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Radiometric age, thermobarometry and mode of emplacement of the Totalp peridotite in the Eastern Swiss Alps*

by *Tj. Peters*[1] and *A. Stettler*[2]

Abstract:

Using the Ar^{39}/Ar^{40} technique, primary phlogopite from two pyroxene pyroxenites yielded a retention age of 160 ± 8 M.a. This age is interpreted as the time of upwelling of the mantle and coincides with the rifting phase producing the coarse breccias of the lower Austroalpine nappes. From the composition of clino- and orthopyroxenes in the peridotites and pyroxenites, using different geothermometers and geobarometers an equilibrium temperature of 830 to 975 °C and pressure of 10 ± 3 kb was estimated. Bulk chemical analyses of the phlogopite-bearing garnet and spinel pyroxenites are consistent with melt compositions calculated from lherzolite.

The Totalp peridotite is interpreted as fragment of a subcontinental mantle being detached as part of a small, short-lived ocean in a rift system between the Eurasian and African continental plates.

Keywords: Thermobarometry, mantle upwelling, Ar^{39}/Ar^{40} method, peridotite, Swiss Alps.

Introduction

Although ultramafic bodies – peridotites and serpentinites with associated rocks in orogenic mountain belts – have received much scientific attention within the scope of global tectonics, origin, time and mode of emplacement are still not completely clear. Many investigators like DAL PIAZ (1974) for the western Alps, CORTESOGNO et al. (1970) for the Apennines and DIETRICH (1975) for the eastern Alps interpret these ultramafic masses and associated gabbros and basic volcanics as "model" ophiolite sequence. Hereby, the term ophiolite is used in the sense of the Penrose Conference 1972. Ophiolite formation is related to mid-oceanic ridges. However, it should be kept in mind that undisturbed model sequences, like those of the Oman mountains (ALLEMANN and PETERS, 1972) are very scarce. In most cases "ophiolite" sequences have to be reconstructed from isolated members of the ophiolite series. Even if the basic volcanics can be dated stratigraphically, it would be speculative to apply this age to the ultrabasic rocks normally being in tectonic contact with the volcanics. Furthermore, in most orogenic belts regional metamorphism has obliterated possible evidences of their original age.

Because of chemical and mineralogical reasons, the use of isotopic age determinations is limited. Minerals suited for application of absolute age determinations like phlogopite, hornblende and zircon, if present at all, occur only in very minor quantities. However, for the Ar^{39}/Ar^{40}-technique, only a few milligrams are required. This method was applied to a phlogopite concentrate, handpicked from magnetic enrichments of a 30 kg sample of pyroxenite layer in the Totalp ultrabasic mass (PETERS, 1963).

During Alpine metamorphism, the Totalp region was only affected by a metamorphism in prehnite-pumpellyite facies. Towards the

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* Dedicated to Professor Ernst Niggli on the occasion of his 70th birthday.

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south, the ultrabasic rocks underwent higher degrees of metamorphism and lower greenschist facies conditions are reached in the Upper Engadine. The position of the Totalp region favours the assumption that the argon 39/40 isotopic ratios have not been changed during alpine metamorphism.

Geologic Situation

The Totalp serpentinite occurs within the Arosa Zone, tectonically the highest Penninic unit. It consists of superposed sheets, composed of Mesozoic sediments, basic volcanics and scarce Paleozoic crystalline basement rocks. The Arosa Zone can be linked with the Platta Nappe. It can be tracked from the Engadine through the Oberhalbstein, the Davos region and the Vorarlberg to the Allgäu. The facies of the Triassic and Lower Jurassic sediments is similar to that of the Lower Austroalpine units, but thicknesses are more reduced. Marls, interbedded with breccias, are probably age-equivalent with those in the higher tectonic units of Middle Jurassic age. They are overlain by radiolarian cherts and Aptychus limestones, representing Upper Jurassic and part of the Lower Cretaceous. Pillow lavas and meta-hyaloclastites are interbedded and overlie the radiolarian cherts. According to DIETRICH (1970) basic volcanics are also interbedded with gray marls and fine-grained limestones of Aptian-Albian age. In the next higher datable strata (Cenomanian), no evidence of volcanic activity is seen.

