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Lower Palaeozoic Amphibolite to Granulite Facies Metamorphism in the Ivrea Zone (Southern Alps, Northern Italy)

by *J. C. Hunziker*[1] and *A. Zingg*[2]

Abstract

Microprobe and isotopic studies were carried out on high grade metapelites from the Ivrea zone, Southern Alps. Temperatures ranging from 580 to 820 °C and pressures from 4.7 to 10 kb were obtained using various calibrations in a profile going from amphibolite to granulite facies. A geothermal gradient between 25-38 °C/km was calculated for this crustal section. Rb-Sr total rock determinations on 9 metapelite samples of 30 to 50 kg define an isochron of 478 ± 20 m.y. with an initial $^{87}Sr/^{86}Sr$ ratio of 0.7086 ± 0.0008 which is interpreted as the age of the regional metamorphism of the Ivrea zone.

The evolution of the Ivrea zone can be summarized as follows:

1) Sedimentation of a mainly pelitic series 480-700 m.y. ago.
2) Burial of these sediments down to a depth of about 30 km.
3) Intrusion of mafic and ultramafic magma into this pelitic sequence probably causing the peak of metamorphism during Caledonian time.
4) Reequilibration of the magmatic assemblages of the mafic and ultramafic rocks under deep crustal conditions during very slow cooling of the whole complex.
5) Tilting of the rock series from the original horizontal position into its present day subvertical position presumably during Hercynian time. The time of the thrusting of the Ivrea zone over rocks of lower density is still discussed.

1) Geological Setting

The Southern Alps, west and southwest of Lago Maggiore consist of the Ivrea zone and the Strona-Ceneri zone (Fig. 1).

According to geophysical investigations the Ivrea zone is a segement of the lower crust thrust over rocks of lower density. (Fig. 1c and e.g. German Research Group for Explosion Seismology 1968, GIESE 1968).

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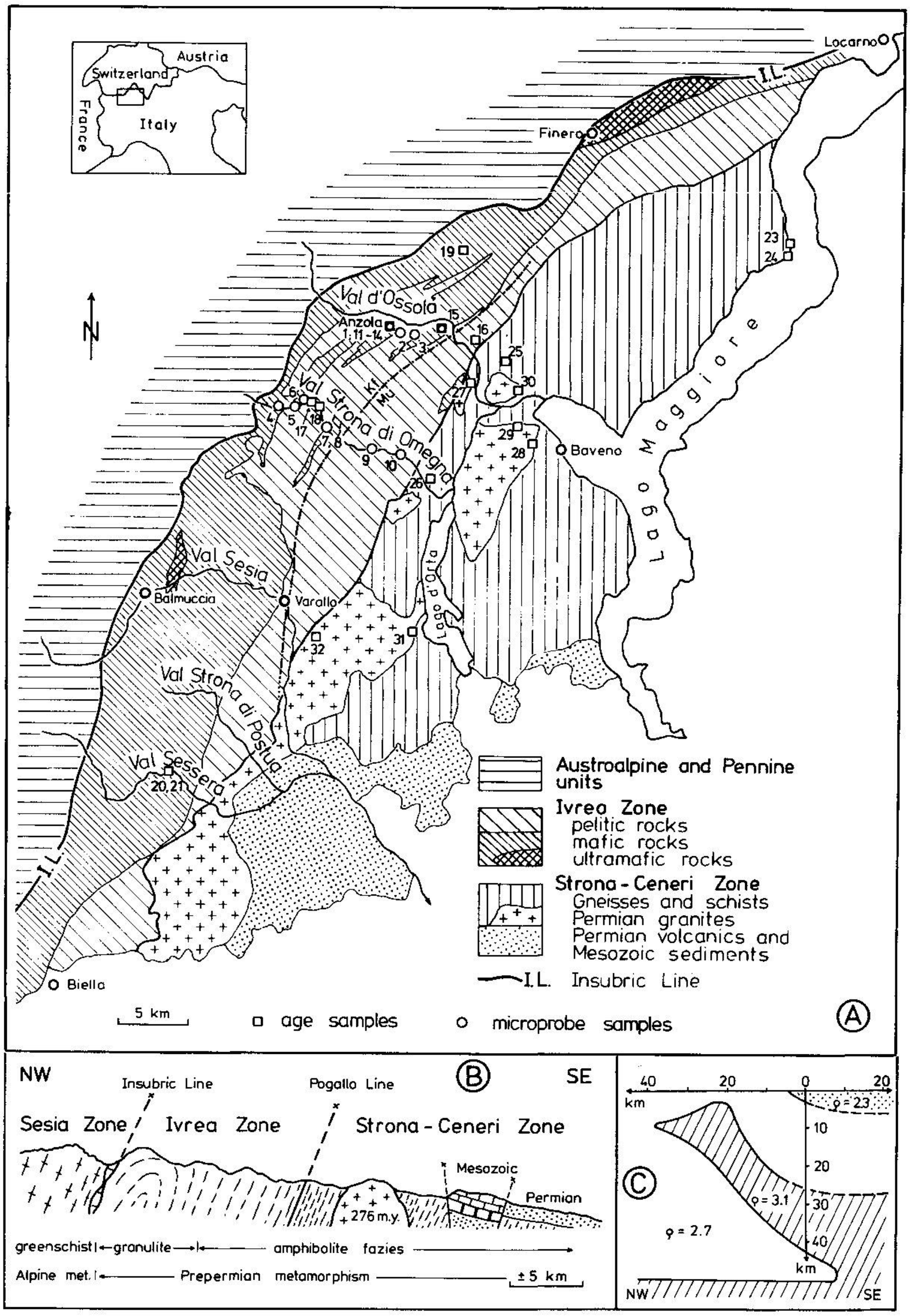

Fig. 1 (A) Sketch map of the Southern Alps west of Lago Maggiore and location of the samples. (B) Schematic composite section. (C) Structure of the crust in the Ivrea zone region deduced from seismic data according to the German Research Group for Explosion Seismology (1968).

The Ivrea zone is made up of a steeply dipping sequence of metapelitic, metabasic and minor metacarbonate and meta-ultrabasic rocks. The metamorphic grade increases from the amphibolite facies in the SE to the granulite facies in the NW. Most rocks show a single phase of crystallization. Reaction textures have been observed only in few samples (PEYRONEL PAGLIANI and BORIANI 1967, SCHMID 1967, BERTOLANI 1968, ZINGG this volume). A later overprint of the granulite facies rocks under amphibolite facies conditions is restricted to areas affected by penetrative deformation after the peak of metamorphism (KRUHL and VOLL 1976, STECK and TIÈCHE 1976).

The Strona-Ceneri zone represents a more shallow part of the crust and consists mainly of granitic gneisses and micaschists in amphibolite facies (REINHARD 1964, BORIANI et al. 1977). These basement rocks are cut by Permian intrusives and discordantly covered by non metamorphic Permian volcanics and Mesozoic sediments. The age of metamorphism therefore has to be older than 280 m.y.

The contact between the Ivrea- and Strona-Ceneri zone is normal in the NE and tectonic in the SW (Pogallo line and other faults) according to BORIANI and SACCHI 1973 and BORIANI et al. 1977).

For the Strona-Ceneri zone concordant U-Pb monazite ages and slightly discordant U-Pb zircon ages around 450 m.y. were obtained by KÖPPEL and GRÜNENFELDER (1971, 1978/79) and KÖPPEL (1974). The monazite from the NW part of the Strona-Ceneri zone, close to the Ivrea zone give ages between 290 and 330 m.y.

The ages from zircons of the Ivrea zone are discordant and point to a lower intercept with the concordia around 280 m.y. where the monazite are located. From these results, the metamorphism of the Ivrea zone was interpreted either as a two phase event or as a one phase event followed by very slow cooling.

The aim of the present study was to establish the age and PT conditions of the amphibolite to granulite facies metamorphism in the Ivrea zone using metapelite samples. Some samples of the pre-Permian basement rocks and of the Permian intrusives from the Strona-Ceneri zone were included in this study.

2) Sample Description and Geological Location of the Samples (Fig. 1)

2.1. IVREA ZONE (TABLES 1 AND 2)

The petrology of the area investigated by us has been studied by PEYRONEL PAGLIANI and BORIANI (1967), BERTOLANI (1968), BERTOLANI and GARUTI (1970), SCHMID (1967), SCHMID and WOOD (1976) and ZINGG (this volume). The whole sampling area lies within the sillimanite field. However, andalusite was found as relic SW of the Val Sesia. Cordierite is restricted to the SW part of the Ivrea zone.

The muscovite-K-feldspar isograd is fairly parallel to the compositional banding of the zone in the Val d'Ossola region but turns in a N-S direction in the Val Sesia region (Fig. 1).

Partial melting of the metapelites is observed in many places. This is documented by the lateral transition of a banded metapelite (sample KAW 1789)

Table 1 Modes (volume per cent from point count) of the microprobe analyses samples from the Ivrea zone. The opaque minerals are graphite, magnetite, ilmenite, rutile and in rare cases pyrrhotite. Abbreviations for tables 1 to 4: Qtz: quartz, Kf: K-feldspar, Plg: plagioclase, Ga: garnet, Bi: biotite, Mu: muscovite, Sil: sillimanite, Opq: opaque minerals, Opx: orthopyroxene, Spi: spinel, Chl: chlorite, Am: amphibole.

	Samples		Qtz	Kf	Plg	Ga	Bi	Mu	Sil	Opq	Locality
●	1	Iv 533	39	14	15	14	3		15	<1	SW Anzola, Val d'Ossola
■	2	Iv 530	32	6	38	15	8			1	SE Anzola, Val d'Ossola
◆	3	Iv 528	30	8	9	19	16		16	2	W Gabbio, Val d'Ossola
▼	15	KAW 85	46	18	13	3	19		1	<1	Teglia, Val d'Ossola
○	4	VS 41	25	15	5	35	<1		18	2	E Campello Monti, Val Strona di Omegna
□	5	Iv 491	35	20	20	18	<1		5	2	Sant Anna, W Forno, Val Strona di Omegna
◇	6	Iv 490	41	3	23	21	<1		10	2	Santa Lucia, W Forno, Val Strona di Omegna
▽	7	Iv 485	72	5	12	3	7		1	<1	Grampi, Val Strona di Omegna
△	8	KAW 508	45	25	15	3	10		2	<1	Grampi, Val Strona di Omegna
+	9	Iv 45	4		4	6	45	10	30	1	W Strona, Val Strona di Omegna
×	10	Iv 458	48		13	1	31	3	4	<1	SE Strona, Val Strona di Omegna

Table 2 Modes (estimated volume per cent) of the age samples from the Ivrea zone.

Samples	Qtz	Kf	Plg	Ga	Bi	Sil	Opq	others	Locality
11 KAW 447	30	15	5	33	1	15	1		Quarry of Anzola, Val d'Ossola
12 KAW 448-454	30	25	10	20	1	12	2		Quarry of Anzola, Val d'Ossola
13 KAW 472	2	25		40	<1	30	3		Quarry of Anzola, Val d'Ossola
14 KAW 468	25	25		25	4	20	1		Quarry of Anzola, Val d'Ossola
15 KAW 85	46	18	13	3	19	1	<1		Teglia, Val d'Ossola
16 KAW 506	30		30		22	8	<1	Mu: 10	Candoglia, Val d'Ossola
17 KAW 599			60	20	2		2	Opx: 16	Forno, Val Strona di Omegna
18 KAW 509	40	10	26	20	2		2		Forno, Val Strona di Omegna
19 KAW 1067	40	20		24	<1	15	1		L'Arca, Val Grande
20 KAW 1787	40	35[1)]	24		<1			Opx:1, Spi<1	NE Alpe Balmello, Val Sessera
21 KAW 1789	20	25	35	18	1		1		NE Alpe Balmello, Val Sessera

into a homogeneous gneiss which has completely lost the compositional banding (sample KAW 1787) as shown in Fig. 2.

