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Reference

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Introduction

The Piemont zone in the Western Alps, a penninic structural unit, is considered as being made up of two parts. The first has an oceanic origin, and includes ophiolitic rocks and pelagic sediments, probably upper Jurassic and lower (?) Cretaceous. The second part is made up of trench deposits, including pelitic and carbonate sediments, interlayered basic rocks, and some ultrabasics, starting in the lower Jurassic or upper Triassic, and possibly going up to the lower Cretaceous. The first part has been called the Zermatt-Saas zone, extending Bearth’s (1967) denomination to the whole of the Western Alps. The second part is the Combin zone (Dal Piaz, 1971; Bearth, 1974; Dal Piaz, 1974; Frey et al., 1974). Both zones are metamorphic. The assignment of an outcrop to one of these zones is sometimes not unequivocal owing to the mélange structure mixing both zones in the southern area, away from the Zermatt-Aosta region where they have been defined.

Radiometric age determinations [Bocquet (Desmons) et al., 1974; Hunziker, 1974] demonstrated three Alpine metamorphic events in the internal part of the
Western Alps. The Eoalpine event, late Cretaceous in age, is connected mainly with high pressure, eclogitic and glaucophane schist assemblages. The so-called Alpine, or Lepontine, event is late Eocene-early Oligocene in age, and applies, in the Western Alps, to glaucophane greenschist, and greenschist assemblages. As yet little is known about the exact nature and extent of the third, late Oligocene-early Miocene (late Alpine) event, which has so far been obtained on the amphiboles. Each of these three metamorphic events coincides with orogenic phases, and also with the assumed timing and direction of plate motion in the Atlantic-Tethyan area [see discussion: Bocquet (Desmons), 1974; Desmons, in prep.].

As inferred from the metamorphic assemblages displayed in both the Zermatt-Saas and the Combin zones, and from structural arguments, the latter zone seems to have escaped the Eoalpine event.

The aim of the present study is to answer the following question: Does the isotopic composition of the white micas from both parts of the Piemont zone support the assumed timing and distribution of the metamorphic events?

**Sampling**

The location, metamorphic assemblage, structural origin and mineral species of the 14 collected samples are listed on Table 1. The geographic distribution extends from the vicinity of the Mt Rosa

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**Table 1. Mineral assemblage of the rock, location and mineral species of the mica samples**