According to their stratigraphic position, the basic extrusives cannot be younger than Upper Cretaceous nor can they be older than Upper Jurassic. However, so far only the Lower Cretaceous age of certain volcanic strata is palaeontologically ascertained.

For the gabbros and ultramafic rocks there is only indirect evidence for the time of emplacement. The upper limit is given by the presence of Cr-spinel grains in Cenomanian flysch deposits. The absence of gabbro components in Middle Jurassic sediments indicates a position preventing them from being eroded during this time.

Recently, WEISSERT and BERNOULLI (1984, 1985) have discussed the origin of the ophicalcites that build up a large part of the serpentinites of the Totalp and the Oberhalbstein. Their findings of breccia and sandstone textures in the ophicalcites led them to the conclusion that they were formed through fragmentation of oceanic basement and filling by pelagic sediments. To account for the high temperature minerals, hydrothermal processes are considered, although their own O^{18}-work (WEISSERT and BERNOULLI, op. cit.) did not show any signs for such an activity. This is in accord with the mineral paragenesis of the ophicalcites which are quite different from those in hydrothermally affected serpentinites. Although there is no doubt about the occurrence of fault breccias in the ophicalcites, they cannot explain the origin of the main mass of ophicalcites. Their mineral paragenesis combined with the chaotic textures rather indicate a process of hydraulic fracturing that occurred when the still relatively hot (300 °C) peridotitic mass came into contact with wet sediments, as envisaged earlier (PETERS, 1963).

Petrography

The ultramafic mass of the Totalp consists mainly of spinel-lherzolites (cpx + opx + fo + Cr-spinel) in varying degrees of serpentinisation. Less than 2% consists of peridotite mylonites occurring as 2 to 80 cm wide zones or masses of several m^2 which have been the most resistant to serpentinisation and have thus preserved their original mineralogy. In many outcrops parallel banding due to enrichment of clinopyroxenes in layers is conspicuous. No folding of these bands as described in other areas like Arami (MÖCKEL, 1969) is observed. Pyroxenite dikes of between 5 to 40 cm thickness are parallel to the layering. Among the pyroxenites different ratios of pyroxene to olivine, accompained by changing Cr-content of the spinel-phase are encountered. In Totalp, pyroxenites (cpx + opx + gr + sp) with pyrope-rich garnet are also found, whereas in the Platta region rodingitised gabbros occur. In some pyroxenites pargasitic hornblende and phlogopite were detected.

The peridotites and pyroxenites are medium- to coarse-grained and have an anhedral granular texture. Especially in the unserpentinised pyroxenites a second generation of fine-grained (5–50 micron) pyroxene aggregates occurs between the larger grains (0.5–2 cm). Pargasitic hornblende (10–100 micron), phlogopite (0.1–1 mm flakes), spinel and ore minerals are found in these interstitial aggregates. The larger pyroxene grains often exhibit exsolution lamellae: cpx in opx, opx in cpx, green spinel in cpx and opx and in the garnet-pyroxenites garnet in opx. As mentioned before, the rocks have been affected by serpentinisation of which the phenomena have been described elsewhere (PETERS, 1963).

Mineral and Rock Chemistry

Although a number of rock and mineral analyses were presented by PETERS (1963, 1968), additional rock analyses were performed by X-ray fluorescence. Mineral analyses were determined by microprobe.

ROCK CHEMISTRY

Bulk analyses of major elements of samples in which the minerals were analysed were carried out and listed in Table 1. The composition of the peridotite-mylonites is believed to be representative of the main mass of the original spinel-lherzolite. Strongly serpentinised samples are depleted in elements like Ca that probably have been mobilised and are now present in calcite veins and in rhodingites.

The composition of the spinel-lherzolites is comparable to the estimated average upper mantle compositions of RINGWOOD (1966) and CARTER (1970) which differ mainly in their K_2O-content.

Generally, in the western Mediterranean the peridotites are lherzolitic (NICOLAS and

Tab. 1 Chemical composition of some rocks from the Totalp serpentinite. Analysis of fired (1050°C) samples. In weight %.