Textures of our samples suggest that equilibrium was reached. The only reaction texture observed is the retrograde replacement of garnet by sillimanite and quartz (sample KAW 508/7, compare Fig. 6 of SCHMID and WOOD 1976). Late to postmetamorphic deformation has generated mylonites in several places of the granulite facies domain, but the original mineral assemblages were not affected (samples KAW 447, 448–454, 472).

KÖPPEL (1974) has determined U-Pb zircon and monazite ages on samples from the same localities. The corresponding samples are: Anzl = KAW 447, 448–454, 472, Candoglia = KAW 506, Stro3 = KAW 599, Stro2 = KAW 508 for the Ivrea zone and Stro1 = KAW 507 for the Strona-Ceneri zone. The U-Pb ages are discussed in chapter 5.

2.2. STRONA-CENERI ZONE: BASEMENT ROCKS (TABLE 3)

All samples with the exception of KAW 507, come from the region NW of Lago Maggiore which was remapped recently by BORIANI et al. (1977). Table 3 gives the modes and the corresponding lithological units.

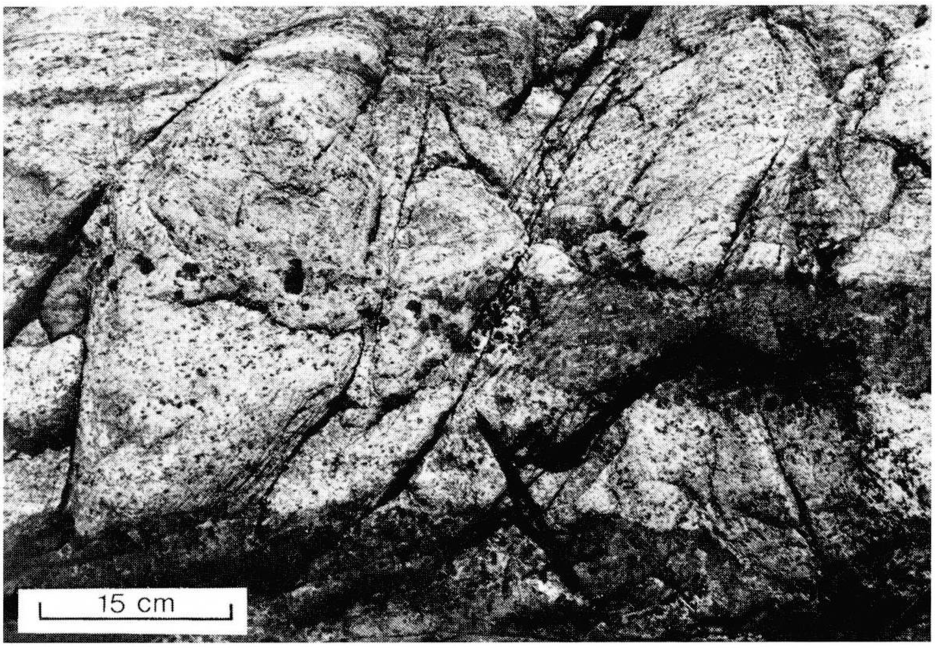

Fig. 2 Transition from a banded metapelite to a homogeneous quartz-feldspar rock. The original banding of the metapelite is nearby erased by anatectic melting forming the light coloured rock. Single garnets within the light rock mark the continuation of the compositional banding of the metapelite. Locality: NE Alpe Balmello, Val Sessera.

2.3. STRONA-CENERI ZONE: PERMIAN INTRUSIVES (TABLE 4)

The Permian granites cut the steeply dipping series of the Strona-Ceneri zone and produce contact aureoles (GALLITELLI 1937). More basic intrusives occur at the limit of the Ivrea zone (SCHILLING 1957, BORIANI and SACCHI 1973). These more basic rocks are often deformed and a weak schistosity may develope. Sample KAW 80 is such a quartzdiorite with quartz recrystallized to fine grained aggregates. The other samples are unalterated granites taken from various localities. The monazite U-Pb ages of the diorite vary between 295–310 m.y. and are distinctly older than those of the Mt. Orfano granite which vary between 270–275 (KÖPPEL and GRÜNENFELDER 1978/79).

3) Mineral Chemistry: Microprobe work

3.1. OPERATING CONDITIONS

The analysis were performed as single point measurements on a fully automatic ARL SEMQ microprobe with 6 wave length dispersive spectrometers using natural minerals as standards. ZAF corrections were applied. The operating conditions were 15 kV acceleration voltage, 20 second counting time, 20 nA beam current on brass for biotite and feldspars and 40 nA for garnet.

3.2. GARNET ZONING AND EQUILIBRIUM AMONG GARNET AND BIOTITE

Garnets from samples collected *below the muscovite-K-Feldspar* isograd are occasionally weakly zoned in a range of ½ mole per cent of Fe and Mn. Mn is replaced by Fe towards the rim of the garnets and this zonation can be attributed to prograde metamorphism.

Above the Muscovite-K-Feldspar isograd the garnets are homogeneous as long as they are isolated from biotite. Garnets in contact with biotites are zoned with respect to Fe and Mg, Fe being enriched within some 10 µm of the common grain boundary. This zonation can be described by the exchange reaction

$$\text{Mg-garnet} + \text{Fe-biotite} = \text{Fe-garnet} + \text{Mg-biotite}\ (1)$$

The distribution coefficient K_D (1)[1] increases towards the common grain

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[1] $$K_D(1) = \frac{X_{Fe}^{Ga} \cdot X_{Mg}^{Bi}}{X_{Mg}^{Ga}\ X_{Fe}^{Bi}}\ ;\ X_{Mg} = \frac{N_{Mg}}{N_{Mg} + N_{Fe}}$$

$$X_{Fe} = \frac{N_{Fe}}{N_{Fe} + N_{Mg}}$$

N is the number of moles and the total Fe was calculated as Fe^{2+}.

Table 3 Modes (estimated volume per cent) of the age samples from the basement of the Strona-Ceneri zone. Lithological units after 1) Boriani et al. (1977), 2) Bertolani (1968).

Samples	Qtz	Kf	Plg	Bi	Mu	Ga	Opq	Others	Lith.	Locality
23 KAW 504	40		35	15	5	<1	<1	Chl: 5	ms 1)	Carmine Inferiore
24 KAW 564	30		45	25			<1		og 1)	La Punchetta, S Carmine Inferiore
25 KAW 572	40		35	10	10	2	1	Chl: 2	ms 1)	Mergozzo
26 KAW 507	25	10	30	20				Am: 15	Mi 2)	Germagno, Val Strona di Omegna

Table 4 Modes (volume per cent from point count) of the age samples from the Permian intrusives of the Strona-Ceneri zone.

Samples	Qtz	Kf	Plg	Bi	Opq	Others	Locality
27 KAW 80	10		57	27	<1	Ga: 6	Camponi, S Ornavasso
28 KAW 598	47	27	23	3	<1		Baveno, Cava Oltrefiume
29 KAW 907	29	45	24	2	<1		Baveno, Madonna della Scarpia
30 KAW 906	27	36	33	4	<1		Mont Orfano
31 KAW 565	35	32	27	6	<1		Alzo, Cava Madonna del Sasso
32 KAW 1204	34	2	48	11	<1	Mu: 5	Rocca Pietra, near Varallo

boundary of the garnet-biotite pairs. Therefore, the temperature decreases (THOMPSON 1976b, FERRY and SPEAR 1978) and the zonation must be attributed to retrograde Fe-Mg exchange during cooling (compare HESS 1971).

Biotites in contact with garnets are not zoned because of the higher diffusion rate of Fe and Mg in this mineral. Their X_{Mg}, however is significantly higher than X_{Mg} of the isolated biotites, in agreement with the retrograde exchange reaction (1).

A systematic study of garnet, biotite and cordierite from the Val Strona di Postua region suggests, that the composition of mineral pairs is affected by retrograde Fe-Mg exchange not only within a contact zone of some 10 μm but over the whole mineral grains (ZINGG 1978). Therefore, isolated garnets and biotites are used for the temperature estimations of the peak of the metamorphism, assuming that they were in equilibrium through a fluid phase. Garnet-biotite pairs give temperatures of some cooling stage of the amphibolite to granulite facies metamorphism.

Biotite inclusions in garnet have also been affected by this retrograde Fe-Mg exchange and have higher X_{Mg} than isolated biotites. TiO_2 contents of these inclusions are lower than those of isolated biotites (Fig. 4 and appendix table 4) as they were trapped during rising metamorphism. Therefore biotite inclusions in garnet do not represent higher grade conditions than isolated biotites, in contrast to the assumption of SCHMID and WOOD (1976).

Looking at the equilibrium between garnet and biotite we can consider the evolution of K_D (1) with increasing metamorphic grade as shown in the profile of Fig. 5. In the amphibolite facies equilibrium among these two phases is suggested by the more or less regular decrease of K_D (1) with increasing metamorphic grade. Although further increase of the metamorphic grade is shown by the mineral assemblages and the mineral compositions the K_D (1) value of samples from the granulite facies domain increase and scatter irregularly indicating disequilibrium among garnet and biotite. From this region some late hydrothermal activity is known from the sericitization of the feldspar and from secondary aqueous fluid inclusions (personal communication, DeNegri, Modena). This hydrothermal activity seems to be responsible for the disequilibrium observed, as biotite inclusions in garnet are apparently not affected by it. (Fig. 5).

3.3. MINERAL COMPOSITIONS

The analyses are reported in the appendix and X_{Mg} of garnet and biotite are plotted on Fig. 5 along a WNW-ESE profile with increasing metamorphic grade. For additional analyses see SCHMID and WOOD (1976).

Feldspars (appendix: table 1 and 2)

The end member compositions were calculated from the K : Na : Ca : Ba proportions. The alkali-feldspar is a microcline. Perthitic and antiperthitic exsolutions of the microcline and plagioclase, respectively were observed in several samples. The composition of the plagioclase from our samples show no systematic change with increasing metamorphic grade. Most of the plagioclase are slightly zoned, the rims are about 1 mole per cent more albitic than the centers. Garnet, the only additional Ca bearing phase, shows no equivalent zonation.

Garnet (appendix: table 3, Fig. 3)

For the calculation of the structural formulae no ferric iron had to be assumed in agreement with wet chemical analyses of SCHMID (1972) and SCHMID and WOOD (1976). Reducing conditions are also suggested by the presence of graphite in most samples.

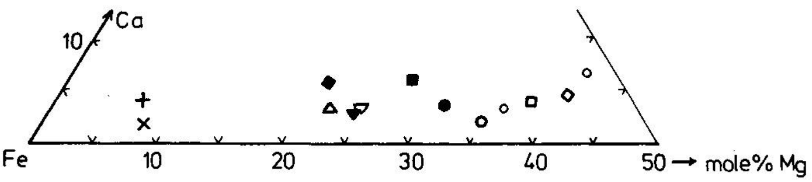

Fig. 3 Composition of isolated garnets (mean values) in a Fe-Mg-Ca plot. Mn ranges from 0.6 to 6.0 mole per cent. Small open circles: samples not used for PT estimations, for the other symbols, see table 1 or 5.

