Pressure—Temperature—Time Paths of Regional Metamorphism II. Their Inference and Interpretation using Mineral Assemblages in Metamorphic Rocks

Pressure–Temperature–Time Paths of Regional Metamorphism II. Their Inference and Interpretation using Mineral Assemblages in Metamorphic Rocks

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ABSTRACT

A companion paper (England & Thompson, 1984*a*) investigates the pressure–temperature–time (*PTt*) paths followed by rocks undergoing burial metamorphism in continental thickening events. This paper discusses problems involved in inferring such paths from the petrological data available in metamorphic rocks and—once such paths are determined—how they may be interpreted in terms of the thermal budgets of metamorphism. Each of the principal facies series (glaucophane–jadeite, andalusite–sillimanite and kyanite–sillimanite) may be encountered by rocks involved in the thickening and erosion of continental crust in a regime of average continental heat flow. The inference of a minimum thermal budget required for a given metamorphism depends strongly on a knowledge of the *PTt* paths followed by rocks during the metamorphism. Discrimination between possible thermal regimes is greatly enhanced if portions of *PTt* paths, rather than single *PT* points, are available, and additional constraint is possible if these paths are supplemented by geochronological, structural and heat flow data.

I. INTRODUCTION

One of the standard inverse problems of metamorphic petrology is to infer the thermal regime operating during a metamorphic episode from the mineral assemblages and compositions in rocks that reach the surface after the episode is finished. In fact, orogenic processes are sufficiently complicated, and petrological data sufficiently limited that no formal inverse is ever attempted; interpretation usually depends on the calculation of pressure–temperature–time (*PTt*) paths for model tectonic settings and on using these with metamorphic petrological data to give estimates of thermal regimes operating during metamorphism.

The previous paper (England & Thompson, 1984*a*, hereafter called Part I) outlined some of the model pressure–temperature–time (*PTt*) paths that would be followed by rocks under a variety of conditions resulting from crustal thickening, and addressed the question of what class of information is recoverable from *PT* data obtained from metamorphic rocks. A major conclusion of Part I is that we expect the overall thermal development of a regional metamorphic terrain to depend on relatively few large scale parameters and, consequently, that if it is possible to extract information from mineral assemblages about regional *PTt* paths, we should be able to make estimates of the magnitudes of the parameters controlling the regional metamorphic budget—notably the total heat supply to the orogenic belt, the lengthscales of thickening, the length and time scales of erosion and the extent of magmatic involvement in the metamorphic episode. Smaller scale phenomena (intrusions, transposition of isograds by folding or faulting, etc.) will affect observations made in the field, but cannot be uniquely interpreted in terms of larger scale phenomena.

In this paper we discuss the problems of relating petrological *PT* points or, more usefully, portions of *PTt* paths to regional thermal regimes. In principle such data, particularly in conjunction with heat flow, isotopic and structural studies, can provide valuable information on the tectonic setting and thermal budget of metamorphism. Unfortunately, it is often not possible to gather all the information necessary for a unique interpretation of a given terrain; nonetheless *PTt* data may be used to place constraints on metamorphic processes, and we hope that the considerations presented here will help in the interpretation of such data.

2. *PT* POINTS AND *PTt* PATHS

Many of the inferences of conditions during the evolution of orogenic belts that have been made using *PT* data from metamorphic rocks are based on relatively simple conceptual models of the metamorphic environment and, while these models serve a useful purpose, it is important to recognize their limitations as well as their merits. In this section we consider how these conceptual models need modification in view of the nature of the *PTt* paths that are followed by rocks undergoing regional metamorphism. For consistency, most of the paths discussed here are taken from section 5 of Part I, but many of the arguments may readily be applied (with the appropriate adjustments) to other forms of *PTt* paths.

