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Oligocene, Permian and Panafrikan zircon ages from rocks of the Balmuccia Peridotite and of the Lower Layered Group in the Ivrea Zone*

by Dieter Gebauer^{1,2}, Rolf Schmid¹, Albrecht von Quadt¹ and Peter Ulmer¹

Abstract

U–Pb dating (SHRIMP) of magmatic zircons from a meta-gabbrodiorite dike within the Balmuccia peridotite yields an Oligocene age (31.1 ± 1.5 Ma). Similarly, a meta-gabbrodiorite within the Lower Layered Group (LLG), adjacent to the Balmuccia peridotite, yields an age of 26.0 ± 0.8 Ma. Based on petrological data and microstructural observations both rocks must have been emplaced into, and equilibrated within, deep crustal levels. Sm–Nd whole-rock data of these two rocks plot on the 600 Ma isochron of PIN and SILLS (1986) established for similar rocks of the LLG. This isochron must thus be reinterpreted as an inherited mantle mixing line possibly resulting from Panafrikan mantle melting. The initial ϵ_{Nd} -values of +7.7 and +11.1 as well as the geochemical data suggest a heterogeneous depleted mantle source for the two gabbroic rocks, possibly created by ESE subduction of Mesozoic oceanic crust (Valais trough?). Such a model is in line with S-directed, Oligocene subduction of similar oceanic crust of the Valais trough inferred for the Lepontine Alps (GEBAUER et al., 1992, this volume).

A Panafrikan magmatic event at 670 ± 36 Ma, probably related to mantle melting, as well as a Permian high T-event ($265 \pm 4/-5$ Ma), interpreted to be the result of underplating during successive Permo-Triassic rifting episodes, was detected in zircons of a meta-pyroxenite layer from the LLG.

Keywords: U–Pb zircon dating, Oligocene, continental crust, dikes, peridotite, Ivrea Zone, Southern Alps.

Geological Setting

The Ivrea Zone, situated in the westernmost part of the Southern Alps – in between the Strona Ceneri Zone metamorphosed in pre-Alpine time and the Alpine Sesia Zone involved in the Alpine orogenesis – reveals a tilted cross section through the lower and middle continental crust. It exposes very high proportions of mafic-ultramafic rock sequences which have a protracted evolution with an exceptionally large amount of documented geological events that find no counterpart in any other part of Hercynian or Alpine Europe.

The rocks from which zircons were extracted were sampled in the Sesia Valley, from a locality above Bottorno at a height of 700 m where the

rocks of the Lower Layered Group (LLG) are in igneous contact with the Balmuccia Peridotite. The LLG forms one of the deepest units of the Ivrea zone, reaching a thickness of up to 600 m (RIVALENTI et al., 1981). It consists mainly of pyroxene-rich gabbroic-dioritic rocks and intercalated layers or lenses of pyroxenites, peridotites and granulite-facies metasediments.

Oligocene zircon data (SHRIMP) from two meta-gabbrodiorites

Magmatic, multiply zoned zircons were extracted from two meta-gabbrodiorites rich in clinopyroxene and orthopyroxene, containing plagioclase of

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An 39–50. One of them (ISO 1) occurs as a 1 m thick dike within the peridotite, 5 m from its contact to the LLG. The other one (ISO 2) originates from the LLG, in a distance of 100 m to ISO 1. In the literature, the dike has been named "gabbroic pod" or gabbro-dike whereas the rocks of the LLG have been referred to as gabbros or meta-gabbros (e.g. RIVALENTI et al., 1981). They are differently LREE-depleted and their bulk rock compositions support their cumulate character (PIN and SILLS, 1986).

Both studied rocks display layering and a planar to linear fabric subparallel to the layering. The dike (ISO 1) strikes E–W, approximately perpendicular to the elongation axis of the peridotite, and dips with 60° towards the S. Its foliation runs parallel to the walls of the dike. Between the dike and the peridotite a pyroxenite layer of a few mm thickness, formed by contact-metasomatism, is typically present. The dike belongs to the youngest of the various dike suites within the peridotite (e.g. RIVALENTI et al., 1981), crosscutting both the Cr-diopside as well as the Al-augite dikes. The foliation of ISO 2 follows the general NNW–SSE trend of the main foliation of the LLG, and the rock bears no special features which would distinguish it from the other gabbroic-dioritic rocks of this group. It is interesting to note that at their igneous contact with the Balmuccia peridotite the rocks of the LLG form a similar pyroxenitic rim of ~ 1 cm thickness towards the peridotite as ISO 1 does.

