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BURIAL METAMORPHISM

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ABSTRACT

Burial metamorphic rocks, as defined by Coombs (1961), are mainly of low grade and have been metamorphosed without being affected by penetrative deformation, *i.e.*, without development of schistosity. Many burial metamorphic rocks show zeolitic assemblages, whereas others have the clay mineral-carbonate association. The validity of these two trends is confirmed by a study of modern active metamorphism in geothermal areas and in deep sedimentary basins where progressive diagenesis can be observed. The primary control of mineral association with depth appears to be temperature rather than pressure. The difference between the two trends may be controlled in part by the ratio of μ_{CO_2}/μ_{H_2O} of the associated fluid phase, and in part by the nature of the protolith — volcanic and volcanogenic rocks tend to become zeolitic.

Experimental data on the phase equilibria of laumontite and analcime give the maximum stability fields of these two typical zeolites; in nature other factors may enter, all of which tend to restrict the stabilities of these zeolites to even lower pressures. Under such low-pressure conditions massive volcanic rocks and possibly also greywackes are not likely to develop penetrative schistosity. It may not be valid to use zeolitic assemblages and lack of schistosity in a rock to deduce simple burial metamorphism. With this in mind, the occurrences of zeolitic rocks in orogenic belts can be simply explained as metamorphism at shallow depths.

INTRODUCTION

Just two decades ago, Coombs (1954) published a report on the metamorphic petrology of Triassic volcanic and volcanogenic rocks of the New Zealand geosyncline from the Taringatura district. In this study, he noted that the mineralogical changes are apparently correlated with the depth of the samples in the stratigraphic section. Because the rocks are not greatly deformed, these depths can be related directly to the depths of burial during metamorphism. In a subsequent paper, Coombs (1961) formally proposed that metamorphism of rocks under conditions of burial, without deformation as evidenced by the development of schistosity, be given a separate niche, and be called "burial metamorphism". According to Coombs (1961, p. 214), in burial metamorphism there is "reconstitution without obvious relation to igneous intrusions and commonly of regional extent; incipient, extensive or complete. The fabric of silicate-rich rocks is not modified by the development of schistosity. The metamorphism appears to follow burial of the sediments or volcanic rocks and is not accompanied by significant penetrative movements. Diagenetic processes in the most restricted sense of that term, that is, occurring essentially at the temperature of deposition, are excluded. Burial metamorphism has been observed to produce mineral assemblages conventionally ascribed to the zeolite facies, the greenschist facies, and perhaps the glaucophane facies. Many slates and sheared greywackes . . . mark transitions from the products of burial metamorphism to those of regional metamorphism."

Since 1954 and especially since 1961, the petrologic literature has included many accounts of low-grade metamorphism of rocks which show little or no evidence of schistosity; these rocks have therefore been designated as products of burial metamorphism. The accounts come from many places: Puerto Rico (Otalora 1964; Jolly 1970), the Andean and Coastal Range geosynclines of Chile (Levi 1969, 1970; Thompson 1971a), the Alps of Switzerland and France (Martini 1968, 1972; Martini & Vuagnat 1965, 1970), northern New Brunswick (Mossman & Bachinski 1972), the Tamworth area, New South Wales, Australia (Packham & Crook 1960), the Tanzawa Mountains, Japan (Seki *et al.* 1969a), and Oregon (Brown & Thayer 1963), to mention but a few examples.

Many of the mineral assemblages attributed to burial metamorphism carry zeolites, mainly analcime or the calcic zeolites such as heulandite and laumontite. This observed association is so common that it is clearly not accidental: Coombs' definition, quoted above, specifically recognizes it and most petrologists will probably accept this. This article intends to examine this association and their possible causal relationship.

TWO TRENDS OF BURIAL METAMORPHISM

The active geothermal areas provide insight to the important variables for mineralogical reactions during burial metamorphism. True, in his classification of metamorphism, Coombs (1961) excluded hydrothermal metamorphism from "burial metamorphism", for obvious reasons. Active geothermal areas tend to be regionally limited, and the thermal centers are distinct and localized. Pressures obtaining in geothermal areas range from low to very low (depth of direct drilling generally not exceeding a few km). The measured geothermal gradients are anomalously high; thus at the Salton Sea field, the Wairakei field, and the Broadlands field, the average geothermal gradients are all around 200-300°C per km (values calculated from Muffler & White 1969, Ellis 1967, and Mahon & Finlayson 1972); these gradients are about an order of magnitude greater than those of normal continental crust and thus, presumably, of most burial-metamorphosed terranes.

Despite these caveats, active geothermal areas remain instructive for our understanding of ancient burial metamorphic terranes, simply because these areas are the rare cases where rock metamorphism can be studied not only as fossil specimens but, with a dash of faith, as an ongoing process. Here it is possible to attempt direct measurement or calculaton of the numerical values for the various physiochemical variables that lead to the observed alterations. Thus, study of active metamorphism even in aberrant areas affords an opportunity to assess the validity of theoretical considerations of controls of metamorphic assemblages by actual examples as well as to calibrate our yardsticks for fossil areas.

