

THE POLYCYCLIC LITHOSPHERE: AN ATTEMPT TO ASSESS ITS OROGENIC MEMORY

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ABSTRACT

The orogenic memory of the continental lithosphere is evaluated. Lithostratigraphic, tectonic, metamorphic, plutonic and geochemical criteria for polycyclicality are reviewed. Their mutual relations, significance and resolving power are discussed, and the possible causes of fading of the orogenic memory are explored. Emphasis is placed on criteria such as metamorphosed molasse deposits, interference patterns of superimposed folds, basement upthrusts and basement nappes, mylonitized mylonites, diaphoretic features, contrasting metamorphic facies series, polymetamorphic marker formations, thermal domes, diapirs and the convecting basement effect. The geochemical polarity of granites and granulites is reviewed in the light of thermal basement fractionation, and its possible reflections on the evolution patterns and dating powers of some radiogenic isotopes are considered. It is concluded that the memory of rocks for orogenic cycles, although finite, can be stretched considerably when judicious use is made of carefully weighted parameters. No single criterion for polycyclicality can as yet be applied unequivocally.

Noli turbare circulos meos
ARCHIMEDES.

INTRODUCTION

Sea floor spreading and plate tectonics have convinced many of us that the oceanic crust and upper mantle are almost continuously being created and consumed for the last 200 million years and probably well beyond. But among the many problems these elegant hypotheses have not been able to solve so far, the question as to whether the continental lithosphere was also subject to quasi-continuous recycling of some sort, looms large. In other words have the continental crust and its complementary upper mantle been uniquely created, and added on to by orogenic accretion, or are they also being partially consumed and re-created by intermittent geodynamic and geothermal processes leaving unmistakable traces of their polycyclic nature in their various products? Any attempt to answer this question must be based on a survey of the available criteria for polycyclicality and

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on an assessment of their significance for past dynamothermal regimes and their efficacy in preserving these critical features. In other words: we may ask ourselves how good is the orogenic memory of the continental lithosphere? This raises further questions such as: how far can we rely on unconformities and conglomerates, on interference patterns of folds, on polymetamorphic features, on the distribution of lithophile elements, on radiometric age determinations, or on lead and strontium evolution patterns for the identification of a previous orogenic cycle and for an estimate of the role its products played in a subsequent orogenic event?

The lower continental crust

But before we engage in such a survey, let us briefly review the history of the argument for polycyclicality of the continental lithosphere. In Werners's 18th century conception of primitive rocks, followed by transitional strata, stratified rocks, alluvial deposits and volcanic rocks, each being unique in character and in position in space and time, there obviously was no place for polycyclicality in the formation of the earth's crust. Recognizing the significance of plutonic intrusions, uplift and angular unconformities, James Hutton was the first to introduce a cyclic notion into his concept of mountain building by successive revolutions. Hall, while supporting Hutton's theory with ingenious experiments, invented the orogenic cycle, but the reality of sedimentary, tectonic, magmatic and geochemical cycles became firmly established only through the works of Bertrand (1894), Barrell (1917), Stille (1940), Umbgrove (1947), Goldschmidt (1937), and Wegmann (1965, 1966).

Along with the ideas of crustal development by alternating evolutions and revolutions, the distinction between cratons, shields and platforms on the one, and mobile zones, geosynclines and orogenes on the other hand, became part of the geological establishment. The notion of continental accretion through transformation of peripheral orogenic belts and their welding on to a central craton, set forth by Dana in 1873, was a logical consequence of the *prima facie* situation in the North American shield and its platforms. Appropriate though it may be for the ocean-based or ensimatic geosynclines, such as occur off volcanic islands arcs and go to constitute the outer parts of paired orogenic belts, it is most unlikely to apply to the ensialic or craton-based geosynclines which are known to encroach considerably, if not wholly, on the continental slopes and platforms (P. Michot, 1963, 1968; Wynne-Edwards, 1970).

For the predominantly Precambrian shields it was not realized until the advent of refined field and laboratory techniques that substantial portions have repeatedly been re-tectonized, re-metamorphosed and re-mobilized, but for the Phanerozoic orogenes the presence of recycled cratonic basement was recognized comparatively early.

Taking the Western Alps as an example of the latter group, we find that we owe the first mention of a pre-Triassic orogenic cycle to Charles Lory (1860). Today, the metamorphic map of the Alps (Zwart *et al.*, 1974) shows the Alpine arc to consist, for at least a good quarter of its metamorphic surface area, of polycyclic basement ranging in age from Variscan to Precambrian (fig. 1). For the Scandinavian Caledonides Wilson and Nicholson have shown that the Precambrian basement outcrop covers about half the orogenic terrain, with Caledonian structures predominantly overprinted on Precambrian ones (fig. 2).

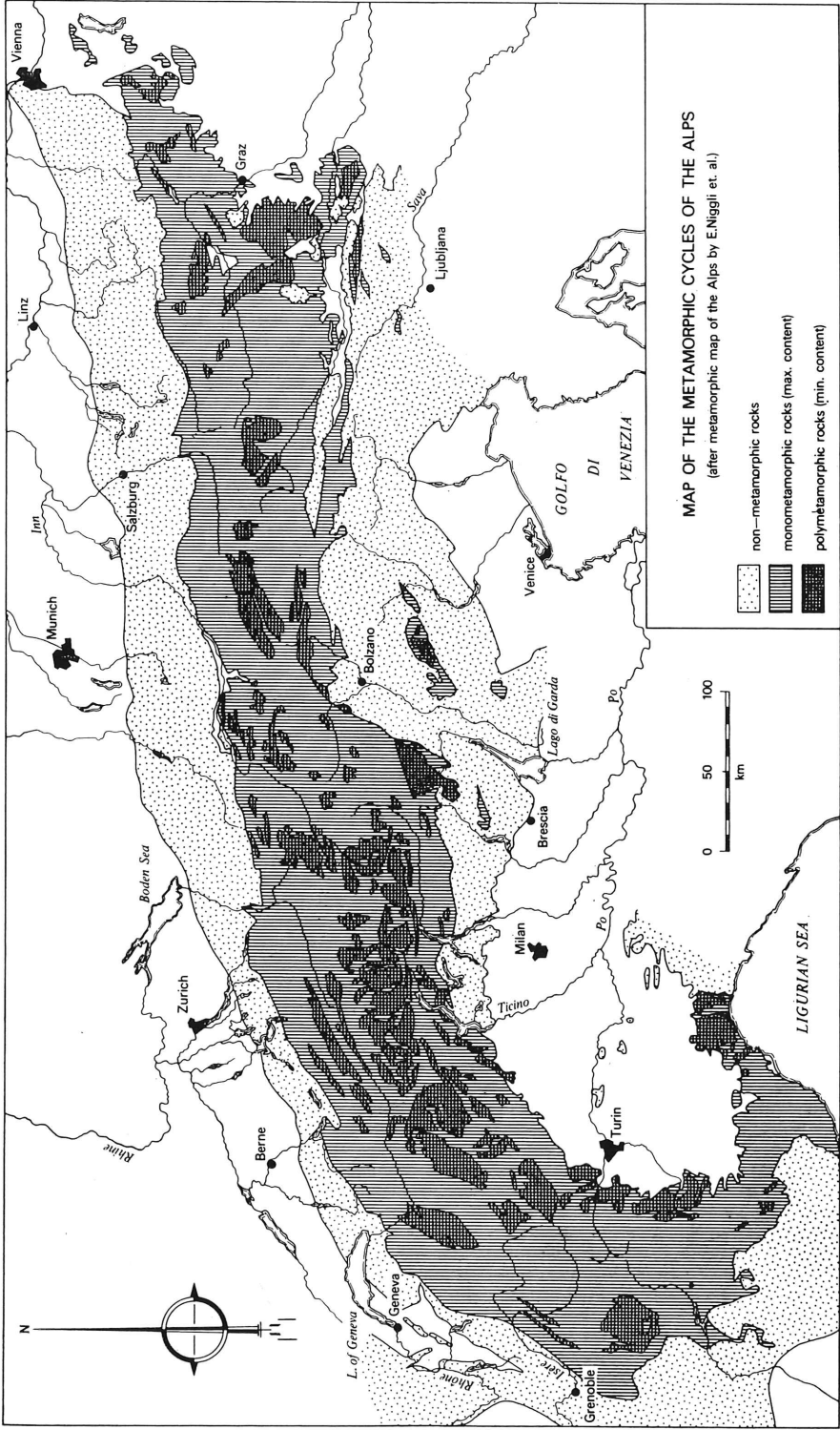


FIG. 1. — Map of the metamorphic cycles of the Alps, simplified from Zwart, Sobolev, Niggli et al. (1974). At least one quarter of the metamorphic terrain is polymetamorphic.

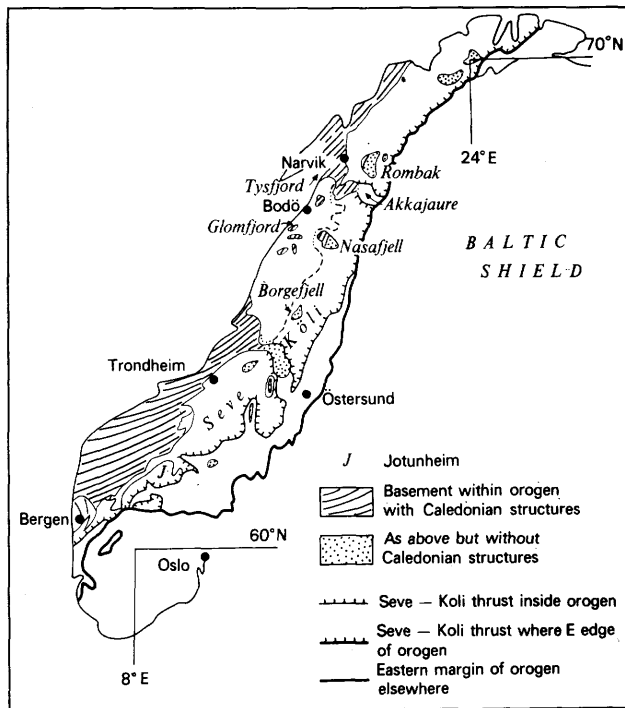


FIG. 2. — Outline map of the Scandinavian Caledonides showing the relative areal proportions of polycyclic and monocyclic rocks (after Wilson and Nicholson, 1973, fig. 1).

The late-Proterozoic Grenville province of eastern Canada is even more significant for the first group. Considered until recently to be a peripheral accretion to the Superior proto-continental craton, it is now known to overlap by as much as 90 % on the Superior, the Churchill and the Nain provinces, consisting as it does for the greater part of very old gneisses, amphibolites, quartzites, marbles, banded iron formations and granitic rocks (Wynne-Edwards, 1969) recycled through the Kenoran (2.5 b.yr.), the Hudsonian (1.7 b.yr.) as well as the Elsonian (1.35 b.yr.) orogenies (Davis *et al.*, 1967; Wynne-Edwards, 1969).

Stille's (1940) attempt to re-inforce the contrast between orogenes and cratons by dubbing the recycled basement of younger orogenes quasi-cratonic was not realistic, since the greater part of the cratons and about half the surface area of the orogenic belts thus becomes quasi-cratonic. Moreover, the orogenic belts of the North American continent and the Baltic Shield were shown by Wynne-Edwards and Zia-Ul Hasan (1970) to intersect pre-existing ones and sometimes to cut right across older cratonic provinces, refuting the principle of lateral accretion. Vertical accretion, proposed by P. Michot (1963, 1968) as the most prominent growth factor of continents is an attractive compromise.

The upper mantle beneath the continents

While the polycyclic nature of the lower continental crust is slowly gaining recognition, the continental upper mantle is still regarded as homogeneous and monocyclic by many of its students. Yet, indications of its recycling are rapidly accumulating. For instance if the alpine-type harzburgites and dunites are assumed to be fragments of the upper mantle (de Roever, 1957), then their contents of Na, Ca, Al and so-called incompatible elements (K, Ti, P, Ba, Sr, Rb, Ir, Hf, U, Th, Pb and light REE) are too low to allow them to be the parent rocks of oceanic basalts, while their $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are too high to warrant their interpretation as either infusible residues or early crystallized fractions of such basaltic parent rocks. Therefore, Roe (1964), Stueber and Murthy (1966), and Bonatti (1971) suggested two-stage strontium evolution models for their development, consisting first of a subcontinental stage during the fractionation of the primitive crust (1 to 3.5 b.yr. ago) during which the lithophile and incompatible elements (including Rb) were strongly depleted, and a subsequent suboceanic stage when further fractional fusion, yielding basalts particularly low in potash and other incompatible elements, became possible. Hurley (1967) suggested that the Alpine-type peridotite pattern of low Rb/Sr and high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios could result from accumulation of partially fused fractions of old crust dragged down into the mantle.