	Peridotite Mylonite			Picotite-Pyroxenite	Spinel-Pyroxenite	Pyrope-Pyroxenite	
Mineral Assemblage	ol + opx + cpx + sp			opx+cpx+sp +pa+ph	opx+cpx+sp	opx+cpx+gr+sp	
Sample Nr.	TP 357	TP 49	TP 462	TP 87	TP 340	TP 460	TP 63b
SiO_2	44.42	44.29	43.83	44.64	44.40	45.89	44.76
Al_2O_3	4.16	2.97	3.71	5.11	14.05	12.92	11.13
Fe_2O_3*	8.25	8.99	9.45	7.24	6.16	5.29	5.01
MgO	37.44	41.09	39.69	33.92	22.48	15.204	17.03
CaO	3.82	1.31	2.22	6.70	10.23	18.14	19.48
K_2O	0.03	0.02	0.04	0.02	0.02	0.09	0.03
MnO	0.14	0.11	0.13	0.13	0.12	0.13	0.15
TiO_2	0.20	0.09	0.13	0.23	0.22	0.13	0.20
P_2O_5	0.03	0.03	0.03	0.03	0.03	0.02	0.03
Na_2O	0.25	0.17	0.19	0.28	0.63	0.64	0.60
Cr_2O_3	0.371	0.269	0.326	0.909	0.367	0.142	0.188
NiO	0.304	0.33	0.316	0.254	0.146	0.050	0.137
Total	99.42	99.67	100.06	99.46	98.85	98.65	98.75
Ignition loss	6.99	11.38	3.92	10.10	2.22	1.22	2.02

*Total iron as Fe_2O_3.

Abbreviations: ol = olivine; opx = orthopyroxene; cpx = clinopyroxene; sp = spinel; gr = garnet; ph = phlogopite; pa = pargasite.

JACKSON, 1972), in contrast to the harzburgitic and dunitic peridotites of the eastern Mediterranean and Near East. Either the mantle under these regions has a different primary composition or it has been depleted in calcium by partial melting to different degrees. ROCCI et al. (1980) explain this difference with a normal oceanic sea floor origin for the western Mediterranean ophiolites and a more diversified origin for the eastern Mediterranean ophiolites which are younger.

The chemical composition of the pyroxenites is partly explained by segregation of clinopyroxenes as supported by field evidence, showing cpx-rich layers grading into pyroxenites and partly by the addition of the low melting fraction of lherzolite. The composition is—apart from CaO—similar to the partial melt from anhydrous lherzolite (KUSHIRO, 1972). A very small amount of H_2O, incorporated into phlogopite and pargasite, must have formed part of this low melting fraction. In a variation diagram of CaO, Al_2O_3 and Na_2O against MgO the peridotitic analyses show a clear correlation and the pyroxenites lie in a region that coincides with melt compositions calculated, if sample 357 is taken as a source and the other peridotites as residue. The procedure is similar to the one applied by FREY et al. (1985) for the Ronda peridotite.

MINERAL CHEMISTRY

In a number of samples representing different rock types, the compositions of coexisting minerals were determined. The results are listed in Table 2, in which the analyses represent averages of 5 to 10 neighbouring spots. Olivine grains are homogeneous and show only a very small variation in Fo-content (Fo_{90}–Fo_{89}) in different samples. Apart from exsolution lamellae, the clinopyroxenes as well as the orthopyroxenes are homogeneous and show no zoning but their composition varies from sample to sample. The fine-grained interstitial pyroxenes have the same composition as the areas between the exsolution lamellae in the larger grains. In the spinels the Mg/Fe-ratio is rather constant, the Cr_2O_3-content, however, varies significantly from sample to sample. Also presented in Table 2 are analyses of the phlogopite dated by Ar^{39}/Ar^{40} and of the associated amphibole. The amphibole is pargasitic with an $Al_2Mg_{-1}Si_{-1}$ exchange and an $NaAlSi_{-1}\square$ exchange of about 0.75 on the base of 23 O (anhydrous). The high TiO_2-content of this phlogopite is striking. Besides the normal exchange $AlKSi_{-1}\square$ some additional Si is substituted by Al, indicating the possibility of the substitution 2 $SiMg_{-2}$ AlTi as proposed by ROBERT (1976). This would account for two thirds of the titanium. For the remaining Ti the substitution of 2 Mg by 1 Ti, creating one octahedral vacant site (FORBES and FLOWER, 1974), is envisaged and supported by the relatively small b-parameter. Part of the increased octahedral layer charge due to $Mg_{-1}Ti$ exchange seems to be compensated by a reduction of interlayer cations, as indicated by the analyses.