| Sample | Mineral Assemblage                                      | Location                          | Mineral Species     | Paragonite (%) |
|--------|--------------------------------------------------------|-----------------------------------|---------------------|----------------|
| 162    | white mica-quartz-lawsonite-chlorite schist.           | Combin zone (?). St-Véran, carrière de marbre; | Phengite, 6%        |                |
|        |                                                        | x = 963.1; y = 274.2; z = 2,330.     |                     |                |
| 734    | white mica-quartz-lawsonite-chlorite schist.           | Zermatt zone (?). Bardonecchia, col de Fréjus; | Phengite, 7%        |                |
|        |                                                        | x = 941.2; y = 323.3; z = 2,380.     |                     |                |
| 735    | calcite-quartz-lawsonite-white mica-chlorite schist.   | Zermatt zone (?). Bardonecchia, col de Fréjus; | Phengite, 8%        |                |
|        |                                                        | x = 941.3; y = 323.2; z = 2,350.     |                     |                |
| 842    | albite-calcite-white mica-lawsonite-chlorite schist.   | Combin zone (?). Molines-en-Queyras, vallon du Longet. | Phengite, 2%        |                |
|        |                                                        | x = 962.9; y = 276.6; z = 2,400.     |                     |                |
| 844    | calcite-quartz-white mica-lawsonite-chlorite schist.   | Zermatt zone. Ristolas, lac Baricle; | Phengite, 10%       |                |
|        |                                                        | x = 967.9; y = 279.5; z = 2,390.     |                     |                |
| 957    | white mica-garnet-calcite-quartz-albite rock.          | Combin zone. Val d'Isère, sources de l'Isère; | Phengite, 21%       |                |
|        |                                                        | x = 970.9; y = 361.6; z = 2,350.     |                     |                |
| 959    | white mica-calcite-albite-quartz-chlorite schist.      | Combin zone. Val d'Isère, sources de l'Isère; | Phengite, 18%       |                |
|        |                                                        | x = 970.8; y = 361.5; z = 2,340.     |                     |                |
| 1,009  | paragonitic lens in serpentinite.                      | Zermatt zone. Cervières, col de Péas; | Phengite, 24% [as determined by flame photometry, contains 6.50 Na2O and 1.40 K2O: Bocquet (Desmons), 1974]. |                |
|        |                                                        | x = 953.2; y = 289.2; z = 2,600.     |                     |                |
| Ta.99  | paragonitic lens in serpentinite.                      | Zermatt zone. Ristolas, arête de la Taillante; | Phengite, 19%       |                |
|        |                                                        | x = 969.2; y = 278.0; z = 2,800.     |                     |                |
| C.88   | white mica-quartz-garnet-chloritoid schist.            | Zermatt zone. Dondena, colle Fussi. | Phengite, 18%       |                |
|        |                                                        | Phengite, 8%                        |                     |                |
| C.89   | amphibole-clinozoisite-albite-green mica-calcite metagabbro | Zermatt zone. Dondena, M. Delà. Mariposite. | Phengite, 2%        |                |
| C.97   | white mica-calcite-quartz-chlorite schist.             | Combin zone. Busson, valle di Frudiere. | Phengite.           |                |
| C.99   | white mica-calcite-quartz-chlorite schist.             | Combin zone. Gressoney-la-Trinité, stazione superiore della funivia. | Phengite, 4%        |                |
| C.104  | quartz-white mica-calcite-garnet-clinozoisite schist.  | Zermatt zone. Cogne, valle d'Urtier. | Phengite, 2%        |                |
massif in Italy down to the Queyras region in France. The host rocks are calc-schists, micaschists ± chloritoid or garnet, metagabbro, and micaceous lenses in serpentinites.

Technical Methods

Ar extraction, Ar and K measurements are described in Delaloye and Wagner (1974).

The nature of the white micas has been determined from X-ray powder diagrams. The obtained (002) values lie between 9.92 and 9.95 Å for potassic white mica, and from 9.63 to 9.66 Å for paragonite. The intensity (peak height) ratio of these peaks has been used to determine the paragonite content of the samples. The (060) reflection ranging from 1.504 to 1.507 Å points to phengitic nature for all studied potassic white micas (Cipriani et al., 1968).

Discussion of the Results

The results (Table 2) may be divided into 3 groups:

1. Ages of 172 and 120 m.y. of two paragonites; the error limits do not overlap;
2. Eoalpine ages of 62, 73 and 67 m.y. of two phengites and one paragonite;
3. Alpine ages, from 37 to 54 m.y., of eight phengites and one mariposite.

1. The first two paragonite samples have been collected in pure micaceous lenses in antigorite serpentinites. One interpretation is that the mica crystallization may be connected with one of the serpentinization phases. Sample no. 1,009, of middle Jurassic (172 ± 25 m.y.), could date a first primary phase of serpentinization, considered as an ocean-floor, or sub-ocean floor, thermal metamorphism, or as a deep-seated process in the lower oceanic crust (Honnorez and Kirst, 1975). In the case of such a deep-seated process one should not speak of serpentinization phase if the ultrabasite alteration has to be considered as a