Biotite (appendix: table 4, Fig. 4)

Because of analytical difficulties the anhydrous oxide sums are relatively low and are scattered between 94.2 and 97.8%. SCHMID (1972) obtained H_2O contents up to 2.3 weight% by DTA. Wet chemical analysis by the same author gave 0.0 to 6.0 weight per cent Fe_2O_3. For X_{Mg}^{Bi} all Fe has been assumed to be ferrous iron. The structural formulae were calculated for 22 oxygen (and the tetrahedral sites were filled with Al to 8.00).

3.4. VARIATION OF GARNET AND BIOTITE COMPOSITION WITH INCREASING METAMORPHIC GRADE

Garnet and biotite are involved in different reactions below and above the muscovite-K-feldspar isograd (e.g. THOMPSON 1976a).

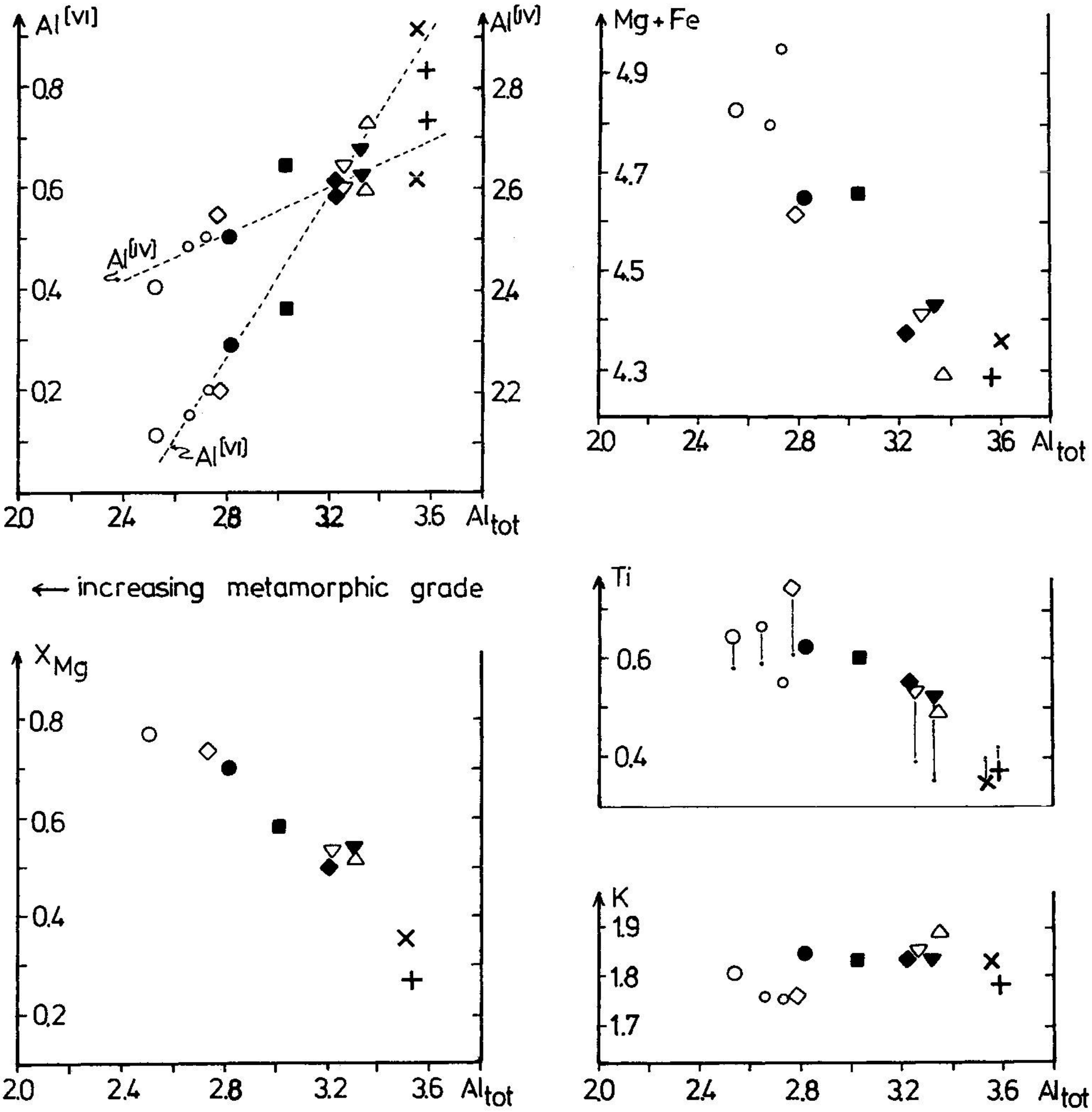

Fig. 4 Composition of isolated biotites normalized to 22 oxygens. The samples from the K-feldspar zone (all symbols except + and x) show a systematic variation of the composition with increasing metamorphic grade. Small open circles: samples not used for PT estimations, dots: Ti contents of biotite included in garnet, for the other symbols, see table 1 or 5.

a) In presence of muscovite: Garnet and biotite show decreasing X_{Mg} with increasing metamorphic grade (Fig. 5). Biotite inclusions in garnet of the two samples have higher X_{Mg} than isolated biotite although no retrograde Fe-Mg exchange could be detected. Both observations would suggest the garnet consuming reaction

Garnet + Muscovite = Biotite + Sillimanite + Quartz (2)

(compare THOMPSON 1976a), but note, that the trend of X_{Mg} was deduced from only 2 samples.

b) In presence of K-feldspar: X_{Mg} of garnet and biotite increase as does the modal ratio of garnet: biotite with increasing metamorphic grade, (Fig. 5) suggesting the continuous reaction

$$\text{Biotite} + \text{Sillimanite} + \text{Quartz} = \text{Garnet} + \text{K-feldspar} + H_2O \ (3)$$

With increasing metamorphic grade the composition of biotite changes as follows (Fig. 4): Si, Ti, Mg + Fe and X_{Mg} increase while Al^{IV} and Al^{VI} decrease.

4) Geothermometry and Geobarometry

4.1. PREVIOUS WORK

Most of the PT estimates were done on the ultramafic and associated rocks of the granulite facies domain close to the Insubric line (CAWTHORN 1975, MEDARIS 1975, ÇAPEDRI et al. 1976, ENGI 1978, GARUTI et al. 1978/79, SHERVAIS 1979). A temperature range of 630 to 1230 °C was obtained using various calibrations of the orthopyroxene-clinopyroxene, olivine-pyroxene-spinel and olivine-spinel thermometer. T around 1200 °C were considered to give the conditions of the magmatic crystallization of these rocks, T around 800 °C were interpreted as subsolidus reequilibration to granulite facies conditions and for the lower temperature retrograde, cation exchange during cooling was assumed.

Pressure estimations range from 12 to 35 kb for the first crystallization and from 8 to 13 kb for the reequilibration under granulite facies conditions within the crust.

From Mn-silicate bearing marbles of the granulite facies in the Val Strona di Omegna a temperature of about 620 °C was obtained by ABRECHT et al. (1978/79), which is too low for granulite facies conditions and was therefore considered by the authors as the blocking temperature for the rhodonite-bustamite system.

For the metapelite the only data available are those of SCHMID and WOOD (1976) for the Ossola region and a single P estimate by SHERVAIS (1979) from a metapelite sample close to Balmuccia.

SCHMID and WOOD (1976) used for pressure determinations the Ca-distribution between garnet and plagioclase according to the reaction.

$$\text{Anorthite} = \text{Grossular} + \text{Sillimanite} + \text{Quartz} \ (4)$$

This reaction is the only geobarometer that can be applied to the metapelites of the whole Ivrea zone. However P estimations using reaction (4) are subject to large uncertainities because of the low grossular content of the Ivrea zone garnets that is the garnet composition plots on the assymptotic limb of the equilibrium curve in a P-X_{Ca}^{Ga} diagramm. (e.g. ZINGG, 1978, Fig. 67). Moreover these uncertainities are enlarged as no experiments were performed in the compositional range of interest to determine the activity of grossular ($a_{Grossular}^{Garnet}$).

For their temperature determination SCHMID and WOOD (1976) used the dehydration reaction (3), assuming an activity of H_2O such that the metapelite were exactly at the point of first melting during metamorphism. PT estimates obtained in this way are very uniform over the Ossola profile and vary around 700 °C and 9 kb. P_{H_2O} decreases from 0.6 to 0.4 P_{tot} from the amphibolite to the granulite facies domain.

4.2. PT ESTIMATES OF THE PRESENT STUDY

To be independent of the activity of H_2O temperatures were calculated from the Fe-Mg exchange between garnet and biotite (reaction 1) using the calibrations of THOMPSON (1976b) and of FERRY and SPEAR (1978).

Pressures were estimated from reaction (4) for want of a better geobarometer. The numerical term derived by SCHMID and WOOD (1976) and the γ_{Ca}^{Plg} of ORVILLE (1972) were adopted. As the extrapolation of the available $a_{Grossular}^{Garnet}$ (HENSEN et al. 1975, CRESSEY et al. 1978) to the garnet compositions of interest is very uncertain minimum pressures were calculated assuming ideal mixing of the grossular component in garnet and maximum pressures using a regular solution model with the Margules parameter of HENSEN et al. (1975) for the grossular-pyrop series ($W_{CaMg} = 7460 - 4.3\,T$, $W_{CaFe} = 0$, $W_{FeMg} = 0$). The PT data obtained this way are given in table 5 and plotted in Fig. 5. These data are affected by incertainties difficult to estimate. Both THOMPSON (1976) and FERRY and SPEAR (1978) give an error of ± 50 °C on their calibration. The error for the thermodynamic calculation of reaction (4) by SCHMID and WOOD (1976) gives an uncertainity of the pressure estimation of ± 1.2 kb for the T of our interest.

4.3. DISCUSSION OF THE PT RESULTS

A regular increase of P and T in the Ivrea zone towards the NW is observed except in the highest grade part where garnet-biotite equilibrium appears to have been affected by late hydrothermal activity (Fig. 5). The regular increase of PT contrasts with the rapid replacement of biotite by garnet (e.g. by reaction 3) at the beginning of the K-feldspar field (Fig. 5, see also Fig. 3 of SCHMID and WOOD, 1976). This rapid replacement must be related to the drop of the activity of H_2O as suggested by fluid inclusions (DE NEGRI and TOURET 1978/79) and by the common occurrence of migmatites in this region. Extensive anatexis

Table 5 PT estimations from the metapelite samples of the Val d'Ossola and the Val Strona di Omegna. The mineral analyses and the definitions of the various x values are given in the appendix. I: isolated garnet or biotite measured at a distance of 20 μm from the edge, P: garnet-biotite and garnet-plagioclase pairs measured at 40–50 μm and 20 μm respectively from their common grain boundary. Isolated garnet and biotite give temperatures close to the peak of metamorphism, garnet-biotite pairs give temperatures of some cooling stage of metamorphism because of retrograde Fe-Mg exchange. F+S: calibration of FERRY and SPEAR (1978) for reaction (1), ABT: calibration of THOMPSON (1976b) for reaction (1), S+W: numerical term for lnK (4) derived by SCHMID and WOOD (1976), P_{min}: minimum pressure calculated from reaction (4) with $(T_{Ga\text{-}Bi})^I_{ABT}$, assuming ideal solution of the grossular component in garnet, P_{max}: maximum pressure calculated from reaction (4) with $(T_{Ga\text{-}Bi})^I_{F+S}$, using the activity data of the grossular-pyrope series of HENSEN et al. (1975). For samples with disequilibrium between garnet and biotite (see chapter 3.2) temperatures were extrapolated from Fig. 5.