In a review of phase relations in low-grade metamorphism, Zen & Thompson (1974) summarized the petrologic data for three active geothermal areas: Salton Sea, Wairakei, and Broadlands. These areas illustrate two metamorphic trends which have already been noted by Coombs (1971) and which seem to be most important for burial metamorphic rocks. One is the clay mineral-carbonate trend, the other is the zeolite trend. The literature and data on mineral assemblages for the three areas do not need repetition; instead, additional examples will be given in an attempt to bridge the gap between contemporary hydrothermal metamorphism and ancient burial metamorphism.

The mineral assemblages of the Salton Sea area (Muffler & White 1969) may be regarded as the archetype of the clay mineral-carbonate trend. These data are consistent with those from the Gulf Coast sedimentary rocks, where information is available through oil field drilling, and where nonvolcanic clastic sediments are also involved. Burst (1969) and Perry & Hower (1970) reported on the clay mineral transitions in the Gulf Coast. Though the changes are more diagenetic than metamorphic, they nonetheless provide data on the beginning of a mineralogical trend in a nondeformed area. The clay-fraction minerals undergo systematic changes with depth (and temperature). According to Burst (1969), the depths of onset of mixed-layer clay and the loss of its expandability are respectively about 2.6 and 3.8 km, and the data of Perry & Hower (1970) generally agree with these values. These depths are the maximum depths as there is no eroded overburden. The geothermal gradient is close to 0.1°C per bar, and the temperatures at the two depths cited are about 100 and 135°C, respectively, close to the actually measured in-hole temperatures. These temperatures are lower than those for comparable mineralogical changes at Salton Sea. Here again zeolites have not been reported, though no independent data on the CO_2 fugacity is known to me.

As a final example of burial metamorphism in the nonzeolitic trend we may cite the Belt sequence of clastic sedimentary rocks in western Montana and northern Idaho. The micaceous minerals of these rocks were studied by Maxwell & Hower (1967), who found that the *2M* polymorph of muscovite and mixed-layer minerals are absent in areas where the reconstructed original section of Belt rocks is thin (about 1.5 km near Little Belt Mountain). In the areas of thicker Belt section (as much as 5-8 km in the Glacier National Park region and 12 km at Pend Oreille Lake area, Idaho), more *2M* polymorph obtains and the amount also increases with depth at a given area. The lowest sample from the thickest section, at Pend Oreille Lake, also yielded biotite. My own *x*-ray study of scattered samples from Glacier National Park and nearby areas show albite, chlorite, muscovite, and quartz, and a mafic rock from near Logan's Pass carries as metamorphic minerals epidote, chlorite, and actinolite; prehnite is suspected though not confirmed. Eslinger & Savin's (1973) measurements of oxygen isotope temperatures from this area indicate values from about 200° to a little over 300°C. Thus the temperatures as well as mineral assemblages from this area resemble the deeper samples from the Salton Sea area, whereas the Gulf Coast samples described by Burst and Perry & Hower resemble the upper parts of Salton Sea cores. Despite the high geothermal gradient at Salton Sea, the mineralogy suggests that the transitions here can indeed be used as a guide to understanding the factors that control the mineralogical reactions in more "normal" areas of burial metamorphism.

The second trend in burial metamorphism is exemplified by the Wairakei geothermal region in New Zealand, described, among others, by Steiner (1953, 1968), Coombs *et al.* (1959), and Ellis (1967), and summarized elsewhere (Zen & Thompson 1974). Wairakei typically represents the zeolitic trend of metamorphism. The most obvious differences between it and the Salton Sea area are the nature of the protolith (volcanic and volcanogenic rocks instead of clastic sediments), and the fugacity of CO_2 (a fraction of a bar instead of about 10 bars). The data from Wairakei can be duplicated by very similar observations elsewhere, for example the Onikobe area in northern Honshu, Japan (Seki *et al.* 1969b), where similar protolith and comparable zeolitic metamorphic assemblages are found. I computed a rough value of the CO_2 fugacity for the Onikobe area of a fraction of a bar on the basis of the data on fluid composition (Seki *et al.* 1969b).

The Broadlands geothermal field in New Zealand provides data suggesting the area to be intermediate between the Salton Sea trend and the Wairakei trend. Calcic zeolites are found, but only mordenite and wairakite, and these are less abundant than at Wairakei. This difference may be correlated with the higher calculated CO_2 fugacity for the Broadlands area: Mahon & Finlayson (1972) gave a value of about 9 bars from water chemistry. However, a recent direct determination, based on the composition of fluid inclusions in hydrothermal quartz crystals, provided an estimate of CO_2 pressure of about 0.25 bar, and a salinity considerably greater than the modern value (Browne *et al.* 1974).

