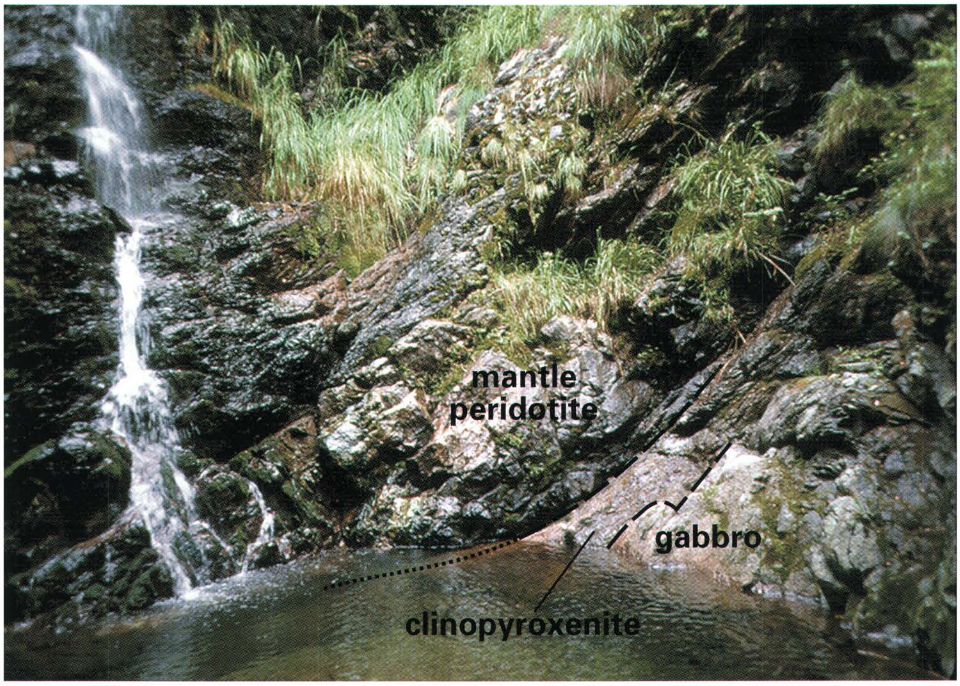


Figure 16.—View to the north of the contacts of the mantle peridotite, altered pegmatitic clinopyroxenite (pc), and gabbro (mc2) of the Mafic Complex approximately 800 m west-northwest of the village of Meula. Contact dips at moderate angle to the west and flattens to subhorizontal orientation under the shallow pool. Relation is consistent with exposure of the lower contact of a lensoidal body of peridotite. See fig. 3, on map sheet, or geologic map for location.

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