Geothermometry and Geobarometry

Mineral assemblages and compositions of coexisting minerals were used to estimate temperature and pressure conditions of their formation and equilibrium.

The assemblages fo + opx + cpx + sp are characteristic for the spinel peridotite facies limited towards lower pressures by the plagioclase-lherzolite/spinel-lherzolite reaction and towards higher pressures by the spinel-lherzolite/garnet-lherzolite reaction. The possible P,T-region is further restricted by the assemblage opx + cpx + sp + gt to the high pressure side of the spinel-gabbro/garnet pyroxenite reaction. The experimental data on natural systems (O'HARA et al., 1971) and on the pure MgO-Al_2O_3-SiO_2-system (DANCKWERTH and NEWTON, 1978) as well as thermodynamic calculations (OBATA, 1976) show very little pressure variations with temperatures below 1000 °C. For temperatures of around 900 °C a pressure range of 9 to 12.5 kb from Obata's data (op. cit.) and 11 to 16 kb from those by HERZBERG (1978) is inferred. With respect to the Al-content of the orthopyroxene coexisting with garnet (sample 63b), a pressure of 16 kb is calculated using the geobarometer of HARLEY (1984).

As practically all the Totalp rocks contain two pyroxenes, the pyroxene "solvus" method can be used.

For the equilibrium

$$\underset{opx}{Mg_2Si_2O_6} \rightleftharpoons \underset{cpx}{Mg_2Si_2O_6}$$

Tab. 2 Microprobe analyses of coexisting minerals in serpentinised peridotites (TP 357, 49, 462) and pyroxenites (TP 87, 340, 63b).
n.d. = not detected by probe.