In both rocks a more or less pronounced deformation of a cumulate to granoblastic structure is observed on the micro-scale, with plastic deformation of plagioclase and pyroxene grains involving twin gliding, bending and sometimes dynamic recrystallization of the plagioclase. This is usually assumed to indicate temperatures of > 500 °C during deformation (e.g. KIRBY, 1983).

In applying various geothermometers and -barometers to microprobe data of co-existing minerals, one has to conclude that the rocks were completely equilibrated at around 750–800 °C and ≥ 7 kbar (see Appendix).

Ionprobe U–Pb dating (SHRIMP) carried out at the Australian National University in Canberra reveals that the two zircon populations have formed during the Oligocene, at 31.1 ± 1.5 Ma (95% confidence limit) in ISO 1, and at 26.0 ± 0.8 Ma in ISO 2. When viewed at by means of cathodoluminescence, the measured zircons display the same features as they are typical for zircons of magmatic rocks. We therefore interpret the SHRIMP data as intrusion ages of the two meta-gabbrodiorites. The two rocks are not cogenetic as the age difference is about 5 Ma. The

different initial ϵ_{Nd} -values of + 7.7 and + 11.1 support this observation.

Reinterpretation of Sm–Nd whole-rock data from the LLG

Similar rock types including also pyroxenites and one peridotite from the LLG have previously been analysed for Sm–Nd whole-rock by PIN and SILLS (1986) who obtained an isochron age of 596 ± 35 Ma with an initial ϵ_{Nd} -value of 1.7 ± 1.1 . The two samples dated for zircon also fall onto this isochron. Although the authors were cautious about the age significance of their ca. 600 Ma isochron they nevertheless favoured the interpretation that this age represents the magmatic formation of the LLG which must thus have formed from a time-integrated, slightly depleted mantle source. The emplacement of the rocks of the LLG into their present position was thought to have occurred at a later stage together with the residual Balmuccia peridotite. If the zircon ages are assigned to the intrusion of at least part of the rocks of the LLG, this interpretation cannot be upheld any more. Instead, it can be shown that the 600 Ma isochron reflects a mixing line, possibly revealing an inherited age from previous mantle melting. Unfortunately, Sm–Nd whole-rock isochrons can give too old ages as has been shown for example by GÖPEL (1981) on Permian basalts of the Saar-Nahe basin (SW-Germany) yielding an isochron age of 1580 Ma. Similarly, too old Sm–Nd isochrons were obtained, for example, by DE-PAOLO (1981) or CHAUVEL et al. (1985), and by STILLE and BULETTI (1987) who obtained an isochron age of 341 Ma for the Permian Lugano volcanics (Southern Alps) exceeding their true age by ca. 30%.

Are the meta-gabbrodiorites deep seated intrusions?

Although originally unexpected and surprising, the Oligocene ages of the two gabbrodiorites fit very well within the age pattern of other magmatic rocks along the Periadriatic Line. For its western segment Oligocene magmatic rocks are known to exist within the Sesia zone (plutonic rocks of Biella and Traversella), at the top of the Sesia basement (andesites) and even within the Ivrea zone (Miagliano, 31 Ma; CARRARO and FERRARA, 1968). It surprises, however, that both the geobarometric/geothermometric data as well as the plastic deformation behaviour of the meta-gabbrodiorites indicate that they were emplaced

at a great depth. A minimum estimate of this depth is given by the temperature from which such rocks start to deform by crystal plasticity. Assuming that this temperature is 500 °C and that the thermal gradient was less than 50 °C/km the minimum depth would be 10 km. It is, however, more likely that the calculated re-equilibration conditions (750–800 °C, ≥ 7 kbar) reflect the ambient temperatures and pressures of the rocks during their emplacement. These indicate depths of ≥ 20 km. Both estimates imply that the Balmuccia peridotite and the LLG represent an isolated slice, or part of such a slice within the Ivrea Zone, exhumed in Late to post-Oligocene time. The borders of this slice should be discernable in the field and in fact various shear zones are present to the W of the Balmuccia peridotite as well as to the E of the LLG (BRODIE et al., 1992, and personal communication). It is, however, not yet established if they qualify for the later proposed exhumation processes.