Moreover the structural and metamorphic history of most alpine-type peridotites and of certain types of ultramafic nodules in basalts is turning out to be polyphase, if not polycyclic. Not only the lherzolites, but also the non-stratiform harzburgites, dunites, wehrlites and websterites appear to have been repeatedly deformed and recrystallized in the predominantly solid state, frequently revealing a subcrustal history older than that exhibited by the surrounding crustal rocks (Collée, 1962-1963; Avé Lallemant, 1967; Möckel, 1969; Nicolas *et al.*, 1971). Complex metamorphic histories have also been unravelled in these rock types usually starting off with mantle-type P/T conditions gradually developing into mineral associations characteristic of the lower and upper crust (Green, 1964; den Tex, 1969; Kornprobst, 1969; Wyllie, 1970; Forestier, 1971; Rost, 1971). It appears that some Alpine-type peridotites, such as the garnet peridotite of Alpe Arami in the Swiss Alps, are derived from the continental upper mantle, and that it is potentially just as polycyclic as the continental crust. As constituents of the continental lithosphere they may have belonged to an approximately closed system during orogenic recycling. However, subcontinental rifting and other oceanization processes create oceanic crust from continental upper mantle, while subduction and obduction mix-up oceanic upper mantle with continental crust. Thus it cannot be excluded that orogenic recycling affects the sum total of the lithosphere in the course of geological time.

THE CRITERIA OF POLYCYCLICITY

In principle the geologist owes his ability to recognize recycling processes in actual rocks to the general inefficacy of such processes in overcoming the inertia of crystalline rocks, i.e. their inherent resistance to complete re-shaping and reconstitution. But the capacity of rocks to retain a memory of their orogenic past diminishes rapidly with the intensity of the orogenic processes and with the

number of cycles to which they were subjected. The criteria of polycyclicality are numerous, but only rarely is it possible to use more than a few in a specific instance, so that the evidence for recycling is mostly of a circumstantial nature only.

Stratigraphic glimpses of preceding cycles

Since James Hutton, angular unconformities with basal conglomerates carrying pebbles of various crystalline rocks constitute the classical evidence for the closure of an orogenic cycle from the stratigrapher's point of view. However great thicknesses of rapidly accumulated volcanodetrital deposits of terrestrial or shallow water derivation such as the Alpine Molasse, the Variscan Rotliegendes and Verrucano, and the Caledonian Old Red Sandstone have become equally, if not more useful in identifying the final stages of uplift and erosion of the Phanerozoic orogenic belts.

When deformed and metamorphosed, these so-called molasse formations present evidence for a second orogenic cycle to have affected the products of a previous cycle, much more convincingly than does a separate meta-conglomerate without angular unconformity. Unfortunately, angular unconformities tend to be blurred by metamorphism and tectonic transposition, while the pebbles of polygenic conglomerates are liable to have been altered at the surface, remetamorphosed or stretched beyond recognition (Mehnert, 1938; den Tex, 1950).

It is true that, from 1854 onward, more or less metamorphosed conglomerates have been discovered in the Saxothuringian and Moldanubian basement inliers of the Variscan orogene of central Europe, but seldom have they been found to contain discordantly oriented pebbles of crystalline schists, or to be accompanied by angular unconformities (Mehnert, 1938).

Although there is no strict necessity for every orogenic cycle to sustain a period of uplift and exhumation sufficiently important to expose its high-grade crystalline schists, or even to allow a polygenic conglomerate or molasse deposit to be formed, the evidence for re-cycled molasse is accumulating rapidly in spite of the difficulties of recognizing deformed and metamorphic molasse deposits. Indeed in the Precambrian mobile belts such evidence is frequently poor or wanting. However, the argument over the correlation of the Moine Schists with the Proterozoic Torridonian molasse in the foreland of the Scottish Caledonides, settled positively after a century and a half of vigorous dispute, shows, that the search for re-cycled molasse can be successfully extended into the more distant geological past (McCulloch, 1819; Peach and Horne, 1930; Read, 1934; Watson, 1963; Sutton, 1968).

The Variscan orogene of the Iberian peninsula is an interesting case-in-point. Angular unconformities involving Precambrian strata have been reported from various localities (Bard *et al.*, 1972), but the underlying rocks are hardly ever polytectonic, and never polymetamorphic. On the other hand this belt contains abundant more or less metamorphic molasse deposits of various ages. Of these the most ancient and most important is the so-called "Ollo de Sapo" formation: a belt of infra-Cambrian arkosic and rhyolitic rocks containing large feldspars and bluish quartz crystals, which forms the NW limb of the geanticlinal axis of the Hesperian Massif (Parga Pondal *et al.*, 1964; Bard *et al.*, 1972). It probably grades upwards and outwards into more flyschoid semipelitic rocks, such as the uppermost Precambrian Series of Villalba, while the equivalent "Seria Negra" of the SW limb contains Ollo de Sapo-like rocks in its lower levels (fig. 3). This characteristic

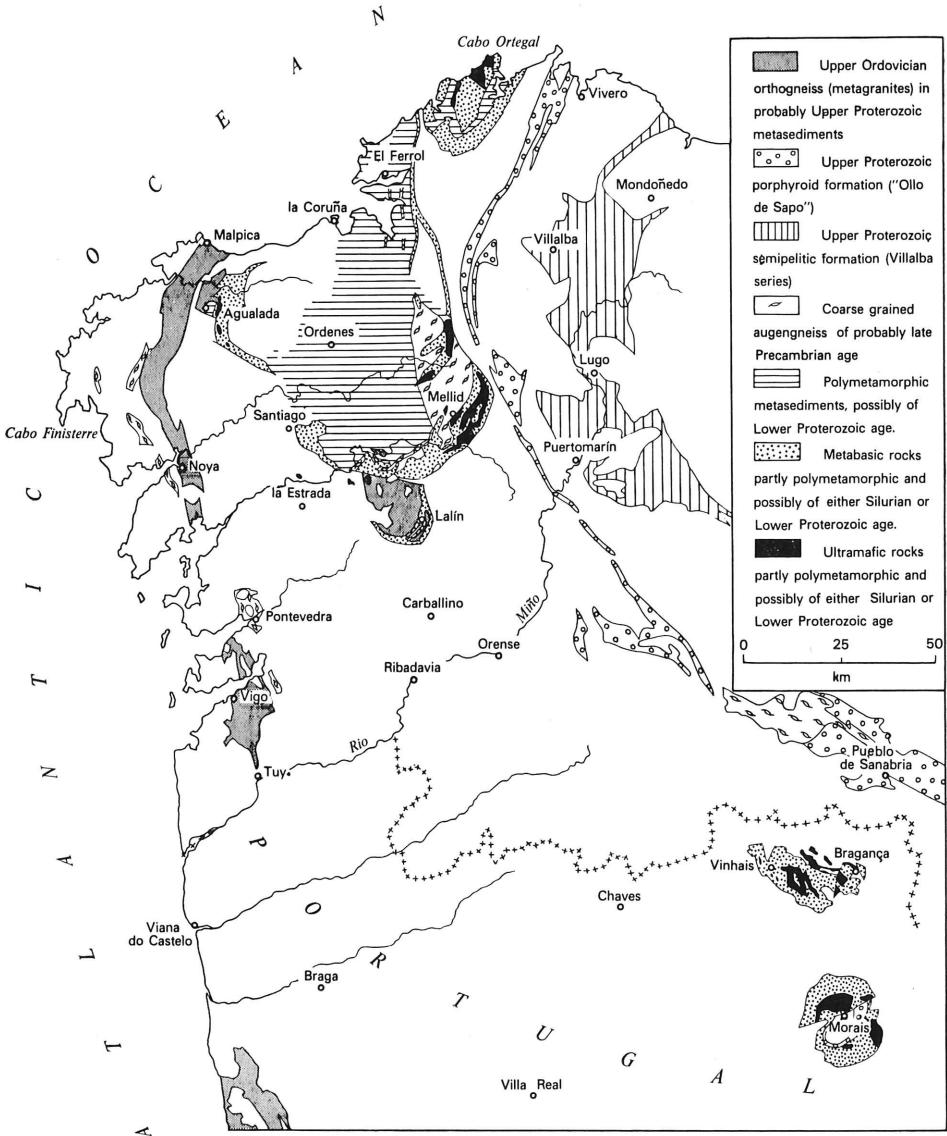


FIG. 3. — Map showing the distribution of polymetamorphic rocks, of metamorphosed molasse and flyschoid deposits of Upper Proterozoic age and of the (poly-)metamorphic marker formation in the Variscan orogene of Galicia (NW Spain). Data from den Tex and Floor (1972) and Bard et al. (1972).

molasse deposit, which can be traced from low-grade metamorphic into migmatitic facies, lies either autochthonous on meta-igneous augengneisses (Sierra Segundera, NW and SE of Puebla de Sanabria: Anthonioz and Ferragne, 1967) or allochthonous on polymetamorphic Precambrian ophiolites and paragneisses (Morais-Lagoa, NE Portugal: Anthonioz, 1966, 1969).

Less important or doubtful horizons of molasse-type occur in the Lower Cambrian "Schisto-grauwackico", Candana and Herreria formations of NW Portugal, eastern Galicia and the western Asturias (Matte, 1968; Walter, 1968; Bouyx, 1970; Bard *et al.*, 1972), and in the almost ubiquitous "Armorican Quartzite", marking the transgression from the SE of the Ordovician sea, and in the Upper Ordovician/Lower Silurian volcano-detrital formations of Vilaflor, Truchas-Rio Sil, and Celanova (Ribeiro *et al.*, 1960; Matte, 1964; Ferragne, 1966, 1972; for reviews cf. Parga Pondal *et al.*, 1964; Walter, 1969). These horizons generally repose disconformably or unconformably on older strata, but only in very few instances, such as the Ordovician underlying the Silurian porphyroids in the synform of Verin, is it possible to infer a nonconformable sequence in the sense that the overlying beds fail to show the effects of metamorphic or tectonic phases that have affected the strata below the unconformity. Even so, the stratigraphic gap is generally too small to allow an orogenic cycle to have intervened, except in the case where the older polytectonic and polymetamorphic Precambrian is involved, but then—as at Cabo Ortegal—the contact with Silurian strata is always of a tectonically inverted nature (den Tex and Vogel, 1962; Anthonioz, 1969).

It follows that the older Precambrian orogenic belt of the Iberian peninsula, which constitutes the backbone of the central geanticlinal zone of the Variscan orogene, was uplifted, exhumed, covered and flanked by molasse and flyschoid deposits in a number of consecutive stages, the most important of which occurred in uppermost Precambrian time giving rise to the "Ollo de Sapo" molasse. It shows how carefully the argument of metamorphosed molasse should be handled, if the correct number and dates of orogenic cycles are to be established in polycyclic terrains.

The structural palimpsest: can it be deciphered?

It has become a well-established fact that an orogenic cycle can be resolved into several tectonic phases each marking a stage in the dynamic development of the orogene, which can be correlated with phases in its thermal history as evidenced by the waxing and waning stages of metamorphism, plutonism and volcanism. Not necessarily are all these phases developed, nor is their evidence always preserved, but in a very general way a low-temperature subsidence or subduction phase is followed by the piling-up of thrust sheets, fold nappes or recumbent folds under a regime of rising temperatures, increasing metamorphism, granite emplacement and andesitic volcanism, which—in turn—may be succeeded by one or more compressive phases causing upright fold systems of parallel or cross-orientation and oblique wrench fault systems to develop, while the temperature is generally falling. Finally, the isostatic uplift of the orogene is usually accompanied by block faulting of the normal type, and acid to intermediate volcanism.