This last point raises the spectre, that contemporary physical and chemical conditions may not correspond exactly to the conditions under which the mineral alteration took place, even for the active geothermal areas. One could bracket fairly closely the total pressure that existed during geologically recent past for these areas, and perhaps even place reasonable limits on possible variations in temperature, but the constraints on the compositions of fluid phases are more difficult to extrapolate back in time. Contemporary geothermal areas may not be so contemporary, after all! Clearly, much detailed study of the fossil fluid phases, such as through fluid inclusions, is needed to clarify this point, even though the general features of the two metamorphic trends seem reasonably clear.

Factors Controlling Burial Metamorphic Trends

What are the factors that might control or at least affect the fate of the metamorphic rocks, and determine whether they will become zeolitic rocks or clay mineral-carbonate rocks? If the active metamorphic areas are any guide at all, then it seems reasonable to draw these conclusions:

1. Temperature strongly affects the mineral assemblages. Figure 1, based on the data of Muffler & White (1969), strikingly illustrates this point for the Salton Sea area.

2. Total pressure has only a minor effect, at least in the range of pressures encountered, on the order of 1 kb. This is emphasized by the data of Salton Sea, by those of the other geothermal areas, as well as by the Gulf Coast region (Perry & Hower 1970). This conclusion is hardly surprising, for most low-grade metamorphic reactions involve a volatile phase, and thus a large entropy effect, whereas the change of volumes of the solids is relatively small. For burial metamorphism involving several kb of pressure, such as proposed for the Chilean Andes or the Taveyannaz greywacke of the Alps, the effect of total pressure may be considerably greater.

3. The different metamorphic trends are not controlled solely by differences in the thermal gradients. Both the Salton Sea area and the New Zealand geothermal areas have temperature gradients on the order of a few hundred degrees per km, yet their mineralogy differs more than, say, between Salton Sea and the Gulf Coast area or the Belt rocks, or between Wairakei and the areas where more normal gradients must have obtained, *e.g.*, in the Chilean rocks.

4. The differences in the solid part of the bulk rock chemistry probably were not the major control. Comparison of the chemical analyses of the Salton Sea sediments (Muffler & White 1969) and the zeolitized volcanic rocks of Taringatura (Coombs 1954) shows that there is at least as much variation from sample to sample in the same area as between the two areas. The major exceptions are the H_2O and CO_2 contents of the solid rock compositions: high H_2O/CO_2 ratio for the zeolitic rocks, and both low ratio and high CO_2 content for the Salton Sea rocks. However, these differences reflect the zeolitic versus carbonate mineralogy, and cannot be used to explain the differences.

5. On the whole, zeolitic assemblages come from areas where the CO_2 fugacity, if known, is about an order of magnitude below the values in the nonzeolitic rocks (see also Thompson 1971b).

6. There seems to be an over-all correlation between the occurrence of zeolitic assemblages and the constitution of the protolith; volcanic/volcanogenic rocks tend to be zeolitic. This is

true of many areas of burial metamorphism, as well as of active geyser and hot spring regions, which of course tend to occur in volcanic fields. There are obviously exceptions, and zeolitic assemblages do occur in the matrix or as cement in rocks having no obvious volcanic component (Deffeyes 1959). However, the association of zeolite rocks and volcanic rocks seems too recurrent to be dismissed as coincidental.

What might be the cause of the association? I know of no obvious answer. Previously, Zen & Thompson (1974) suggested that natural glasses, with the framework structure involving oxygen-bridged tetrahedra of silicon and aluminum, might allow ready nucleation of zeolitic minerals, which also possess such Si-Al tetrahedra framework; growth of calcic zeolites as alteration products of plagioclase might be similarly rationalized. However, not all zeolites form from glass or plagioclase, nor do all natural glasses form zeolite upon devitrification. Our knowledge of the relative stabilities of alternative as-