		Sample	$(X^{Ga}_{Mg})^I$	$(X^{Ga}_{Mg})^P$	$(X^{Bi}_{Mg})^I$	$(X^{Bi}_{Mg})^P$	$(X^{Ga}_{Ca})^P$	$(X^{Plg}_{Ca})^P$	$(T_{Ga\text{-}Bi})^I$	$(T_{Ga\text{-}Bi})^I$	$(T_{Ga\text{-}Bi})^P$	$(T_{Ga\text{-}Bi})^P$	$P^{min}_{Ga\text{-}Plg}$	$P^{max}_{Ga\text{-}Plg}$
									F+S	ABT	F+S	ABT	S+W	S+W
●	1	Iv 533	0.321	0.292	0.692	0.692	0.036	0.303	640	610	590	570	6.7	9.7
■	2	Iv 530	0.286	-	0.578	-	-	-	780	710	-	-	-	-
◆	3	Iv 528	0.216	0.196	0.496	0.498	0.057	0.489	760	700	700	650	5.9	8.1
▼	15	KAW 85	0.252	0.207	0.537	0.554	0.027	0.221	780	710	640	610	5.6	8.1
○	4	VS 41	0.355	0.348	0.769	0.717	-	-	560	540	640	610	-	-
□	5	Iv 491	0.395	-	-	-	0.041	0.326	-	-	-	-	6.6	10.0
◇	6	Iv 490	0.423	0.408	0.735	0.740	0.046	0.417	730	670	700	650	5.8	9.3
▽	7	Iv 485	0.251	0.203	0.514	0.524	0.035	0.293	820	740	680	630	5.9	8.4
△	8	KAW 508	0.225	0.199	0.519	0.532	-	-	740	680	660	620	-	-
+	9	Iv 45	0.072	-	0.264	-	0.040	0.298	650	610	-	-	4.7	5.7
×	10	Iv 458	0.097	-	0.358	-	0.020	0.104	610	580	-	-	5.6	6.6

and degranitization as suggested by SCHMID (1972, 1978/79) may drastically lower the activity of H_2O in the restites.

The uniformity of temperatures obtained by SCHMID and WOOD (1976) is mainly caused by their assumption about the activity of H_2O. Their data are in good agreement with ours if the temperatures are calculated with the exchange reaction (1) and are represented by the small filled circles in Fig. 5.

The precariousness of the pressure estimations is illustrated in Fig. 5 were the minimum and maximum pressure are plotted. Tentatively extrapolated grossular activities from the data of HENSEN et al. (1975) and CRESSEY et al. (1978) suggest that the effective pressures are reasonably approached by the mean value of the minimum and maximum pressure.

Using the mean pressures, an increase of 0.29 kb/km horizontal distance in the direction perpendicular to the strike of the rock series can be calculated. After the correction for the average dip of 80° a pressure gradient of 0.286 kb/km is obtained. The lithostatic pressure calculated from the mean density of the surface rocks of 2.9 g/m^3 (KISSLING 1980) is 0.284 kb/km. This coincidence suggests that the series of the Ivrea zone were in a horizontal position during metamorphism and tilted afterwards to their present subvertical position. The same conclusion is inferred by current bedding observed in the layered group of the Mafic Formation in Val Sesia (Fig. 7a in Rivalenti 1978/79). This is in contradiction to the conclusions of SCHMID and WOOD (1976) who postulate that the Ivrea zone was already in the present position during metamorphism and the intrusion of the mafic rocks because of their uniform temperatures.

For the crustal segment represented by the Ivrea zone a geothermal gradient of 25°/km (calibration of THOMPSON 1976 for the temperature) or 38°/km (calibration of FERRY and SPEAR 1978) during metamorphism can be calculated.

5) Rb-Sr Age Determinations

5.1. METHODS

11 metapelite samples from the Ivrea zone (table 2), 4 gneiss samples from the Strona-Ceneri zone (table 3) and 6 samples from the Permian Baveno granite suite (table 4) were analyzed for Rb and Sr.

The mechanical and chemical preparation of the samples is described in JÄGER and FAUL (1960), JÄGER (1962) and WÜTHRICH (1965) for the laboratory blanks, see STILLE (1979). Rb and Sr were measured on an Ion Instruments 35 cm radius 90° deflection solid source mass spectrometer, using a triple filament. For further details see BRUNNER (1973). For data measured before 1972, the Varian MATCH 4 mass. spectrometer was used. Isotopic ratios, mean values and errors were calculated using an on-line PDP8 computer.

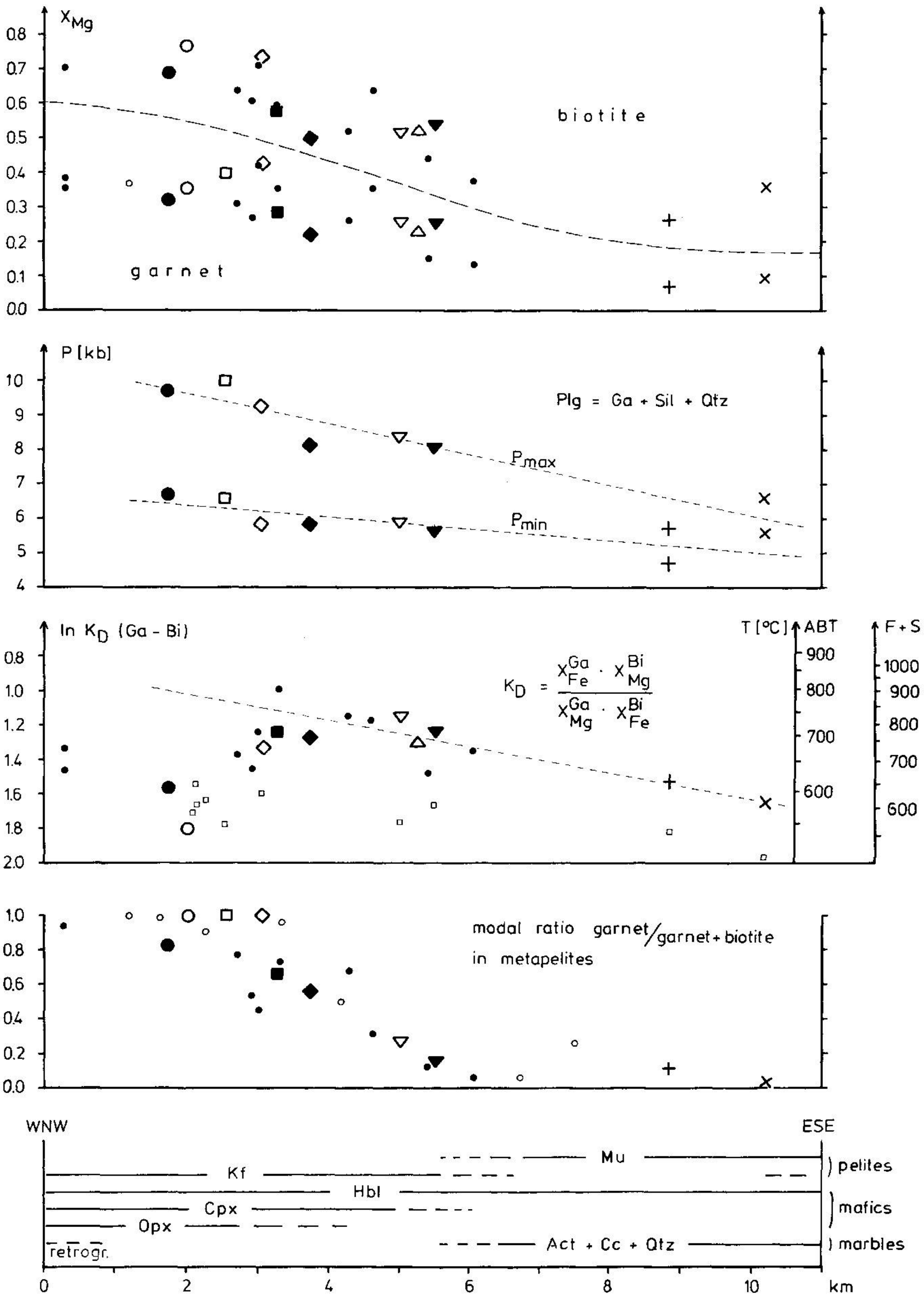

Fig. 5 Variation of P, T, X_{Mg}^{Ga} and X_{Mg}^{Bi} compared with the change of the mineral assemblages in a composite profile perpendicular to the strike of the Ivrea zone. For the PT estimations compare chapters 3.2 and 4.2. The Val d'Ossola and Val Strona di Omegna profiles were adjusted using the muscovite-K-feldspar isograd as a fixpoint. A small error is introduced by this procedure as this isograd is neither an isotherm nor an isobar. 0 km correspond to the Insubric Line in Val Strona di Omegna. Small filled circles: data of Schmid and Wood (1976) recalculated according to reaction (1), small open circles: data not used for PT estimations, small squares: T obtained from biotite inclusions in garnet. These inclusions, except for the two samples of the muscovite zone, have suffered a strong retrograde Fe-Mg exchange during cooling but are still in equilibrium with a small portion of the garnet around the inclusion. For the other symbols, see table 1 or 5.

Table 6 Rb-Sr data and age results on rocks of the Southern Alps.