Panafrican and Permian zircon data (SHRIMP) from a meta-pyroxenite of the LLG

Only very few zircons could be separated from a coarse-grained, 5 m thick meta-pyroxenite layer (ISO 3) occurring within the LLG, only a few meters E of ISO 2. Cathodoluminescence studies show that there exist both magmatic as well as partially or completely altered metamorphic domains within one crystal. As only two small grains could be analyzed by ionprobe, only 4 spots, 2 each, could be dated and the corresponding data fall on a discordia trajectory from 670 ± 36 Ma to $265 +4/-5$ Ma. In fact, only one data point containing both magmatic and metamorphic domains is about 77% discordant. Two data points obtained from metamorphic domains are concordant at 265 Ma and one data point from a magmatic domain is concordant at 670 Ma. As neither the age relation of the meta-pyroxenite to the enclosing pyroxene-rich gabbrodiorites nor the origin of the zircons, inherited or co-magmatic, are clearly established, the obtained ages cannot be assigned safely to the history of the zircon's host- or mother rock. However, due to the difficulties in establishing the relative intrusion sequence of the coarse-grained pyroxenite and the meta-gabbrodiorites of the LLG (ISO 2) in the field, a more or less simultaneous formation of the two rocks at ca. 26 Ma seems most probable. This would imply that the dated zircons are not co-magmatic but xenocrysts from either the source or country rocks. Irrespective of the true age of the protolith of the meta-pyroxenite, however, the age data

suggest partial mantle melting at the end of the Precambrian and Permian HT-metamorphic overprinting, probably under granulite facies conditions. The latter deduction is based on the observation that metamorphically induced rearrangement of trace and minor elements in zircon is so far only observed under granulite- and eclogite facies conditions (GEBAUER et al., 1988 and unpublished data).

A Permian event has been inferred to by a number of previous age data. Unfortunately, these ages on Permian to Permo-Carboniferous metamorphic overprinting within the Ivrea zone are not yet constrained in both time and/or in regional distribution. Based on two zircon suite data arrays as well as on monazite data from the same samples, KÖPPEL (1974) interpreted the regional granulite to upper amphibolite facies metamorphism to have occurred before 295 ± 5 Ma. In contrast, TEUFEL and SCHÄRER (1989) believe that they have dated the amphibolite to granulite facies event at 280 Ma. This is based on U–Pb monazite ages on single grains and small size-fractions and agrees with the best estimate on the age of the main gabbro-diorite body which, according to conventional zircon data of PIN (1986), is $285 +7/-5$ Ma. As the corresponding rocks can be interpreted to have crystallized magmatically at ambient conditions of approximately 700° and 6 kbar (SILLS, 1984), this age would also be related to metamorphism. An imprecise Sm–Nd garnet-whole-rock age of 271 ± 22 Ma (VOSHAGE et al., 1987) can be taken as support of this.

On the other hand, there are indications for granulite facies metamorphism at around 230 Ma to 250 Ma which are based on Sm–Nd mineral isochrons on acid granulites and a garnet-bearing gabbro (POLVÉ, 1983, or VOSHAGE et al., 1987).

The $265 +4/-5$ Ma age of metamorphic redistribution of trace elements in the zircons from the meta-pyroxenite of the LLG is thus significantly different from previous hypotheses on the, or better on one age of granulite facies transformation of various rocks within the Ivrea zone. We do not think that this new date is in contradiction to the previous hypotheses on the timing of the lower crustal metamorphic overprint. Successive magmatic underplating during rifting episodes could be responsible for such a long lasting and pulsed thermal evolution. In fact, we now know (STÄHLE et al., 1990; GEBAUER, 1992b, in press) that magmatic activities in the Ivrea zone extended well into the Triassic.

The 670 ± 36 Ma age of magmatic crystallization of zircon extracted from the meta-pyroxenite reflects the as yet oldest event in the history of the

mafic and ultramafic Ivrea rocks, respectively their mantle source(s). It conforms very well with a mantle heating event as detected in the garnet-peridotite of Alpe Arami in the southern steep belt of the Central Alps (GEBAUER et al., 1991 and 1992, this volume). Similarly, a Panafrican age for partial melting of mantle material could be detected from a peridotite of the Pyrenees at Lac de Lherz, the type locality of lherzolite (GEBAUER, unpubl.).