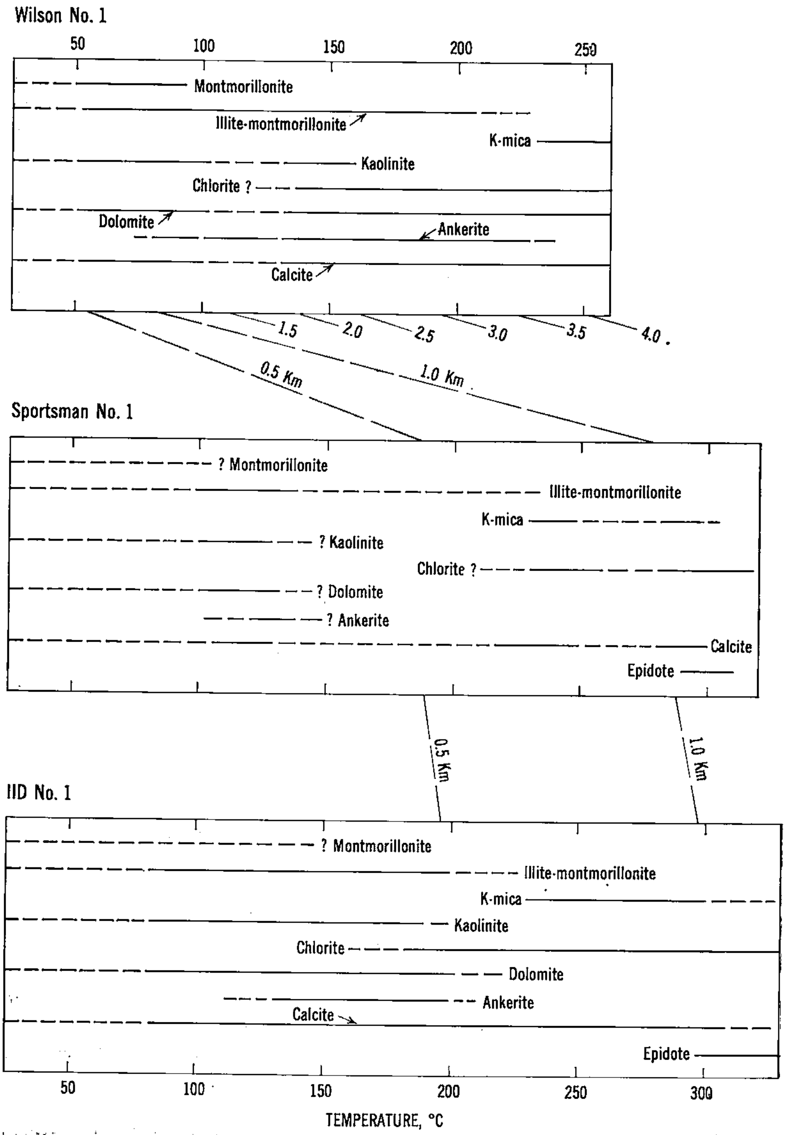

FIG. 1. Mineral assemblages of three drill cores from Salton Sea geothermal area, California. From Muffler & White (1969). Dashed lines indicate interpolation and uncertain identifications. The "correlation lines" connecting the three cores are lines of constant depth.

semblages for given external conditions, or of the kinetics of nucleation and crystal growth, or their control by the nature of the rocks such as previous thermal history of the glass, composition of the glass, permeability of the glass, *etc.*, is so skimpy that a great deal of basic research is needed before anything definite can be said.

The permeability of rocks as control of metamorphic assemblage must be significant. Levi (1970) made the same point. Certainly this would be true if the assemblages depend on ionic exchange equilibria (Fisher 1970; Eugster 1970; Surdam 1973). Many volcanic and volcanogenic rocks are massive even though the associated pelites may be shaly or even slaty because these rocks yield differently under stress. Impeded flow of groundwater through massive volcanic rocks may provide a setting for devitrification approximating reaction in a closed system. High alkali/H ratio and high silica activity might be expected as a result; these conditions are favorable to the formation of zeolites. On the other hand, these chemical conditions are not commonly expected in clastic sediments.

Surdam (1973) reported on some preliminary experimental results of devitrifying natural basaltic glass in simulated sea water. He found that with suitable manipulation of the experimental conditions, he could obtain a sequence of reaction events involving dissolution of the glass, formation of an alumino-silicate gel (presumably retaining the tetrahedral framework) and crystallization of zeolite from the gel. These results are consistent with the suggestion given above, even though the events Surdam observed may not be universally applicable to natural devitrification. It would also be interesting to measure the relative diffusion coefficients of H_2O and CO_2 through and along cracks in natural glasses of various compositions, for if massive glasses, in which material transport must largely depend on molecular diffusion to the sites of reaction, show membrane effect, then a strong case can be made for control of the metamorphic assemblages by the physical state of the rock.

Some Problematic Areas of "Burial Metamorphism"

Three areas of proposed burial metamorphism present special problems that need to be considered.

