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Alpine Metamorphism of the Alps

A Review

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With 1 figure in the text and 2 plates

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This paper was initiated while working for the "Metamorphic Map of the Alps" (Sheet 17 of the Metamorphic Map of Europe, Scale 1 : 1 000 000; General coordinator: Prof. H. J. Zwart, Leiden). A knowledge of the structural and paleogeographic units is essential for an understanding of the Alpine metamorphism. It is impossible to review here all the work dealing with the relationships between structural and metamorphic petrology; instead, this paper gives an example of a particularly well studied area. In Chapter IV, a summary of the geochronologic data essential to an understanding of Alpine metamorphism, is presented. References are only given for general and recent papers. In the paper, extensive use is made of the terms Western, Central, and Eastern Alps. This subdivision is based mainly on geographical and traditional criteria: the Western Alps comprise the Alpine arc between Genova and the border between France and Switzerland; the Central Alps comprise mainly the Swiss Alps; the Eastern Alps are located mainly in Austria and northeastern Italy.

I. STRUCTURAL AND PALEOGEOGRAPHIC OUTLINE OF THE ALPS (PLATE I)

ARGAND, 1934; CADISCH, 1953; CLAR, 1965, 1973; DEBELMAS and LEMOINE, 1970; GWINNER, 1971; OBERHAUSER, 1968; RAMSAY, 1963; TOLLMANN, 1963, 1971, 1973; TRÜMPY, 1960; TRÜMPY and HACCARD, 1969.

The Alps may be subdivided into four main paleogeographic and structural domains: the Dauphiné-Helvetic domain on the external margin; the Penninic domain; the overthrust Austroalpine units; and the Southern Alps on the inner side.

1. Dauphiné-Helvetic Domain

This domain is formed by a pre-Westphalian crystalline basement consisting of paraschists, granitic gneisses, migmatites, amphibolites, marbles, and rare slices of ultrabasic rocks, with abundant Hercynian granites. The following massifs can be recognized from SW to NE: Argentera, Pelvoux, Belledonne, Mt. Blanc and Aiguilles Rouges, Gotthard and Aar. The post-Hercynian cover commences with transgressive shales and fine-grained clayey sandstones of Stephanian-Westphalian age, and some Permian arenaceous-pelitic material; these deposits are generally now restricted to narrow synclines pinched into the crystalline basement.

The Permo-Mesozoic to Paleogene sequence, mainly carbonate rocks, still adheres locally to the substratum, but has in part slid outwards from the chain to form the Helvetic nappes. The Ultrahelvetic nappes, superposed on the Helvetic nappes, originated in a more internal and deeper area of the common basin of deposition. The Ultrahelvetic nappes are in turn tectonically overthrust by the Prealps, a sequence of nappes of internal, in part South Alpine origin.

Continuous to the Eastern Alps, the Helvetic domain forms a narrow unit at the front of the orogen which is overthrust by the Austroalpine Flyschzone and the Northern Limestone Alps.

2. Penninic Domain

This domain comprises a pre-Triassic crystalline basement and a Mesozoic cover. This Mesozoic cover is overthrust partly by the crystalline basement, thus proving the Alpine nappe structures. The following tectonic units can be distinguished from bottom to top and from external to internal parts of the belt in the *Western and Central Alps:*

a) The lower Penninic (or Lepontine) units, which are restricted to the Ossola and Ticino culminations.

b) 1. In the west of the Lepontine: the Bernhard-Briançonnais unit, and the Monte Rosa–Gran Paradiso–Dora-Maira unit. A paleogeographic reconstruction of the Penninic domain, representing the end of the Paleozoic, indicates a single crystalline basement, formed by the Bernhard-Briançonnais unit on its external side, and the Monte Rosa–Gran Paradiso–Dora-Maira unit on its internal side.
   2. In the east of the Lepontine the Penninic units of the Grisons are exposed.

The pre-Triassic *basement* in all Penninic units is poly-metamorphic. The oldest recognizable rocks are predominantly paraschists, with minor amphibolites, migmatites and marbles. In Hercynian time, widespread granites intruded into this basement. The intensity and regional extent of the Hercynian metamorphism is not yet clear. There is evidence in the Western Alps that the discordant Permo-Carbonifereous cover of the Hercynian basement was metamorphosed during Permian time. This is also probably true for the Hercynian basement.

The numerous *Mesozoic units* of the Penninic domain are differentiated by their lithologic composition and depositional environments. From the exterior toward the interior of the belt these units are:

a) The *north-Penninic Calcschists or Valais Zone* is mainly developed in the Lepontine and the Grisons. Towards the west, the Valais Calcschists narrow progressively; their continuation into the Western Alps is still a matter in dispute. Some authors consider the Sion-Courmayeur Zone to be the prolongation of this zone. The Valais Zone comprises Mesozoic detrital and carbonate sedimentary rocks, with associated ophiolites. This sequence probably represents sedimentation in a geosynclinal trough with a sialic substratum.

b) The *Subbriançonnais and Briançonnais Zones* are mainly developed in the Western Alps with minor equivalents in the Grisons. These zones correspond with the external and internal sides of a single basin, respectively. The Subbriançonnais Zone has the characteristics of open sea sedimentation; the Briançonnais Zone has those of platform sedimentation presumably on the Bernhard-Briançonnais basement. In the Briançonnais Zone the cover consists of a Mesozoic-Paleocene sequence of predominantly carbonate rocks.

The *Piemont Zone* with calcschists ("schistes lustrés") and ophiolites is present mainly in the Western Alps with minor equivalents in the Grisons. The rocks of this zone originally consisted of marls, carbonate and arenaceous rocks, shales and radiolarites, closely associated with submarine ophiolitic extrusions (diabases, pillow-lavas, etc.), and possibly with some sills. These extrusions are linked with basaltic magmas, discharged along fractures during the late Jurassic – early Cretaceous phase of tectonic extension. The Mesozoic series was probably deposited in part on an ocean floor of gabbros and ultrabasic rocks and in part on a thin sialic margin above a Triassic – lower Jurassic non-ophiolitic sequence. During Alpine movements the Piemont Calcschist nappe was thrust toward the external part of the belt, partly overriding the Briançonnais Zone.

In the *Eastern Alps*, the Penninic domain includes the Austroalpine Flyschzone, and most of the rocks in the Engadine, Tauern and Rechnitz-Bernstein Windows.

c) 1. There is general agreement that the Austroalpine Flyschzone is derived from the Penninic domain, but from which part is still a matter of dispute. East of Vienna this zone trends oblique to the continuation of the Helvetic domain and forms the outermost unit of the Carpathians.

2. The Penninic domain within the Engadine Window is composed mainly of Mesozoic calcschists and minor ophiolites and is probably the continuation of the Valais Zone in the Grisons.

3. The Penninic domain of the Tauern Window can be subdivided into two major paleogeographic units. The northern part is now formed by the "Central gneiss domes", a pre-Alpine basement composed of high-grade metamorphic paragneisses, and low-grade early Paleozoic ophiolites (Habachserie). Both paragneisses and ophiolites are intruded by large bodies of Hercynian granitic rocks (partly of very late Paleozoic age). There is no proof for a Caledonian metamorphism at present, but the Hercynian metamorphism was fairly intense (low pressure – high temperature). It is possible that the transgressive rocks of Lower Paleozoic to Permian age were metamorphosed by a second Hercynian event.

   The metamorphosed Mesozoic cover of these gneiss domes consists of Permoskythian quartzitic rocks, Middle Triassic carbonates, pelitic-quartzitic Keuper schists, Jurassic rocks with a pronounced clastic influence (Brennkogelfacies), and, in the NW, of transgressive Upper Jurassic carbonates (Hochstegenkalk). To the south, in post-Triassic time, marly rocks and ophiolites were deposited, at least in part, on an oceanic crust (Glockner Series). This zone was then overthrust as a huge nappe system onto the former frontal barrier.

   There is no general agreement on how to correlate these units with the Central and Western Alps. If the Tauern gneiss domes are considered as the continuation of the Briançonnais, the Glockner Nappe would correspond with the Piemont Zone. If a correlation with the continuation of the external massifs is preferred, the internal Penninic Briançonnais platform could pinch out eastward.

4. Together with some Paleozoic and Triassic rocks the typical Mesozoic calcschists and ophiolites occur again in the Rechnitz-Bernstein Window at the eastern end of the Alps.

3. Austroalpine Domain

As indicated by the name, the Austroalpine units occur mainly in the Eastern Alps. In the *Western Alps* the Austroalpine domain is represented by the Dent Blanche nappe system with adjacent klippen and the Sesia-Lanzo Zone. The Dent Blanche nappe s. l. can be subdivided into two independent structural units. The lower element is formed by the Arolla Series consisting mainly of granites, granodiorites and diorites, with minor paraschists. The upper element comprises the Valpelline Series formed by a complex of pre-Alpine high-grade crystalline rocks. The tectonic contact between these two structural elements is emphasized by a well-developed mylonite and blasto-mylonite horizon.

The Sesia-Lanzo Zone can be subdivided into two independent structural units. The lower element, called the Sesia-Lanzo Zone s. str., is made up mainly of gneisses and eclogite-like rocks of granitic to dioritic composition ("Micascisti eclogitici" of the Italian geologists) and minor paraschists. Southeastward from the Sesia-Lanzo Zone a gradational transition toward the Ivrea Zone of the Southern Alps seems to exist. The upper element, called "seconda zona diorito-kinzigitica", contains the same rocks as the Valpelline Series and overlies the Sesia-Lanzo Zone s. str. tectonically as several klippen.

From a paleogeographic view the Arolla Series corresponds to the Sesia-Lanzo Zone s. str., whereas the Valpelline Series and the "seconda zona diorito-kinzigitica" correspond to the Ivrea Zone.

In the *Eastern Alps* (including also the eastern part of the Grisons) the Austroalpine domain is divided into Lower, Middle and Upper Austroalpine units.