Table 2. K – Ar data and age results

| Sample n° | Lab n°   | Mineral   | %K | Ar\(^{40}\) rad. moles/g | %Ar\(^{40}\) rad. | Ar\(^{40}\)/Ar\(^{36}\) | K\(^{40}\)/Ar\(^{36}\) | age (m.y.) |
|-----------|----------|-----------|----|--------------------------|-----------------|-------------------|------------------------|------------|
| 1,009     | KA 804   | paragonite| 1.08| 3.453 \times 10^{-10}   | 21.5            | 3.195 \times 10^{2} | 6.554 \times 10^{1}   | 172 ± 25    |
| Ta.99     | KA 810   | paragonite| 1.07| 2.385 \times 10^{-10}   | 19.5            | 3.181 \times 10^{2} | 8.576 \times 10^{1}   | 120 ± 20    |
| 844       | KA 813   | paragonite| 1.06| 1.399 \times 10^{-10}   | 31.2            | 4.013 \times 10^{2} | 2.897 \times 10^{2}   | 73 ± 6      |
| 734       | KA 815   | paragonite| 2.95| 3.566 \times 10^{-10}   | 39.6            | 4.842 \times 10^{2} | 4.841 \times 10^{2}   | 67 ± 4      |
| 735       | KA 805   | phengite  | 7.37| 8.251 \times 10^{-10}   | 87.5            | 2.295 \times 10^{3} | 5.470 \times 10^{3}   | 62 ± 0.5    |
| C.104     | KA 812   | phengite  | 7.32| 6.142 \times 10^{-10}   | 76.2            | 1.037 \times 10^{3} | 2.470 \times 10^{3}   | 54 ± 0.8    |
| C.97      | KA 809   | phengite  | 7.38| 6.510 \times 10^{-10}   | 89.8            | 2.410 \times 10^{3} | 7.477 \times 10^{3}   | 49 ± 0.4    |
| 842       | KA 807   | phengite  | 7.79| 6.063 \times 10^{-10}   | 89.8            | 2.777 \times 10^{3} | 9.777 \times 10^{3}   | 43 ± 0.3    |
| C.88      | KA 806   | phengite  | 6.08| 4.599 \times 10^{-10}   | 87.5            | 2.262 \times 10^{3} | 7.973 \times 10^{3}   | 42 ± 0.1    |
| 162       | KA 816   | phengite  | 6.87| 5.084 \times 10^{-10}   | 83.6            | 1.744 \times 10^{3} | 6.008 \times 10^{3}   | 41 ± 0.4    |
| 959       | KA 814   | phengite  | 8.07| 5.706 \times 10^{-10}   | 85.5            | 1.970 \times 10^{3} | 7.268 \times 10^{3}   | 39 ± 0.4    |
| 957       | KA 808   | phengite  | 6.33| 4.390 \times 10^{-10}   | 84.5            | 1.694 \times 10^{3} | 6.298 \times 10^{3}   | 39 ± 0.4    |
| C.99      | KA 811   | phengite  | 6.95| 4.638 \times 10^{-10}   | 86.0            | 2.036 \times 10^{3} | 8.008 \times 10^{3}   | 37 ± 0.3    |
| C.89      | KA 828   | mariposite| 8.11| 5.385 \times 10^{-10}   | 72.3            | 1.047 \times 10^{3} | 3.477 \times 10^{3}   | 37 ± 0.6    |
"continuous process" (Coleman, 1972) during the whole ocean crust formation. Such an alteration is likely to have taken place in the Western Alpine ophiolitic rocks. Middle Jurassic is the age of some ophiolites of the Western Alps and Corsica (Pusztaszeri, 1969; Bertrand, 1971; Loomis, in Gianelli and Principi, 1974; Bertrand and Delaloye, in press).

Sample no. Ta. 99, 120±20 m.y., could have been formed either during the Eoalpine, and so have retained some radiogenic argon from an earlier phase, or at a hypothetical serpentinitization phase in the early Cretaceous. This would involve a continuous formation of oceanic crust and ultramafic alteration from late Jurassic until late Cretaceous time. Finally, an Alpine formation age with inherited argon cannot be precluded.