Although Panafrican continental crust is only rarely exposed throughout the European Hercynides it is strongly represented as recycled material in either post-Panafrican (meta)sediments or granitoids. This applies to metasediments of the Ceneri Zone (GRÜNENFELDER et al., 1984), to metasediments and orthogneisses of the Gotthard Massif (GEBAUER and QUADT, 1991 and GEBAUER, unpubl.), to recent river sand of the Po-delta (GEBAUER, 1992a, in press) as well as to a series of further sediments, metasediments and granitoids from the non-Alpine part of the European Hercynides (GEBAUER et al., 1989). It supports the conclusions that about 15% of the Precambrian continental crust of Hercynian Europe has been extracted from the mantle during the Panafrican polycyclic orogeny (GEBAUER, 1992b, in press). Thus, the 670 Ma magmatic event detected in the Ivrea Zone might well be related to this widespread production of continental crust.

Geodynamic implications

The unexpected intrusion of Oligocene mafic rocks into deep crustal levels argue for Alpine granulite facies conditions in the vicinity of the dated rocks. Thus, the Ivrea Zone seems to be a segment of the lower and middle continental crust which suffered granulite facies overprinting as early as in the Ordovician (HUNZIKER and ZINGG, 1980) and as late as in the Oligocene. Permo-Carboniferous to Permo-Triassic lower crustal re-equilibration of various isotopic systems is suggested here to be a consequence of successive magmatic underplating during quasi-continuous rifting episodes of the continental crust overthickened at the end of the Hercynian (Carboniferous) collision orogeny.

The time integrated depleted mantle source as well as the present day depleted nature of the two meta-gabbrodiorites argues for a derivation of the melts from a suboceanic mantle source. At this stage of investigation we feel that subducted peridotitic mantle of Mesozoic oceanic crust is the best candidate for the source of the two Oligocene rocks. Such a scenario would imply ESE

subduction of the Mesozoic oceanic crust (Valais trough?) below the earlier subducted and uplifted mainly continental Sesia Zone, different Oligocene melting events of its peridotitic parts, ascent of the melts into its present country-rocks and rapid exhumation.

Considering possible mechanisms for the juxtaposition of a deep crustal Ivrea segment with cooler Ivrea rocks as well as with the adjacent cooler Sesia crust, the following speculations might serve for a working hypothesis. Following S.M. SCHMID et al. (1989), the Ivrea Zone was thrust over the relatively cool Sesia Zone after the Oligocene. Such a thrusting was suggested by the authors in order to explain the structural position of Oligocene andesites between the erosional surface of the Sesia zone and their mylonite belt 1 formed during pre-Oligocene underthrusting of the Sesia Zone underneath the Ivrea Zone. Elaborating on figure 5c of S.M. SCHMID et al. (1989), we propose the existence of thrust planes within the Ivrea Zone in order to exhume lower crust over a vertical distance of more than 10 km. Subsequent extensional collapse of the Ivrea Zone and probably also of the Strona-Ceneri Zone along normal faults is then the most probable mechanism to bring successively cooler rocks adjacent to the Balmuccia-LLG slice. Part of the shear zones in the surroundings of the Balmuccia peridotite and elsewhere within the Ivrea Zone might be related to such post-Oligocene tectonic movements. Further differential uplift of the Balmuccia area is likely to have occurred during backthrusting and verticalization of both the Sesia and the Ivrea Zone, which is also confirmed by palaeomagnetic data (S.M. SCHMID et al., 1989).

From K-Ar- and Rb-Sr ages it is generally concluded that the Ivrea zone as a whole reached shallow crustal levels already in the Early Mesozoic. Petrological and geochronological data indicate, however, that parts of the Ivrea Zone were still, or again, at about 250–500 °C in the Alpine time (STÄHLE et al., 1986; BÜRGI and KLÖTZLI, 1990), resulting e.g. in partial rejuvenation of biotites (BÜRGI and KLÖTZLI, 1990).

None of the existing mineral ages contradict a Late to post-Oligocene exhumation of the Balmuccia peridotite and the LLG. They all originate from other units of the Ivrea zone, from localities which are more than 7 km apart, with the exception of Sm-Nd data of plagioclase and garnet from a gabbro of the Upper Layered Group which, along with the whole rock point, plot on a 248 Ma isochron (VOSHAGE et al., 1987). From chemical analyses, PIN and SILLS (1986) emphasized, however, that the gabbros in that area should better be grouped with those of the Main

Gabbro. As the corresponding rock sample originates from a locality which is in a 600 m distance to the E of the LLG, it is not surprising that the analyzed garnet was not reset to an Oligocene age (e.g. COHEN et al., 1988). This holds all the more if the above suggested exhumation model is correct as it implies that cooler and older rocks were brought into contact with hotter and younger rocks during extensional faulting caused by gravitational collapse of the overthickened crust.