The *Lower Austroalpine* unit was originally situated at the northern front of the Austroalpine domain but was later dragged to its present position by the overlying Middle Austroalpine thrust sheet. It comprises the Bernina-Err unit in the Grisons and forms the uppermost tectonic part of the Engadine and Tauern Windows. The pre-Permian basement consists mostly of granites in the Bernina-Err unit and low-grade metamorphic Paleozoic rocks in the Tauern Window. The Triassic and part of the Jurassic are composed of platform carbonate rocks of similar sedimentary facies to rocks in adjoining Austro-alpine areas. The part of the lower Austroalpine unit which was originally northernmost in the Tauern area shows a pronounced clastic environment in the Jurassic and Lower Cretaceous.

The Lower Austroalpine pre-Alpine metamorphic basement with abundant Hercynian granites and minor Mesozoic rocks is found in the Semmering-Wechsel Window. The question of whether the deepest tectonic unit, the Paleozoic rocks of the Wechsel, can be considered as part of the Penninic or as part of the Lower Austroalpine unit is still under discussion.

The main mass of the Austroalpine rocks, the actual overriding plate, can be subdivided into the Middle and the Upper Austroalpine units. They are separated by a continuous thrust plane. Their original relationship is still a matter of dispute.

The *Middle Austroalpine* unit is composed of extensive pre-Alpine meta-morphic basement rocks which underwent intense pre-Alpine medium- to high-grade metamorphism and intrusion by granites. It comprises the Silvretta-Ötztal-Campo mass to the west, the crystalline rocks south of the Tauern Window, and to the east, the crystalline rocks of the Schladminger-Seckauer Tauern, Sau-Koralps, Gleinalm and other localities. The geological evidence for a progressive Hercynian metamorphism is mainly derived from the Saualpe area and sufficient proof exists that most of the mineral parageneses in other areas have probably also been formed or affected by this polyphase event.

Belts composed of lower-grade, mostly phyllitic Paleozoic rocks cover this basement in some areas. These rocks are now connected with the basement, e. g. in the Schneebergzug and Gurktal mass, or separated from the basement by a thrust plane, e. g. the Paleozoic of Graz. The relationship of these belts with the basement before the onset of Alpine tectonics is in discussion. A transgressive Permo-Mesozoic sequence was preserved in the Engadine Dolomites, Gailtal Alps, Stangalm (west of Gurktal) and other areas.

The *Upper Austroalpine* thrust sheet consists of the Paleozoic Grauwackenzone and the transgressive Mesozoic sequence of the Northern Limestone Alps. Both units have a distinct nappe structure but Alpine metamorphism is only developed in their basal parts.

4. Southern Alps

This tectonic unit is described very briefly here because it was only slightly affected by Alpine metamorphism. The Southern Alps are separated from the other Alpine complexes described above by a major tectonic feature, the Insubric Line, which is replaced eastward by the Giudicaria Line and the Alpine-Dinaric Line.

The polymetamorphic crystalline basement consists of acid and basic meta-igneous rocks and metasedimentary rocks. There is a general decrease in pre-Alpine metamorphic grade from the Ivrea Zone in the west (granulite facies) towards the Adamello in the east (greenschist facies). Along the northeast border of the Southern Alps, early Paleozoic rocks in low-grade greenschist facies and anchizone are present (Carnic Alps and Karawanken). This basement is discordantly overlain by unmetamorphosed Upper Carboniferous sediments which are intruded by Permian granites and associated extrusives. The thick sedimentary sequence ranges in age from Carboniferous to Tertiary.

II. ALPINE METAMORPHISM (PLATE II)

Plate II was constructed following the recommendations of the Commission for the Geological Map of the World (Zwart et al., 1967), with some minor modifications. Metamorphic facies are divided into three facies series dependent mainly on pressure (low, intermediate and high pressure series) and into four facies groups dependent mainly on temperature (laumontite and prehnite-pumpellyite facies group, greenschist facies group, amphibolite facies group, and two pyroxene (granulite) facies group). In this classification the glaucophanitic (blueschist) facies is included in a "greenschist facies group", a procedure which is not followed by all current facies schemes. Problems dealing with facies boundaries in specific areas are discussed in the text.

1. Western Alps

In the western part of the Alps the Alpine metamorphism produced three main groups of different mineral associations. An early high-pressure group (eclogites, glaucophane ± jadeite ± lawsonite schists, eclogitic schists and kyanite-bearing rocks) was followed by a second group of low- to intermediate-pressure belonging to the amphibolite facies in the Lepontine area and the greenschist facies from western Monte Rosa to the Ligurian Sea. A third paragenetic group is characterised by zeolites, prehnite and pumpellyite, and is restricted to the external part of the belt. The grade of Alpine metamorphism increases from the external to the internal parts of the belt.

a) High-pressure Facies: Eclogites, Glaucophane Schists ± Jadeite, Eclogitic Schists Kyanite-bearing Rocks

Bearth, 1962, 1970, 1974; Bocquet, 1971, 1974; Bocquet et al., 1974; Chatterjee, 1971; Compagnoni and Maffeo, 1973; Dal Piaz, 1974a, 1974b; Dal Piaz et al., 1972, 1973; Fry and Fyfe, 1971; Goffe et al., 1973; Mottana, 1971; Saliot, 1973; Steen, 1972; Viterbo and Blackburn, 1968.

High-pressure – low-temperature mineral assemblages are typical of the earliest Alpine metamorphic event. On the map, the boundary defining the area affected by this metamorphism has been drawn using the distribution of *sodic amphiboles*. Recent age determinations have shown that these amphiboles are not all the same age (Bocquet et al., 1974). These minerals are abundant in the Piemont Zone (except the Combin Zone s. l.[1])), in the internal portion of the Sesia-Lanzo Zone ("Micascisti eclogitici"), in the Austroalpine klippen, in the Bernhard-Briançonnais unit, the Sion-Courmayeur Zone (see p. 250) and in parts of the Monte Rosa–Gran Paradiso–Dora-Maira unit. They are present locally in the "seconda zona diorito-kinzigitica". In the Ligurian

[1]) This represents the autochthonous to parautochthonous Mesozoic cover series of the Upper Pennine basement from Monte Rosa, Gran Paradiso and Dora-Maira.

Alps glaucophane occurs in the "Gruppo di Voltri" and the schists of the Savona area NW of Genova.

In pelitic, marly and carbonate rocks, the main minerals formed during this high-pressure stage are: sodic amphibole, garnet, phengite and/or paragonite, chloritoid, lawsonite, jadeite, omphacite, rutile, and sphene. In the internal area the Na-amphiboles are partly glaucophane, in the external area it is glaucophane, ferroglaucophane or crossite in the micaschists, and crossite, magnesioriebeckite or riebeckite in the calcschists and carbonate rocks. Mesozoic or pre-Triassic basic and ultrabasic rocks contain mainly: glaucophane or crossite, lawsonite, pumpellyite, jadeite, aegirine, and occasionally kyanite and chloritoid.

Eclogites and glaucophanic eclogites are shown on the map by solid triangles. Although they do not form large continuous outcrops, their distribution is regional. They are principally located in the internal portion of the glaucophane zone (eastern part of the Piemont Zone, Monte Rosa–Gran Paradiso–Dora-Maira unit, and parts of the Sesia-Lanzo Zone (SCHEURING et al., 1974)). Their highest concentration is in the pre-Triassic basement of the southwestern part of the Sesia-Lanzo Zone ("Micascisti eclogitici"). Some eclogitic rocks display a partial metamorphic alteration into the greenschist facies.

In the Piemont Zone the eclogites were developed at the expense of the ophiolites. The eclogite-like parageneses occuring within the crystalline basement of the Monte Rosa–Gran Paradiso–Dora-Maira unit and the Sesia-Lanzo Zone originated from pre-Triassic basic to acid (!) rocks.

The Alpine high-pressure metamorphism which produced the eclogites is subdivided into at least two main stages: the first one with garnet, jadeite and/or omphacite, rutile; the second with glaucophane, phengite and/or paragonite, associated with the first distinctive Alpine schistosity. In some cases two or three separate stages of Na-pyroxene development are apparent.

Kyanite is occasionally present in both the pre-Triassic basement and the Mesozoic cover, sometimes together with chloritoid. In the basement, kyanite occurs often as pseudomorphs after sillimanite.

Paragonite is regionally distributed in the calcschists and occasionally in ophiolites, but also occurs in the pre-Mesozoic paraschists of the basement.

It is interesting to note the presence of large *jadeite* porphyroblasts in the presence of quartz in the Upper Paleozoic rocks of the internal Briançonnais Zone (Ambin, Acceglio, Vanoise). Jadeite also exists in the Sesia-Lanzo Zone, principally in the eclogitic metagranites, in the Versoyen area, in some ophiolites of the Piemont Zone, in Gran Paradiso, and in other areas. Its distribution may be wider than is currently known.

Lawsonite is mainly restricted to the southwestern part of the Alpine belt, and also occurs in the "Gruppo di Voltri" west of Genova and in the internal, southwestern part of the Sesia-Lanzo Zone. Pseudomorphs of zoisite, white mica and calcite after lawsonite are widespread. *Deerite* and *Ferrocarpholite* have been reported occasionally from very iron- and iron-aluminum-rich rocks, respectively.

It is important to note that glaucophane, paragonite, and lawsonite were not only formed during the earliest Alpine metamorphic event. Stratigraphic, petrographic and geochronologic evidence indicate that there are several later generations of these minerals.

b) Low- and Intermediate-pressure Facies: Greenschist Facies

Bearth, 1958, 1962, 1967; Bocquet, 1971, 1974; Bortolami and Dal Piaz, 1970; Chatterjee, 1971; Michard, 1967; Nicolas, 1969; von Raumer, 1971, 1974; Vialon, 1966; Wetzel, 1974.

This facies includes all the subsequent alterations to which the pre-Alpine and early Alpine metamorphic assemblages have been subjected, contemporaneous or not, and regardless of whether the modifications result from decreased pressure, increased temperature, variations in the chemical environment, or deformation.