	Clinopyroxenes					Orthopyroxenes					Olivines			Spinels			Pyrope	Phlogopite	Pargasite
	TP 49	TP 357	TP 462	TP 340	TP 63b	TP 49	TP 357	TP 462	TP 340	TP 63b	TP 49	TP 357	TP 462	TP 49	TP 357	TP 462	TP 63b	TP 87	TP 87
SiO_2	52.49	51.44	50.89	51.76	52.39	56.22	55.48	53.75	55.59	55.04	42.11	41.38	40.31	0.00	0.00	0.00	42.39	38.25	42.06
Al_2O_3	6.01	6.42	6.13	7.06	6.28	3.27	2.90	4.36	3.63	3.49	0.00	0.00	0.00	60.54	59.01	58.04	24.54	16.11	14.83
TiO_2	0.68	0.81	0.64	0.41	0.26	0.09	0.13	0.13	0.07	0.04	0.00	0.00	0.00	0.00	0.00	0.03	0.03	4.28	1.94
Cr_2O_3	0.58	0.54	0.60	0.22	0.21	0.21	0.19	0.34	0.09	0.09	0.00	0.00	0.00	6.89	7.38	9.61	0.26	0.77	1.01
FeO	2.23	2.55	2.49	2.55	2.34	6.35	6.61	6.64	6.87	7.26	9.83	9.71	10.14	11.40	11.69	12.26	9.76	3.80	3.53
MgO	14.68	14.58	14.76	14.67	14.89	33.49	33.52	33.02	33.43	33.19	48.94	49.09	49.07	19.67	19.75	19.40	19.41	21.91	16.66
MnO	0.07	0.09	0.11	0.10	0.08	0.13	0.16	0.17	0.17	0.13	0.14	0.14	0.16	0.11	n.d.	0.11	0.37	n.d.	n.d.
CaO	21.77	22.14	21.42	21.91	22.46	0.37	0.28	0.40	0.28	0.32	0.02	0.02	0.02	0.03	n.d.	n.d.	5.02	n.d.	11.83
Na_2O	1.50	1.50	0.69	1.37	1.27	0.01	n.d.	0.01	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.64	2.89
K_2O	0.01	0.10	n.d.	tr	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	9.41	0.66
Total	100.02	100.17	98.73	100.05	100.18	100.14	99.27	98.82	100.13	99.56	101.04	100.34	99.70	98.64	97.83	99.45	101.78	95.17	96.41
	Cations based on 240																	220	230
Si	7.59	7.48	7.49	7.49	7.57	7.74	7.69	7.54	7.68	7.66	6.11	6.05	5.96	0.00	0.00	0.00	5.94	5.44	6.19
Al^{IV}	0.41	0.53	0.51	0.51	0.43	0.26	0.24	0.36	0.32	0.28	0.00	0.00	0.00	0.00	0.00	0.00	0.05	2.56	1.81
Al^{VI}	0.62	0.58	0.55	0.70	0.64	0.27	0.24	0.36	0.27	0.29	0.00	0.00	0.00	11.11	10.96	10.71	4.00	0.13	0.70
Ti	0.07	0.09	0.07	0.09	0.03	0.01	0.02	0.01	0.01	0.04	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.46	0.21
Cr	0.07	0.06	0.07	0.03	0.02	0.02	0.02	0.04	0.01	0.01	0.00	0.00	0.00	0.85	0.92	1.19	0.03	0.09	0.11
Fe	0.27	0.31	0.31	0.31	0.28	0.73	0.77	0.78	0.79	0.85	1.19	1.19	1.25	1.48	1.54	1.61	1.14	0.45	0.42
Mg	3.17	3.16	3.24	3.17	3.21	6.87	7.04	6.90	6.88	6.89	10.58	10.70	10.01	4.56	4.64	4.52	4.05	4.64	3.57
Mn	0.01	0.01	0.01	0.01	0.01	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.01	0.00	0.02	0.04	0.00	0.00
Ca	3.37	3.45	3.38	3.40	3.48	0.06	0.04	0.06	0.04	0.05	0.00	0.00	0.00	0.00	0.00	0.00	0.75	0.00	1.82
Na	0.42	0.42	0.48	0.39	0.35	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.10	0.80
K	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	1.71	0.12

equations for the equilibrium constant from WELLS (1977), MORI (1977) and HERZBERG (1978) were applied to our data using WOOD and BANNO (1973) site assignment (see Table 2). The resulting temperatures vary between 800 and 1050 °C. The values calculated with the equation of WELLS are about 150 °C lower than those obtained from HERZBERG'S equation. With MORI'S equation intermediate temperatures result midway.

If the equilibrium

$$\underset{opx}{CaMgSi_2O_6} \rightleftharpoons \underset{cpx}{CaMgSi_2O_6}$$

is used as suggested by EVANS and TROMMSDORFF (1978) temperatures between 800 and 900 °C are obtained.

OBATA (1976) has shown, that the aluminium-content of orthopyroxenes is only a function of temperature in spinel lherzolites, and this has been confirmed experimentally by DANCKWERTH and NEWTON (1978). With the equilibrium constant for the equilibrium

$$\underset{en}{Mg_2Si_2O_6} + \underset{sp}{MgAl_2O_4} \rightleftharpoons \underset{Mg\ TS}{MgAl_2SiO_6} + \underset{fo}{Mg_2SiO_4}$$

using OBATA'S data, temperatures between 930 and 1070 °C were obtained. HERZBERG (1976, 1978) showed a similar temperature dependence of Ca-Tschermaks in clinopyroxenes. Using his equation for the equilibrium constant from our data, and assigning all Al to Ca-Tschermaks, temperatures between 1050 and 1200 °C result. However, if jadeite is formed (the cpx contain appreciable amounts of Na) first, then temperatures between 910 and 960 °C are obtained. These different temperature estimates are summarised in Table 3. The different samples show the same range of temperatures indicating that they were equilibrated under the same physical conditions. At present, it is difficult to decide which set of data is correct. Thus, only a range of 830 to 975 °C and a pressure of 10 ± 3 kb can be given for the equilibrium temperature.