Such polyphase deformation naturally gives rise to interference patterns, but so do deformations belonging to successive orogenic cycles. However, polycyclic interference patterns of fold systems are likely to be of lower symmetry than monocyclic interference patterns, the strain ellipsoids of which are strongly interrelated. Ramsay (1967) has provided us with an exhaustive survey of the interference patterns caused by superimposed folding. Of the nine varieties of three types of interference patterns, recognized by him, the least symmetrical ones (i.e. B_1 , D_1 , E_1 , F_3 and H_2) should be expected to occur most abundantly in polycyclic terrains (fig. 4). In fact, asymmetrical eyed and mushroom-type folds have been reported from such complex terrains as the Penninic nappes of the Swiss Alps, the Moine schists of the Scottish Caledonides (cf. Ramsay, 1967, fig. 10-10,11,12) and the older Precambrian inliers of the NW Iberian peninsula (Vogel, 1967; Engels, 1972). In the latter instance asymmetrical mushroom fold hinges appear on scales ranging from a 1:50 000 map to a thin section (figs. 5 and 6). They are due to Variscan refolding of recumbent Precambrian folds, the polycyclic nature of the terrain having been confirmed by lithostratigraphic, metamorphic and radiometric data (Priem *et al.*, 1970; Vogel and Abdel Monem, 1971; Hubregtse, 1973).

By itself, fold interference patterns are certainly not a very reliable criterion for polycyclicly, but there may be other structural evidence to support it. Thus portions of continental and oceanic crust, that were previously subjected to orogenic processes, should in the average contain a higher proportion of crystalline rocks than newly added geosynclinal deposits. When incorporated in a subsequent

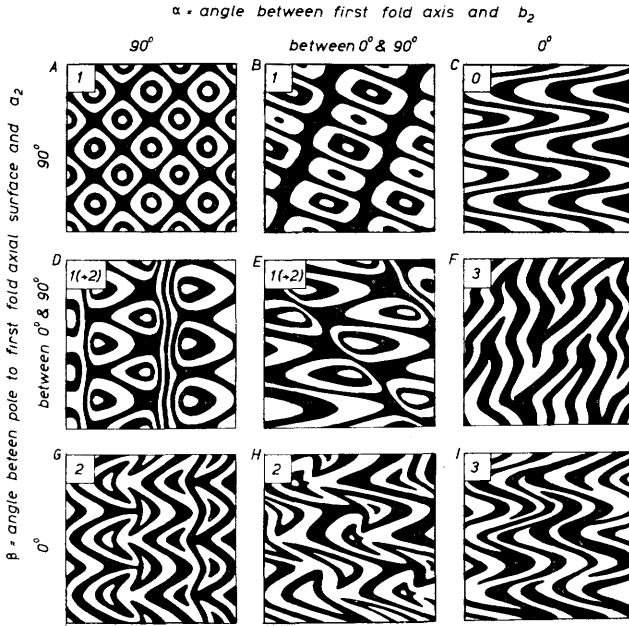


FIG. 4. — Interference patterns of superimposed folds from Ramsay (1967, figs. 10-13). The least symmetrical patterns are B_1 , D_1 ($\rightarrow 2$), E_1 ($\rightarrow 2$), F_3 and H_2 (cf. figs. 5 and 6).

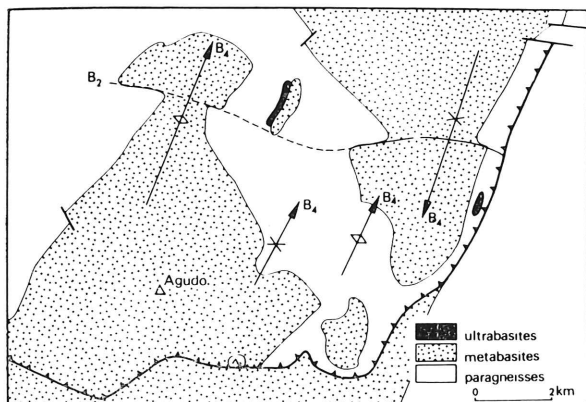


FIG. 5. — *Mushroom-type outcrop pattern of recumbent Precambrian folds (B_2) refolded by upright Variscan folds (B_4) in the basement complex at Cabo Ortegal (from Engels, 1972, fig. III-21). Cf. figs. 3 and 4 (G_2 , H_2).*

orogenic cycle such almost entirely crystalline bodies should behave, at least during the first tectonic phase, in a much more brittle manner than do their nonmetamorphic sedimentary counterparts. This leads to a structural style with emphasis on imbricated thrust sheets and basement wedges, on upright parallel folds with fractured limbs, and on movement concentrated in shear belts and mylonite zones. Now, this is the predominant tectonic style encountered in the



FIG. 6. — *Asymmetrical mushroom fold, due to F_4 refolding F_2 , in the Precambrian amphibolites at Cabo Ortegal (NW Spain) (from Engels, 1972, fig. III-27). Cf. figs. 3 and 4 (H_2).*

geanticlinal axial zone of the Variscan orogene of NW Spain, in contradistinction with the large-scale recumbent folds that were formed simultaneously in the neighbouring geosynclinal zones of eastern Galicia and Portugal (Matte, 1968; den Tex and Floor, 1971; Bard *et al.*, 1971).

But a different situation arises when crystalline rocks are subjected to long-sustained pressures at depths where about half their melting temperature is reached (i.e. $\sim 350^\circ\text{C}$ for water bearing felsic gneisses). In Carey's (1954) terms such rocks could become rheid on passing the threshold value: $\eta/\mu \cdot 10^3$, where a thousand times the ratio of viscosity (η , in poises) to rigidity (μ , in dynes $\cdot \text{cm}^{-2}$), gives the time (in seconds) required for rheid behaviour, i.e. for yielding predominantly by steady state creep or pseudoviscous flow. The resulting structural style has an emphasis on similar flow folds, on fold nappes of the Penninic type and on more or less diapiric basement domes of mantled granitic gneiss, and anorthosite-norite-granulite associations (Eskola, 1949; J. and P. Michot, 1960; and Martignole and Schrijver, 1970). Carey (1954) gave examples of rheid or flow folding including tear-drop synforms from the Precambrian terrain at Broken Hill (N.S.W., Australia), where the grade of metamorphism was amphibolite-granulite facies, while Wynne-Edwards (1963) produced evidence supporting the case for pseudo-fluid behaviour of rocks under equally high-grade conditions in the Grenville province of Canada.

The Penninic nappes of the Swiss Alps exhibit a similar style of deformation, not only in their Mesozoic envelope rocks, but also in the gneissic, migmatitic and granitic core masses of the nappes. A controversy over the age of these core masses and their evolution has arisen in the early part of this century and has not yet led to a consensus of opinions. E. Wenk (1948, 1962, 1970) regards the core masses, or rather their granitization and structural imprint, as exclusively Alpine features, but others (Nabholz, 1954; Zwart, 1967; Milnes, 1969; Niggli, 1970; Hunziker, 1970) have pointed out that the migmatites and granites of the core masses are partly Variscan or even older in age, and that the metamorphic zones and isograds of the Lepontine (or late Alpine phase) cut right across the structures, requiring the nappes to have been formed well before the thermal climax. Most workers agree, however, that the nappes were piled up in the time span between Upper Cretaceous and Lower Oligocene (i.e. between 70 and 40 m.yr. ago), while the thermal climax of the Lepontine metamorphism has been dated radiometrically at 35-38 m.yr. (Jäger, 1973). We may also infer that the age of approximately 70 m.yr., obtained for the Eo-alpine high-pressure metamorphism probably marks the climax of subduction, and therefore the latest possible commencement for severe loading and stressing of the crystalline basement rocks. This leaves roughly 30 m.yr. for their rheid behaviour to become dominant. Now, if the viscosity of the lower crustal rocks may be estimated at 10^{22} poises and their rigidity is in the order of 10^{11} dynes $\cdot \text{cm}^{-2}$ (Clark *et al.*, 1966), the minimum time required for rheid behaviour should be 10^{14} seconds, or approximately 10 m.yr.⁽¹⁾ This would mean that there was sufficient time available for the Penninic nappes, and their flow folds, to develop before the Lepontine metamorphism reached amphibolite facies conditions with temperatures close to the minimum melting point of felsic gneisses and granites.

(1) In 10 m.yr. a quarter to a third of the estimated temperature interval of 400°C between Eo-alpine and Lepontine metamorphic climax may have been recovered.

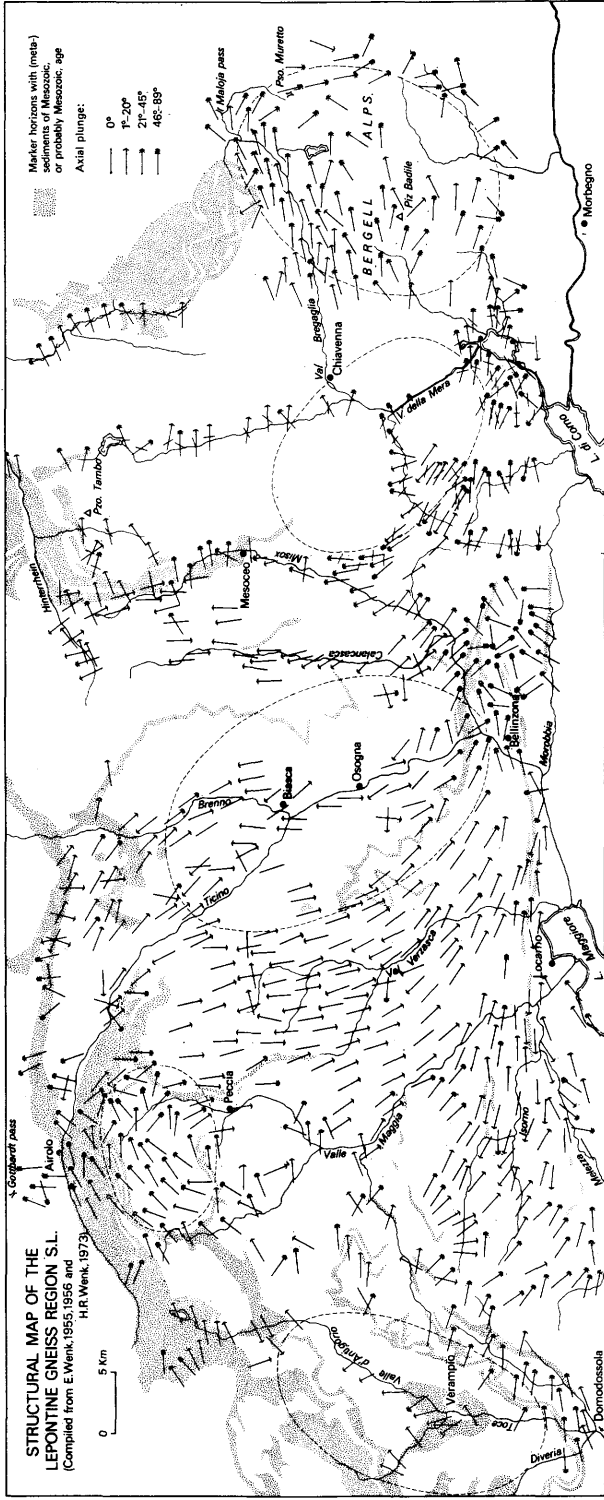


FIG. 7. — Lincation map with the inferred subdome patterns in the Lepontine Gneiss Region of the Swiss Alps. Subdomes approximately outlined with dashed lines. The Maggia cross-zone links the Ticino-culmination (Osogna-Biasca) with a possible dome near Airolo-Peccia. Data from E. Wenk (1955, 1956) and H. R. Wenk (1973).

Such a sequence of events in the Penninic realm would reconcile the apparently conflicting evidence of Variscan, Eo-alpine and intermediate whole rock ages, preserved in the frontal parts of the Penninic nappes, which were relatively cold when they were first emplaced, as against the predominantly late-Alpine ages recorded from the thermal dome structures in the Lepontine gneiss region and the Bergell Alps, where the pre-existing basement nappes and their envelopes were probably thermally activated to become convective, partially fused and geochemically recycled. Thus the orogenic memory of the Penninic nappes, with their well-established pre-Alpine granites and migmatites, may have faded considerably in the Lepontine gneiss region and the Bergell Alps, owing to late-Alpine thermal recycling. But this latter part of the Penninic story will be dealt with more amply in the next section (figs. 7 and 8).