Between Monte Rosa and the Ligurian Sea the early Alpine assemblages were overprinted and in part obliterated by a greenschist facies metamorphism, which generated chlorite, green biotite, blue-green amphibole, actinolite, stilpnomelane, paragonite and chloritoid, but also glaucophane (Hunziker, 1974; Bocquet et al., 1974). Some basic rocks were changed to "prasinites" or "ovardites"[2]), the eclogites to albite amphibolites.

The greenschist facies is mainly developed in the Penninic domain, but extends into the Helvetic domain. In addition, it is also developed in the external part of the Sesia-Lanzo Zone and in parts of the Austroalpine units. Towards the Ossola and Ticino culminations it grades from greenschist to amphibolite facies assemblages. The grade of the greenschist facies decreases progressively from the internal to the external portions of the belt. The episode of greenschist metamorphism generally appears to post-date the overthrusting of the middle Penninic crystalline units by the Calcschists nappe and of the Calcschists nappe by the Austroalpine nappe. It was roughly coeval with complex and variable deformation on all scales, aligned both parallel and normal to the arc of the Western Alps. One or more new schistosities developed in conjunction with it. The crystallization of certain minerals continued after the last main phase of deformation.

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[2]) Petrographical words used in Alpine geology to indicate amphibole-albite-epidote and chlorite-albite-epidote schists, respectively, which originated mainly from basalts or tuffs.

Albite is a common mineral in this facies. It is abundant in the paraschists (more or less phyllitic calcschists, micaschists, gneisses and marbles), metabasites (amphibole-albite-epidote schists, metadiabases, metagabbros, albite amphibolites), and the metamorphic derivatives of the Hercynian granites. In the latter, chessboard albite commonly substitutes for K-feldspar porphyroblasts. The growth of albite porphyroblasts, especially in phyllitic beds, is a late, frequently post-kinematic event. In the highest-grade part of the greenschist facies, albite commonly has an oligoclase rim.

c) Zeolite and Prehnite-pumpellyite Facies. Anchimetamorphism

Aprahamian, 1974; Artru, 1972; Artru et al., 1969; Dunoyer de Segonzac, 1969; Dunoyer et al., 1974; Kübler, 1969, 1974; Kübler et al., 1974; Martini, 1968, 1972; Martini and Vuagnat, 1970; Poty, 1969; von Raumer, 1974; Saliot, 1973; Sawatzki, 1974; Sawatzki and Vuagnat, 1971; Siddans, 1971.

In the Western Alps the boundary representing the transition from diagenesis to zeolite facies metamorphism does not always coincide with the boundary based on illite crystallinity which defines the beginning of anchizone metamorphism. The lower limit of the zeolite facies on the map is based on the presence or absence of laumontite where the appropriate rocks occur. In the absence of these rocks illite crystallinity has been used to locate the boundary. The question of the zeolite facies/anchizone criteria is discussed on p. 264.

A zeolite facies is known in the Helvetic domain in lower Tertiary greywackes containing volcanic material, lying west of Mont Blanc and south of the Pelvoux massif. The minerals formed from the outer part of the belt inwards are: heulandite, laumontite, prehnite and pumpellyite. Pumpellyite seems to post-date laumontite. The assemblages in the crushed zones are slightly higher grade than in the surrounding rock or its fissures: pumpellyite in the crushed zones of laumontite-bearing rocks; laumontite in the crushed zones of heulandite-bearing rocks. Coombs (1971) has proposed the following explanation for this phenomenon: in the crushed zones, where the solutions circulate in a system open to the surface, the water pressure would be less than lithostatic pressure and hence the less hydrous assemblages would appear.

Small pockets or veins containing pumpellyite, prehnite, heulandite and laumontite occur in the Dauphiné-Helvetic crystalline massifs. In basic rocks of the Aiguilles Rouges Massif laumontite, prehnite and pumpellyite also occur as rock-forming minerals. Stilpnomelane could, at least partly, belong to this prehnite-pumpellyite facies.

Prehnite and pumpellyite reappear in the vicinity of the Sestri-Voltaggio line and, farther to the east, in the Apennine ophiolites.

The outer boundary for the anchimetamorphism may be drawn just west of the Dauphiné-Helvetic crystalline massifs, and west of the French Helminthoid Flysch nappe (between Pelvoux and Argentera). South of the Durance river, the available data are sparsely distributed. Consequently, the proposed limit must be regarded as provisional. Its position may change when additional information is obtained.

The boundary between the anchizone and greenschist domain has been extrapolated from the available illite crystallinity data and is considered to be extremely provisional at this time. A few illite crystallinity measurements have been made around the Pelvoux Massif and in the Briançonnais Zone. Data are lacking in the northern (NE of the Belledonne Massif) and southernmost parts (Argentera Massif and Ligurian Alps) of the Western Alps. The first appearance of minerals of the Alpine greenschist facies, such as green biotite, blue-green amphibole, iron-rich chloritoid, etc., occurs further toward the internal side of the Alpine arc than the boundary located from illite crystallinity data.

2. Central Alps

Four different groups of mineral associations will be described from the Central Alps:

a) The metamorphic sequence in the Lepontine area ("Lepontinische Gneissregion", Wenk, 1956) and adjacent areas represents the classical Alpine metamorphic event in the Central Alps.
b) This intermediate-pressure facies series grades towards the east into a low-pressure facies series in the Bergell area.
c) On either side of the Lepontine area remnants of an earlier high-pressure metamorphic event are present. The eastern area in the Grisons will be described only, the western area was already described on p. 254–256.
d) A zone of very low-grade metamorphism occurs in the outermost parts of the Central Alps.

a) The Regional Metamorphic Sequence in the Lepontine and Adjacent Areas

Bearth, 1958; Blattner, 1965; Chadwick, 1968; Chatterjee, 1961; Egli, 1966; Evans and Trommsdorff, 1970; Fox, 1974; J. D. Frey, 1967; M. Frey, 1969, 1974; Frey and Orville, 1974; Graeser and Niggli, 1967; Hänny, 1972; Keller, 1968; Keusen, 1972; Möckel, 1969; Niggli E., 1960, 1970; Niggli and Niggli, 1965; Poty et al., 1974; v. Raumer, 1971; Schwander and Wenk, 1967; Sharma, 1969; Steck, 1966; Steck and Burri, 1971; Steiger, 1962; Stern, 1966; Thompson, 1974; Trommsdorff, 1966, 1968, 1972; Trommsdorff and Evans, 1974; Wenk, 1956, 1962, 1968, 1970; Wenk and Keller, 1969.

The first mineral zones in the Central Alps were described in 1960 by NIGGLI. Despite the complicated tectonic situation these mineral zones are arranged in a very simple manner. The mineral zones described below in some detail are arranged with increasing metamorphic grade.

The outermost *stilpnomelane zone* is covering parts of the Helvetic nappes, the northern part of the Aar Massif as well as the eastern part of the Grisons. Stilpnomelane is mainly restricted to rocks of granitic composition and iron- and calcium-rich, alumina-poor sediments (e. g. glauconite-bearing limestones, iron oolites). While the higher-grade boundary is fairly well known, its lower-grade boundary is still poorly defined. More details are given on p. 264.

Although almandine-rich garnets do not appear in pelitic rocks until the highest-grade chloritoid zone, *in granitic rocks* of the Aar Massif *garnets* with the unusual composition $\frac{1}{3}$ almandine, $\frac{1}{3}$ spessartine and $\frac{1}{3}$ grossularite appear already within the stilpnomelane zone (STECK, 1966). Very similar garnets were also described from the Mont-Blanc Massif (VON RAUMER, 1971). STECK and BURRI (1971) followed these garnets into zones of higher metamorphic grade. In granites of the Gotthard Massif (chloritoid zone), the Alpine garnets are slightly impoverished in the spessartine component but still grossularite-rich. Some of these garnets show atoll structures which may indicate the polyphase nature of Alpine regional metamorphism. In Pennine gneisses of the northern Lepontine region the garnets become impoverished in Ca due to the production of an An-richer plagioclase.

The *chloritoid zone* covers an area more or less outlined by the Gotthard Massif with its remnants of a sedimentary cover and an outlier to the east. Chloritoid is almost exclusively restricted to Mesozoic pelitic and marly rocks. The low-grade boundary of the chloritoid zone on the northern side of the Gotthard Massif is defined by rocks of unsuitable composition and not by metamorphic grade. This is clear from the fact that chloritoid is found north-east of the Gotthard Massif at the greenschist facies – anchizone boundary at the southern end of the Vättis Window. The low-grade chloritoid is found mainly with assemblages containing K-white mica, chlorite, paragonite and margarite with or without calcite and dolomite. At higher grade chloritoid-kyanite assemblages are quite common. Chloritoid-staurolite bearing meta-sediments occur over a range of several kilometers; from textural evidence it seems that staurolite is always later than chloritoid, therefore documenting the progressive character of this regional metamorphism.

The *kyanite zone* is limited to the north by the southern border of the Gotthard Massif, to the west and south by the Simplon-Centovalli fault and the Insubric line, respectively, and extends towards the east to a point 10 km west of the Bergell Granite. Kyanite persists into the stability field of sillimanite. The kyanite occurrence in the northern part of the Adula Nappe (working area of VAN DER PLAS) belongs presumably to the early, high-pressure Alpine metamorphic event and may be compared with the kyanite occurrences in the Western Alps (DAL PIAZ, 1971).

The staurolite zone coincides almost exactly with the kyanite zone. The greenschist/amphibolite facies boundary was drawn according to the staurolite isograd except southwest of the Simplon-Centovalli fault, where staurolite is not present, probably for chemical reasons. In this region, the albite-oligoclase transition was used to define the upper limit of the greenschist facies (BEARTH, 1958). Staurolite seems to coexist typically with muscovite, biotite, garnet, kyanite, plagioclase, paragonite and more rarely also with hornblende, chloritoid and margarite. A relatively narrow zone along the staurolite isograd on the southern border of the Gotthard Massif is remarkable for the large (5–10 cm) porphyroblasts in Mesozoic metasediments. It may be that the relatively large amount of volatiles produced at this metamorphic grade is the reason for this feature.