First is the thick sequence of "burial metamorphic" rocks of the Chilean Andes and Coastal Range, reported by Levi (1969, 1970) and Thompson (1971a). The rocks are volcanic and volcanogenic, and according to Levi show little evidence of internal deformation. The mineral assemblages show increase in grade from top down, from laumontite-bearing (and analcime-bearing ? Levi 1969, p. 40) to prehnite-pumpellyite-bearing, to albite-epidote-actinolite-chlorite-bearing. However, the pattern is repeated several times, each repeat series of increasing metamoprhic grade being interrupted by unconformities above and below. Levi (1969, 1970) interpreted the repeated series as caused by episodic metamorphism, the episodes being defined by the unconformities. Each series of burial metamorphic assemblages is self-sealed so that the assemblages are not affected by subsequent metamorphic events. Zen & Thompson (1974) pointed out that the calcite and zeolite cements may not be effective seals, so that the applicability of the hypothesis needs demonstration[1]. Moreover, for rocks immediately above the Early Cretaceous Patagua unconformity, the stratigraphic section is as much as 18 km thick (Levi 1970), corresponding to a pressure of about 5 kb if the entire section was at one time complete and flat-lying. Under such circumstances, laumontite, which is the calcic zeolite reported here, would not be stable; instead, it should decompose to form lawsonite + quartz + H_2O according to the experimental data of Liou (1971a). Since Liou's data are for the condition of $P_{total} = P_{H_2O}$, the hypothesis of self-sealing cannot be used to explain the nonoccurrence of this reaction.

Inasmuch as obviously the 15-28 km of stratigraphic section (Levi 1970) does not now rest in a complete flat-lying sequence, the relative timing of metamorphism and deformation needs documentation. Interpretation of the geologic relations in the framework of "burial metamorphism" with its implied static conditions and following a simple regional gradient seems a topic that bears further investigation.

The second area is that of Silurian and Devonian volcanic rocks of northern New Brunswick, described by Mossman & Bachinski (1972). The rocks have analcime-bearing assemblages. They show little schistosity and, because of this fact and because of the nature of the mineral assemblages, Mossman & Bachinski suggested a burial mode for the rock

[1]Thompson (1971a) suggested that the "unconformities" may actually be large-scale low-angle thrust faults. If so, the interpretation of the genesis of the mineral assemblages in these rocks will have to be drastically modified.

metamorphism. It seems rather strange that these rocks, which are older than the Acadian orogeny and which lie in the core part of the Appalachian foldbelt, should have been affected only to the extent of simple burial metamorphism. An alternatve explanation that does not preclude some deformation, and thus better conforms to the geologic framework, would be more acceptable.

The third problematic group of rocks is the Lower Tertiary Taveyannaz greywacke of Switzerland and France. The rocks are volcanogenic and contain laumontite-bearing and prehnite-pumpellyite-bearing assemblages. These rocks, studied by Martini (1968, 1972) and Martini & Vuagnat (1965, 1970), constitute part of the autochthonous cover rock of the external massifs as well as part of the subhelvetian and Helvetian nappes. The massive greywackes show no cleavage whereas the associated pelites show weak cleavage (D. S. Coombs 1973, written communication). According to Martini & Vuagnat (1970), the burial metamorphism took place during the late Tertiary neoalpine metamorphic event (beginning in late Oligocene) and was associated with the overriding of the prealpine nappe piles. The thickness of the nappe piles was estimated (Martini & Vuagnat 1970) to be at least 5-6 km but probably no more than 10 km. The temperature of metamorphism was estimated to be around 200°C. The implied thermal gradient was a typical continental gradient.

It is puzzling that rocks of the Helvetian and associated terranes, be they autochthonous or allochthonous, escaped deformation during the metamorphic event, despite the lack of penetrative schistosity. As is for the rocks of the Appalachian foldbelt in New Brunswick, geologic data suggest otherwise, and alternative explanations seem desirable. The consequences of the idea that the rocks were metamorphosed during loading by a pile of nappes 6 or more km thick can be crudely considered, by a rough estimate of the thermal relaxation time after loading of the nappes. I took the surface temperature to be 0°C, an initial geothermal gradient of 28°C/km, uniform down to an initial compensation depth of 32 km, where the temperature is held constant at 900°C. A 8-km thick allochthon was rapidly (instantaneous relative to the associated process) emplaced, carrying with it the same geothermal gradient and the boundary condition $T\ (z=0) = 0$. Along the thrust surface, the temperature must be initially at least as high as the average temperature of the two piles, or just over 100°C. A new thermal steady state leads to a gradient $\Delta T/\Delta z = 900/40 = 22.5$°C/km (Fig. 2). To estimate the time required to establish this new steady