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APPENDIX: Mineral-chemistry and thermo-barometry

MINERAL CHEMISTRY

All important mineral constituents of samples ISO 1, ISO 2, and ISO 3 have been analyzed with a Cameca SX50 Electron Microprobe, equipped with 5 wavelength dispersive spectrometers. The data were on-line corrected with PAP correction procedure. Acquisition parameters were 15 KeV acceleration voltage and 20 nA sample current. Natural and synthetic oxides were used as standards.

Representative analyses of analyzed phases are given in table 1.

ISO 1 contains unzoned cpx and opx, both containing exsolution lamellae; plagioclase with inverse zoning ranging from An₃₉ in the core to An₄₄ at the rim; slightly zoned hercynitic spinel ($x_{Mg} = 0.614$ in the core, 0.585 at the rim); and late interstitial paragonitic to kaersutitic amphibole.

ISO 2 consists of strongly exsolved cpx and opx, plagioclase, ilmenite, spinel and late garnet. Figure 1 shows a profile across a cpx grain from its contact with a late garnet to a neighbouring ilmenite grain. The profile displays strongly decreasing Na which is correlated with a decrease in Al^{VI} and can therefore be assigned to a decrease in jadeitic component. Decreasing jadeitic component is clearly associated with decreasing pressure, whereas decreasing Ti is correlated with decreasing Al^{IV} (not shown) and is predominantly temperature sensitive. Ca and Na decrease slightly from core to rim, indicating slight temperature decrease. Plagioclase shows inverse zoning from An₄₁ in the core to An₄₉ at the rim; ilmenites contain a high amount of Mn and are therefore probably late or secondary; small spinel inclusions in garnet or ilmenite are hercynitic spinels.

ISO 3 contains cpx, opx, amphibole and spinel. No plagioclase has been detected. Cpx and opx both show exsolution lamellae. Cpx shows zoning similar to ISO 2. Na content in cpx is lower than in samples ISO 1 and ISO 2. Spinels and amphiboles are similar to ISO 1, except x_{Mg} is lower for all Fe-Mg phases.

GEOTHERMOMETRY

Table 2 shows the results of thermo-barometric calculations using pyroxene, garnet-pyroxene and garnet-pyroxene-plagioclase equilibria.

Cpx-opx thermometers have been applied to both coexisting opx-cpx grains as well as to exsolution lamellae inside single grains. In both cases

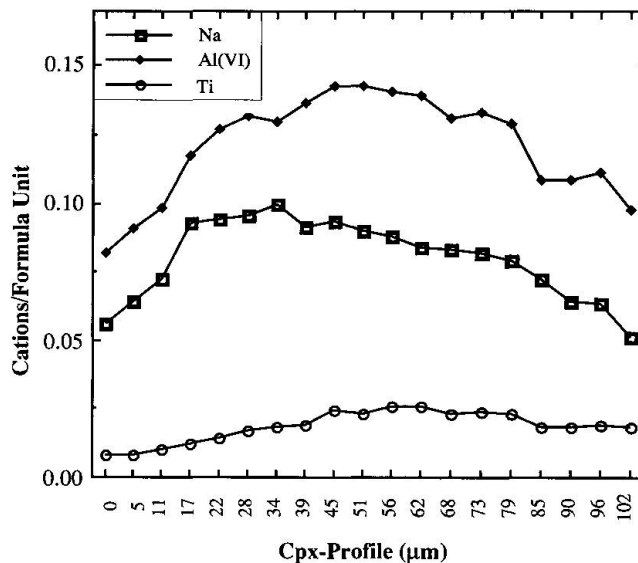


Fig. 1 Profile through a cpx grain from sample ISO 2. The profile ranges from garnet (to the left of the profile) across cpx to an ilmenite grain to the right of the profile. Bottom numbers indicate spacing between data points in μm . The ordinate gives cations calculated on the basis of 12 oxygens and $\text{Fe}^{3+} = \text{acmite component}$. $\text{Al}^{\text{VI}} = \text{Al}^{\text{tot}} - \text{Al}^{\text{IV}}$ and $\text{Al}^{\text{IV}} = 2 - \text{Si}$. For discussion see text.

similar temperatures of 750–800 °C are calculated using the most reliable pyroxene thermometers of WELLS (1977), HERVIG and SMITH (1979) and BREY and KÖHLER (1990). Temperatures calculated from most of the other thermometers represent maximum values, because the presence of Na in cpx is neglected (for details see BREY and KÖHLER, 1990). As no significant temperature differences between exsolution lamellae and inter-grain cpx-opx pairs can be observed, the calculated temperatures most probably reflect subsolidus equilibration temperatures close to the ambient conditions of the intrusives.