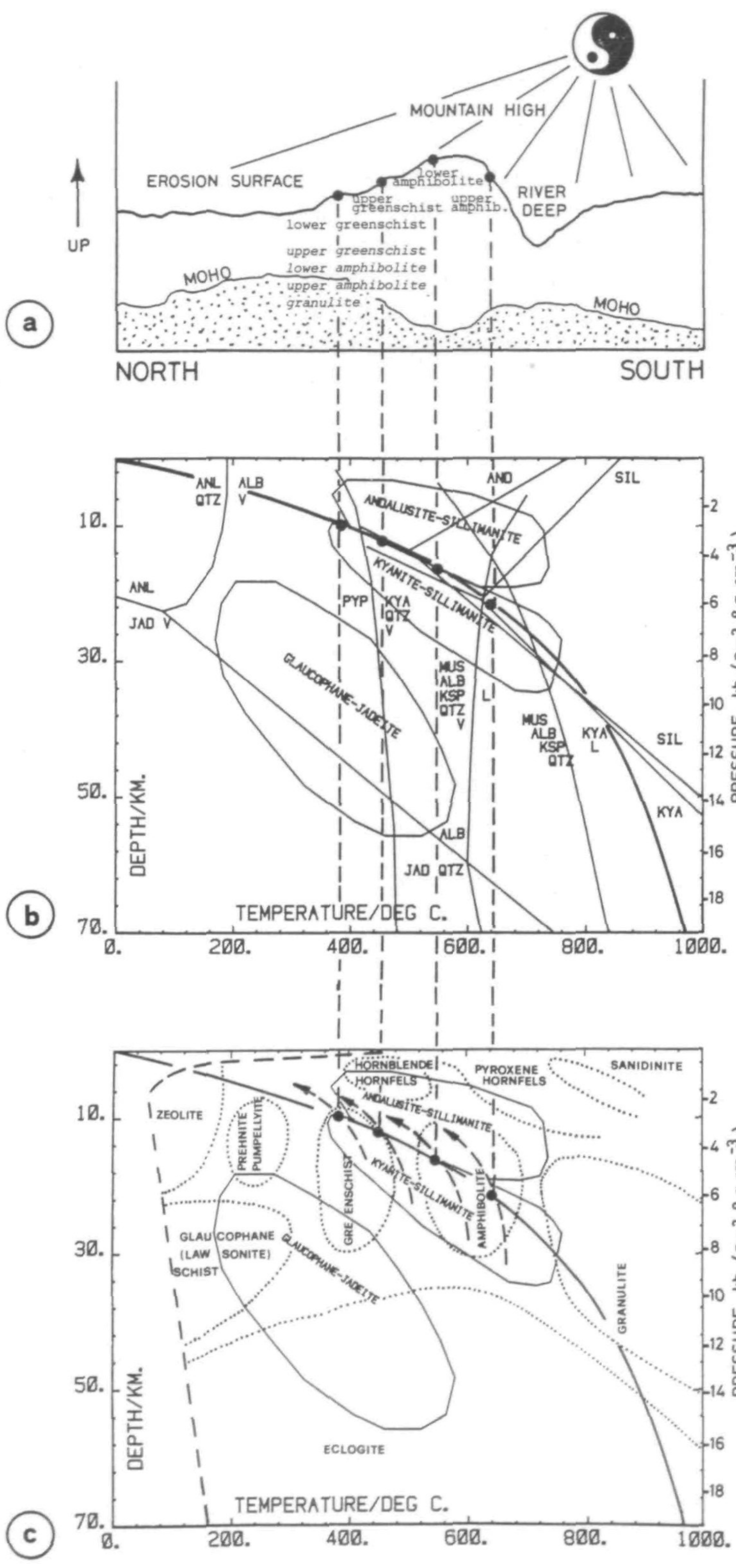
The first set of problems encountered in inferring thermal and tectonic regimes from metamorphic data are concerned with deciding where on a *PTt* path the points that make up a *PT* array are recorded, and the rest of this section is largely concerned with them. We consider how the kinds of *PTt* paths calculated in Part I influence *PT* data recorded in mineral assemblages on the surface, through their effects on reaction kinetics, chemical diffusion and through the control exerted by fluids on the mineral assemblages preserved. In particular, we emphasize the means whereby information may be gained on parts of *PT* paths, rather than on discrete *PT* points.