The *sillimanite zone* is restricted to the highest grade part of Alpine metamorphism, that is, an area located directly north of the Insubric line and extending from the Bergell to the west as far as 20 km west of Locarno. Sillimanite is fibrolitic everywhere except east of the Mera valley toward the Bergell Massif where it occurs as prismatic crystals.

The third alumosilicate, *andalusite*, is not a rock-forming porphyroblast in the Lepontine area but occurs in quartz nodules (sometimes together with kyanite and/or sillimanite) in topographically higher areas within the staurolite zone (KELLER, 1968; WENK, 1970; THOMPSON, 1974). However, rock-forming andalusite (together with cordierite) is common in a small area east of the Bergell Massif (WENK et al., 1974).

Cordierite in an unusual paragenesis with biotite, plagioclase, kyanite and quartz, is known from the Verzasca valley only (WENK, 1968) but becomes important in the Bergell area (WENK et al., 1974).

The staurolite and sillimanite zones may be further subdivided with aid of the *An-content in plagioclases coexisting with calcite* in Mesozoic Bündnerschiefer and marbles (WENK, 1962) and in epidote-bearing amphibolites (WENK and KELLER, 1969). The albite-oligoclase boundary is well defined, but for plagioclases with An $>$ 30 mol.-% the zones become less well defined and the zone boundaries cannot be regarded as isograds. It may also be mentioned here that plagioclase in margarite-bearing phyllites and schists of a Liassic black-shale formation around the Gotthard Massif shows an unusual feature: already the first formed plagioclase in the chlorite zone is oligoclase-andesine, not albite as one would expect (FREY and ORVILLE, 1974).

Siliceous carbonate rocks are well suited for mapping zones of progressive regional metamorphism in the Lepontine. With increasing metamorphic grade the following phases appear (TROMMSDORFF, 1966): talc, tremolite, diopside and forsterite. The tremolite-calcite isograd corresponds approximately to the first appearance of staurolite and kyanite in pelitic rocks, the paragenesis andesine-calcite and scapolite-calcite in calcareous rocks. The diopside-calcite and the forsterite-calcite isograds surround the sillimanite field. With a single exception (TROMMSDORFF, 1968), the assemblage quartz-calcite remains stable throughout the Lepontine, but wollastonite is common in the Bergell area.

A sequence of seven critical mineral assemblages has been recognized in regional metamorphosed *ultramafic rocks* of the Lepontine and Rhetic Alps (EVANS and TROMMSDORFF, 1970; see also TROMMSDORFF and EVANS, 1974). Unfortunately, this type of rock is not common enough to delineate a pattern of isograds. With increasing metamorphic grade these assemblages are as follows: diopside-serpentine-(brucite), diopside-serpentine-forsterite, tremolite-serpentine-forsterite entering approximately at the staurolite isograd, tremolite-talc-forsterite, tremolite-anthophyllite-forsterite, tremolite-enstatite-forsterite, and diopside-enstatite-forsterite.

The occurrence of *migmatites* in the Lepontine area is a hot subject and has produced many controversial discussions. Their distribution covers not only the entire sillimanite zone, but shows also some outliers to the north (e. g. Maggia valley). E. WENK believes that these migmatites were formed during the climax of Alpine metamorphism in the Lepontine. However, radiometric determinations by HÄNNY (1972, footnote p. 43) and GULSON (1973) give a Hercynian age for the migmatites west of the Bergell. The migmatites undoubtedly recrystallized during Alpine regional metamorphism and some mobilisation may have occurred. However, it is still an open question to what extent these migmatites formed originally during Hercynian or Alpine orogenesis.

Young *pegmatites* are scattered throughout the sillimanite zone.

Relicts of pre-Alpine structures and metamorphism are still preserved at grades as high as the lower-grade amphibolite facies. The eclogite occurrence of Alpe Arami may be a relict of the Early Alpine high-pressure event.

b) The Bergell Area

CORNELIUS, 1915; GYR, 1967; MOTICSKA, 1970; STAUB, 1958; TROMMSDORFF and EVANS, 1972; H. R. WENK, 1973; WENK et al., 1974.

The Bergell area, located between the Lepontine area to the west and the Rhetic Alps to the east and north, occupies a key position for the unravelling of the Alpine metamorphic history in the Central Alps (see e. g. MILNES, 1969; H. R. WENK, 1973).

Approaching the Bergell area from the Lepontine the Barrovian sequence of regional metamorphism grades into a low-pressure facies series with some characteristics of contact metamorphism (TROMMSDORFF, 1966, p. 445; NIGGLI, 1970; WENK, 1970, p. 37). 10 km west of the Bergell Granite staurolite and kyanite become rare while cordierite, prismatic sillimanite, andalusite and wollastonite are abundant. In Val Codera sapphirine occurs in an assemblage containing orthopyroxene, sillimanite and cordierite. ACKERMAND and SEIFERT (1970) suggest on the basis of hydrothermal experiments that this rock was probably formed under granulite facies conditions. These authors conclude that these rocks were emplaced tectonically into the low-pressure terrain.

The granitic rocks of the Bergell Massif (Bergell granodiorite and tonalite, Novate granite) have certainly been formed at some time during the Alpine orogeny. There is, however, a rigorous controversy still going on with respect to the origin, the evolution, the age relations of the crystallization with tectonics and metamorphic phases, and the genetic relationships between the various members of the Bergell suite. Even some published data obtained with different methods are contradictory.

Recent geochronological work indicates that the granitic rocks of the Bergell Massif may be younger than the peak of the "Lepontine phase"[3]) of Alpine metamorphism (see p. 271). This possibility was already suggested by GYR (1967) as well as by NIGGLI (in JÄGER et al., 1967, p. 46). TROMMSDORFF and EVANS (1972) found in serpentinites in the aureole of the Bergell tonalite a contact metamorphism superimposed on an older regional metamorphism; this latter may, however, belong to an older, Cretaceous phase of Alpine metamorphism.

The radiometric age determinations are in agreement with the view that the granitic rocks of Bergell intruded a pile of already emplaced nappes (CORNELIUS, 1915; STAUB, 1958; GYR, 1967). Some of the Bergell rocks, however, show a strong foliation, particularly the Novate granite and the Melirolo tonalite-gneiss, the western continuation of the Bergell Massif. Some deformation may have occurred during or after the emplacement.

After H. R. WENK (1973) the Bergell "granite" is a deformed concordant plate which was emplaced together with the nappes. Its primary formation would therefore pre-date nappe formation. WENK believes that the "granite" was mobilized early in the Alpine orogeny and intruded sedimentary rocks of the Alpine geosyncline. However, the contacts between the Bergell "granite" and the country rock are also compatible with a lopolith or ethmolith shaped intrusive body. WENK interpreted Rb-Sr mica ages from west of the Bergell as representing the climax of metamorphism. JÄGER et al. (1967) concluded that these ages represented cooling ages.

Although the relationship between the low-pressure amphibolite facies

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[3]) The term "Lepontine metamorphic phase" as used in this paper includes only those rocks affected by the mid-Tertiary metamorphic event which lie within the area enclosed by the mineral zone boundaries of NIGGLI (1970). The term is not restricted to rocks which occur within the Lepontine gneiss region as defined by WENK (1956).

metamorphism, the Bergell "granite", and the young pegmatites is not yet clear, some genetic connection is very probable.

c) "The Grisons"
(eastern part of the Central Alps)

Bucher and Pfeifer, 1973; Dietrich, 1969; Dietrich and Peters, 1971; Gansser, 1937; Niggli, 1970; Niggli and Niggli, 1965; Peters, 1963; Peters et al., 1973; van der Plas, 1959; Thum and Nabholz, 1972; Trommsdorff et al., 1970.

In some parts of the Grisons there is evidence for an early Alpine, high-pressure event. In the northern Adula Nappe the following minerals are typical for this event: glaucophane, crossite, sodium-pyroxene and garnet. Phengite and kyanite probably also belong to this event. The same minerals (except kyanite) are also present in the northern part of the Tambo Nappe. In the Oberhalbstein glaucophane is widespread.

Absolute age determinations suggest that this early event in the Grisons can be correlated with the high-pressure event in the Western Alps (see p. 274). It may be speculated that the Tessin area was also affected by this early Alpine event but all evidence was completely obliterated by the later "Lepontine phase" of metamorphism.

Parts of the Grisons were later affected by a second or even third Alpine metamorphic phase, probably related to the "Lepontine event". This is demonstrated, for example, by the zoning of amphiboles in the northern Adula Nappe. The early sodium amphiboles are first rimmed by a blue-green amphibole and then by actinolite. In both the Pennine units and the Austroalpine nappes the metamorphic grade of this later phase(s) belongs to the zeolite facies and greenschist facies, generally with metamorphic grade increasing towards the south and southwest. However, the possibility that some of these metamorphic effects belong to the above mentioned early Alpine high-pressure event cannot be excluded. The progressive character of this low-grade metamorphism is documented in manganiferous radiolarian cherts (Trommsdorff et al., 1970; Peters et al., 1973), in serpentinites and associated rocks (Peters, 1963; Dietrich, 1969; Dietrich and Peters, 1971) and in argillaceous marly rocks and sandstones of pre-flysch and flysch type (Thum and Nabholz, 1972).

d) Helvetic Nappes, Prealpes

Durney, 1974; Frey, 1969, 1970; Frey and Niggli, 1971; Frey et al., 1973; Jäger et al., 1961; Kübler, 1970, 1974; Martini, 1972; Martini and Vuagnat, 1965, 1970; Niggli et al., 1956; Niggli and Niggli, 1965.

Since a very low-grade metamorphism in the most external parts of the Alps has been detected only recently, our knowledge is limited. Detailed studies have been made in the Savoy Alps west of the Mt. Blanc and Aiguilles Rouges Massifs (see p. 257) and the Glarus Alps. However, in the region between these two areas the proposed limits must be regardes as provisional.