In ISO 2 cpx-garnet-temperatures range around 750 °C for adjacent garnet-cpx-rims and around 700 °C for core compositions of neighbouring garnet and cpx grains; this apparent inconsistency is due to inverse zoning of Mg in cpx, related to decreasing Tschermak-molecule in cpx from core to rim.

GEOBAROMETRY

The opx-garnet-pressures for ISO 2 of ≈ 8 kbar have to be considered with care because no unal-

Tab. 1 Representative Microprobe Analyses of Minerals.

Sample	iso1 cpx1 core	iso1 cpx3 rim	iso1 opx2	iso1 plg6 core	iso1 plg8 rim	iso1 amp1	iso1 spl1	iso2 cpx1 core	iso2 cpx4 rim	iso2 opx7	iso2 plg6 core	iso2 plg4 rim
SiO ₂	49.28	49.14	52.47	58.79	57.60	41.33	0.01	48.96	49.56	52.46	56.95	55.41
TiO ₂	1.66	1.91	0.24	0.04	0.01	4.50	0.02	1.25	1.01	0.16	0.00	0.00
Cr ₂ O ₃	0.08	0.10	0.05	0.00	0.00	0.09	1.12	0.07	0.03	0.00	0.03	0.01
Al ₂ O ₃	8.37	8.81	5.82	26.81	27.56	14.09	63.15	7.85	6.46	2.49	26.37	27.36
Fe ₂ O ₃	0.00	0.00	0.66	0.00	0.10	0.00	2.20	0.00	0.00	2.01	0.52	0.12
FeO	5.83	6.03	13.50	0.01	0.00	9.11	17.86	9.03	7.76	18.55	0.00	0.00
MnO	0.13	0.15	0.29	0.00	0.04	0.08	0.10	0.24	0.22	0.45	0.00	0.03
MgO	12.18	11.69	27.19	0.00	0.00	14.17	15.69	10.87	12.18	24.27	0.00	0.00
CaO	21.11	21.36	0.29	8.15	9.56	11.95	0.00	20.58	21.81	0.48	9.18	10.27
Na ₂ O	1.47	1.68	0.07	7.03	6.51	3.64	0.03	1.52	0.92	0.00	6.70	5.96
K ₂ O	0.01	0.00	0.00	0.02	0.04	0.06	0.00	0.02	0.02	0.00	0.04	0.11
Total	100.11	100.87	100.59	100.84	101.43	99.03	100.17	100.38	99.98	100.88	99.79	99.27
Cations assuming stoichiometry and charge balance*												
Si	1.8125	1.7984	1.8646	2.6058	2.5462	5.9561	0.0002	1.8217	1.8446	1.9143	2.5579	2.5077
Ti	0.0460	0.0526	0.0065	0.0012	0.0004	0.4876	0.0004	0.0350	0.0283	0.0045	0.0000	0.0000
Cr	0.0023	0.0029	0.0013	0.0000	0.0000	0.0107	0.0230	0.0021	0.0010	0.0000	0.0011	0.0004
Al	0.3627	0.3799	0.2438	1.4006	1.4356	2.3929	1.9341	0.3444	0.2835	0.1071	1.3959	1.4593
Fe ₃	0.0000	0.0000	0.0176	0.0000	0.0035	0.0000	0.0430	0.0000	0.0000	0.0553	0.0177	0.0042
Fe ₂	0.1792	0.1846	0.4011	0.0002	0.0000	1.0986	0.3882	0.2809	0.2416	0.5661	0.0000	0.0000
Mn	0.0042	0.0045	0.0088	0.0000	0.0015	0.0098	0.0022	0.0076	0.0068	0.0138	0.0000	0.0011