Ar^{39} / Ar^{40} determinations

EXPERIMENTAL PROCEDURE

A phlogopite separate of 10.4 mg was irradiated along with three CC-27 hornblende monitors in a single Harwell can, which was placed in our preferred position in the core of the FR-2 reactor, Gesellschaft für Kernforschung, Karlsruhe.

The sample assemblage was subjected to an integrated neutron flux of 5.60×10^{18} cm^{-2} (E > O, 1 MeV). The sample preparation prior to and after irradiation as well as the characteristics of the hornblende monitors have been described elsewhere (STETTLER et al., 1973, 1974). Based on the Co^{58} activity of Ni-wires attached to each sample container a horizontal and vertical flux gradient of < 2.8 % was measured and corrected for. The argon composition of the CC-27 hornblende monitors allowed to calculate a V-value of 0.02434 ± 0.00010. Adopting K = 2600 ppm and Ca = 8.5 % for CC-27 monitors the conversion factors for K and Ca inferred from decay corrected Ar^{39}_K and Ar^{37} turned out to be C_{39} (K) = 1.83×10^{-4} cm³ STP Ar^{39}_K/g K and Ca_{37} (Ca) = 0.925×10^{-4} cm³ STP Ar^{37}/g Ca.

Notations and relations used throughout this paper

Ar^{40}_r	=	radiogenic Ar^{40}
Ar^{39}_K	=	potassium-derived Ar^{39}
V	=	$(e^{tm}-1)/(Ar^{40}/Ar^{39}_K)_M$ where M denotes monitor
λ	=	5.305×10^{-10} y^{-1}, λ_ε = 0.585×10^{-10} y^{-1}
K^{40}/K	=	1.19×10^{-4}
λ_{39}	=	7.2×10^{-6} day^{-1}
λ_{37}	=	0.01975 day^{-1}
$\frac{Ar^{40}}{Ar^{36}}_{atm}$	=	295.5

The phlogopite sample was degassed stepwise and the expelled argon analysed on-line with a double sector magnetic mass spectrometer at the Physikalisches Institut Bern (SCHWARZMÜLLER, 1970). The procedural blank (in 10^{-8} cm³ STP) was 0.15 at temperatures below 1100 °C rising up to 0.30 at 1700 °C.

Interfering isotopes like Ar^{36}_{Ca}, Ar^{39}_{Ca} and Ar^{40}_K produced by neutron reactions on Ca and

Tab. 3 Temperature estimates from the chemical composition of pyroxenes, using the equilibrium constants from different authors. In °C.

Method		Solvus cpx - opx				Al in opx	Al in cpx	
Author	Wood & Banno 1979	Wells 1977	Mori 1977	Herzberg 1978	Evans & Trommsdorff 1977	Obata 1977	Herzberg 1978	Herzberg 1978*
Peridotite Mylonite TP 49	943	827	909	950	900	960	1130	960
Peridotite Mylonite TP 462	934	826	901	960	900	1070	1100	
Peridotite Mylonite TP 357	902	781	825	900	775	930	1110	940
Pyrop Pyroxenite TP 63b	928	870	996	1050	825		1050	
Spinel Pyroxenite TP 340	944	832	924	1000	830		1200	910

*with Jd correction

K have been subtracted using the reaction yields established previously for our reactor positions (STETTLER, 1973).

RESULTS

The data of stepwise released argon from the phlogopite sample are summarised in Table 4. Fig. 1 gives the Ar^{39}-Ar^{40} release pattern. More than 90% of the Ar^{39}_{K} are released between extraction temperatures of 800 °C and 1700 °C. In this temperature range the Ar^{39}-Ar^{40} age curve appears slightly saddle-shaped. Comparable, and even much more pronounced saddle-shaped Ar^{39}-Ar^{40} release patterns have often been observed in terrestrial igneous (LANPHERE and DALRYMPLE 1971, BRERETON 1972, OZIMA and SAITO 1973, COWPERTHWAITE et al., 1972, KANEOKA 1974) and metamorphic (DALLMEYER 1975, ALEXANDER 1975) rocks and minerals. Apparently those samples contained extraneous argon and the minima of the age curves approached but did not reach the known ages of the samples. According to this interpretation $(156 \pm 2) \times 10^6$ y would be the best value for the phlogopite age.