2.1. Geotherms, facies series and PT paths

The material of these papers concerns the perturbing and relaxation of temperatures in continental lithosphere, and the *PT* conditions experienced by rocks moving relative to these evolving geotherms. As the word geotherm is used in at least two distinct senses in the petrological literature it is worth including definitions here to avoid possible ambiguities.

A geotherm is simply the relation between temperature and depth within the earth at any one time; this relation may be a steady state one or a transient one, but in either case its

FIG. 1. Metamorphic mineral assemblages at the erosion surface in (*a*) are assigned through experimentally calibrated mineral reactions to metamorphic facies series (*b*) and facies (*c*). The equilibrium data in (*b*) are shown as examples only; the sources are given by Thompson & Thompson (1976, fig. 13) and by Thompson & Tracy (1979, fig. 2) and transferred to this figure using a crustal density of 2·8 Mg m^{-3}. The equilibria include two versions of the relative stability of Al_2SiO_5 polymorphs (from Holdaway, 1971 and Richardson *et al.*, 1969; where andalusite = AND; kyanite = KYA; sillimanite = SIL), and the breakdown of albite (ALB) to jadeite (JAD) *plus* quartz (QTZ). The dehydration reactions producing almost pure H_2O (vapor = V) are experimentally investigated in the presence of excess water ($a_{H_2O} \approx 1$). Other abbreviations: muscovite (MUS), analcime (ANL), pyrophyllite (PYP), potash-feldspar (KSP). Mineral assemblages assigned to mineral facies (*c*) at the erosion surface in (*a*) are shown with normal type to distinguish them from facies representing present day conditions in the underlying crust, shown by italic type. The *P–T* locations of metamorphic facies (*c*) and facies series (*b*) are based upon correlation with experimental studies of many more mineral equilibria than shown here (for example see Fyfe *et al.*, 1978, fig. 6.8). Because of the transitional nature of mineral assemblages in the field and the displacement of calibrated reactions through crystalline solution in natural minerals, the *P–T* boundaries of facies and facies series are approximate and gradational. One common practice is the assigning of mineral assemblages at the erosion surface (solid points in (*a*)) to a steady-state geotherm, shown by the heavy curves in (*b*) and (*c*). The dashed lines crossing the geotherm in (*c*) represent schematic *PTt* paths experienced by rocks during exhumation (see Part I).


MOUNTAIN HIGH
EROSION SURFACE
RIVER DEEP
UP
lower amphibolite
upper greenschist
upper amphib.
lower greenschist
upper greenschist
lower amphibolite
upper amphibolite
granulite
MOHO
MOHO
a
NORTH
SOUTH
ANL QTZ
ALB V
AND
SIL
ANDALUSITE-SILLIMANITE
KYANITE-SILLIMANITE
ANL
JAD V
PYP
KYA QTZ V
GLAUCOPHANE-JADEITE
MUS ALB KSP QTZ V
L
MUS ALB KSP QTZ
KYA L
SIL
KYA
ALB
JAD QTZ
DEPTH/KM.
PRESSURE, kb (p=2.8 g cm^{-3})
TEMPERATURE/DEG C.
b
HORNBLENDE HORNFELS
PYROXENE HORNFELS
SANIDINITE
ZEOLITE
PREHNITE PUMPELLYITE
GREENSCHIST
AMPHIBOLITE
GLAUCOPHANE (LAWSONITE) SCHIST
GRANULITE
ECLOGITE
c

reference level is the earth's surface. A transient geotherm may be the result of transient heat sources, of motions due to erosion, overthrusting, rifting etc., or of thermal relaxation from a previously perturbed regime; all three of these cases occur during the evolution of the systems discussed in this paper.