Samples	Locality	^{87}Rb	$Sr_{com.}$	$^{87}Sr/^{86}Sr$	$^{87}Rb/^{86}Sr$	Isochron age m.y.	Initial ratio
IVREA ZONE							
11 KAW 447	Anzola	23.6	105	0.72014	1.6773		
12 KAW 448-454	Anzola	17.6	198	0.71636	0.9065		
13 KAW 472	Anzola	22.7	118	0.72185	1.9568	453 ± 54	0.7098 ± 0.0017
14 KAW 468	Anzola	33.2	95.8	0.73309	3.5446	460 ± 30	0.7095 ± 0.0013
15 KAW 85	Teglia	34.8	101	0.73325	3.5292		
16 KAW 506	Candoglia	50.8	101	0.74317	5.1621	478 ± 20	0.7086 ± 0.0008
19 KAW 1o67	Val Grande	29.8	132	0.72338	2.3093		
17 KAW 599	Forno	1.73	661	0.70729	0.0268		
18 KAW 509	Forno	4.60	280	0.71024	0.1682		
20 KAW 1787	Val Sessera	1.38	527	0.71228	0.0268	parallel to 478m.y.	0.7120 ± 0.0007
21 KAW 1789	Val Sessera	16.4	251	0.71781	0.6678		
STRONA-CENERI ZONE: BASEMENT ROCKS							
23 KAW 504	Carmine Inf.	24.0	203	0.72051	1.2045		
24 KAW 564	La Punchetta	39.9	162	0.72395	2.5186		
25 KAW 572	Mergozzo	40.9	182	0.72766	2.2985	473 ± 29	0.7107 ± 0.0018
26 KAW 507	Germagno	51.4	59.1	0.77116	8.8924		
STRONA-CENERI ZONE: PERMIAN INTRUSIVES							
27 KAW 80	Camponi	35.3	450	0.71144	0.80246		
28 KAW 598	Baveno	62.4	20.1	0.83341	31.688		
29 KAW 907	Baveno	69.1	22.4	0.83040	31.531		
30 KAW 906	Mt. Orfano	56.9	85.6	0.73590	6.7961	276 ± 5	0.7087 ± 0.0009
31 KAW 565	Alzo	56.1	78.7	0.73938	7.2885		
32 KAW 1204	Rocca Pietra	37.9	196	0.71516	1.9785		

On the same computer age calculations and isochrons were established using the programme of Brooks et al. (1972). For Rb and Sr the isotopic ratios and constants were used, following the convention of the IUGS subcommission on geochronology, compiled by Steiger and Jäger (1977):

$^{86}\mathrm{Sr}/^{88}\mathrm{Sr} = 0{,}1194$ (atomic ratios)
$^{84}\mathrm{Sr}/^{86}\mathrm{Sr} = 0{,}056584$ ”
$^{85}\mathrm{Rb}/^{87}\mathrm{Rb} = 2{,}59265$ ”
$\lambda\,^{87}\mathrm{Rb} = 1{,}42 \cdot 10^{-11}\,\mathrm{y}^{-1}$ ”

5.2. ANALYTICAL RESULTS

9 paragneiss samples of at least 30 to 50 kg of rock from the Ivrea zone (see Fig. 1 and table 6), define an isochron of 478 ± 20 m.y. with an initial $^{87}\mathrm{Sr}/^{86}\mathrm{Sr}$ ratio of 0,7086 ± 0,0008 (see Fig. 1 and 6 and table 6).

The sample localities can be subdivided into 3 different regions; Valle d'Ossola, Val Strona di Omegna and Val Sessera, defining 3 profiles through the

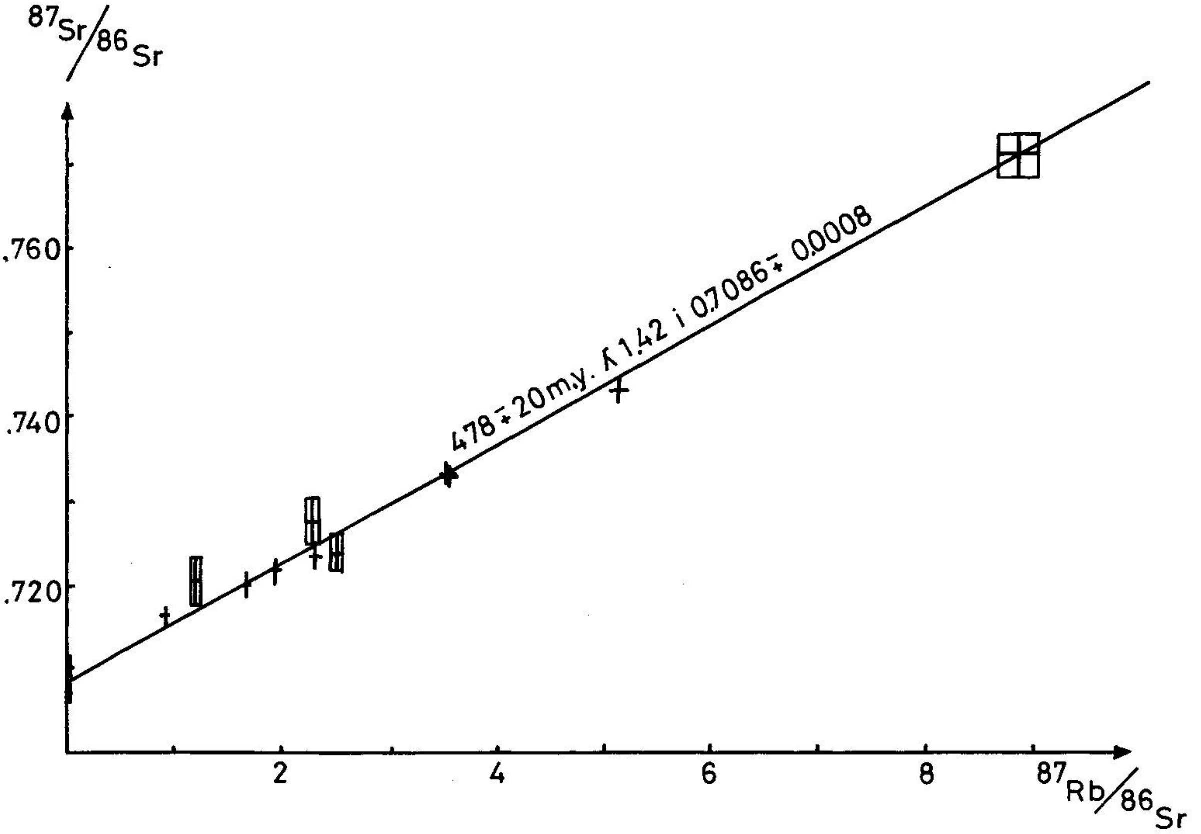

Fig. 6 Rb-Sr whole rock data on amphibolite to granulite facies rocks of the Southern Alps. + Metapelites of the Ivrea zone:
478 ± 20 m.y., $^{87}\mathrm{Sr}/^{86}\mathrm{Sr}$ initial: 0.7086 ± 0.0008
⊞ Basement rocks of the Strona-Ceneri zone: 473 ± 23 m.y., $^{87}\mathrm{Sr}/^{86}\mathrm{Sr}$ initial: 0.7107 ± 0.0018.

Ivrea zone. The Valle d'Ossola profile is best represented with seven samples. 4 of these 7 samples were collected within 300 m distance near the quarry of Anzola; the other 3 samples with increasing distance up to 8 km from the locality of Anzola.

The 7 Valle d'Ossola samples define an isochron of 460 ± 30 m.y. with an initial $^{87}Sr/^{86}Sr$ ratio of 0,7095 ± 0,0013, see table 6. If a separate isochron is calculated for the 4 Anzola samples the uncertainty is bigger but within the error margins, the same result is found (453 ± 54 m.y. initial $^{87}Sr/^{86}Sr$ 0,7098 ± 0,0017). Adding the 2 samples from the Val Strona leads to a better defined isochron of 478 ± 20 with an initial ratio of 0,7086 ± 0,0008. The slight increase in age (within the error limits) is due to a closer definition of the initial $^{87}Sr/^{86}Sr$ value by the last samples, as seen in Fig. 6.

The two samples from the Val Sessera define a line parallel to the 478 m.y. isochron with a slightly higher initial ratio (see table 6). Slight differences in the original $^{87}Sr/^{86}Sr$ ratio of the sediments can be expected if we consider that these two samples were collected at a distance between 20 and 40 km from the other samples.

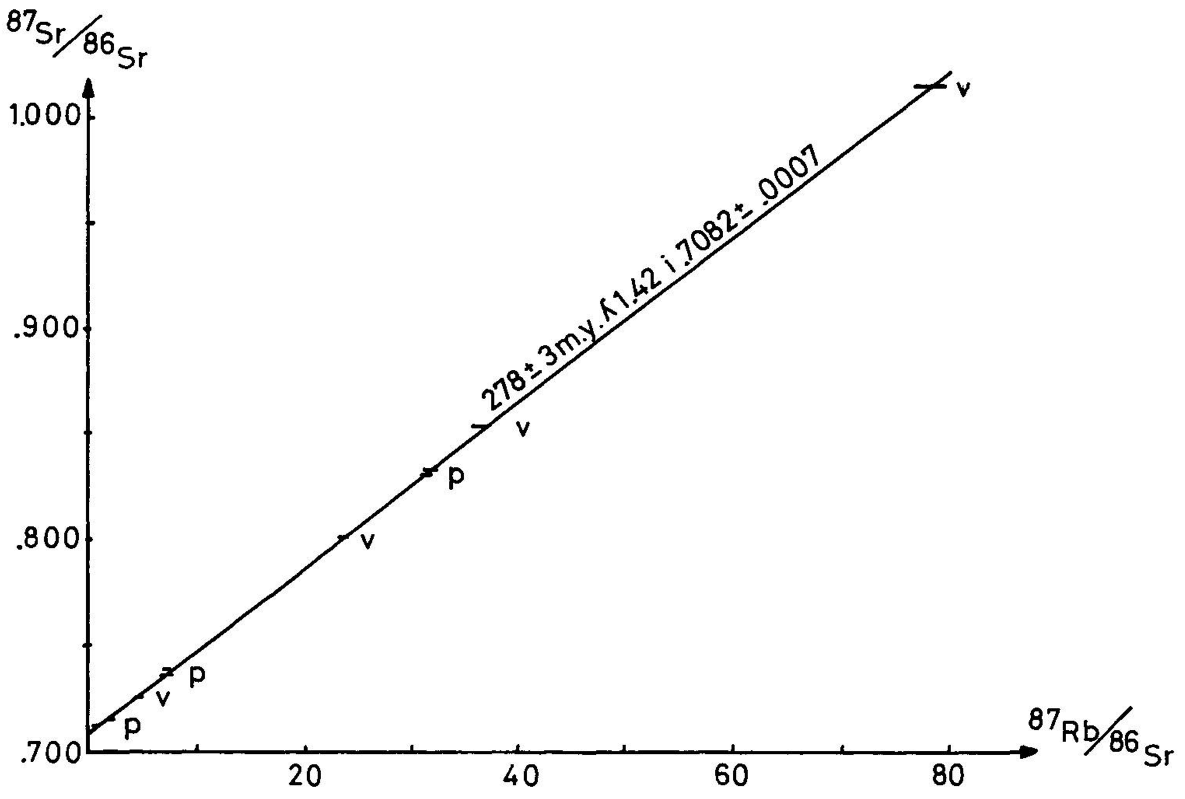

Fig. 7 Rb-Sr whole rock data on Permian magmatic rocks from the Strona-Ceneri zone. P: Granites to quartz diorites from the Baveno granite suite. V: Rhyolites and rhyolitic tuffs from the Permian quartz porphyry province.
P isochron: 276 ± 5 m.y., $^{87}Sr/^{86}Sr$ initial: 0.7087 ± 0.0009.
V isochron: 280 ± 5 m.y., $^{87}Sr/^{86}Sr$ initial: 0.7071 ± 0.0013.
P+V isochron: 278 ± 3 m.y., $^{87}Sr/^{86}Sr$ initial: 0.7082 ± 0.0007.

From the analytical point of view paragneiss samples are rather difficult to measure and never yield high precision Sr measurements, so that the error on the fourth decimal is rather high. Both the given analytical error and the possible differences in intial $^{87}Sr/^{86}Sr$ ratios between the different localities nevertheless did not lead to scattering above the analytical uncertainities of ± 0,2% and thus allow the calculation of a geologically meaningful age.