Glass-rich lunar samples also exhibited saddle-shaped Ar^{39}-Ar^{40} release patterns. This has been ascribed to neutron irradiation effects inducing considerable redistribution of argon isotopes or changes of lattice site retentivity in the samples (DAVIS et al. 1971, ALEXANDER et al. 1973, TURNER and CADOGAN 1974, HORN et al. 1975, HUNEKE and SMITH 1976). For such samples the average high temperature age was assumed to reflect the true age. Following this interpretation an age of $(165 \pm 6) \times 10^6$ y could be inferred for the phlogopite.

Since we cannot decide which one of the two different interpretations cited above is valid, we adopt an average value of 160×10^6 y with a somewhat higher uncertainty of $\pm 8 \times 10^6$ y for the phlogopite Ar^{40} retention age. The small plateau formed by the retention steps between 600 and 790 °C indicate an influence caused by a geologic event. It could be due either to the alpine metamorphism that reached prehnite-pumpellyte facies or the serpentinisa-

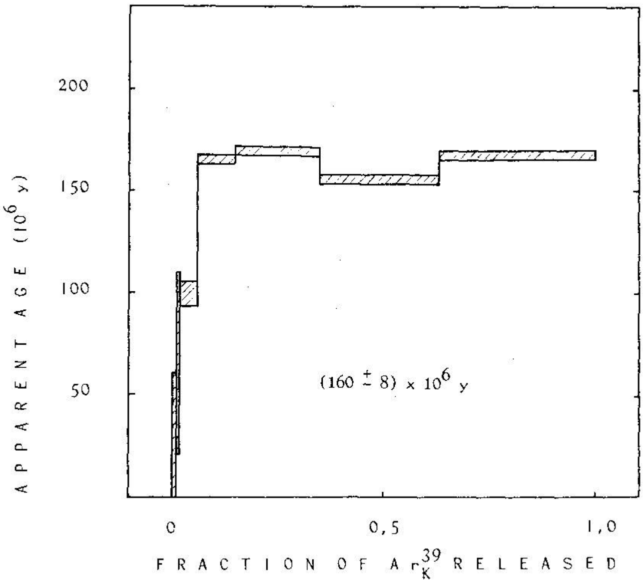

Fig. 1 Ar^{39}-Ar^{40} release pattern of the phlogopite separate.

Tab. 4 Argon results from stepwise heating of a 10.4 mg neutron activated phlogopite separate. Concentrations determined from ion beam intensities.

Temp. °d	Ar^{39}_K $10^{-8}cm^3STP/g$	Ar^{40}_r/Ar^{40}	Ar^{40}/Ar^{36}	Ar^{38}/Ar^{37}	Ar^{39}_K/Ar^{37}	Ar^{36}/Ar^{38}	Ar^{40}/Ar^{39}_K	Apparent age(10^6y)
350	0.655 ±.070	0.19 ±.10	365 ±25	0.173 ±.020	1.60 ±.18	3.78 ±.30	132.0 ± 8.0	890 ±350
570	5.25 ±.60	0.25 ±.10	392 ±20	0.1815 ±.0080	2.310 ±.050	0.602 ±.035	20.1 ±1.1	210 ± 70
640	2.9 ±1.4	0.20 ±.20	370 ±180	0.203 ±.016	3.73 ±.18	0.185 ±.070	3.7 ±1.1	30 ± 30
700	6.3 ±1.0	0.25 ±.18	390 ±60	0.332 ±.025	6.05 ±.25	0.265 ±.035	5.90 ±.60	65 ± 45
790	43.0 ±2.5	0.545 ±.030	645 ±25	0.490 ±.050	0.1563 ±.0030	0.208 ±.025	5.00 ±.10	120.5 ± 6.0
890	85.5 ±4.5	0.835 ±.015	1790 ±100	1.10 ±.17	55.2 ±1.6	0.105 ±.017	4.530 ±.045	166.0 ± 2.5
960	198 ±10	0.9645 ±.0080	8300 ±1000	2.05 ±.45	150.0 ± 5.0	0.0205 ±.0050	4.012 ±.025	169.5 ± 1.5
1150	275 ±15	0.893 ±.025	2750 ±250	0.84 ±.19	66.9 ±1.4	0.0345 ±.0080	3.975 ±.035	156.2 ± 2.0
1700	347 ±18	0.967 ±.012	9000 ±2000	0.260 ±.060	19.22 ± .30	0.031 ±.010	3.962 ±.035	168.0 ± 2.0
Total	963	0.869	2250	0.460	28.4	0.0925	4.260	162.5