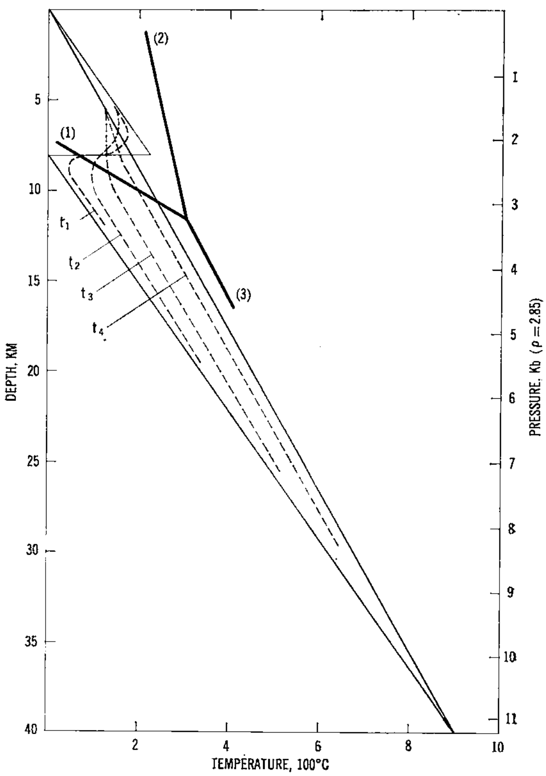

FIG. 2. Schematic representation of the temperature profile after an instantaneous emplacement of an 8-km thick nappe pile. Fine solid lines, initial temperature distribution; fine long-dashed line, the final steady state where the boundary temperatures are fixed both above and below as specified in the text. Fine dotted lines, general appearance of temperature profiles at intermediate times, marked t_1, t_2, t_3, t_4. Heavy solid lines are for the univariant phase equilibria, (1) laumontite = lawsonite + 2 quartz + 2 H_2O; (2) laumontite = wairakite + 2 H_2O; (3) wairakite = lawsonite + 2 quartz, according to Liou (1971a).

state, I took a flux of 1 HFU, a thermal conductivity of 3×10^{-3} cal/sec. cm. k, Cp of 0.24 cal/g, and density of 2.85 g/cm^3. It would take about 10 million years for the thermal profile to return to a linear gradient. This value does not take into account any convective heat loss along the thrust plane, or frictional generation of heat. Oxburgh & Turcotte (1974) showed that the heat generated by friction is probably a short-term factor in the thermal evolution of the thrust mass; in any event, the effect should not extend beyond a few meters of the thrust surface.

One could also use a lower boundary condition of constant heat flux instead of constant temperature, so that at the new steady state the original thermal gradient is reestablished. This would roughly increase the relaxation time to a little more than twice the first value. This set of boundary conditions is in fact that used by Oxburgh & Turcotte (1974), who worked on a very similar problem, for which the temperature profile as a function of time after initial thrusting is explicitly given. My estimate of relaxation time is within about 10% of their value when due adjustment is made for different assumptions on the thickness of thrust sheet, initial gradient, *etc.*

Interest in this exercise lies in that, for a load of 8 km and density of 2.85 g/cm^3, during the initial part of the heating-up cycle the rocks at the top of the "autochthon", which includes much of the Taveyannaz greywacke, lie in the stability field of lawsonite rather than of laumontite, according to the data of Liou (1971a; see Fig. 2), which applies to the condition of $P_{total} = P_{H_2O}$, or maximum stability of laumontite. Pending detailed solution of the transient heat flow problem, and pending availability of data on the rates of nucleation of minerals such as lawsonite in the appropriate host material, we cannot say whether there would have been time for laumontite to break down and for lawsonite to nucleate and grow in the Taveyannaz greywacke according to the reaction: laumontite = lawsonite + 2 quartz + 2 H_2O. However, near the thrust surface, the time of residence of the rocks in the lawsonite field probably would be hundreds of thousands, if not a few million years, which seems enough time for lawsonite to form if it was stable. It would appear that the 6 km of "minimum load" may also be close to the maximum load, or else the formation of the zeolitic assemblage might not have been causally related to the emplacement of the nappe piles. Anything more than 6 km of load would obviously compound the problem; thermal gradients larger than about 30°/km would, for comparable load, bring the rocks into the field of stability of wairakite according to Liou's (1971a) data, and wairakite is not known in these rocks, even as relicts.

The transition of laumontite to lawsonite involves a decrease in the volume of the solid phases; ΔV_s = 1.35 cal/bar per mole of laumontite. Isothermally, therefore, an independent variation in the value of μ_{H_2O} must be compensated for by a change in the total rock pressure, if equilibrium is to be maintained for the reaction. The relation between the variables is

$$\Delta V_s \Delta P_{total}(T') = 2\Delta\mu_{H_2O}(T', P)$$

where T' is a fixed temperature. Suppose the chemical potenital of H_2O is lower than that for a full l oad of rocks because of fracture in the rocks, or because of communication with a permeable bed that controls the hydrostatic head, or any other means. The equilibrium boundary for the dehydration of laumontite would be shifted to lower pressures by an amount given by the above equation (see also Coombs *et al.* 1959). As an example, let the new value of μ_{H_2O} correspond to 1.5 kb of H_2O. We readily calculate, using the data in Fisher & Zen (1971), that the equilibrium rock pressure would be shifted by as much as 1 kb at 200°C, to the position indicated in Figure 3 (the new univariant curve must meet the curve for $P_{total} = P_{H_2O}$ at 1.5 kb, of course, and meet it discontinuously). Similar consideration leads to the shifted position for the reaction laumontite = wairakite + 2H_2O; the curve wairakite = lawsonite + 2 quartz is independent of H_2O and so the new pseudo-invariant point lies on the metastable prolongation of the univariant curve for the last-cited reaction.