The 4 gneiss samples from the adjoining Strona-Ceneri zone (Fig. 1) define an isochron of 473 ± 29 m.y. with an initial $^{87}Sr/^{86}Sr$ ratio of 0,7107 ± 0,0018, see Fig. 6. The close coincidence of initial ratio and age of the Strona-Ceneri with the Ivrea-Verbano samples demonstrates the similarity in the older history of both Southern Alpine domains. 6 granitic samples of the Baveno granite suite (see Fig. 1), ranging from quartz diorites to granites yield an Rb/Sr isochron of 276 ± 5 m.y. with an initial Sr ratio of 0,7087 ± 0,0009, see Fig. 7 and table 6. A similar age and initial ratio was found for the widespread Permian rhyolites of the Southern Alps, (HUNZIKER, 1974 and Fig. 7) showing that both rhyolites and granites have a comagmatic origin as already pointed out by field relations. When combined on one isochron, rhyolites and granites define an isochron of 278 ± 3 m.y., with an initial $^{87}Sr/^{86}Sr$ ratio of 0,7082 ± 0,0007, see Fig. 7.

5.3. DISCUSSION

Ivrea zone

Isochrons on metapelites can be interpreted in different manners. In unmetamorphic sediments the isochron may date the time of sedimentation or of diagenesis (CLAUER 1976). In metamorphic terrains, during dehydration of the rock, already at a low grade stage, Rb-Sr systems may reequilibrate and might not be affected by subsequent higher grade metamorphism until partial melting occurs.

As partial melting of the Ivrea zone metapelites is observed on a regional scale (SCHMID 1978/79) our isochron of 478 m.y. can be interpreted as the time of amphibolite to granulite facies metamorphism. This interpretation is supported by the two samples of the migmatites from Val Sessera (KAW 1787 and 1789) shown in Fig. 2. From field relations and petrological studies this partial melting took place close to the thermal peak of granulite facies metamorphism. Such a melting process would most likely have opened the Rb-Sr system of the rock. As from this sample both palaesome and neosome define a line parallel to the 478 m.y. isochron, we can safely conclude that 478 ± 20 m.y. can be interpreted as the time of granulite to amphibolite facies metamorphism.

Previous work by GRAESER and HUNZIKER (1968) on banded granulites showing a post metamorphic mylonitisation gave evidence for a 310 m.y.

isochron. (Recalculated with the new constants 338 ± 41 m.y. with an initial $^{87}Sr/^{86}Sr$ value of 0,7128 ± 0,0020). These samples also are from the locality of Anzola. All bands taken together as one big sample (448–454) plot on the Caledonian isochron, see Fig. 6.

The samples of GRAESER and HUNZIKER (1968) are 1–4 cm thick slabs of 3–5 kg in weight from a mylonitic metapelite. Our new samples, however, are 30–50 kg in weight, both with mylonitic and equilibrium textures, so that we think that the crucial difference consists in sample size, in other words, in exchange dimensions. The locality of Anzola yields both Hercynian and Caledonian ages depending on sample size. Big samples give Caledonian ages, small samples mylonitic slabs of the banded bigger samples furnish a Hercynian isochron. Taking samples of intermediate size, a possibility that can never be ruled out, especially in migmatic rocks, leads to analytical points that neither fall on the 478 nor on the 338 m.y. isochron. Such points enlarge the scatter on both events. These problems will be studied in greater detail in a further work.

Strona-Ceneri zone: basement rocks

The four gneisses from the Strona-Ceneri zone were measured before 1972 and show bigger uncertainities. Within the error limits these data fall on the same isochron as our Ivrea-samples (See table 6 and Fig. 6). In analogy to the Ivrea zone we interpret the 473 ± 29 m.y. as time of amphibolite facies metamorphism. This interpretation is in agreement with PIDGEON et al. (1970) reporting Caledonian U-Pb ages of zircon from the Strona-Ceneri zone and also with monazite and zircon data of KÖPPEL and GRÜNENFELDER (1971) reporting an age of 450 ± 10 m.y. for the Caledonian metamorphism.

ALLEGRE et al. (1969) and (1974) in an attempt to extend the Cadomian event to the Alps, reinterpreted all the U-Pb data using multi stage and/or multi episodic models and came to the conclusion, that in the Ivrea- as well as in the Strona-Ceneri zone, the dominating event was a metamorphism around 520 to 580 m.y. followed by a Hercynian event around 300 m.y. Reinforcement of this new interpretation was given by HAMET et ALBAREDE (1973) reporting a Rb-Sr isochron on some of KÖPPEL and GRÜNENFELDER's paragneiss samples from the Strona-Ceneri zone, with a calculated age of 543 m.y. (λ 1,42) and an initial $^{87}Sr/^{86}Sr$ ratio of 0,7135. From the 7 reported Rb-Sr analyses, only 5 were used to calculate this isochron. Errors are quoted neither on the age nor on the initial value. The $^{87}Rb/^{86}Sr$ as well as the $^{87}Sr/^{86}Sr$ ratios are given to only 2, 3 sometimes 4 decimal places. Recalculating the data points of HAMET et ALBAREDE and omitting the two samples lying off the isochron we get 539 ± 55 m.y. (λ 1,42) with an initial 87/86 ratio of 0,7134 ± 0,0014. Within analytical error these data do not exclude a Caledonian event and are still in agreement with our data. We think that sample size played an important role in the scatter of HAMET et ALBAREDE's data and explains the 2 discordant whole rock samples. Taking all the existing data into consideration, we do not see the necessity of a Cadomian event in the southern Alps, but rather favour a strong Caledonian metamorphism 478 ± 20 m.y. ago.

The Rb-Sr whole rock ages are the same in the Ivrea zone and Strona-Ceneri zone but the mineral ages differ. This can best be explained looking at the post Caledonian history of these two adjoining crustal segments.

In the Strona-Ceneri zone which represents a more shallow part of the crust (amphibolite facies metamorphism) the U-Pb system of monazite and zircon equilibrated at about the same time as the Rb-Sr whole rock system (450–500 m.y.). Only the Rb-Sr and K-Ar mica and amphibole ages were reset during Hercynian time at temperatures around 500 °C. After the Hercynian cooling and erosion, sedimentation of Westphal D occured in the southern part of the Strona-Ceneri zone.

In the Ivrea zone which represents a segment of deep crust (amphibolite to granulite facies, mafics associated with ultramafics in the deepest part) a high temperature regime lasted presumably from Caledonian to Hercynian time (Fig. 10). Very slow cooling after the peak of the amphibolite to granulite facies metamorphism is suggested by an extended retrograde Fe-Mg exchange between garnet, biotite and cordierite (ZINGG 1978) and temperatures of around 550–600 °C still prevailed during Hercynian time. The closing temperature of 300 ± 50 °C for K-Ar biotite was reached about 180 m.y. ago (HUNZIKER 1974). The high temperature of 600 °C and more maintained over 150 m.y. obviously affected zircon and monazite U-Pb systems, shifting the points down from 450–500 m.y. towards 285–300 m.y. with concordant monazite at 275 m.y. (KÖPPEL 1974). The NW most part of the Strona-Ceneri zone in the Val d'Ossola region show the same age pattern as the Ivrea zone.

Strona-Ceneri zone: Permian magmatism

The Baveno suite has been envisaged as the product of degranitisation of the metapelites from the Ivrea zone during Caledonian metamorphism. The isochron of 278 ± 3 m.y. together with the rather high mean $^{87}Rb/^{86}Sr$ value of 11 ± 3 draw strong constraints on the life time of this acid magma, in other words on the time of degranitisation. Fig. 8 shows that this Permian magma had only a short history prior to emplacement, reaching no further back than 310 m.y. Thus the magma generation for the Baveno suite is Hercynian in age. Anatectic melting of metapelites from the Ivrea and Strona-Ceneri zone during Hercynian time would lead to rocks with high $^{87}Rb/^{86}Sr$ ratios above 0,7140. Therefore

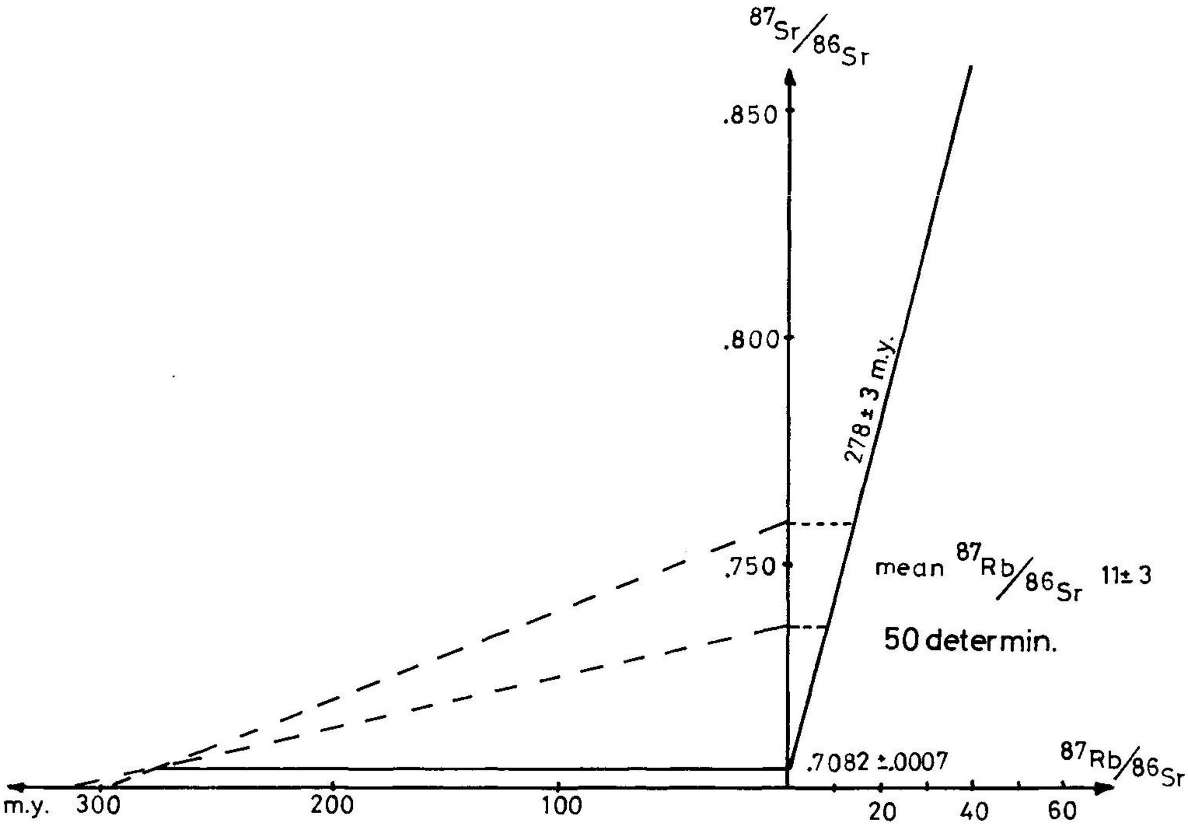

Fig. 8 Combined Compston/Jeffery/Nicolaysen diagramm, showing the evolution of the Permian magmatism in the Strona-Ceneri zone. The short lifetime of the magmachamber prior to extrusion or intrusion is not compatible with a Permian magmatism as a consequence of a Caledonian degranitisation process.

we assume that the magma was generated from less differentiated parts of the crust through partial melting controlled by pressure release during the Hercynian event. This event could correspond to the tilting of the crustal segment represented by the Ivrea and Strona-Ceneri zone from its original position into subvertical position, with the deepest part of the crust (highest grade metamorphic rocks) exposed in the NW, as observed today.

The widespread Orthogneiss bodies in the Strona-Ceneri zone could more likely represent products of a Caledonian degranitisation.