As in the Western Alps, the lower-grade limit of the zeolite facies is drawn according to the presence of laumontite in the Taveyanne sandstone. In the absence of this formation measurements of the illite crystallinity in shales and slates have been used to define the boundary. However, the boundaries based on these two different criteria do not represent the same metamorphic grade. Illite crystallinity measurements in shales intercalated with laumontite-bearing graywackes show that the beginning of the zeolite facies lies well within the diagenesis zone (= unmetamorphosed zone) and therefore outside the anchizone as defined by the illite crystallinity. The boundary between the prehnite-pumpellyite facies and the greenschist facies coincides approximately with the higher-grade limit of the anchizone.

In the *Taveyanne sandstone*, a detrital series of Eocene-Oligocene age which includes graywackes of andesitic composition, the following mineral assemblages can be observed with increasing metamorphic grade: laumontite + albite + chlorite + quartz + corrensite, prehnite + pumpellyite + albite + chlorite + quartz ± epidote, pumpellyite + actinolite + epidote + albite + chlorite + quartz.

Spilitic rocks belonging to the *Verrucano of the Glarus Alps* (60 km SE of Zürich) do not show any diagnostic assemblages from the zeolite and prehnite-pumpellyite facies. Combined with illite crystallinity data these rocks are placed within the greenschist facies. The Permian Verrucano is separated from the underlying anchimetamorphic Cretaceous and Tertiary Flysch by the Glarus overthrust. The implications of this inversion in metamorphic grade will be discussed on p. 277.

In *shales and slates* of the anchizone, in addition to illite and chlorite, common mineral phases include pyrophyllite, mixed-layer paragonite/muscovite, and paragonite. Besides the formation of new minerals in the anchizone, other changes in pelitic and marly rocks observed include: 1 Md illite changes to 2 M_1 muscovite; the slates change colour from red to pink due to the increasing solid solution of TiO_2 in hematite; and textural changes occur due to reactions between clastic quartz and clay cement.

Glauconite- and quartz-bearing limestones of the Glarus Alps show the following mineral sequence with increasing grade within the anchizone and the beginning of the epizone: glauconite ± chlorite, glauconite + stilpnomelane + k-feldspar ± chlorite, stilpnomelane + k-feldspar ± chlorite, and biotite + stilpnomelane + k-feldspar. The first stilpnomelane appears near the middle of the anchizone while the first biotite (brown and/or green) enters at the boundary between the anchizone and greenschist facies. Riebeckite may be an additional phase in all assemblages mentioned above.

3. Eastern Alps

While the description of Alpine metamorphism in the Western Alps was subdivided on the basis of different mineral assemblages, in the Central Alps the metamorphism in each region was described. In the Eastern Alps the available knowledge favours a subdivision of the description based on tectonic units and then discussion of each tectonic unit in different regions.

a) Middle and Upper Austroalpine Units

The metamorphism of the main Austroalpine thrust mass can be considered to be of early Alpine age. Only locally near the Penninic units, however, can evidence of a younger event be found. The early Alpine metamorphism belongs to the intermediate-pressure series and took place when the Austroalpine domain was still located south of the Penninic domain.

1. Upper Austroalpine

BAUER et al., 1969; CORNELIUS, 1952; METZ, 1954; MOSTLER, 1968, 1972.

Only the very base of the *Northern Limestone Alps* (the Permoskythian rocks), particularly on the southern side, was affected by Alpine metamorphism. Metamorphic grade reaches the anchizone and the lower-grade greenschist facies. Only rarely do higher stratigraphic levels show noticeable alteration (e. g. N Landeck, E Tennengebirge). The knowledge of the low-grade metamorphism is highly provisional at present.

The greenschist facies metamorphism of the *Grauwackenzone* is usually considered to be a pre-Alpine event. However, in metamorphic Paleozoic rocks chloritoid post-dating Alpine (?) cleavage planes (Mitterberg) and the slightly metamorphic base of the Limestone Alps are evidence for an Alpine event which cannot always be distinguished from the older metamorphism.

2. Middle Austroalpine

The variable grade of Alpine metamorphism of the Middle Austroalpine is reflected in the Permo-Mesozoic cover. The effects of this event in the polymetamorphic crystalline basement are uncertain in many cases.

W of the Tauern Window

ANDREATTA, 1954; BAUMANN et al., 1971; DE VECCHI et al., 1971; GRAUERT, 1969; GREGANIN and PICCIRILLO, 1972; LANGHEINRICH, 1965; PURTSCHELLER, 1971; PURTSCHELLER et al., 1972; SASSI, 1972; SCHMIDEGG, 1964; SCHMIDT, 1965; ZANETTIN, 1971; ZANETTIN and JUSTIN-VISENTIN, 1971.

In the northwestern part of the Silvretta-Ötztal mass there is no evidence for a distinct Alpine recrystallization. Southeastward, the grade of Alpine metamorphism increases from anchizone in the west to greenschist facies with biotite in graphitic phyllites in the Stubai Mesozoic rocks. Metamorphic grade further increases (phengite, biotite, garnet, microcline) toward the southern part of the Ötztal mass where the Schneeberg belt presents a controversial problem. Mineral assemblages in Paleozoic rocks of this prominent micaschist belt are interpreted to be newly formed during an Alpine metamorphic event or partly affected by this event. This event has also affected the neighbouring pre-Alpine crystalline basement and formed new biotite, garnet and plagioclase in some zones. The western limit of this strong Alpine influence and why this zone is relatively narrow is not known.

S of the Tauern

Baggio et al., 1971; Borsi et al., 1973; Paulitsch, 1960; Sassi and Zanferrari, 1972; Sassi et al., 1973.

The Alpine metamorphism within the crystalline basement is recognizable only along the northern border zone, where late Alpine biotite ages are known. An early Alpine thermal effect had a wider distribution along this zone, but then terminates rapidly towards the south and the Permo-Mesozoic of the Gailtal Alps is unmetamorphosed. Along the Alpine-Dinaric Line a weak Alpine thermal effect is present.

E of the Tauern

Angel, 1924; Beck-Managetta, 1970; Becker and Schuhmacher, 1972; Clar, 1965; Exner, 1967, 1972; Exner and Faupl, 1970; Formanek, 1964; Formanek et al., 1962; Fritsch, 1962, 1965; Heritsch, 1963; Hinterlechner-Ravnik, 1971; Metz, 1953, 1958; Metz et al., 1964; Neugebauer, 1970; Pilger and Weissenbach, 1970; Richter, 1973; Thurner, 1970; Tollmann, 1971.

Most of the greenschist facies overprinting on the crystalline basement is considered to be an Alpine event. This event is reflected in the Mesozoic cover (phengite, sometimes biotite) as well. Similar to the case in the Stubai area (see above), in the Stangalm area early Alpine and late Alpine thrust movements have disturbed the original metamorphic pattern in the sedimentary cover. Alpine metamorphism is not found in the uppermost sedimentary units. The maximum temperatures attained during the early Alpine metamorphism are not known, but it is possible that temperatures typical of the lower amphibolite facies were reached in the deepest parts of the crystalline basement. Similarities in rock sequence and mineral assemblages suggest that the micaschists of Radenthein may represent a continuation of the Schneeberg belt.

b) Lower Austroalpine and Penninic Units

1. Semmering-Wechsel-Rechnitz Window

Evren, 1972; Faupl, 1972; Wiesenseder, 1971.

The penetrative character of the Alpine overprinting on older metamorphic and granitic rocks is well developed in this area. Abundant phengite in Permo-Mesozoic rocks and older granites indicates that Alpine facies series may be the intermediate pressure type. Metamorphic grade does not exceed the greenschist facies. A slight increase in grade with depth is recognizable.

In the Mesozoic Penninic calcschist-ophiolite sequence of the Rechnitz-Bernstein Window glaucophane-bearing rocks are common and glaucophane was also found in Lower Austroalpine rocks. Remnants of a transgressive unmetamorphosed Upper Eocene indicate the end of erosion in this region when metamorphism was still continuing in the same unit further to the west.

It seems possible that the imprint of an early Alpine metamorphism is also present in parts of the Lower Austroalpine unit exposed in the other windows (especially in the Tauern Window) but no data are available to differentiate the effect of the early and late Alpine phases. Stronger evidence for the presence of the early phase exists in the Lower Austroalpine Err-Bernina unit between Eastern and Central Alps.

2. Tauern Window

Ackermand and Karl, 1972; Angenheister et al., 1972; Baggio and De Vecchi, 1966; Beran, 1969; Bianchi and Dal Piaz, 1934; Cliff et al., 1971; Cornelius and Clar, 1939; De Vecchi et al., 1971; De Vecchi and Picirillo, 1968; Exner, 1957, 1971, 1974; Exner and Faupl, 1970; Fischer and Nothaft, 1954; Frasl, 1958; Frasl and Frank, 1964, 1966; Friedrichsen et al., 1973; Häberle, 1969; Höck, 1974; Hoernes, 1973; Hoernes and Friedrichsen, 1974; Hörmann et al., 1973; Karl, 1959; Karl and Schmidegg, 1964; Koark, 1950; Meixner, 1958; Miller, 1974; Morteani, 1971; Oxburgh, 1968; Raith, 1971; Sander, 1929; Sassi, 1972; Scharbert, 1954; Schmidegg, 1961.

In the Tauern Window a mineral zone sequence indicating increasing Alpine metamorphism is well developed. Stilpnomelane-bearing rocks occur only in its external parts. The metamorphic grade rises toward the interior and in the vicinity of the gneiss domes staurolite appears in Mesozoic rocks. In the area of the gneiss domes epidote and oligoclase coexist over large areas. Clearly, physical conditions corresponding to the amphibolite facies were attained. The exposed parts of the Tauern Window have nowhere been subjected to Alpine anatexis. Phengite is the abundant white mica. Graphitic schists and quartzites, where kyanite sometimes appears even before biotite, correspond to the greenschist facies. A typical assemblage in the carbonate-poor Bündnerschiefer phyllites is: potassic white mica – paragonite – chlorite – quartz – albite ± garnet ± calcite. The chemistry of these rocks does not favour the appearance of biotite (see Höck, 1974). Chloritoid is widespread in rocks of Keuper-, Liassic- and also Lower Permian age and at higher temperatures often associated with garnet. In siliceous dolomites only tremolite occurs in the central part of the Tauern Window as a common mineral. The assemblage of the basic rocks show a transition from actinolite – epidote – albite in the external parts to blue-green hornblende – epidote – albite to blue-green hornblende – epidote – plagioclase ± almandine in the internal parts. The crystallization of all these assemblages has occurred after the deformation. In the uppermost tectonic levels of the Pennine in the Tauern Window syn- to pretectonic crystallization prevails in connection with a further tectonic movement of the Austroalpine nappes. This intermediate-pressure type Alpine metamorphism appears to be uniformly progressive. Detailed investigations on the mineral chemistry revealed the effects of a late heating phase in the central part. Geologic evidence suggests an estimated thickness of cover not greater than 15 km corresponding to a pressure of 5 kb and petrologic evidence, a temperature of 500–550° C at the base of the Mesozoic sequence.