Relicts and products of polymetamorphism: Successive stages in basement re-activation

Almost imperceptibly we have descended into the infernal realm of metamorphism and plutonism, where one of the principal energy sources of orogenic processes is located. Classical legend has it that mortals lose their memory when crossing the Lethe into Pluto's realm. But does it apply to rocks as well?

PLURIFACIAL VERSUS POLYMETAMORPHISM

Polymetamorphism, a term coined by Read in 1949 was designed to denote the presence of a record in rocks of "two or more unified acts that are separable from one another and present no obvious genetic connexion." As such it was meant to express the metamorphic effects of superimposed orogenic cycles. However, as Read clearly perceived, its opposite number: monometamorphism, may also consist of episodes or phases separable from one another by acts of incomplete replacement or recrystallization, or by their interference with tectonic phases belonging to a single orogenic cycle. Since, in monocyclic metamorphism, several metamorphic facies are usually involved but not always visibly preserved, de Roever and Nijhuis (1963) introduced the term plurifacial metamorphism and de Roever (1972) declared it applicable to polymetamorphic rocks as well.

To the present author it seems more appropriate to reserve plurifacial metamorphism for monometamorphic rocks only, and to retain Read's polymetamorphism for polycyclic rocks, a principle which is also not universally adhered to (e.g. Johnson, 1961; Zwart, 1962). Read (1949) rightly observed that "as we can work only with the material that the rocks provide, a decision on polymetamorphism or monometamorphism is often impossible to make." This is largely due to the fact that both plurifacial metamorphism and polymetamorphism are visibly preserved only by virtue of the affected rocks failing to re-equilibrate completely under changing conditions of pressure, temperature and other variables.

If a distinction can be made at all, it is because monometamorphism is essentially prograde throughout. In the subsidence or subduction stage temperatures remain very low, while the pressure is increasing rapidly (fig. 9). Mineral associations formed at this stage are of the zeolite-, glaucophaneschist-, or greenschist-facies. When the orogenic climax of crustal shortening, folding and mantle diapirism has been reached, temperatures are rising steeply, while the pressure increase lags behind. This is the stage when mineral associations of the amphibolite-

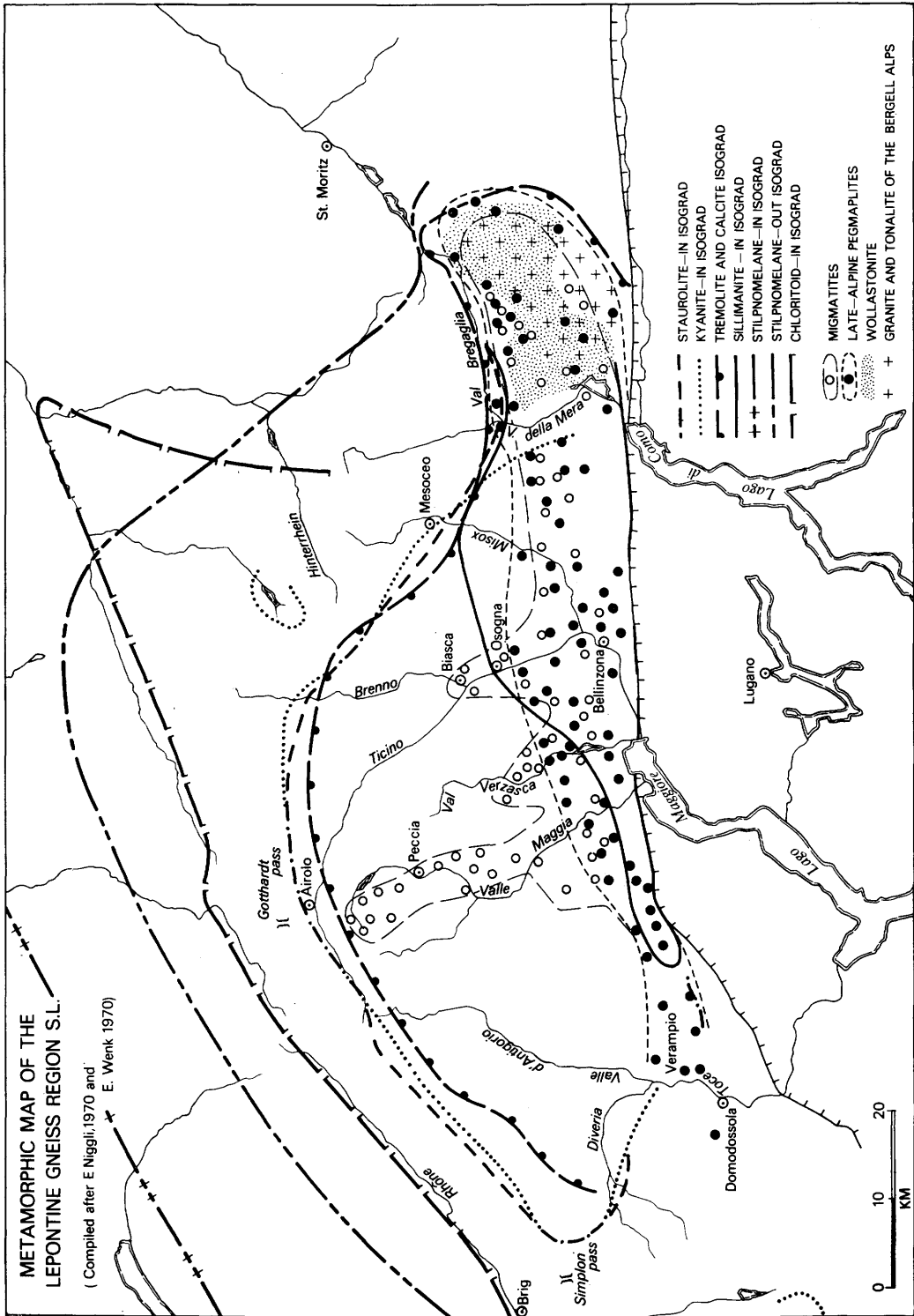


FIG. 8. — Metamorphic map of the Lepontine Gneiss Region in the Swiss Alps (after E. Wenk, 1970 and E. Niggli, 1970).

and granulite facies are formed and partial melting may set in. Now, if the minerals critical of the first stage (such as lawsonite, jadeite and phengite) are not everywhere completely destroyed to yield associations critical of the second stage (e.g. albite, anorthite, andalusite, cordierite and sillimanite) either through low reaction kinetics and low diffusion rates in solid phases (zoning in plagioclase and garnet porphyroblasts), or through the formation of locally closed systems by way of reaction rims, armoured relicts, etc. we are faced with evidence of plurifacial prograde metamorphism, the facies series of which may be either of a single type throughout (for instance of the low-pressure type, as in the Central Pyrenees: Zwart, 1962), or composite, starting off in the high- and ending up in the intermediate or low-pressure facies series as in Japan, the Betic Cordillera of Southern Spain and the Swiss Alps (Miyashiro, 1961; de Roever and Nijhuis, 1963; Niggli, 1970). Unloading and cooling of such rocks, as is bound to occur at the end of an orogenic cycle, generally does not cause retrogradation of the climactic mineral assemblages, since uplift and erosion take place along discrete surfaces leaving the bulk of the crystalline rocks undisturbed and impenetrable for aqueous solutions (fig. 9).

But when these metastable mineral assemblages become involved in a superimposed orogenic cycle, they will be penetratively deformed at low temperatures during the subduction and early folding phases of the new cycle, thus giving access to aqueous solutions, and increasing the kinetics of low-grade re-equili-

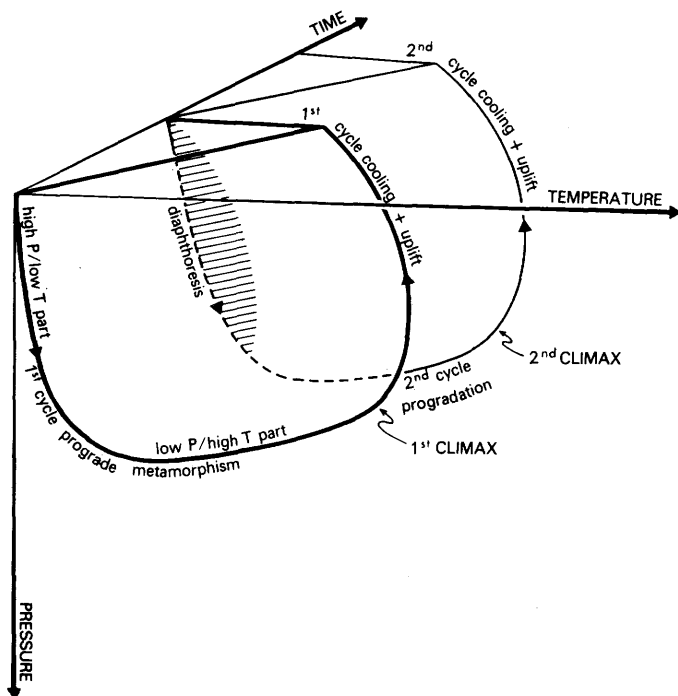


FIG. 9. — *P-T-time graph of two superimposed orogenic cycles showing the overlap (hatched area) in which retrogradation (diaphthoresis) occurs.*

bration reactions. In this manner the result of prograde metamorphism in a preceding cycle is undone by an initial phase of retrogradation in the superimposed cycle. Perhaps without fully realizing these implications, the Austrian geologist Becke introduced the term *diaphthoresis* for retrograded crystalline rocks in 1909, but it never succeeded in gaining the recognition it deserved outside the German literature.

Read (1949) was among the first geologists to recognize the proper orogenic significance of retrogradation, to wit as a juncture between cycles in polymetamorphism. Unfortunately, retrogradation is by no means an infallible criterion of polymetamorphism. Its intensity and, in fact, its very presence depend on the amount of uplift of the preceding orogene, and on the extent and localization of the subductive or obductive movements of the superimposed cycle. If these effects are insufficient for pronounced and locally concentrated retrogradation to occur, we cannot be sure of polycyclicality. Slightly retrograde plurifacial metamorphism is indeed observed to have taken place within a single orogenic cycle, for instance during second or later phase folding and flattening under a cooling regime with incipient unloading only, or as a precursor to contactmetamorphism, superimposed on regional metamorphism by late or postorogenic granite intrusions (Atherton and Edmunds, 1966).

MYLONITIZED MYLONITES

Inequivocal evidence of diaphthoretic polymetamorphism comes almost exclusively from thrust belts and mylonite zones in otherwise high-grade crystalline basement terrain. At the onset of a new orogenic cycle such portions of rigid rock from the lower crust, while rapidly being loaded at low temperatures and subjected to high strain rates, are ideally suited to subgrain formation, cataclasis, penetration by aqueous fluids and retrogradation.

The classical ground for features of this kind lies in the NW Highlands of Scotland where Lapworth (1885) coined the terme mylonite in the Moine thrust belt, which—it was then believed—grossly delimits the Precambrian basement of the foreland from rocks of the Caledonian orogene *sensu stricto*. Later detailed structural and petrological work, mainly by Christie (1960) and Johnson (1960), revealed the presence of two generations of mylonite at the Caledonian front to wit:

- 1° primary mylonitic rocks or blastomylonites;
- 2° secondary mylonites, phyllonites, kakyrites or cataclasites.

Of these only the primary mylonites, affecting high-grade mineral assemblages of the Lewisian basement, are markedly retrograde, having a matrix recrystallized in the greenschist facies. They are the products of the earliest Caledonian movements which were followed by at least two further tectonic and prograde metamorphic episodes before the secondary mylonites, constituting the actual Moine thrust and associated high-level movement planes, were formed. These secondary mylonites sometimes contain unaltered fragments of the primary mylonites. Since the latter have remained substantially in a low-grade state in this peripheral zone of the Caledonian orogene, they could hardly be retrograded by the movements on the Moine and related thrust planes. Thus it cannot be altogether excluded that the high-level movements of the Moine thrust were the predecessors of a third Variscan (?) orogenic cycle affecting its distant Precambrian and Caledonian foreland.