Mg	0.6677	0.6379	1.4402	0.0000	0.0000	3.0443	0.6075	0.6027	0.6757	1.3199	0.0000	0.0000
Ca	0.8318	0.8374	0.0112	0.3870	0.4529	1.8460	0.0000	0.8205	0.8695	0.0189	0.4418	0.4980
Na	0.1050	0.1193	0.0049	0.6038	0.5575	1.0163	0.0014	0.1094	0.0665	0.0000	0.5835	0.5230
K	0.0005	0.0000	0.0000	0.0013	0.0024	0.0107	0.0000	0.0008	0.0009	0.0000	0.0023	0.0064
xMg	0.788	0.776	0.786			0.735	0.610	0.682	0.737	0.700		
An				39.0	44.7						43.5	48.5
Sample	iso2 gar2	iso2 spl2 core	iso2 spl1 rim	iso2 ilm5	iso3 cpx7 core	iso3 cpx1 rim	iso3 opx9	iso3 amp2	iso3 ilm1			
SiO ₂	38.57	0.03	0.01	0.03	49.05	47.85	50.87	40.21	0.01			
TiO ₂	0.04	0.72	0.78	52.69	1.13	1.35	0.10	4.83	0.10			
Cr ₂ O ₃	0.04	0.20	0.22	0.05	0.13	0.14	0.11	0.20	2.58			
Al ₂ O ₃	21.80	60.70	59.77	0.00	7.74	7.57	4.31	14.41	55.51			
Fe ₂ O ₃	2.72	0.74	0.08	0.22	0.00	0.00	2.42	0.00	7.09			
FeO	20.05	24.31	26.46	44.70	7.59	8.06	18.73	11.18	23.21			
MnO	1.22	0.13	0.18	2.25	0.19	0.15	0.37	0.07	0.08			
MgO	9.36	11.09	9.40	0.15	11.81	11.49	23.11	11.72	11.36			
CaO	6.39	0.01	0.00	0.02	22.01	22.22	0.48	11.59	0.00			
Na ₂ O	0.00	0.06	0.05	0.03	1.02	0.73	0.00	3.08	0.00			
K ₂ O	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.46	0.00			
Total	100.19	97.98	96.95	100.14	100.67	99.58	100.49	97.74	99.94			
Cations assuming stoichiometry and charge balance*												
Si	2.9395	0.0008	0.0003	0.0008	1.8131	1.7966	1.8690	5.9493	0.0004			
Ti	0.0023	0.0148	0.0163	0.9974	0.0314	0.0382	0.0028	0.5373	0.0022			
Cr	0.0024	0.0043	0.0048	0.0010	0.0037	0.0041	0.0032	0.0230	0.0561			
Al	1.9581	1.9525	1.9629	0.0000	0.3372	0.3351	0.1866	2.5124	1.7956			
Fe ₃	0.1560	0.0151	0.0017	0.0041	0.0000	0.0000	0.0668	0.0000	0.1463			
Fe ₂	1.2780	0.5548	0.6166	0.9411	0.2347	0.2532	0.5755	1.3837	0.5328			
Mn	0.0788	0.0030	0.0042	0.0480	0.0061	0.0048	0.0116	0.0089	0.0019			
Mg	1.0632	0.4511	0.3904	0.0056	0.6506	0.6429	1.2656	2.5854	0.4646			
Ca	0.5218	0.0003	0.0000	0.0005	0.8716	0.8939	0.0191	1.8380	0.0000			
Na	0.0000	0.0032	0.0027	0.0015	0.0731	0.0531	0.0000	0.8833	0.0000			
K	0.0000	0.0000	0.0000	0.0000	0.0000	0.0004	0.0000	0.0863	0.0000			
xMg	0.454	0.448	0.388	0.006	0.735	0.717	0.687	0.651	0.466			