Individual *PT* points obtained from mineral assemblages are often interpreted in terms of geothermal gradients (Fig. 1); it should be clear from Part I that while an individual *PT* point may have been set on a geotherm, its relation to the heat source distribution giving rise to that geotherm is usually unclear. The problem becomes worse if an array of *PT* points from one terrain is to be interpreted; such arrays have been referred to, misleadingly, as 'metamorphic geotherms' (e.g. England & Richardson, 1977). In fact these arrays bear no relation to any one geotherm that existed during metamorphism, because they consist of a set of *PT* conditions inferred from assemblages that were preserved at different times and different places during metamorphism (e.g. England & Richardson, 1977, fig. 1; section 3.1 below). We prefer the designation '*piezothermic array*' (Richardson & England, 1979) for the set of peak metamorphic conditions experienced in a thickened pile, or '*PT array*' for a suite of metamorphic conditions preserved in a metamorphic terrain.

Certain rock types and minerals are often regarded as characteristic of distinct pressure (*P*) and temperature (*T*) regimes, and the indications these give of changing metamorphic grade on a regional scale lie behind the facies series concept of Miyashiro (1961). The three principal facies series, glaucophane–jadeite (Gla–Jad), kyanite–sillimanite (Kya–Sil) and andalusite–sillimanite (And–Sil) are taken to be representative of, respectively, high *P*– low *T*, intermediate *T* and *P*, and high *T*–low *P* metamorphism. In addition, several workers (e.g. Heitanen, 1967) specify intermediate facies series (see Fig. 1).

In principle this approach acknowledges only that the rocks involved have passed through some region of *PT* space that is characterized by the stability fields of the diagnostic minerals, although the facies series concept, in its application, is often used to infer a tectonic setting for the metamorphism. We shall refer in this paper to facies, but we shall do so only in defining areas of *PT* space that rocks may *pass through* on the *PTt* paths of regional metamorphism.

Figs. 1 and 2 indicate the approximate extent in *PT* space of the facies that are used in this paper, and show some of the many mineral reactions that are used to define them. Also shown are the reaction lines for the onset of melting in wet and dry pelites and amphibolites. Estimates of metamorphic conditions more precise than these can be made for individual rock samples using laboratory calibrations of mineral equilibria and of mineral compositions. Estimates of precision range from an optimistic ± 5 °C and ± 200 bars to a realistic ± 50 °C and ± 2 kb. We do not discuss the various current methods of mineral barometry and thermometry, as they have been extensively reviewed elsewhere (e.g. Essene, 1982), but we certainly expect precision of and consistency among barometers and thermometers to improve as experimental techniques evolve.

Fig. 2 also contains the results of a geobarometric and geothermometric study of Perkins & Newton (1981), using the assemblage Opx + Cpx + Plg + Gar + Qtz from a worldwide selection of mafic granulites, and the results of Tracy *et al.* (1976) from a study of amphibolite facies metapelites in central Massachusetts. In each case the results of two barometric estimates are connected by vertical bars, and indicate a pressure range of around 2 kb, while temperature uncertainties are unknown, but may be as large as 50 °C. These are shown, first to illustrate the relation of these rocks to the Kya–Sil facies series and the granulite facies chosen in Fig. 1, and secondly to emphasize that the major problem associated with the use of such *PT* data is not just to increase the precision of the measurements, but to know when on a *PTt* path an individual *PT* point is established.