From these calculations it would appear that the stability boundary of laumontite (and of wairakite) is markedly shifted to lower pressures if during metamorphism the system was permeable, so the rock pressure exceeded water pressure, or if the fluid phase contained appreciable amounts of other components. Likewise, if in a complex chemical system laumontite reacts to form some other mineral assemblage upon increase of pressure (*e.g.*, prehnite or pumpellyite-bearing assemblages), rather than to form lawsonite directly, then the stability field of both laumontite and of lawsonite would be reduced. Thus the actual field of stability of laumontite very probably is considerably more restricted than indicated by the experimental data of Liou (1971a) and is in the region of lower pressure. This conclusion clearly has geologic implications on the validity of associating laumontite-bearing reactions to specific geologic processes, as is done for the Taveyannaz greywacke and for the Chilean rocks. It also affects our notion of the environment of "burial metamorphism".

"Burial Metamorphism": An Assessment

What was just said regarding the stability field of laumontite in chemically complex systems, or in systems where the H_2O fugacity is less than that of pure H_2O at lithostatic pressure (either by dilution or by external control of

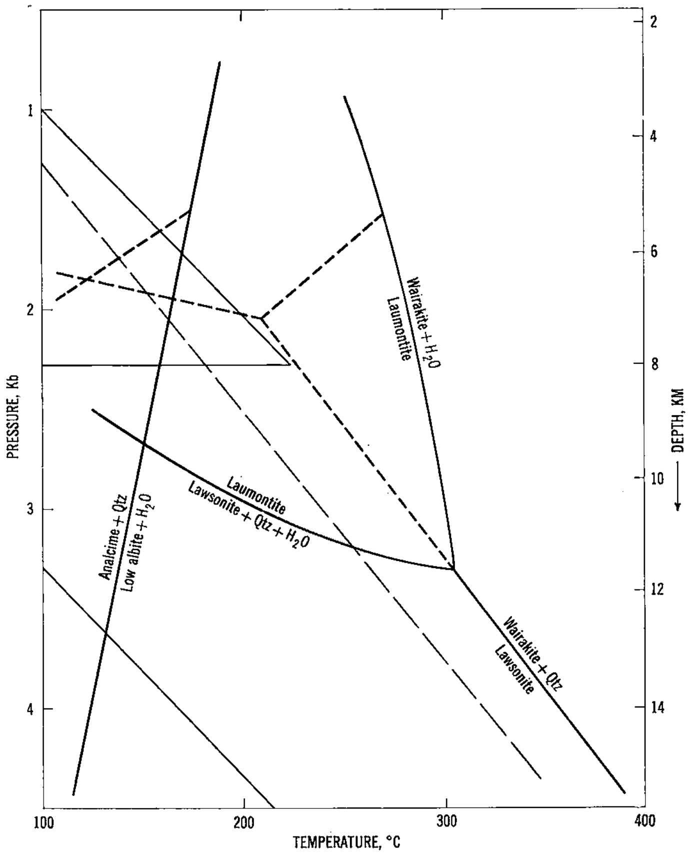

FIG. 3. Details of a part of Figure 2, showing the calculated effect on the phase equilibria if the fugacity of H_2O corresponds to $P_{H2O} = 1.5$ kb either by chemical dilution or by a head less than lithostatic. The initial and final temperature profiles of Figure 2 are sketched in as light lines.

the hydrostatic head), can be said for analcime stability as well, on the basis of the experimental work of Liou (1971b), Thompson (1971c) and others; the effects are depicted in Figure 3. It seems reasonable to suggest, therefore, that the zeolitic "burial metamorphic" mineral assemblages indicate burial to depths considerably shallower than is permitted by a simple reading of the equilibrium phase diagrams experimentally derived under the condition of $P_{\text{total}} = P_{H_2O}$.

Under such low confining pressures (at most 1-2 kb), and at low temperatures (200-300°C), basaltic volcanic rocks under stress will fail by brittle fracture rather than by ductile yielding (see Griggs *et al.* 1960; Handin 1966; Robertson 1972). One might presume that the same may be said of the more siliceous volcanic rocks such as andesitic, dacitic or rhyolitic flows which so characteristically host zeolitic assemblages. Development of penetrative schistosity, which is part of the criterion for metamorphism in a non-burial environment, can hardly be expected for the deformation and concomitant metamorphism of massive volcanic rocks under these conditions. Perhaps the same consideration can be applied also to at least some of the volcanogenic sandstones and greywackes. I do not mean by this to say that all zeolitic rocks heretofore attributed to burial metamorphism have merely been deformed in a brittle regime; however, I do suggest that the presence or absence of penetrative schistosity may be controlled by the physics of the rocks rather than by the tectonic environment, and therefore it may not be valid to use the lack of schistosity to deduce geologic history, as study of rock metamorphism must necessarily lead up to. Pillow lavas bearing the prehnite-pumpellyite assemblage in eastern New York (Zen 1974, fig. 5) are entirely massive and show no effect of deformation; the interbedded greywackes are likewise undeformed, but the interbedded pelites are good slates.