* Cations are calculated on the basis of stoichiometry and charge balance, except for cpx, where the model of Cawthorn and Collerson has been applied (12 charges, Fe³⁺ exclusively assigned to the acmite-component). The following normalization values have been used: opx: 4 cations - 12 charges, plagioclase: 5 cations - 16 charges, garnet: 8 cations - 24 charges, spinel: 3 cations - 8 charges, ilmenite: 2 cations - 6 charges, amphibole: Σ cations = 13+Ca+Na+K - 46 charges.
 $xMg = Mg / (Mg + Fe^{2+})$; $An = Ca / (Ca + Na + K)$

Tab. 2 Geothermobarometry.

Thermobarometer	ISO1 σ	ISO2 σ cum	ISO2 σ prof	ISO2 σ gar	ISO3 σ
Cpx - Opx Thermometer (10kbar)					
Wells (1977)	792 \pm 23	813 \pm 22	814 \pm 32	815 \pm 19	805 \pm 16
Wood & Banno(1973)	850 \pm 19	834 \pm 16	832 \pm 24	836 \pm 14	823 \pm 12
Nehru & Wyllie (1974)	785 \pm 33	864 \pm 36	871 \pm 52	867 \pm 31	865 \pm 25
Nickel & Brey (1984)	798 \pm 26	897 \pm 13	910 \pm 29	914 \pm 11	897 \pm 25
Nickel et al. (1985)	838 \pm 7	820 \pm 32	853 \pm 25	789 \pm 17	778 \pm 16
Hervig & Smith (1979) (Na)	795 \pm 54	777 \pm 50	762 \pm 54	795 \pm 33	798 \pm 38
Brey & Köhler (1990) (solvus)	750 \pm 30	773 \pm 18	790 \pm 43	722 \pm 15	796 \pm 30
Brey & Köhler (1990) (Na)	747 \pm 80	720 \pm 50	700 \pm 82	748 \pm 42	753 \pm 57
Cpx - Garnet Thermometer (10kbar)					
Mori & Green (1978)			705 \pm 20	768 \pm 30	
Raheim & Green (1974)			720 \pm 40	763 \pm 21	
Ellis & Green (1979)			760 \pm 17	803 \pm 24	
Brey & Köhler (1990)			650 \pm 17	700 \pm 25	
Opx - Garnet Thermometer (10kbar)					
Harley (1984)				780 \pm 31	
Brey & Köhler (1990)				738 \pm 38	
Opx - Garnet Barometer (750°C)					
Wood & Banno (1973, 74)				8.7 \pm 0.9	
Lane & Ganguly (1980)				7.5 \pm 0.9	
Harley (1984)				8.2 \pm 0.5	
Opx - Garnet -Plagioclase Barometer (750°C)					
Newton & Perkins (1982)				9.8 \pm 0.2	
Cpx - Garnet -Plagioclase Barometer (750°C)					
Newton & Perkins (1982)				7.2 - 8.6	

All temperatures are calculated using a fixed pressure of 10 kbar. Pressure dependence for the cpx-opx solvus thermometer is less than 3 °C/kbar for the pyroxenes analyzed (BREY and KÖHLER, 1990). The cpx-garnet thermometer has a pressure dependence of \approx 4 °C/kbar, the opx-garnet thermometer \approx 7 °C/kbar. The barometers are all calculated at a temperature of 750 °C; the pyroxene-garnet-plagioclase barometers show a temperature dependence of \approx 12 bars/°C.

Temperatures and pressures are all mean values of at least 5–10 individual pairs, the σ -Errors indicated in the table are the standard deviations of the averaged temperatures (or pressures) for the individual pairs in one rock (or set for ISO 2).

The 3 columns for ISO 2 represent: cum – major cumulate phases pyroxenes, for thermometry the small rims, having low Na and Al contents where avoided; prof – profile from garnet (core) to cpx (core), the larger errors are due to the effect of mixing core and rim analyses; gar – directly coexisting garnet – cpx pairs (or triplets including plagioclase).

tered opx grains could be found in immediate contact with garnet. However, the few grains which are unaltered prove to be unzoned. Pressures calculated from the assemblage garnet, cpx and plagioclase range at 7.2–8.6 kbar and those for garnet, opx and plagioclase around 9.8 kbar. Both pressure estimates have to be regarded as maximum ones, because quartz required to balance the reaction is not present in the rock. The activity of quartz is therefore less than 1 and the calculated K_d represents a maximum value. The lowest value of 7.2 kbar for cpx-garnet-plagioclase has been obtained from rim analyses, the highest value of 8.6 kbar from core analyses of these three phases. The highest value probably represents the maximum pressure of garnet formation, because the zoning of all 3 phases reflects pressure decrease (increasing Ca in plagioclase, decreasing Na in cpx and slightly decreasing pyrope and grossular in garnet). A pressure of \geq

7 kbar for the equilibration of ISO 2 is also inferred from the Na-content of cpx. Experiments performed in a nearly identical bulk composition (ULMER, unpublished data) indicate > 7 kbar for Na-contents of up 1.6 wt% in cpx at temperatures of 750–800 °C.

In conclusion, the three samples ISO 1, 2, and 3 have equilibrated under conditions of 750–800 °C and pressures of ≥ 7 kbar. Subsequent cooling to ≈ 700 –750 °C lead to the formation of retrograde garnet. The very flat slopes of possible garnet-forming reactions (either spinel + plagioclase + opx = garnet + cpx, or plagioclase + cpx = garnet + qtz) indicate nearly isobaric cooling in the first step (formation of garnet) followed by decompression at 700–750 °C reflected by inversely zoned plagioclase, and decrease of jadeitic component in cpx.

(References Appendix see p. 117/118)