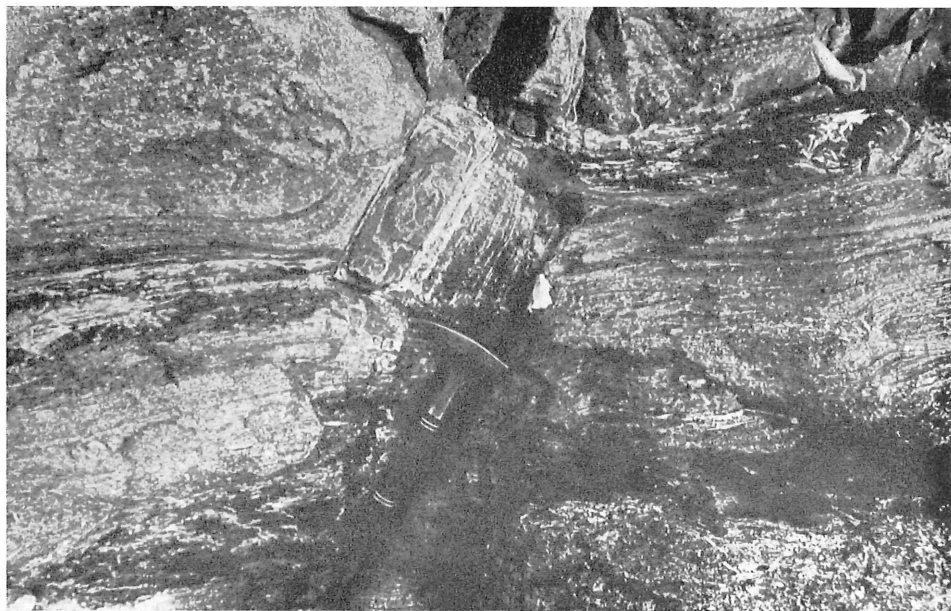


FIG. 10. — *Rotated fragment of ultramafic blastomylonite in re-mylonitized metabasites of the Carreiro zone of tectonic movement in the eastern border zone of the Precambrian complex at Cabo Ortegal (NW Spain).*

Bellière (1971) and Anthonioz (1971) have attempted to classify the mylonitic rocks according to their facies group, in other words their epi-, meso- or catazonal conditions of formation, and to relate them to the plutonic versus metamorphic character of the mono- versus polycyclicality of the terrain in which they occur. The common denominator of their conclusions is, that there are hardly any unequivocal criteria for the polycyclicality of a mylonite other than re-mylonitization and its strongly diaphthoretic character referred to above. The mylonites of the lower crust and upper mantle, which are granulite or eclogite facies rocks with discoid or fusiform quartz or olivine crystals, often arranged in discrete bands, are the highest-grade and most thoroughly recrystallized blastomylonites in the scale. By themselves these catazonal mylonites do not constitute evidence for polycyclicality, but when fragmented and retrograded by non-metamorphic or epizonal mylonites they leave no doubt whatsoever about the incidence of a second orogenic cycle.

The high-grade Precambrian inliers in the Variscan orogene of NW Spain and NE Portugal are also beginning to yield evidence of this calibre. Catazonal and mesozonal blastomylonites of Precambrian age have been recorded for some time (den Tex and Vogel, 1962; Anthonioz, 1968), while Anthonioz (1967, 1969) described basal breccias with a low-grade or non-metamorphic matrix containing fragments of amphibolite from the Precambrian inliers of Bragança and Morais. In the well-exposed Carreiro zone of tectonic movement on the west coast of the Precambrian inlier at Cabo Ortegal, Vogel (1967) and Engels (1972) discovered rotated blocks of ultramafic and mafic granulite facies blastomylonite set in a matrix of amphibolite facies blastomylonites (fig. 10).

These features could also be attributed to pinching, drag, and selective protection from retrogradation of competent horizons within one orogenic cycle, but the recent discovery⁽²⁾ of nonmetamorphic kakyrites and epizonal mylonites, affecting mesozonal blastomylonites of the Punta de Prado formation, which constitutes the north-eastern perimeter of the "Oredenes basin," lends much more weight to the hypothesis of polycyclicality of the terrain in question (figs. 3 and 11).

CONTRASTING FACIES SERIES AND POLYMETAMORPHIC MARKER FORMATIONS

Another metamorphic criterion that has been applied successfully in the Variscan orogene of Western Galicia is the uniqueness of the Variscan and Precambrian facies series here involved. Whereas the Precambrian facies series is of the high-pressure-intermediate-temperature type, carrying kyanite in metapelitic and garnet in metabasic rocks, the Variscan facies series is of the low-pressure (Abukuma) type with andalusite and sillimanite in metapelites and no garnet in uncontaminated metabasites (den Tex *et al.*, 1972).

A useful marker formation to separate the orogenic cycles in Western Galicia is also available. It plays the same role as the basic dyke swarms in the Lewisian of NW Scotland, but here it is provided by a generation of epeirogenic granites, dated by whole-rock Rb/Sr isochron at 460-430 m.yr. (Priem *et al.*, 1970). These orthogneisses did not suffer any of the high-pressure metamorphism and associated tectonization, but were affected by the complete cycle of low-pressure metamorphism and Variscan tectonics (fig. 3). However, Capdevila (1968, 1969) mapped kyanite in undoubted Paleozoic rocks in eastern Galicia and, on this ground, he claimed that kyanite could not be used as an index mineral of the Precambrian metamorphism in NW Spain. Indeed, in isolated occurrences, kyanite may not be significant, but when found throughout a facies series, including partially fused migmatites, it performs a role transcending that of kyanite in the Variscan metamorphism where migmatites and anatectic granites contain sillimanite as the stable aluminium silicate.

THERMAL DOMES, DIAPIRS AND THE CONVECTING BASEMENT EFFECT

The Variscan kyanite, reported by Capdevila, occurs along synformal axes between basement highs such as the Ollo de Sapo ridge and the Lugo dome in eastern Galicia (fig. 3). This poses the question of the influence of basement domes on the geothermal gradient and thereby on the metamorphic facies series ultimately developed in a particular area. Fonteilles and Guitard (1968a, 1968b) have devoted much attention to what they called the "effet de socle" (basement-effect) in metamorphic terrains. Their work in the eastern Pyrenees revealed that regional metamorphic zones, as delimited by mineral isograds, are not only concentrically disposed around basement domes and the "roots" of nappes, but that they are progressively telescoped as the top of the basement rises and the thickness of overburden decreases. If, as Fonteilles and Guitard assume, the thermal conductivity of basement and supracrustal rocks is roughly equal, and the thermal flux at the base of the crust is uniform, this basement effect should be due entirely to the more or less complete absorption of the latent heats of fusion and transformation involved in the mineral reactions defining isograds. In high-

(2) W. F. J. Burgers (pers. comm., 1974).

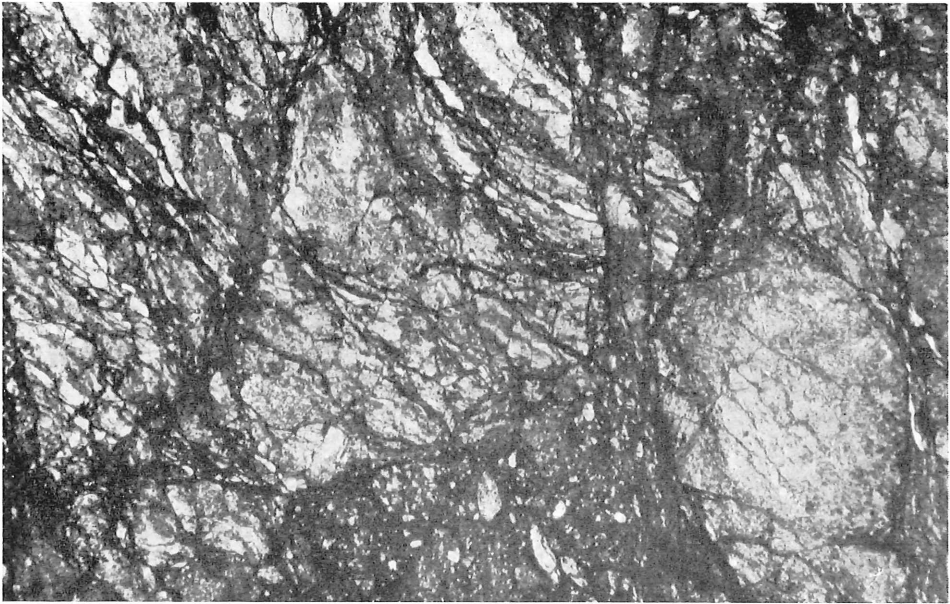


FIG. 11. — *Mylonitized blastomylonite from the Punta de Prado formation in the NW perimeter of the Precambrian complex around Ordenes (NW Spain). Thin section showing fragments of blastomylonite (upper left) in mylonite matrix (lower centre). Crossed polarizers. Lin magn. . . . x.*

grade basement rocks partial fusion only should absorb heat of this type, while the heat produced by other dehydration reactions might be balanced by early retrograde hydration reactions. Steep and rapidly growing temperature gradients could thus be maintained in the basement. In the deeper supracrustals, at high temperatures, much heat would be absorbed by dehydration reactions running to completion between a number of isograds. This would lead to temperature gradients becoming less steep, the more so, the greater their depth to basement. The rate of rise of the geoisotherms in time should also slow down in this zone giving most of the endothermal reactions sufficient time for completion, except at the present level of the isograds themselves where unarmoured relics of lower grade (e.g. staurolite in andalusite) are often observed. According to Fonteilles and Guitard the instability of the thermal regime, at all stages but the very climax, is best seen in the high-level supracrustals where at low temperatures, provoked by slowly rising, gentle gradients, the kinetics of the early dehydration reactions are much slower than the rate of rise of the geoisotherms, yielding many disequilibrium assemblages in vaguely defined metamorphic zones.

This model of a basement effect on the metamorphism of the supracrustal envelope has a number of shortcomings. Firstly, it fails to account for the fact that such an important group of basement domes as the Saxon granulite range, the Münchberger gneiss massif and the NW Iberian complexes of Cabo Ortegal, the Ordenes "basin," Bragança and Morais has had an opposite effect on the

Variscan metamorphism of their envelopes. These basement inliers often contain Variscan kyanite instead of the usual andalusite or sillimanite and they are surrounded by wide, vaguely delimited zones of low-grade Variscan metamorphism. A second defect of the model is that it fails to explain why the basement rose at all to form the domes, why they are frequently later than the associated nappes, and why this uprise started earlier than the regional metamorphic climax (as evidenced by the isograds transgressing obliquely into the dome) and continued later as shown by steepening, inversion, telescoping and even transgression of metamorphic zones by the basement domes.

In fact, Fonteilles and Guitard based their model of the thermal regime entirely on the, allegedly uniform, heat conductivity of the rocks and on their latent heats of transformation and fusion. They disregarded the effective heat flow due to thermal convection arising from gravitational instability in a layered system heated from below. Talbot (1971) incorporated the convection effect in his model for the emplacement of a mantled gneiss dome in Fungwi Reserve, Rhodesia. He showed that under certain reasonable assumptions the Rayleigh number critical for convective overturn in layers of felsic gneiss or granite, covered by supracrustal rocks, may be reached at temperatures well below the solidus. This would lead to at least half a cycle of convection, giving rise to a mantled gneiss dome, or, (if the Rayleigh number is greatly exceeded on the scale employed) to one or more full cycles of convection yielding homogenized granitic domes, diapirs, tabular batholiths and surface flows (fig. 12). Of the various parameters involved, the thickness of the convecting layer is most sensitive since it enters at the fourth power into the Rayleigh number. The local thermal gradient required for convection is much higher than can be accounted for by regional heat conduction alone (270 °C/km in the case of the smallest wavelength domes at Fungwi Reserve), while the viscosity of the gneiss dome could have been as low as 5×10^6 poises, close to its melting temperature, allowing it to yield by pseudo-viscous flow. Talbot expects the local heat flow to be stabilized by an intimate interplay of convection and conduction through thermally anisotropic minerals of specific preferred orientations.