The present rock-forming mineral in the host serpentinite is antigorite; chrysotile is only present in late fractures, and lizardite replaces pyroxene. Antigorite is considered to form in place of chrysotile and lizardite at higher temperature (above 450–500°C, Johannes, 1969; Ishi and Saito, 1973; above 220–460°C, Wenner and Taylor, 1971). If antigorite and paragonite are coeval, the first, ocean-floor serpentinitization phase should be high tempered enough to generate antigorite; however, this is inconsistent with the most commonly observed lizardite in present-day altered oceanic ultrabasics (Aumento and Loubat, 1971), and also with the temperature range of 180 to 400°C inferred from the mineralogy by Spooner and Fyfe (1973) for ocean-floor metamorphism in Ligury. Another hypothetical explanation is that the paragonite retains its isotopic ratio at a higher temperature than potassic white mica, for which the K—Ar isotopic exchange takes place from about 350°C (Purdy and Jäger, 1976). In this case antigorite would not crystallize coeval with paragonite, but later, during another serpentinitization process, in particular during the Eoalpine.

(2) Eoalpine ages, or ages slightly affected by a subsequent Ar-loss, have been found in three samples from the more western and southwestern part, i.e. the outermost part of the Piemont zone, where metamorphic parageneses display a decreasing temperature gradient during Alpine time [late Eocene-Oligocene; Bocquet (Desmons), 1974]. These samples are from the Zermatt-Saas zone, or of unclear relationship due to the mélangé structure. Their isotopic composition thus escaped totally, or almost completely, subsequent alteration during Alpine or Oligo-Miocene (late Alpine) time. They show that in that part of the area the metamorphic grade during the later event has not been strong enough to destroy all relics. One of these samples (no. 734) contains both paragonite and phengite (2.95% K). However, its age (67±4 m.y.) seems to be a true age, as indicated by the similar age obtained on a phengite sample of the same location (no. 735:62±0.5 m.y.).

(3) Comparison of K—Ar with Rb—Sr data of the same samples [Bocquet (Desmons) et al., 1974] has shown that many K—Ar results from the Western Alps are slightly affected by Ar40 inheritance. This also seems to be the case of most samples of the present Alpine (Lepontine) age group. They extend from 35 to 54 m.y. as shown by an Ar40/Ar36 – K40/Ar36 plot (Fig. 1). If some inherited Ar is assumed, the isochrons should be tilted and would cut the Ar40/Ar36 ordinate at higher values than the normal initial ratio.
Some of the samples dated as Eocene are from the Zermatt-Saas zone and have thus been rejuvenated. All these samples come from the northeastern part of the area, i.e. closer to the so-called Lepontine thermal high of the Central Alps (Frey et al., 1974). Thus, in this area, the K–Ar isotopic re-equilibration temperature has been reached, and an approximate 350°C isograd can be delimited for that late Eocene metamorphic phase, running approximately southeast of the Gran Paradiso massif.

In the Combin zone, white micas showing Eocene ages are distributed throughout the whole of the studied area. If the Combin zone did not suffer Eoalpine metamorphism, the slight amount of surplus argon could have been inherited from some detrital phyllitic material. A larger number of samples is necessary to demonstrate any regularity in the distribution of this overpressure in relation to the temperature gradient and to the possible structural position at the time of the thermal high.

Finally, it must be emphasized that the late Alpine (late Oligocene-early Miocene) event does not appear amongst the present data. Taking a re-opening
temperature for K-white mica of 350° C, one can conclude that the late Alpine event, though Na-amphibole producing or isotopically re-equilibrating, has not been high tempered enough to change the isotopic ratio of the micas in the studied area, or to crystallize micas. This would confirm the low temperature generally assumed for Na-amphibole formation and alteration, and is in agreement with the suggestion made by Coleman and Lanphere (1971) that glaucophane would be particularly sensitive to metamorphic or deformational effects.

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