tion. The apparent age of about 100 M.a. has no significant meaning, it only indicates that a disturbance took place between 160 M.a. and today.

From the total amounts of Ar^{39}_K, Ar^{37} and Ar^{38}_{Cl} we infer

K = 5.25 ± 0.25%
Ca = 0.35 ± 0.05%
Cl = 200 ± 30 ppm

using the conversion factor for K and Ca given above and a production yield P_{38} (Cl) = 12.8 × 10^{-6} cm³ STP/g Cl/10^{18} cm^{-2} derived from NaCl irradiated in the same reactor position.

Discussion

The radiometric age of 160 M.a. ± 8 corresponds stratigraphically to the Dogger series, at the Bathonian-Callovian boundary (Geologic Time Scale, VAN HINTE, 1976). A similar age was obtained with zircon from Corsican ophiolitic albitites by OHNENSTETTER et al. (1981). The studies of TRÜMPY (1975) and FINGER (1978) have shown that this was the period of deposition of most of the coarse breccias of the Lower Austroalpine units and the Arosa zone, indicating block faulting on a large scale at this time. Detailed studies generally show transport from the south towards the north where troughs were forming, leading to the bathyal basins where in the late Jurassic radiolarian cherts were deposited. This faulting and deepening of the upper Penninic basins is interpreted as the result of an extension phase (PETERS, 1969). The continental crust was thinned and, according to authors like DIETRICH (1975), an extensive oceanic crust was formed between the Eurasian and Austroadriatic plates. Whether an extended oceanic crust with a circumpacific-like Benioff zone (DIETRICH, 1974, DAL PIAZ, 1974) was formed or only a drifting apart of two continental plates with or without crustal thinning at their borders (PETERS, 1969, HSÜ, 1979) took place, the upper mantle must have domed upwards in this region. During this uprise the isotherms rose also, but less fast than the upper mantle peridotites, so that these cooled. During this cooling the chemical and isotopic exchange equilibria were blocked. For the Ar/K equilibria temperatures of around 300°C are presumed. It is highly probable that the age of 160 M.a. dates this "upwelling" of mantle material. As no minerals with higher blocking temperatures for other isotope equilibria could be dated, the rate of isotherm movement cannot be determined. A conservative rate of 1mm/y and a geothermal gradient of 30°/km may, however, be assumed. In this case, the blocking temperature for the pyroxene geothermometers would have been passed 2 millions years earlier. This is a short time span compared to the uncertainties in radiometric age. It is assumed to belong to the same geologic event. If the basaltic extrusives of the Arosa zone represent the partial melt of the associated ultrabasics as required by the ridge-type ophiolite model (COLEMANN, 1977), than these lavas should be older. The stratigraphic evidence, however, indicates a younger, Upper Jurassic to Lower Cretaceous age for the basic extrusives in the Arosa zone. The K/Ar ages on hornblendes from a diabase-sill in the Platta region of 112 and 115 ± 4 M.a. (DIETRICH, 1969) are not necessarily ages of formation, but could be the result of the influence of the eo-alpine metamorphism.

The non-harzburgitic character of the ultrabasics of the Arosa zone does not favour the ridge-type ophiolite model. The basaltic magmas are believed to have originated from peridotites at deeper levels of the upper mantle that were not brought to the surface during the alpine orogenesis. Applying the Na/Cr-discrimination diagram for clinopyroxenes after KORNPROBST et al. (1981) the Totalp peridotite would be "sub-continental", which supports our model of a small, relatively short lived ocean in a rift system.

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