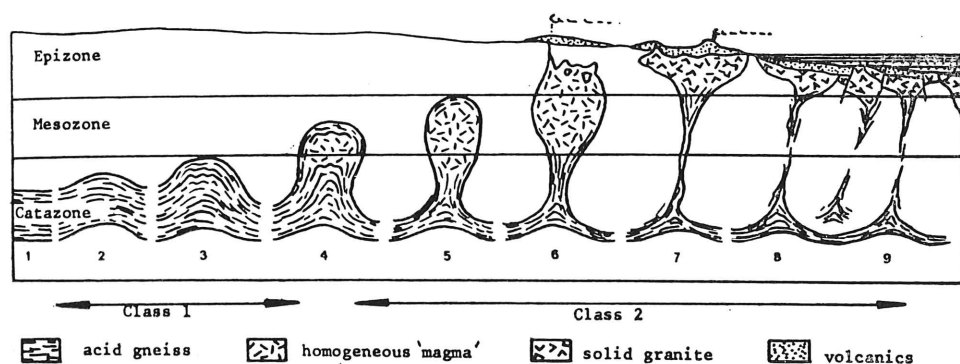


FIG. 12. — Classes and stages of domes and diapirs rising from an acid source layer during orogenesis. The Lepontine gneiss domes *ss.* could be subsolidus gneiss domes of class 1. The Bergell Alps could correlate with a partially fused granitic diapir of class 2 (stages 5-6) (from Talbot, 1971, fig. 8).

Now, if it is true that felsic gneisses under appropriate conditions can reach their critical Rayleigh number while well below their melting temperature, then not only would a mechanism be provided for the early rise of basement domes, but also for the steepening through a convective half-turn of the thermal gradient leading to a rapid rise of the geoisotherms, and to creation by telescoping of narrow metamorphic zones above the basement dome. More often than not such rising gneiss domes would melt progressively by adiabatic expansion, particularly so at the top, where a mantle of watersaturated arkosic erosion products of the gneiss may still be present to promote complete fusion. Examples of mantled gneiss domes and basement diapirs surrounded by zoned metamorphic aureoles are being found in the majority of orogenic belts.

In the Alps, which is probably the most profoundly studied orogene of all, their scarcity and subdued nature has somewhat delayed recognition. Steeply dipping gneissose margins occur in the Lepontine gneiss region or Ticino culmination of the Penninic nappes in the Swiss Alps. This large-scale dome structure, which extends eastward into the Misox-Valle della Mera region of the Grisons (E. Wenk, 1956), seems to consist of a regular grid of smaller scale subdomes, centred roughly on the domains of quaquaversal dip and downdip plunging lineations as mapped by E. Wenk (1955, 1956) in the Verampio and Osogna areas, possibly also in the Airolo-Peccia and Misox-Valle della Mera areas, and certainly so in the Bergell Alps (H. R. Wenk, 1973).

Of these the Bergell complex is a special case. Longtimes considered to be one of the few examples in the central Alps of completely igneous and peripherally flow-layered granite and tonalite, intruded at the very end of the Alpine orogeny, it was later shown to be intimately related to the structural and metamorphic evolution of the Lepontine gneiss region *sensu lato* (E. Wenk, 1956, 1962; Moticska, 1970; H. R. Wenk, 1970, 1973). The more obvious subdome centres appear to be connected by zones of steep foliations and subhorizontal lineations, such as are expected to occur between twin-domes from Ramberg's (1967) centrifuged models (fig. 13). The wavelength between subdomes is in the order of 50 km, putting a constraint on the thickness of the convecting gneiss layer of 25 km—assuming rigid boundaries—, which is a realistic value when compared with recent geological and geophysical estimates (Giese, 1968; Niggli, 1970). The pegmatite, migmatite and sillimanite domains, as figured by Wenk (1970) and Niggli (1970), appear to coincide with the more obvious subdome centres in the Bergell Alps, the Misox and the Osogna areas, and to follow the zone of steep foliations in the so-called Maggia cross-zone between them. According to the subdome model these parts must have been the loci of maximum heat flow. Less clearly related to the subdome patterns, but covering or enveloping the Lepontine gneiss dome *sensu lato*, are the distribution patterns of regional metamorphic minerals such as plagioclase (An_{17-85}) and tremolite in calcite-bearing rocks, kyanite, staurolite, chloritoid and stilpnomelane, as figured by Wenk (1970) and Niggli (1970). Today it is an acknowledged fact that the Alpine thermal dome structure is centred on the Lepontine gneiss region. But it is still widely held that deep burial, heat conduction and subsequent uplift could cause this thermal dome with an estimated average gradient of 30 °C/km to develop. However conductivity, which decreases with temperature and increases with pressure and water content, cannot account for the inward and downward steepening of the thermal gradient that can be deduced from the outcrop of isograds in the Lepontine structure. Also it appears

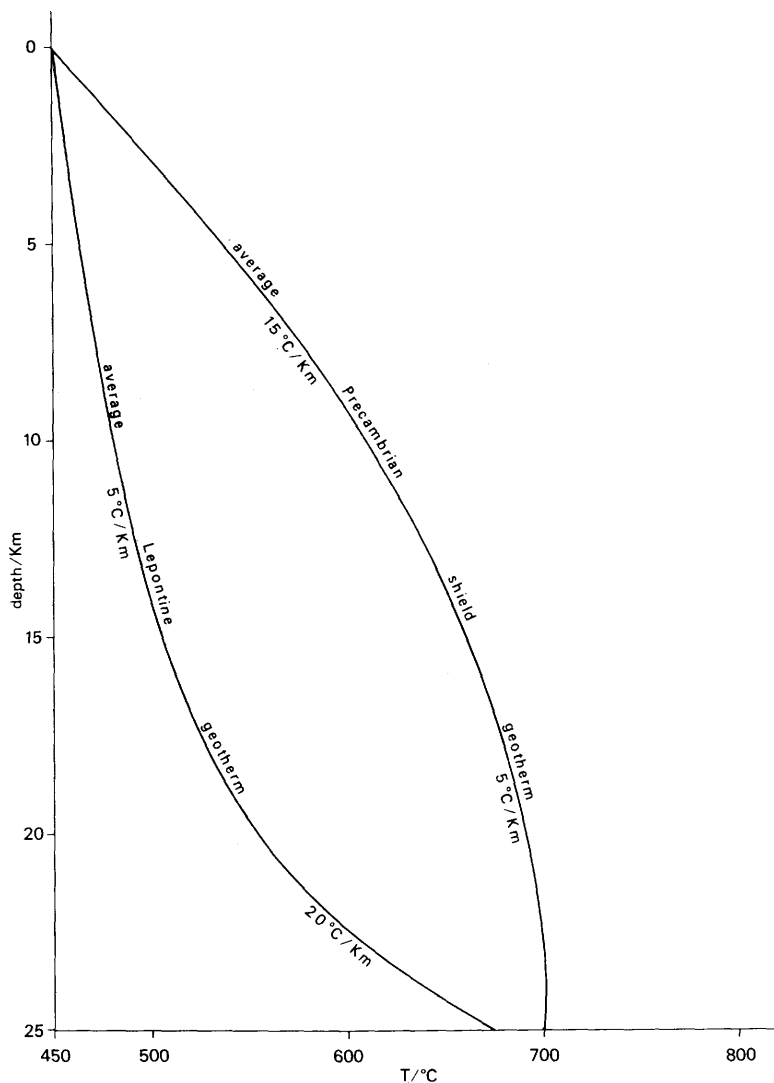


FIG. 13. — Geothermal gradient as deduced from the distances between isograds in the Lepontine Gneiss Region, compared with the average Precambrian shield geotherm, down to a depth of 25 km in the continental crust.

that the deeper parts of the Lepontine subdomes, carrying sillimanite, migmatites and pegmatite veins were at, or slightly above, the minimum melting temperatures of felsic rocks. Using Talbot's (1971) constraints the four or five Lepontine subdomes with average diameters of 2×10^6 cm, wavelength of 5×10^6 cm, thickness of the active layer of 2.5×10^6 cm, viscosity of 10^7 stokes, and average thermal gradient of 4×10^{-5} [$30^\circ\text{C}/\text{km}$ as estimated by Jäger (1973), and $52^\circ\text{C}/\text{km}$ as

quoted by Wenk (1970)] yield a Rayleigh number of 8×10^{14} . This exceeds, by 11 orders Talbot's estimate of the critical value for thermal convection. Even allowing for viscosities ten orders higher than assumed to be effective in the partially molten rocks at the crests of the deeper subdomes, the active gneiss layer of 25 km thickness could still have carried out half a convective overturn, if a million years had been available for its completion. As Talbot pointed out, thoroughly cyclical convection would destroy most of the pre-existing structures in the active layer. E. Wenk (1962) seems to think that such homogenization has indeed been paramount in the Lepontine gneiss dome since he said "Unsere Tiefenzone scheint das Gedächtnis für ihre Vorgeschichte verloren zu haben; es hat eine Wiedergeburt stattgefunden," but other authors (Zwart, 1967; Milnes, 1969; Niggli, 1970) gave evidence that pre-Alpine minerals, structures and isotope ratios have—at least partially—been preserved. Highly refractory bodies, such as the garnet peridotite of Alpe Arami, occurring in the Osogna subdome, have undeniably been subject to several pre-Alpine structural and metamorphic phases (Möckel, 1969; den Tex, 1969).

Again, the Bergell dome differs markedly from the Lepontine subdomes *s.s.* Detailed petrological and structural work by H. R. Wenk (1970, 1973) has revealed that the Bergell complex is conformable and accordant with the Muretto anticline in the overlying Margna and Suretta nappes, and that its predominantly megacrystal and foliated granodiorites and tonalites constitute an asymmetrical structure lying flat, with a blastomylonitic sole, on the Gruf migmatites of the Tambo nappe in the NW, while dipping steeply towards the "root zone" along the Insubric line in the SE. H. R. Wenk's cross-sections (1973) are highly suggestive of a diapiric "hat" in the NW and a pinched-off diapiric "root" in the SE portion of the Bergell Alps. A rim syncline in the overall denser sequence of amphibolites, calcisilicate rocks, marbles and ultramafites belonging to the Suretta and Margna nappes, which surround the tonalitic and granodioritic "hat," is clearly visible on his sections 10-8, in the extreme N of the massif. The radiometric age of the granitic rocks, pebbles of which occur in Oligocene molasse deposits of the Po valley, has not yet been settled satisfactorily, but Gulson's (1973) data indicate that their system was closed after a possibly subsequent event of lead isotope exchange at 30 m.yr., i.e. five to eight m.yr. later than the climax of the Lepontine regional metamorphism (Jäger, 1973). These facts taken in conjunction with the telescoped character of its extremely few, steep and narrow metamorphic zones containing at the highest grade wollastonite instead of the Lepontine association of calcite + diopside or anorthite, and with the temperatures of well over 600 °C that must have obtained during the formation of the gneissic structure and the blastomylonites, suggest that the active gneiss layer under the Bergell Alps was subjected for a longer time to higher temperatures than that in the Lepontine region *s.s.*, leading to increased thermal gradients and density contrasts, and hence to cyclic thermal convection, to more complete homogenization and melting, and to diapirism of the rocks involved. The time lag of 5-8 m.yr. seems ample for the diapiric rise of granodioritic material from the active layer, which would explain the delayed effect of locally increased convective heat flow on a regionally cooling thermal structure. In Talbot's (1971) scheme the Bergell complex would classify as a class 2 dome (fig. 12, Nos. 7-9), whereas the other Lepontine subdomes should be class 1 (fig. 12, Nos. 1-3).

Winkler's (1970) suggestion of a mantle diapir as the ultimate cause of the

Lepontine thermal structure is interesting. Its focus might have been located beneath the Bergell diapir, but it could have been responsible for the uprise of the other Lepontine gneiss domes as well. However, without the intervention of crustal convection I doubt if downward steepening geothermal gradients⁽³⁾ could have been maintained so far above the top of a mantle diapir (fig. 15).

In orogenic belts due to continent-continent or continent-ocean floor collision, where crystalline continental crust underlies thick columns of continental slope deposits, convection of basement may well be the dominant transporting and concentrating agent of heat from the upper mantle upwards, causing regular grids of thermal dome structures to arise with more or less homogenized and diapiric gneisses or granites located in their focal parts. Thus, the notion of a basement effect in orogenic metamorphism seems realistic enough, but since basement may also be upthrust or recumbently overfolded in a much colder rheid or rigid state, I propose to speak of the *convecting* basement effect to express the causal link between regional thermal structures and convecting basement domes and diapirs in ensialic orogenic belts.