In the southern part of the Venediger and Glockner groups a narrow zone of glaucophane-bearing rocks, containing small lenses of eclogitic rocks has been preserved. These rocks have been derived from well-bedded basic tuffaceous metavolcanics and minor lavas and their mineral assemblages and transformation processes resemble those known from the western Pennine regions (see Miller, 1974). The retrogressive replacement of this high-pressure assemblage, especially of clinopyroxene, garnet and glaucophane, is most likely the result of a continuously increasing temperature and decreasing pressure rather than a second episode of both deformation and metamorphism. This is concluded from the fact that both the high-pressure and the intermediate-pressure assemblages are syntectonic to post-tectonic with respect to the pronounced crossfolding event in the middle part of the Tauern. The "white pseudomorphs" in prasinitic rocks, now interpreted as pseudomorphs after lawsonite, suggest a formerly more widespread occurrence of an earlier low-temperature stage of the high-pressure event.

The formation of this relatively early high-pressure event can be correlated with the subduction of the Penninic domain beneath the Austroalpine units. Later, still at high pressure, rising temperature caused the formation of the narrow "eclogite zone". Development of the common mineral assemblages usually post-dates that of the high-pressure mineral assemblages and probably occurred at still high T and lower P.

3. Engadine Window

Kläy, 1957; Thum, 1969.

The Mesozoic rocks of the Engadine window probably only show late Alpine low-grade assemblages corresponding to the chlorite zone in pelitic rocks. Except for the occurrence of pumpellyite + actinolite no evidence for a high-pressure metamorphism is known.

III. RELATIONSHIPS BETWEEN CRYSTALLIZATION AND DEFORMATION

In order to show how much advance has been made in this field, we have chosen, as an example, the border region between the Helvetic and Pennine zones in the Central Alps, where several theses have been completed in the last few years under the guidance of Prof. J. Ramsay.

For the reader who would like to study the relationships between mineral growth and folding in more detail, the following list includes most of the important references (see also Ayrton and Ramsay, this volume).

Western Alps: Ayrton, 1969; Bearth, 1952; Bertrand, 1968; Caron, 1974; Chatterjee, 1963; Cornelius, 1930; Dal Piaz and Sacchi, 1969; Ellenberger, 1958; Gay, 1972; Laduron, 1974; Michard, 1967; Nicolas, 1969; Porada, 1966; Plessmann and Wunderlich, 1961; Siddans, 1971; Tricart, 1974; Vialon, 1967; Wunderlich, 1963.

Central Alps: Chadwick, 1965, 1968; Chatterjee, 1961; Durney, 1972; Durney and Ramsay, 1973; Hall, 1972; Higgins, 1964; Kvale, 1957; Labhart, 1965; Milnes, 1965; Nabholz and Voll, 1963; Plessmann, 1957, 1958; Sibbald, 1971; Steck, 1968; Steiger, 1962; Tan, 1969; Thakur, 1971, 1973; Wunderlich, 1957, 1958.

Eastern Alps (Tauern Window only): Barth et al., 1973; Cliff et al., 1971; Exner, 1957; Frasl and Frank, 1964; Karl, 1952, 1954; Matura, 1967; Nollau, 1969; Sander, 1914, 1942, 1950; Schmidegg, 1961; Schwan, 1965.

Chadwick (1965, 1968), working in the Lukmanier region, distinguished two phases of Alpine deformation, Phase B (earlier) and Phase V (later). Both phases of folding occurred after the main northward transport of the lower Pennine nappes since the Lucomagno Massif thrust sheet, the overthrusted Penninic Bündnerschiefer in adjacent areas and the autochthonous to parautochthonous Mesozoic cover of the Gotthard Massif all underwent the same deformation history. Minor structures related to nappe emplacement were not observed. CHADWICK found the following interesting relationships between phases of deformation and metamorphic mineral growth across a 6 km long cross section. In the north, upper greenschist facies conditions were reached mainly during Phase B deformation while in the south the peak of metamorphism (that is lower-grade amphibolite facies) occurred between Phase B and Phase V. The intensity of both deformation phases increases from north to south parallel with increasing metamorphic grade. These observations were interpreted by CHADWICK in two ways: a) assuming the deformation occurred simultaneously across the area this would imply that the metamorphic maximum migrated from north to south, b) assuming the metamorphic maximum was simultaneous throughout the area this would mean that zones of deformation migrated from south to north. Since both deformation and metamorphic grade are more intense in the south than in the north, CHADWICK favoured possibility b). Alternatively, in the authors' view, the observations could also be explained by a simultaneous northward migration of both deformation and metamorphism, but deformation moving faster than metamorphism.

SIBBALD (1972), working in the Val Piora area directly west of CHADWICK'S area, recognised four episodes of Alpine deformation (D_1–D_4). During the first phase of deformation the allochthonous Pennine series were thrust over the residual autochthonous cover of the Gotthard Massif. Structures formed during phases D_2 and D_3 in the Val Piora area were described as Phase B structures by CHADWICK, those formed during D_4 conform with Phase V structures in the Lukmanier area. SIBBALD found no evidence to support CHADWICK'S hypothesis of migrating deformation episodes. On the other hand SIBBALD gives strong evidence for the progressive nature of Alpine metamorphism. Lower greenschist facies conditions were reached during the first episode of Alpine deformation while lower almandine-amphibolite facies conditions prevailed late in the third and early fourth phases of deformation times.

THAKUR (1973) has made a detailed study of structure and metamorphism in the Molare region, some 12 km south of Lukmanier pass. His results are in good agreement with those of SIBBALD and CHADWICK. THAKUR recorded evidence of a first structural phase associated with the formation of the lower Pennine nappes which were refolded by two later deformation events. Synkinematic mineral growth occurred during the two first deformation phases while porphyroblasts grew in between phase two and three. Correlation of deformation events described by various authors from the Central Alps were given by SIBBALD (1971, table 3,3) and THAKUR (1973, table 1).

THAKUR (Fig. 3) also made an attempt to correlate deformation phases with absolute age determinations from the Central Alps. Although such an attempt may be stimulating, the present authors believe that such a correlation is premature. THAKUR correlates his first nappe-forming deformation phase with whole rock age data ($\approx$ 120 my) from strongly deformed Monte Rosa gneisses. It seems questionable that correlations can be made over a distance of 75 km although similar ages are also known from tectonites of the Suretta Nappe, some 40 km to the east, and at other localities (see below). The second deformation phase is correlated with STEIGER'S (1964) K-Ar data on oriented hornblendes which gave 46 my. Since this early age work no further evidence for such an event has been found and the meaning of these data are still open to discussion. THAKUR (1973, p. 561) assumes that "Rb-Sr ages on biotites probably date the third phase deformation in Pennine zone". These biotite ages show a regular decrease from 15–18 my in the Molare region to < 12 my in the Simplon area (JÄGER et al., 1967) and were interpreted as cooling ages connected with the uplift (CLARK and JÄGER, 1971). They therefore cannot be interpreted as the age of the third deformation phase.

IV. AGE DETERMINATIONS (FIG. 1)

Age data are discussed separately for four different regions where geochronological detailed work has been done.

1. Western Alps

Ages older than 170 my in the Ivrea Zone are interpreted as cooling ages following the Hercynian metamorphism (HUNZIKER, 1974; see also GRAESER and HUNZIKER, 1968; MCDOWELL and SCHMID, 1968; MCDOWELL, 1970). A maximum geochronological age for the Alpine metamorphism is given through the intrusion/extrusion of ophiolites from the Prealps, Savoy Alps, dated with amphiboles (K-Ar) between 140 and 180 my (BERTRAND, 1970). Younger hornblende ages of the ophiolites may be interpreted as an early anorogenic oceanic metamorphic phase (DAL PIAZ, 1974).

Indications for an early Alpine tectonic phase between 110 and 130 my are indicated by total rock ages (Rb-Sr) in the frontal part of the Mt. Rosa Nappe as well as in the Sesia basement (HUNZIKER, 1970, 1974).

Randomly oriented glaucophane and crossite in Mesozoic ophiolites of the Piemont Zone yield K-Ar ages of 80–100 my (HUNZIKER, 1974; BOCQUET et al., 1974). In the "micascisti eclogitici" of the Sesia-Lanzo Zone phengites gave K-Ar ages between 60 and 90 my and Rb-Sr ages between 60 and 80 my (HUNZIKER, 1974). This 60 to 100 my age group of alkali-amphiboles and K-white micas is interpreted to belong to an Upper Cretaceous high-pressure, low-temperature phase of metamorphism.

In the Bernhard-Briançonnais unit and in the Piemont Zone alkali-amphibole and phengite ages (K-Ar) range from 40 to 50 my (Bocquet et al., 1974). From the approximate coincidence of the northern boundary of the staurolite zone with the limits of the field of young muscovites in the Lepontine area Jäger et al. (1967) deduced that the closing of the Rb-Sr system for Alpine K-white micas occurred between 400 and 500° C. Jäger (in Jäger et al., 1967) further deduced that Rb-Sr K-white mica ages from the greenschist-amphibolite facies transition indicate the age of the climax of a regional metamorphic phase. On this basis Hunziker (1969) concluded that the peak of the mid-Tertiary metamorphic phase in the Monte Rosa Nappe occurred approximately 38 my ago. In the Piemont ophiolites of the Zermatt region paragonite ages between 30 and 34 my (K-Ar) were found (Hunziker, 1974).