6) Conclusions and Evolution Model (Fig. 10)

The establishment of an evolutional model is affected by large uncertainities due to incomplete knowledge of the deformational history. Mean Rb-Sr ratios of Ivrea-zone metapelites and the initial $^{87}Sr/^{86}Sr$ ratio 478 ± 20 m.y. ago and today can be used to estimate a maximum age of deposition for these sediments. Using a combined Compston Jeffery/Nicolaysen diagramm (Fig. 9), the line

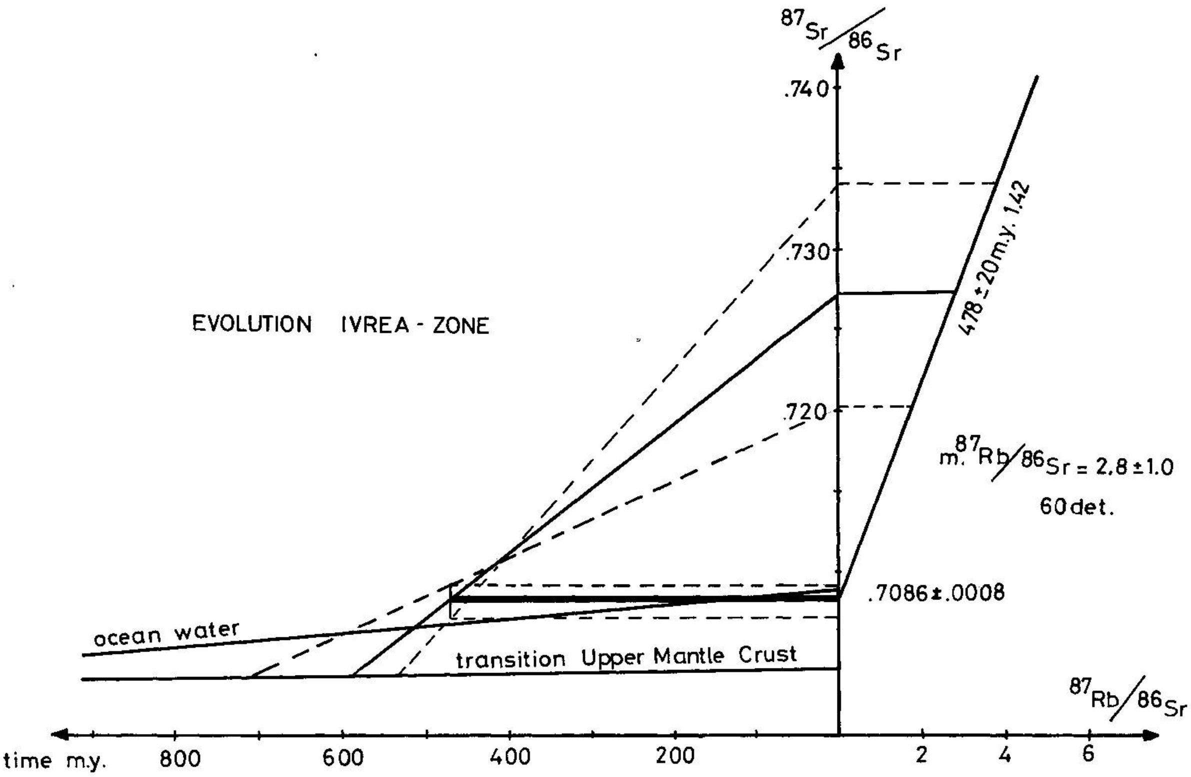

Fig. 9 Combined Compston/Jeffery/Nicolaysen diagramm showing the evolution of the Ivrea zone metapelites. The mean $^{87}Rb/^{86}Sr$ ratio of 2.8 ± 1.0 as derived from 60 X-ray fluorescence determinations together with the measured initial $^{87}Sr/^{86}Sr$ ratio of 0.7086 ± 0.0008 and the age of metamorphism of 478 ± 20 m.y. lead to strong arguments for a sedimentation of the pelites between 700 and 480 m.y.

between the mean $^{87}Sr/^{86}Sr$ and the initial $^{87}Sr/^{86}Sr$ ratio 478 m.y. ago can be extrapolated towards minimum $^{87}Sr/^{86}Sr$ values of the mantle crust transition or of Ocean water. Thus an estimation of the age for the sedimentation leads to values between 480–700 m.y. 700 m.y. corresponds definitely to a maximum age, as all possible geological processes changing $^{87}Sr/^{86}Sr$ ratios or causing incomplete equilibration will only lead to a diminution of this maximum age of sedimentation.

The discordant zircon data pointing to upper intercepts of 1900–2500 m.y. (KÖPPEL and GRÜNENFELDER, 1971) show that the source for the sediments must have been very old continental crust. This old continental crust might be any shield area of Northern Europe or Africa and does not necessarily reflect older crust in the vicinity of the Ivrea zone, as zircons can easily be transported over distances of more than 1000 km. After sedimentation on an unknown substratum, the sediments were buried to depths of about 30 km, slowly warming up through a resetting of the geothermal gradient. Basic and ultramafic material then intruded this predominantly pelitic sedimentary sequence, causing the thermal peak of metamorphism and disturbing the pattern of mineral isograds locally (Val Sesia). During the thermal climax both mafic and pelitic rocks were deformed together for the first time. Most likely our Rb/Sr metapelite whole rock isochron dates this event.

The time after intrusion of the ultramafics/mafics is marked by reequilibration of the magmatic assemblage to deeper crustal conditions. From the pressure gradient obtained by geobarometry and from current bedding observed in the layered group of the Mafic Formation we conclude that the rock series of the Ivrea zone were in a horizontal position during Caledonian time. The extended retrograde Fe-Mg exchange between garnet and biotite (and cordierite in the SW part of the zone) observed by Zingg (1978) suggest that subsequent cooling was extremely slow. Therefore we favour the idea that the crustal segment represented by the Ivrea zone has remained in its original position and under high T regime up to Hercynian time. The Hercynian event, as far as recorded in the rocks of the Ivrea zone, seems to be due essentially to tectonic ac-

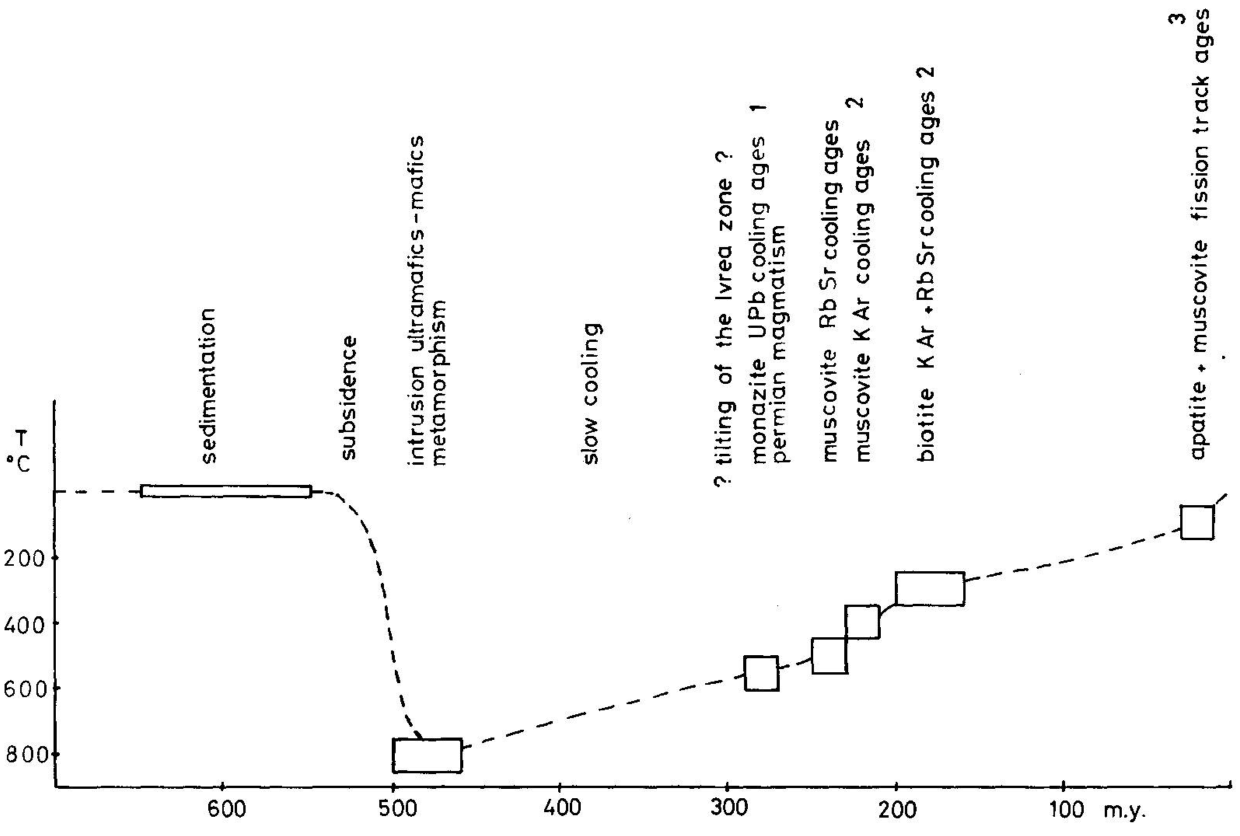

Fig. 10 Thermal evolution of the granulite facies domain of the Ivrea zone derived from the data presented in this paper and from age data of 1) Köppel (1974) and Köppel and Grünenfelder (1978/79), 2) Hunziker (1974), 3) Wagner and Reimer (1972). The closing temperatures for the various mineral systems are those estimated for the Central Alps. Slow cooling between Caledonian and Hercynian time seems more probable than two thermal peaks because of the strong retrograde Fe-Mg exchange between garnet-biotite pairs (and garnet-cordierite pairs in the SW part of the Ivrea zone) indicating a very long cooling period at temperatures ≥ 580°C (Zingg 1978). Note that most of the different ages plotted in this figure were obtained on the same samples and that these samples show in thin section no mineral reactions.

tivity: through the tilting of the rock series to the vertical they were brought into more superficial regions and the high temperature regime was stopped. The closing of the monazite U-Pb system thus marks the abrupt end of the high temperature history of the Ivrea zone, as already suggested by KÖPPEL (1974). In the neighbouring Strona-Ceneri zone the Hercynian tectonic event is documented by discordant Permocarboniferous conglomerates on the basement. The Permian magmatism which originates from deep seated regions as shown by geochemical considerations might be related to this tectonic phase, and already finds the rock sequence in its present position. After the Hercynian event cooling of the Ivrea zone is faster as shown in Fig. 10. For early Mesozoic time an average cooling rate of 5 °C/m. y. was calculated from the muscovite and biotite ages (HUNZIKER 1974). This rate is still very slow when compared with cooling rates of 25 °C/m. y. (CLARK and JÄGER 1969) for the Central alps.

Apatite and Muscovite fission track data (MILLER and JÄGER, 1968, WAGNER and REIMER, 1972) reveal, that during Alpine times the rocks were at around 100 °C, either as a result of slow cooling since Hercynian time or of reheating during the Alpine event.

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APPENDIX

Table 1 Mean composition of potassic feldspar, endmembers given in mole per cent. Or: orthoclase, Ab: albite, An: anorthite, Cs: celsian.