GRANITES AND GRANULITES: WITNESSES OF CHEMICAL RECYCLING WITHIN THE CRUST?

The lower continental crust, or intermediate layer of seismologists, was formerly held to be composed exclusively of gabbroic or basaltic rocks. In the last decade, however, a number of authors (Heier, 1965; Heier and Adams, 1965; Green and Lambert, 1965; den Tex, 1965; Ringwood and Green, 1966; Belousov, 1966) have adduced geochemical, geophysical and petrological evidence that the lower crust should consist of anhydrous granulite- and eclogite-facies rocks, the chemistry of which ranges from acid to ultrabasic but could well become increasingly basic with depth.

Granulites are rocks that have been metamorphosed under conditions of high lithostatic (4-17 kb) and much lower vapour pressures. At depths of 15-60 km, corresponding to these lithostatic pressures, low water vapour pressures can only be due to water deficiency in the fluid phase (Winkler, 1967; Brown and Fyfe, 1970). Water-saturated metasediments and tuffs, subjected to monocyclic metamorphism, obviously do not meet this requirement, but igneous rocks (with water contents between 0.5 and 1.5 wt %), amphibolites and associated gneisses (modal water content 1-1.5 wt %) could qualify, deriving only limited amounts of a free aqueous phase from the dehydration of their muscovite, biotite or amphibole crystals. The crystalline basement is therefore liable to be thoroughly dehydrated during a subsequent orogenic cycle, the more so when a minimum melting fraction that soaks up the available free water phase is formed, and can effectively be separated from the residue. Fluid inclusions in the quartz crystals of many granulites were shown by Touret (1971) to consist for the greater part of

(3) The average geothermal gradient in the Lepontine gneiss dome has been deduced from the true distances and temperature differences between isograds. It works out at 5 °C/km from the "stilpnomelane-out" to the "staurolite-in" isograd, and 20 °C/km from the "staurolite-in" isograd to that of incipient anatexis in felsic gneisses. Towards the Bergell Alps the latter value rises steadily to a maximum of 75 °C/km.

high-density CO_2 . Since carbon dioxide is less soluble than water in granitic melts, its addition to the liquid phase has a tendency to raise the melting point of granitic rocks (Wyllie and Tuttle, 1959). Water should therefore fractionate into the liquid, and carbon dioxide into the residual vapour phase. This might explain why Touret finds a sudden increase in the partial pressure of water vapour when passing from granulite to amphibolite facies rocks in southern Norway. Brown and Fyfe (1970) and Fyfe and Brown (1971) have produced P/T graphs for the incongruent melting under water-deficient conditions, of muscovite, biotite and hornblende to sillimanite or kyanite, garnet and pyroxene plus granitic or dioritic melt. Fyfe (1970, 1973) discussed possible collecting and separation mechanisms of the granitic melt from the granulitic residue. He finds that drops of granitic melt would collect easily owing to viscosity differences in the solid-liquid system of the order of 10^{10} poises, and that drops of melt should not ascend from the melting front until a commensurate bulge is formed by a spontaneous penetration wave in the low-density layer (Ramberg, 1967).

Much geochemical evidence has lately been adduced to show that granites and other felsic amphibolite facies rocks on the one hand, and high- to intermediate-pressure granulite facies rocks on the other, are complementary chemical systems. Compared with crustal averages the said granulites appear to be impoverished in some elements that are enriched in granitic rocks and vice versa. In particular the radioactive elements U, Th, Rb, and to a lesser extent K, are depleted in lower crustal granulites and enriched in upper crustal amphibolite facies rocks, notably so in granites (Heier, 1965, 1973; Fahrig *et al.*, 1967; Sighinolfi, 1969, 1971). This trend is simulated by other lithophile elements, the ionic charge or radius of which is so large as to preclude their incorporation by isomorphous substitution in high-melting crystal lattices, and to induce their fractionation into low-melting crystals. This is the case with Cs, Li and possibly Pb; whereas Ca, Mg, Fe, Ti and Mn are often enriched in granulite and impoverished in amphibolite facies rocks (Lambert and Heier, 1968; Heier and Brunfelt, 1970; Sighinolfi, 1971).

At given levels of K-abundance Rb appears to be more strongly enriched in granitic, and depleted in granulitic rocks, than K at given levels of Rb-abundance. Thus, with increasing potassium content, the K/Rb ratio decreases (Shaw, 1968; Lewis and Spooner, 1972). It is lowest in upper crustal granites (e.g. Red Heemskirk granite, Tasmania: 80 at 5.30 % K) and much higher in lower crustal granulites (e.g. Colton granulite, Adirondacks: 830 at 0.48 % K). The reason for this behaviour could be that Rb is admitted by K in the crystal lattice of micas rather than in those of feldspars and amphiboles. Incongruent melting of muscovite and biotite, and fractionation of the partial melt, should lead to preferred depletion of Rb in granulitic residues and its preferred enrichment in granitic melts.

On the other hand Sr is admitted by Ca and, to a lesser extent, captured by K in minerals such as amphibole, plagioclase and potash feldspar. Sr therefore does not fractionate into the low-melting granitic component to the same extent as Rb does. Consequently the Rb/Sr ratio is significantly lower in granulites (down to 0.007 at Cabo Ortegal⁽⁴⁾) than in granites (up to 70.9 in the White Heemskirk granite, Tasmania), as compared to the average crustal ratio of 0.18 (Faure and Powel, 1972).

(4) P. W. C. van Calsteren (pers. comm., 1974).

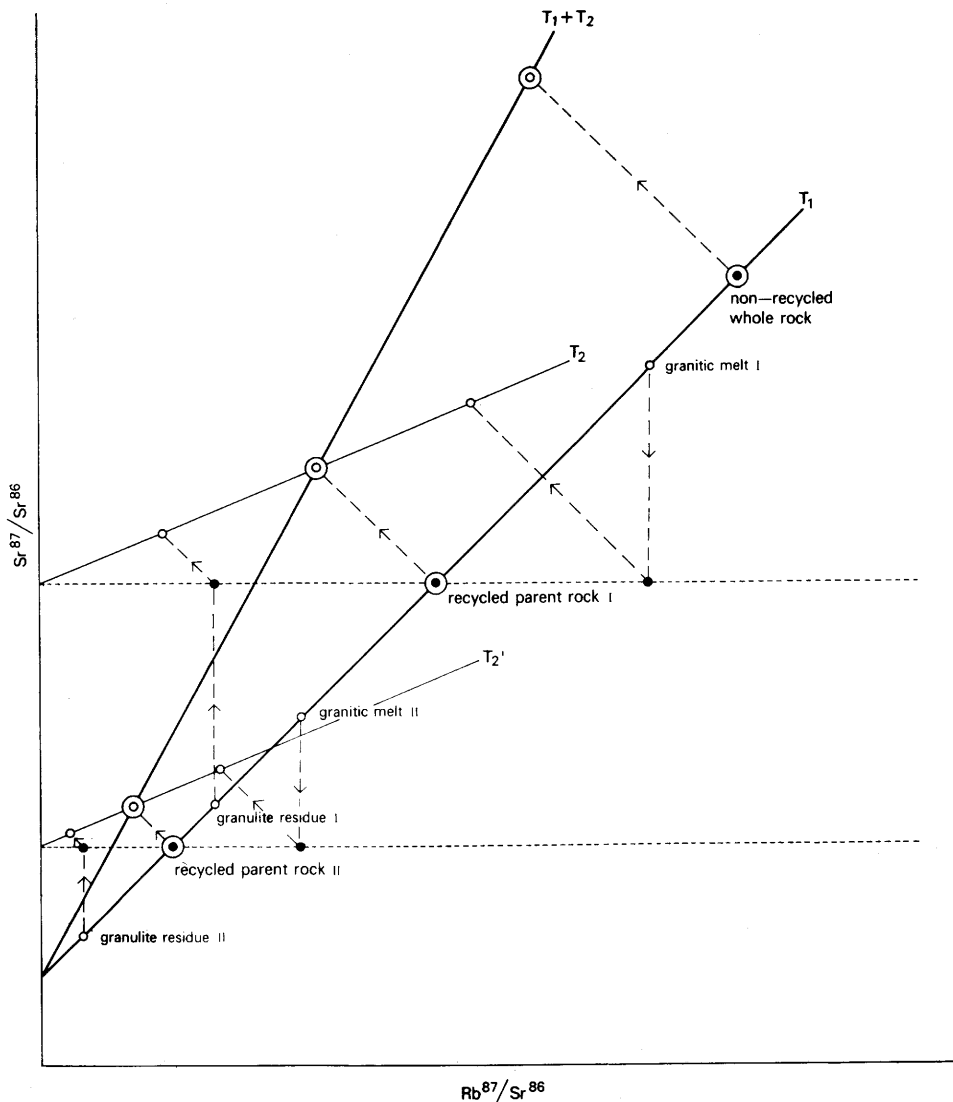


FIG. 14. — Multistage model of the evolution of Rb/Sr and Sr-isotope ratios in parent rocks split into a granulitic residue and a granitic melt and in nonrecycled whole rocks.

If granites and granulites are indeed the counterparts of fractional fusion in a crystalline lower crust, then their Sr-isotope evolution patterns should conform to a multistage model. In a Nicolaysen diagram (fig. 14) whole rocks should at some stage have defined an isochron representing the time interval (T_1) elapsed since their homogeneous Sr-isotope system first became chemically closed. The development at T_1 of fractions of granitic melt and granulitic residue from a single parent rock should cause their separation along the isochron representing T_1 owing to the incongruent melting of mica crystals, richer in Rb than

the parent rock, and the crystallization, in the residue, of sillimanite, kyanite, pyroxene or garnet, poorer in Rb than the parent rock. Renewed homogenization of Sr-isotopes during the process of partial fusion causes the Sr-isotope ratios of the fused and residual fractions to regain the level of the individual parent rock.

After a further lapse of time (T_2) the recycled systems of melt, residue and parent rock should define a new isochron. In the case of large scale, homogeneous recycling the previous history of the parent rock, as represented by the isochron T_1 , is then erased from the Sr-isotope evolution pattern. However, large scale homogeneous recycling of geological materials is not likely to be the rule. Individual parent rocks of different Rb/Sr ratios should, upon recycling, give rise to a set of parallel isochrons T_2 while the parent rocks alone should define a new isochron $T_1 + T_2$. Whole rocks that escaped recycling, owing to their refractory nature or insulated occurrence, should also plot on the isochron $T_1 + T_2$ but without yielding an isochron of the type T_2 . Such heterogeneous recycling has obviously taken place where amphibolite facies rocks occur intimately interstratified with granulite facies rocks, as is often the case in granulite terrains. Samples on the scale of a single layer stand a much better chance of being re-homogenized as regards their Sr-isotopes, than large bulk samples. The latter alone should therefore usually fail to produce clear evidence of isochrons representing the recycling event T_2 . They should exhibit a tendency to remain aligned along the isochron $T_1 + T_2$.

Krogh and Davis (1972, 1973) have shown that some of the layered French River paragneisses from the Grenville Province in Canada were isotopically re-homogenized on the scale of centimeters about a 1000 m.yr. ago, but that larger samples of granitic gneiss still occupy positions on much older isochrons in a Nicolaysen diagram.

Interlayered granulite and amphibolite facies stromalites in the Ivrea-Verbano zone of the Southern Alps were demonstrated by Graeser and Hunziker (1968) to exhibit a very similar pattern of Sr-isotope evolution. Three large scale samples have ratios that can be aligned along isochrons varying between 384 and 1600 m.yr., whereas two of these also plot very close to the isochron of 310 m.yr. defined by eight small-scale samples from individual stromalite layers. The claim by Graeser and Hunziker, that the latter event cannot have been one of degranitization and granulite formation, because the large-scale samples should then also have lost their memory of a previous record, is invalidated by the fact that two of the pertinent large scale samples do indeed plot on or near the 310 m.yr. isochron, while the third one may not have been recycled at all or may have been the parent rock of another set of partial fusion products in the manner shown by figure 14.