The youngest mineral age group was found in the Piemont Zone, the pre-Triassic basement, and the Briançonnais cover of the Western Alps; alkali-amphiboles and biotites yielded ages between 15 and 30 my (K-Ar and Rb-Sr), Bocquet et al., 1974.

2. Central Alps

The intensity of the mid-Tertiary phase of metamorphism obliterated evidence of the early Alpine history in the *Lepontine area.* However, the frontal part of the Monte Leone Nappe provides evidence for a 110–130 my event (total rock, Rb-Sr), see Jäger et al., 1969.

The blocking temperature for the Rb-Sr system in Alpine white micas is considered to be $500 \pm 50°$ C (Jäger, 1973). On this basis Jäger suggests that the peak of the Lepontine metamorphic phase occurred at about 35 my. This value agrees quite well with geological evidence (Trümpy, 1973a, b). Younger ages from biotite and muscovite (Rb-Sr and K-Ar) ranging from 32 to 8 my are interpreted as cooling ages. Three biotite samples from fissures in the Simplon area have the same age as that in the surrounding country rock. As the fissure minerals are known to have crystallized later than country rock biotite, the ages obtained from the country rock must be cooling ages. The regular decrease of these ages from east (Bergell area) to west (Simplon area) and from south (Insubric Line) to north (Simplon area) is concluded to be the result of post metamorphic uplift (Jäger et al., 1967; Purdy and Jäger, in preparation).

Alpine fissure minerals (muscovite, adularia and rarely biotite) gave K-Ar ages older than 10 my (Arnold, 1972; Purdy and Stalder, 1973). Muscovite ages range from 10 to 18 my. Rb-Sr data on some of these micas (Jäger et al., 1967) indicate that older values from adularia (up to 64 my) are probably due to Ar overpressure.

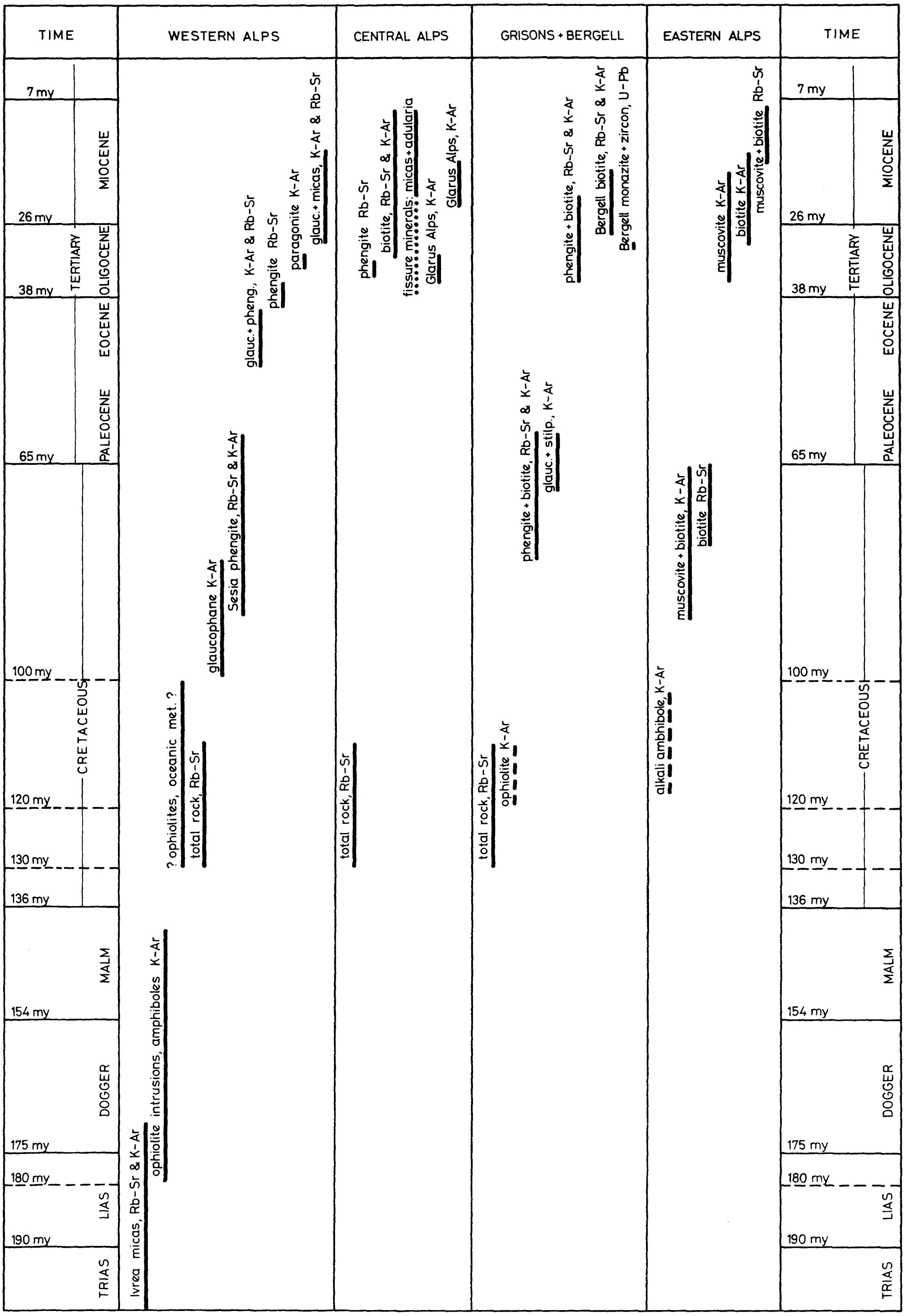

Fig. 1. Alpine geochronology.

The *Glarus Alps* is the only area where an attempt to date lowest-grade metamorphism of the Alps has been made. K-Ar ages of one riebeckite, one stilpnomelane, and one mixture of stilpnomelane and glauconite lead FREY et al. (1973, p. 213) to the conclusion that the peak of metamorphism occurred between 31 and 36 my ago. Several illites of the same region gave similar K-Ar values (unpublished results). It seems possible that these ages are related to the thermal culmination in the Lepontine region. However, stratigraphic evidence indicates that the time span between sedimentation and metamorphism in the Glarus Alps was very small (0–5 my). K-Ar ages on illites and stilpnomelanes between 14 and 26 my point to a second metamorphic (?) event during the Lower Miocene.

3. Grisons and Bergell

Tectonites in the frontal part of the Suretta Nappe indicate a tectonic event between 110 and 130 my (total rock, Rb-Sr), HANSON et al., 1969.

K-Ar data on one amphibole of the Platta ophiolites (112–115 my) was interpreted as the age of emplacement of the ophiolites (DIETRICH, 1969). Since there is no proof that the dated amphibole belongs to the primary mineral assemblage (DIETRICH, 1969, p. 98), the age of 113 my could as well mean the age of an oceanic metamorphism (see p. 271).

In the Upper Pennine nappes north and east of the Bergell K-Ar and Rb-Sr ages on phengites and biotites between 60 and 80 my were found (JÄGER, 1973). One stilpnomelane and one alkali-amphibole yielded K-Ar ages of 63 and 58 my, respectively. These data point to an Upper Cretaceous metamorphic phase between 60 and 80 my.

In the Penninic nappes surrounding the Bergell Rb-Sr and K-Ar ages on phengites and biotites of 22–36 my were found, indicating the "Lepontine phase" of metamorphism with subsequent cooling and in some cases also possible effects of the contact metamorphism. In the Bergell pluton biotite ages of 18–28 my were obtained (Rb-Sr and K-Ar), JÄGER et al., 1967, 1969; JÄGER, 1973. U-Pb data on zircons, sphene and monazites of 30 my (GULSON and KROGH, 1973) were interpreted as the age of crystallization for the Bergell pluton. Rb-Sr and K-Ar data on biotites from Bergell-type boulders in the molasse of the Po plain gave ages of 28 ± 0.5 my (JÄGER, 1973; and unpublished data). U-Pb data on zircons from the same boulders show "that they are undubitably from the Bergell massif" (GULSON and KROGH, 1973, p. 302). The stratigraphic age (fossils!) of this molasse is given as Lower to Middle Oligocene (CITA, 1958). According to currently accepted absolute time scales the sediment containing the boulders would be older than the Bergell boulders! Either the time scale or the paleontological determinations or both must be erroneous.

A poorly defined whole rock Rb-Sr isochron of approximately 25 my was obtained for the Novate granite (GULSON, 1973). This granite is considered to be the youngest rock of the suite also by field evidence.

4. Eastern Alps

Two alkali-amphibole mineral fractions of one sample from a basic dike in the Triassic of the Hallstädter Zone gave K-Ar ages between 102–118 my (unpublished results). According to JÄGER (1973, p. 18) this could be an indication for an early Alpine tectonic phase in the Northern Limestone Alps.

In the Middle Austroalpine crystalline basement as well as in the Schneebergzug, muscovite and biotite K-Ar ages range from 65 to 90 my (OXBURGH et al., 1966; HARRE et al., 1968; BREWER, 1969; LAMBERT, 1970; CLIFF et al., 1971). Rb-Sr ages on biotites of the same regions are grouped more narrowly between 65 and 80 my (OXBURGH et al., 1966; SCHMIDT et al., 1967; MILLER et al., 1967; HARRE et al., 1968; CLIFF et al., 1971). According to CLIFF et al. (1971) these ages represent the cooling after an early Alpine metamorphic phase in the Eastern Alps. The higher K-Ar ages may be due to some Ar overpressure.

Within the Tauern Window muscovite K-Ar ages range from 18–36 my and biotite K-Ar ages from 15–30 my (BESANG et al., 1968; LAMBERT, 1970; CLIFF et al., 1971). Biotite and muscovite Rb-Sr data between 8 and 17 my were obtained (CLIFF et al., 1971; SATIR, 1974). Again the higher K-Ar mica ages can best be interpreted to mean Ar overpressure. The Rb-Sr mica ages represent the cooling after the Tauern crystallization.