	POTASSIC FELDSPAR				
	Or	Ab	An	Cs	phase
Iv 528	89.9	8.8	0.4	0.9	microcline
KAW 85	87.1	11.8	0.4	0.7	microcline
VS 41	85.5	13.5	0.4	0.6	perthite
Iv 491	77.5	19.7	2.1	0.7	microcline
Iv 491	85.2	12.9	1.0	0.9	perthite
Iv 491	7.7	83.1	9.1	0.1	exsolved phase
Iv 490	88.8	9.2	0.9	1.1	microcline
Iv 490	18.9	78.8	1.2	1.1	exsolved phase
Iv 485	76.5	22.5	0.2	0.8	microcline
Iv 485	90.8	8.2	0.2	0.8	perthite
Iv 485	8.7	90.0	1.0	0.3	exsolved phase

Table 2 Mean composition of plagioclase (mole per cent).

	PLAGIOCLASE					
	rim			center		
	An	Ab	Or	An	Ab	Or
Iv 533	30.3	68.4	1.3	31.8	67.2	1.0
Iv 528	48.9	50.5	0.6	48.4	50.8	0.8
KAW 85	22.1	77.2	0.7	23.3	75.8	0.9
Iv 491	32.6	65.8	1.6	35.0	63.7	1.3
Iv 490	41.7	57.6	0.7	41.3	57.7	1.0
Iv 485	29.3	69.8	0.9	29.7	69.3	1.0
Iv 45	29.8	69.6	0.6			
Iv 458	10.4	89.3	0.3	13.1	86.4	0.5

Table 3 Mean composition of isolated garnet (I) and of garnet in contact with biotite (P). (I) was measured at 20 µm from the edge and (P) at 40–50 µm from their common grain boundary. $X_{Mg}{}^* = Mg/Mg+Fe+Ca+Mn$, $X_{Mg} = Mg/Mg+Fe$, $X_{Ca} = Ca/Mg+Fe+Ca+Mn$.
The number of measured grains is indicated below the sample number.

	GARNET								
	Iv 533 3 I	Iv 533 3 P	Iv 530 4 I	Iv 528 4 I	Iv 528 5 P	KAW 85 5 I	KAW 85 6 P	VS 41 6 I	VS 41 3P
SiO_2	38.23	38.32	37.96	37.70	37.58	37.85	37.85	38.56	38.50
Al_2O_3	22.19	22.03	21.75	21.74	21.64	21.64	21.60	22.20	22.36
MgO	7.93	7.14	6.46	4.93	4.50	6.07	4.85	8.89	8.64
FeO	29.90	30.86	28.70	31.93	32.89	32.00	33.21	28.85	28.86
MnO	0.33	0.37	2.01	1.31	1.40	1.53	1.78	0.28	0.29
CaO	1.31	1.30	2.21	2.00	1.94	0.98	1.03	0.88	0.85
Total	99.89	100.02	99.09	99.61	99.95	100.07	100.32	99.66	99.50
Formula per 12 oxygens									
Si	2.98	2.99	3.00	2.99	2.99	2.99	3.00	2.99	2.99
Al	2.04	2.03	2.02	2.03	2.03	2.01	2.02	2.03	2.04
Mg	0.92	0.83	0.76	0.58	0.53	0.71	0.57	1.03	1.00
Fe	1.95	2.01	1.89	2.12	2.19	2.11	2.20	1.87	1.87
Mn	0.02	0.02	0.13	0.09	0.09	0.10	0.12	0.02	0.02
Ca	0.11	0.11	0.19	0.17	0.17	0.08	0.09	0.07	0.07
$X_{Mg}{}^*$	0.307	0.279	0.255	0.197	0.179	0.237	0.192	0.344	0.337
X_{Ca}	0.036	0.037	0.063	0.057	0.055	0.027	0.029	0.024	0.024
X_{Mg}	0.321	0.292	0.286	0.216	0.196	0.252	0.207	0.355	0.348

GARNET

	Iv 491 8 I	Iv 490 5 I	Iv 490 4 P	Iv 485 5 I	Iv 485 6 P	KAW 508 3 I	KAW 508 4 P	Iv 45 8 I	Iv 458 11 I
SiO_2	38.56	38.97	38.91	37.84	37.37	37.65	37.45	36.92	36.75
Al_2O_3	21.94	21.95	22.25	21.75	21.65	21.63	21.58	21.07	21.32
MgO	9.91	10.50	10.25	5.83	4.56	5.21	4.56	1.61	1.90
FeO	27.01	25.52	26.46	30.98	31.85	31.90	32.77	37.69	31.41
MnO	0.32	0.46	0.46	2.39	2.68	2.22	2.39	1.09	7.83
CaO	1.42	1.72	1.56	1.18	1.13	1.20	1.19	1.33	0.65
Total	99.16	99.12	99.89	99.97	99.24	99.81	99.94	99.71	99.86
Formula per 12 oxygens									
Si	2.98	3.00	2.98	2.99	2.99	2.99	2.99	3.01	2.98
Al	2.00	1.99	2.01	2.02	2.04	2.02	2.03	2.02	2.03
Mg	1.14	1.20	1.17	0.69	0.54	0.62	0.54	0.20	0.23
Fe	1.75	1.64	1.70	2.05	2.13	2.12	2.18	2.57	2.14
Mn	0.02	0.03	0.03	0.16	0.18	0.15	0.16	0.08	0.54
Ca	0.12	0.14	0.13	0.10	0.10	0.10	0.10	0.12	0.06
X_{Mg}^*	0.377	0.399	0.387	0.229	0.184	0.206	0.181	0.067	0.077
X_{Ca}	0.039	0.047	0.042	0.033	0.033	0.034	0.034	0.040	0.020
X_{Mg}	0.395	0.423	0.408	0.251	0.203	0.225	0.199	0.072	0.097

Table 4 Mean composition of isolated biotite (I), biotite in contact with garnet (P) and biotite enclosed in garnet (E). (I) was measured at 20 μm from the edge and (P) at 40–50 μm from their common grain boundary. For the ZAF correction no water was introduced. X_{Mg} = Mg/Mg+Fe, where all Fe was assumed to be bivalent.

	BIOTITE										
	Iv 533	Iv 533	Iv 530	Iv 528	Iv 528	KAW 85	KAW 85	KAW 85	VS 41	VS 41	VS 41
	5 I	4 P	6 I	6 I	5 P	8 I	6 P	2 E	8 I	3 P	2 E
SiO_2	37.39	37.10	35.93	36.00	35.14	36.34	35.79	36.44	39.52	37.77	39.19
TiO_2	5.62	5.64	5.34	4.89	4.54	4.71	3.29	3.16	6.03	5.80	5.40
Al_2O_3	16.16	15.85	17.06	17.97	18.06	19.03	19.18	19.60	15.06	14.66	15.20
FeO	11.68	11.95	15.69	17.47	18.27	16.69	16.30	15.87	9.32	11.15	9.49
MnO	0.00	0.00	0.03	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
MgO	14.63	15.10	12.09	9.66	10.17	10.84	11.36	12.25	17.50	15.87	17.78
CaO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00
Na_2O	0.05	0.03	0.01	0.05	0.02	0.05	0.06	0.10	0.09	0.00	0.14
K_2O	9.83	9.92	9.67	9.51	9.44	9.79	9.85	9.33	10.03	9.93	10.00
Total	95.36	95.59	95.82	95.55	95.64	97.45	95.83	96.76	97.55	95.18	97.20
Formula per 22 oxygens											
Si	5.50	5.46	5.36	5.42	5.31	5.33	5.34	5.35	5.60	5.56	5.59
Al	2.50	2.54	2.64	2.58	2.69	2.67	2.66	2.65	2.40	2.44	2.41
Al	0.29	0.21	0.36	0.61	0.53	0.62	0.78	0.75	0.11	0.10	0.14
Ti	0.62	0.62	0.60	0.55	0.52	0.52	0.37	0.35	0.64	0.64	0.58
Fe	1.43	1.47	1.96	2.20	2.31	2.05	2.04	1.95	1.11	1.37	1.13
Mn	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Mg	3.21	3.31	2.69	2.17	2.29	2.38	2.53	2.68	3.70	3.48	3.78
Ca	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na	0.01	0.01	0.00	0.01	0.00	0.02	0.02	0.02	0.02	0.00	0.04
K	1.85	1.86	1.84	1.83	1.82	1.83	1.88	1.75	1.81	1.86	1.82
X_{Mg}	0.692	0.692	0.578	0.496	0.498	0.537	0.554	0.579	0.769	0.717	0.770

B I O T I T E

	Iv 490	Iv 490	Iv 490	Iv 485	Iv 485	Iv 485	KAW 508	KAW 508	Iv 45	Iv 45	Iv 458	Iv 458
	6 I	4 P	3 E	6 I	6 P	3 E	6 I	4 P	8 I	4 E	7 I	3 E
SiO2	38.04	36.25	38.54	36.62	35.90	37.20	36.26	36.07	34.46	34.52	35.79	35.63
TiO2	6.80	5.87	5.67	4.90	4.46	3.42	4.33	3.46	3.86	3.59	3.12	3.32
Al2O3	16.35	16.46	16.46	18.62	18.46	19.22	18.81	19.21	19.61	19.24	19.79	19.34
FeO	10.36	10.19	9.24	17.43	16.67	15.08	16.54	16.70	24.96	24.81	21.75	20.37
MnO	0.00	0.00	0.00	0.00	0.00	0.00	0.05	0.02	0.00	0.00	0.19	0.31
MgO	15.95	16.32	17.63	10.32	10.33	12.50	9.98	10.58	5.01	5.37	6.81	8.18
CaO	0.00	0.00	0.00	0.01	0.00	0.05	0.00	0.02	0.04	0.05	0.02	0.02
Na2O	0.06	0.03	0.05	0.07	0.00	0.12	0.02	0.00	0.09	0.13	0.29	0.24
K2O	9.54	9.11	9.95	9.87	9.67	9.50	9.91	9.88	9.13	9.20	9.51	9.36
Total	97.10	94.23	97.54	97.84	95.49	97.09	95.90	95.94	97.16	96.91	97.27	96.77
Formula per 22 oxygens												
Si	5.46	5.35	5.47	5.37	5.38	5.42	5.41	5.38	5.28	5.30	5.39	5.36
Al	2.54	2.65	2.53	2.63	2.62	2.58	2.59	2.62	2.72	2.70	2.61	2.64
Al	0.20	0.21	0.22	0.59	0.64	0.72	0.72	0.76	0.82	0.78	0.90	0.79
Ti	0.74	0.65	0.60	0.54	0.50	0.37	0.49	0.39	0.37	0.41	0.35	0.38
Fe	1.22	1.26	1.10	2.14	2.09	1.84	2.06	2.08	3.20	3.19	2.74	2.56
Mn	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.02	0.04
Mg	3.38	3.59	3.73	2.26	2.30	2.72	2.22	2.36	1.15	1.23	1.53	1.84
Ca	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.01	0.01	0.00	0.00
Na	0.02	0.01	0.01	0.02	0.00	0.03	0.01	0.00	0.03	0.04	0.08	0.07
K	1.76	1.71	1.80	1.85	1.85	1.77	1.89	1.88	1.78	1.80	1.83	1.80
X_{Mg}	0.735	0.740	0.772	0.514	0.524	0.596	0.519	0.532	0.264	0.278	0.358	0.418