Somewhat similarly banded granulite and amphibolite facies rocks from the Waldviertel in Lower Austria have been analyzed by Arnold and Scharbert (1973). Here the age difference between large and small scale samples is much smaller though still significant (469 ± 11 and 431 ± 35 m.yr. respectively). Jäger and Watznauer (1968) and Watznauer (1974) studied comparable formations in the Saxon Granulite Range and found 6 large scale samples to define an isochron at 437 ± 26 m.yr., while one K and Rb-rich sample plots below the isochron. In fact, its age may be much lower than the 356 m.yr. inferred by Jäger and Watznauer on the assumption that it might have the same initial Sr-isotope ratio as the other samples.

The charnockitic gneisses of Caramany from the Agly massif in the eastern Pyrenees, investigated by Vitrac and Allègre (1971), are suggestive of a two-stage development from the gneisses of Agly and Canigou, which define a 580-550 m.yr. isochron. At least three of these gneisses can also be fitted to a 300 m.yr. isochron, an age which is supported by only slightly discordant U/Pb ratios in zircons from an associated charnockitic granite (Vitrac, 1971, 1972).

The initial isotope ratios of Sr are also critical of recycling. Spooner and Fairbairn (1971) compiled such data for pyroxene granulites from worldwide sources and found a narrow range of low values between 0.700 and 0.707. These ratios are matched by those from the older isochron in the Pyrenees and the Ivrea zone, and approached by that from the Lower Austrian granulites in large samples (0.7095). Recycling leads to significantly higher initial ratios, which is borne out by the younger isochrons of small-scale samples from Lower Austria (0.720), the Pyrenees (0.715) and the Ivrea zone (0.717).

The evidence from Sr-isotope evolution patterns appears to support the heterogeneous multistage model in the granulite terrains of central and southern Europe. Small-scale recycling of an older basement during the Variscan orogeny fits the geological context of that part of the world rather well. In particular the uplifted Ivrea-Strona-Generi zone may well represent the crustal level where the intrusive Variscan granites of the Southern Alps (Mont'Orfano, Baveno, etc.) were collected and drained away as partial melts.

Independent evidence for the companionship of granites and granulites may be forthcoming from integrated studies of the growth patterns of the radiogenic lead isotopes ^{207}Pb and ^{206}Pb with respect to their parent nuclides ^{235}U and ^{238}U . The concordia curve, representing the differential growth with time of the two ratios $^{207}\text{Pb}/^{235}\text{U}$ and $^{206}\text{Pb}/^{238}\text{U}$ in closed systems, is such that loss or gain of the system in uranium as well as loss of lead, cause the ratios to depart from the curve along chords, the upper intercept of which marks the date the system originated, while the lower intercept may indicate the time at which the loss or gain occurred (Wetherill, 1956, 1963).

Now if we assume that a parent rock system is opened by the act of forming two individually closed, fractionated systems: a melt and a residue, U (along with other granitophile elements) should concentrate in the low-melting granitic fraction and be depleted at the same rate in the refractory residue of partially fused basement systems. The behaviour of Pb under such conditions is much less obvious, and its possible loss or gain is neglected for the present purpose. Such fractional systems of granites and granulites with their predominantly new and old generations of U-bearing minerals, such as uraninite, zircon, xenotime, monazite, allanite, pyrochlore, apatite and sphene, should possess radiogenic (U/Pb ratios that plot on the chord intercepting concordia at the date of origin of their parent rock (T_0), and at the time the original system was fractionated by partial melting (T_1). Thus granite zircons, enriched in U relative to Pb, should plot at Q_1 , closer to the lower intercept, whereas residual granulite zircons depleted in U relative to Pb, should be located at Q_1' , beyond the upper intercept with concordia. However, diffusional loss of lead during a later episode, T_2 , could disturb this pattern by shifting the U/Pb ratios of granite and granulite zircons along two separate chords (T_2-Q_1 and T_2-Q_1') towards the new lower intercept: T_2 (Wetherill, 1963). Continuous diffusion of lead from the systems (Tilton, 1960) should have a grossly similar disturbing effect on zircon U/Pb ratios of granites and granulites.

In fact, monazite, apatite, allanite, pyrochlore and uraninite frequently show reverse discordance occupying locations on the upward extension of discordia, as predicted above, but granulite zircons nearly always exhibit normal discordance plotting below concordia, albeit often close to the upper intercept of discordia; whereas granite zircons show a distinct preference for the vicinity of the lower intercept with concordia (Doe, 1970). Loss of lead by episodic or continuous diffusion appears to be a feature especially common in zircon crystals. When following a substantial loss or gain in U, it causes their U/Pb ratios to increase and to revert to positions below concordia along slightly diverging chords, the upper intercepts of which have an erroneous significance on the time-scale. The granulite zircons should yield too high and the granite zircons abnormally low ages for the parent rock closure (fig. 15). Concordia plots of zircons from granites and orthogneisses, recorded in the literature, are mostly concentrated near the lower intercept of discordia chords, sometimes suggesting ages significantly lower than indicated by their Rb/Sr whole rock isochrons or by their geological context.

Zircons from granulites have not been analyzed as extensively yet, but with refined techniques, using fractional samples down to 0.6 mg, more specific data are rapidly becoming available. Their concordia plots are nearly all discordant and often define several parallel or mutually intersecting chords (Krogh and Davis, 1971, 1972).

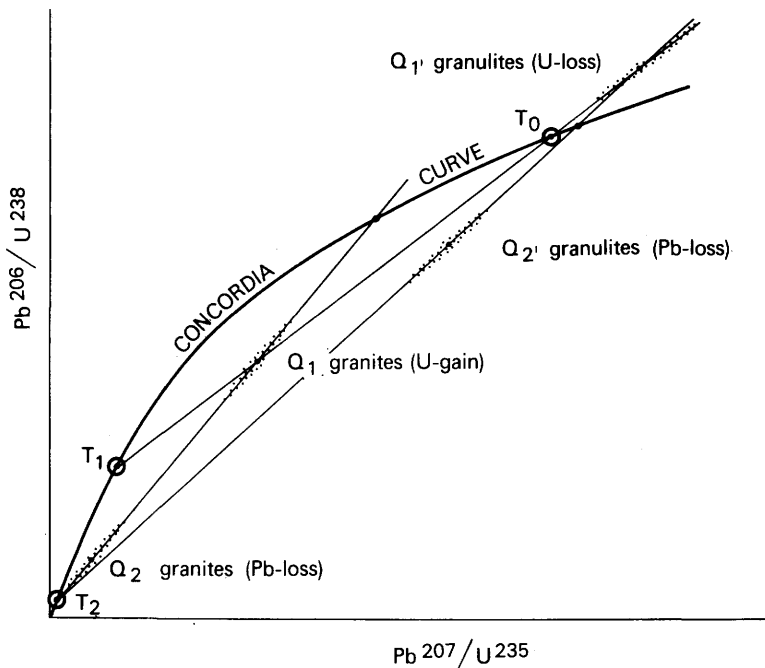


FIG. 15. — Concordia/discordia model of isotopic U/Pb ratios in zircons from granites and associated granulites derived by partial melting (T_1) from a parent rock at T_0 , having suffered a subsequent lead loss at T_2 .

Uranium depletion was shown to have affected zircons from the Lewisian granulitic basement complex in Scotland $2\,900 \pm 100$ m.yr. ago by Moorbath *et al.* (1969). Pidgeon and Bowes (1972) found that zircon samples from the Scourian granulite terrain near Kylesku (Scotland), when analyzed in separate size fractions, define two separate chords with a common upper intercept at $2\,700 \pm 20$ m.yr. and lower intercepts at $1\,715 \pm 15$ and 515 ± 15 m.yr. So far the data on the Lewisian granulite zircons appear to be insufficient to exclude a complex history of possibly repeated U-depletions followed by lead loss during the Caledonian orogeny.

Separate studies of inherited and newly formed zircons and of associated monazite from granulite and amphibolite facies paragneisses, migmatites and granites of the Ivrea-, Strona- and Ceneri-zones of the Southern Alps have been carried out in the Zürich laboratory for isotope geology and mass spectrometry. Köppel and Grünfelder (1971) distinguished essentially new zircons of almost concordant Ordovician age, and predominantly inherited detrital zircons of much higher discordant ages. The first group is distinctly richer in trace elements like U, Y, P and Ca and occurs in K-feldspar bearing granites, granitized gneisses and some paragneisses, whereas the latter, poorer in these trace elements, is essentially confined to K-feldspar free paragneisses and granulites. Köppel and Sommerauer (1974) analyzed the zircon groups and their trace element contents in greater detail and concluded that those of the first group contain more domains of a phase enriched in trace elements and are subject to greater lead loss than those of the second group. Concordant monazite ages of 275 and 295 m.yr. from the Ceneri migmatite and the Mont'Orfano granite were reported by Köppel (1974). These ages bracket the lower intercept of a discordia curve defined by zircons from granulite facies pyroclastics and paragneisses, the upper intercept of which lies at 1 900 m.yr. When considered in the perspective of U-fractionation by partial melting, rather than in that of lead loss alone, these results could also be taken to suggest that partial melting of basement rocks occurred in the southern Alps with a climax at 285 ± 10 m.yr., depleting in U detrital zircons that suffered previous and subsequent lead losses, while enriching in U newly formed zircons that were only subjected to a subsequent lead loss.

Krogh and Davis (1972, 1973) have shown that multistage models should be used to explain the behaviour of U/Pb ratios in zircon fractions from the French River paragneisses, the Sr-isotope patterns of which were discussed alongside. They concluded that zircon crystals are regenerated during high-grade metamorphism and partial melting, and that mixtures of two zircon generations define a single line intersecting concordia at the two different episodes of zircon growth. Although the possibility of loss and gain of U during zircon resorption and growth is admitted, they do not consider U-gain a plausible solution for the problem of the mutually intersecting chords (fig. 15).

Generally speaking I believe that discordant U/Pb ages, as obtained from monazites and specific zircon fractions in granites, granulites and parent rocks, constitute the most promising approach towards an understanding of the intricacies of polycyclic in anatexitic basement terrains. However, the complicated mechanisms of U/Pb fractionation, lead loss and trace element partitioning over several generations of multiphase zircon-type systems are as yet insufficiently known to permit us to rely entirely on the memory of U/Pb isotope ratios in zircons for past metamorphic and magmatic events. Rb/Sr whole rock ages and

initial Sr-isotope ratios of carefully sized and selected samples are equally important sources of information, if judiciously evaluated. They should be given due weight in assessing the geological and geochemical history of the lower continental crust.

CONCLUDING REMARKS

The polycyclic nature of the continental crust is rapidly gaining recognition. Stratigraphic, tectonic, metamorphic, plutonic and geochemical cycles have left their traces in the record of rocks, minerals and structures, but it is often difficult—if not impossible—to assign features characteristic of recycling processes to identifiable orogenic cycles.

Unequivocal criteria for polycyclicality are scarce and their record tends to be obliterated in the course of time. Yet most rocks have a surprisingly retentive memory for geological events, especially at low temperatures. However, under high-grade metamorphic conditions where partial melting sets in, molasse deposits are granitized; folds, faults, mylonites and low-grade mineral associations and their textures are thoroughly disturbed; and even the crystalline basement may be fractionated and recycled beyond recognition. It is clear that, when such conditions have prevailed, no single parameter can suffice to resolve the problems of origin, history and age of the various geological events that affected the rocks in question.

Deciphering the palimpsest of continental history is like piecing together a jigsaw puzzle with most parts missing, or reconstructing a mammoth on the basis of a few tattered phalanges. Yet considerable progress is being made through the philosophy of multiple approaches and multiple working hypotheses. Geological, geochemical and geophysical models of the continental lithosphere are being produced at a staggering rate, but the cyclical aspect of its evolution is difficult to incorporate and even more hazardous to verify with the shreds of evidence that remain.

The model of continental accretion cannot be accepted in its original version of simple lateral growth, while that of sea floor spreading and plate tectonics suffers from similar inadequacies when extrapolated back into early geological time. Fractionation and recycling, probably caused by penetrative convection and diapirism, both in the upper mantle and in the lower crust of the continents must be considered an important counterpart. The views of P. Michot (1963, 1968, 1972), Wegmann (1965, 1966), Belousov (1966, 1970), Ramberg (1967) and van Bemmelen (1972) have a particular perspicacity for the more distant past of the continental lithosphere. Their voices should not be allowed to drown in the often tumultuous noise made by the band-waggoners of "trendy" thinking. Eventually we may arrive at a balanced model of continental evolution, weighting the available evidence according to the quality of its memory for geological events.

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