V. CONCLUSIONS (A GENETIC INTERPRETATION)

On the preceding pages some petrographic and geochronologic data bearing on the metamorphic history of the Alps are presented. However, to decide whether or not the metamorphic events coincide with an orogenic phase the tectonic history and stratigraphy of the Alps must also be considered. Excellent summaries of this data include: for the Western Alps, DEBELMAS (1963a, b); for the Central Alps, TRÜMPY (1973a, b); and for the Eastern Alps, OBERHAUSER (1968) and TOLLMANN (1973).

Radiometric data (more than 500 apparent ages) do not as yet permit the *precise* dating of the individual Alpine metamorphic phases and the various assemblages observed throughout the Alps. These data indicate an age between 60 and 100 my for an early blueschist phase. This *Eoalpine metamorphic event*

(Dal Piaz et al., 1972) is continuous from Genova to the Simplon area. It can be observed again in the Grisons, and further east in the region of the Tauern Window. It may well be that the whole Alps were affected by this blueschist phase but the evidence has been obliterated in some areas by later metamorphic events. Eclogites belonging to this metamorphism are common in the Western Alps and also occur in the Eastern Alps. In the Western Alps the physical conditions for the eclogite-bearing rocks probably ranged between 200 and 400° C and pressures in excess of 8 kb. In the main mass of the Austroalpine thrust sheet in the Eastern Alps, medium pressures prevailed during the Eoalpine metamorphism.

The tectonic loading observed or estimated in the Western Alps is inadequate to account for the high-pressure assemblages. In some of the authors' opinion the most plausible explanation currently available is based on the plate tectonics hypothesis. Several interpretations have been proposed to explain the Alpine tectonism (Dewey and Bird, 1970; Laubscher, 1970; Hunziker, 1971; Ernst, 1971, 1973; Martini, 1972; Dal Piaz et al., 1972; Saliot, 1972). However, no simple model of plate consumption can explain satisfactorily the complexity of the observed geological data.

The classical metamorphic sequences in the Central Alps (Lepontine) and Eastern Alps (Tauern Window) belong to a second metamorphic event which attained amphibolite facies conditions. In other regions of the Alps metamorphic grade during this phase did not exceed that of the greenschist facies. Hunziker (1969, 1970) concluded that the age of the climax of this metamorphic phase in the Monte Rosa region was approximately 38 my.

Two different explanations have been suggested for the *mid-Tertiary metamorphic phase* in the Central Alps. After Wenk (1956, 1962, 1970) the regional metamorphism in the Lepontine was caused by a thermal doming under a cover of 10 to 25 km (Wenk, 1970). Niggli (1970) concluded that, "regional metamorphism in the Simplon-Ticino region is interpreted as a load metamorphism caused by tectonic burial during the formation of the Alpine nappes" in Upper Cretaceous to Early Tertiary time. "At this time, rock thicknesses of 15–30 kilometers must have covered this region". *Suggested* geothermal gradients vary significantly: Zwart (1967) proposes 10 to 17° C/km, Niggli (1970) proposes 20–40° C/km and Wenk (1970) around 50° C/km. All these estimations are derived directly from the particular value assumed for the thickness of the cover. From the thermal model of Clark and Jäger (1969) based on geochronologic and heat flow data, a geothermal gradient in the range from 20 to 40° C is derived. Thompson (1974) determined that the staurolite and sillimanite isogradic surfaces dipped steeply northward in the eastern Lepontine (Val Mesolcina). The isograd pattern was interpreted to mean that at the peak of the Lepontine phase isothermal surfaces also dipped steeply northward, isobaric surfaces were approximately horizontal, and the thermal gradient between the sillimanite and staurolite isograds was approximately 12° C/km. If the temperature range over the entire distance to the earth's surface is considered an average thermal gradient of 35° C/km is obtained for the central part of the area. Extrapolating this work westward THOMPSON suggested that the dip of isogradic surfaces is only 15° N in the Ticino area.

Denudation rates of ½–1 mm/year were derived for the Central Alps by different methods (JÄCKLI, 1957; JÄGER et al., 1967; CLARK and JÄGER, 1969). The impressive relief of the present-day Alps shows that the mean rate of uplift since the mid-Tertiary was slightly larger than the rate of denudation. This is confirmed by the maximum uplift of 1 mm/year obtained by precision nivellements over the last fifty years in the Central Alps (JEANRICHARD, 1972) and in the Tauern Window (SENFTL and EXNER, 1973).

A third Alpine-metamorphic phase of very low-grade can be recognized in the external parts of the Alps. MARTINI and VUAGNAT (1970, p. 62) pointed out that the metamorphism of the Taveyanne sandstone occurred later than the mid-Tertiary metamorphism in the Lepontine. Radiometric evidence in the Glarus Alps suggest a *metamorphic phase at the Oligocene-Miocene boundary* (see p. 274). TRÜMPY (1973a, b) described a tectonic phase of Miocene age in the Helvetic Alps connected to an external subduction zone between the Aar and Gotthard Massifs. In contrast to the Pennine domain, where the metamorphism occurred after the formation of the nappes, post-metamorphic overthrusting caused an inverse metamorphic zonation in the Glarus Alps.

In the Pennine domain of the Western Alps radiometric evidence point to an Oligocene-Miocene event; it is not yet clear whether the rejuvenation of alkali-amphibole and biotite was caused by tectonic movements only (backfolding?) or by metamorphic processes as well. In the Eastern Alps very little is yet known about distribution and age of low-grade metamorphic events.

At present, three Alpine-metamorphic phases can be distinguished in the Alps by radiometric and petrographic evidence: Upper Cretaceous (Eoalpine metamorphic phase), mid-Tertiary, and Oligocene-Miocene. Stratigraphic and tectonic evidence suggests three major orogenic events for the same time periods. It can therefore be concluded that the three Alpine-metamorphic phases are related to orogenic phases.

As pointed out by TRÜMPY (1973b) the present knowledge is more in favour of a paroxismatic than for a continuous evolution of the Alps.

The picture of Alpine metamorphism developed during the last twenty years is much more complex than was assumed initially. In spite of the rapidly increasing amount of new data, many problems remain to be solved.

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SMPM = Schweiz. Mineral. Petrogr. Mitt.

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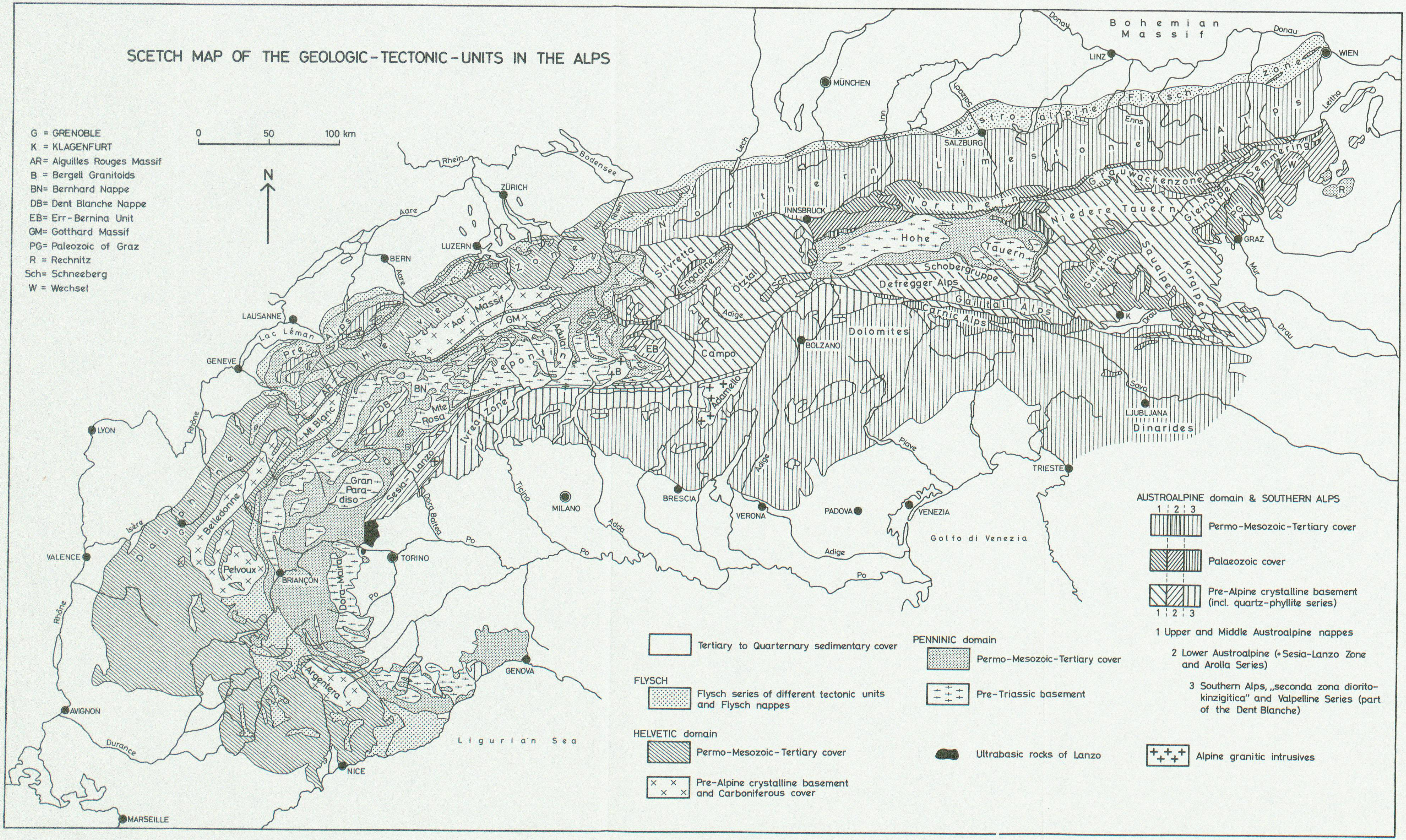

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