There are undoubtedly areas where rocks were metamorphosed under simple burial and at *P, T* values appropriate to the regional geothermal gradient. The Belt rocks of Montana and Idaho and the diagenetically altered rocks of the Gulf Coast are examples. There certainly are zeolitically metamorphosed rocks which fit in the same category. However, each situation must be judged on its own merit, taking into account the geologic knowledge of the area as a whole, rather than depending heavily on the presence of rock schistosity. The by-now standard plot of zeolitic mineral paragenesis versus depth of reconstructed burial (Fig. 4) clearly shows, moreover, that burial as such is not the determining factor, a point made long ago by Packham & Crook (1960). The term "burial metamorphism" presupposes much geologic information, so application of the term can mislead. There is no sharp and readily identifiable boundary between burial metamorphism and regional metamorphism, as Coombs (1961) pointed out; rock texture and the mineral assemblages do not necessarily supply the clue.

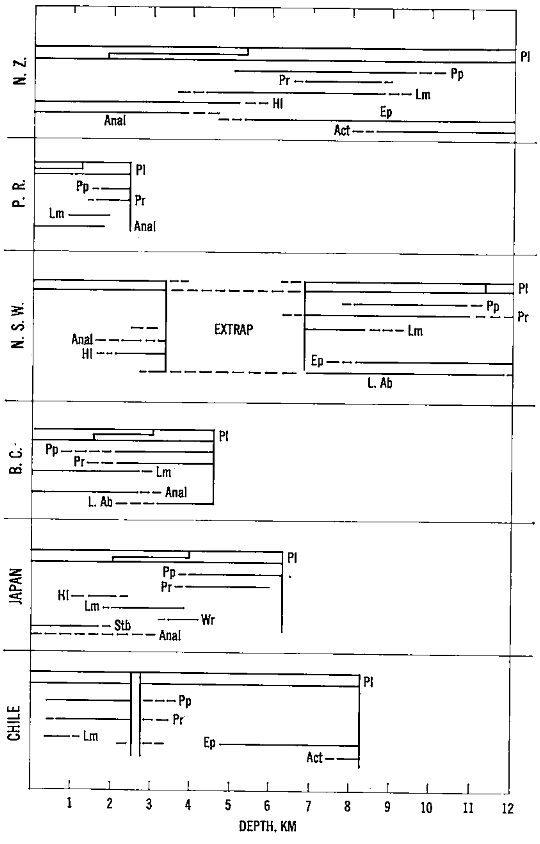

FIG. 4. Comparison of mineralogy of various "burial metamorphic" terranes having zeolitic assemblages. The secondary role of total pressure in determining the assemblages is clear. Sources of data: New Zealand, Coombs 1954; Puerto Rico, Otalora 1964; New South Wales, Packham & Crook 1960; British Columbia, Surdam 1973; Japan, Seki *et al.* 1969a; Chile, Levi 1969, 1970; Thompson 1971a. Based in part on compilation by Thompson 1971a.

Finally, why do we not find zeolitic assemblages more often in old foldbelts, such as the Appalachians and the Caledonides, whereas they seem common in young orogenic belts? One possible answer is simply that zeolitic assemblages form only at very shallow depths and are unlikely to be preserved in deeply eroded, geologically old foldbelts except by accident (as probably happened in northern New Brunswick). One's best hope of finding these assemblages in an old foldbelt might be at the margins, where deep burial by tectonic causes and subsequent excessive erosion might not have occurred. However, at least for some of the ancient foldbelts such as the Caledonides, the Appalachians and the ancient Rockies, the margins in the direction of the craton — where they are best preserved — are underlain by platformal sequences of carbonate rocks and quartzites, typically devoid of volcanic rocks except a few bentonites. These rocks are not good hosts for zeolitic assemblages.

There is, of course, the possibility that the marginal parts of some ancient foldbelts do have zeolitic assemblages preserved despite the ravages of erosion, but simply have not been discovered. As recently as a few years ago, we thought that we at least knew the general nature of metamorphism in the northern Appalachians, even though not all the information has been gathered. The discovery of prehnite-pumpellyite-bearing rocks (Coombs *et al.* 1970; Zen 1974; Richter & Roy and Papezik, this issue) and of zeolitic rocks (Mossman & Bachinski 1972) show how wrong we were. We would be imprudent to suppose that no further surprise of this kind remains.

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