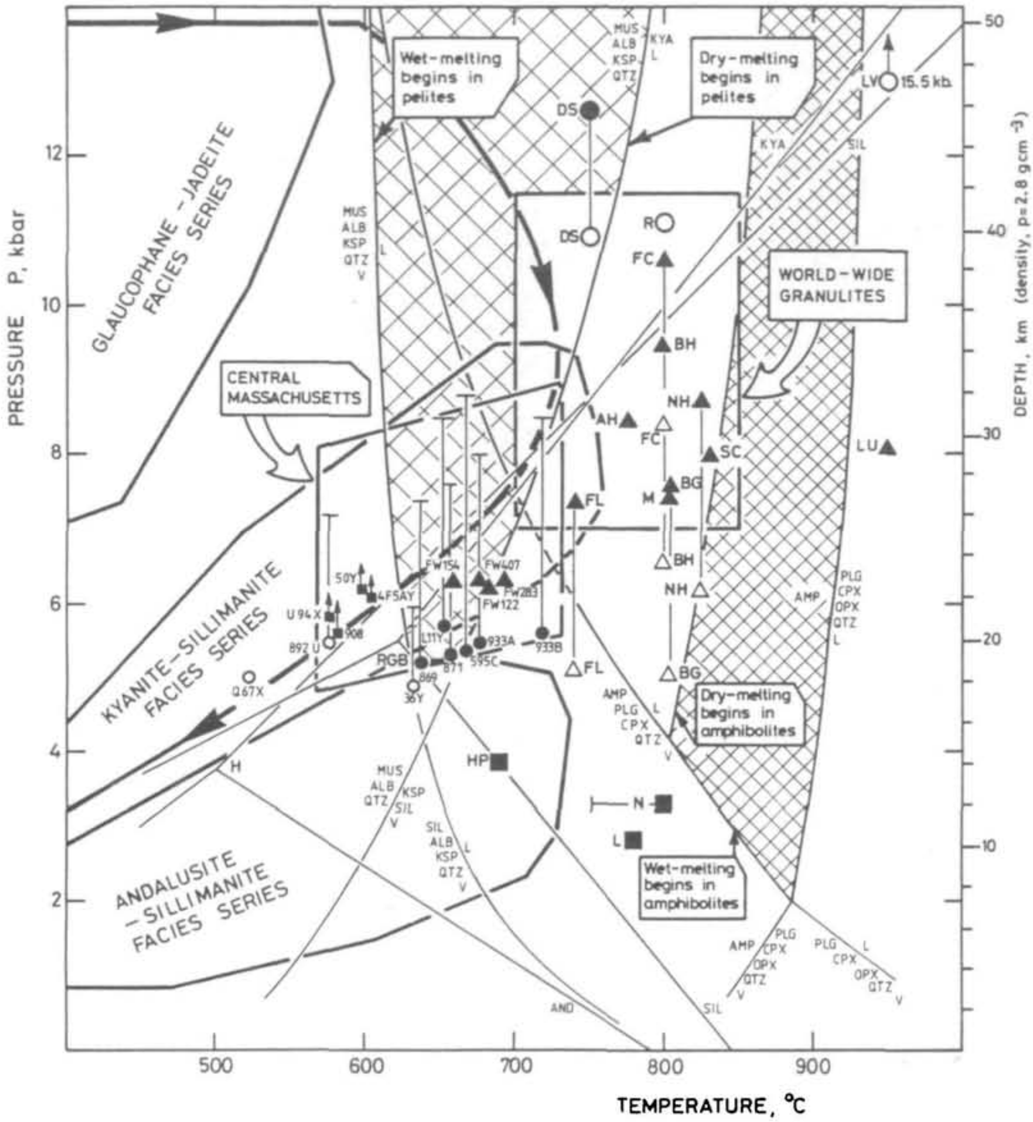


FIG. 2. P–T diagram summarizing pressure and temperatures of metamorphic equilibration for high-grade metapelitic rocks from Central Massachusetts (Tracy *et al.*, 1976, fig. 7) and for various granulites from a worldwide selection (lettered symbols from Perkins & Newton, 1981, fig. 1). Here, the closed and open symbols refer to the respective use of an orthopyroxene or clinopyroxene barometer in an assemblage with plagioclase, garnet and quartz. Two versions of the Al_2SiO_5 phase relations are shown: RGB (Richardson *et al.*, 1969), and H (Holdaway, 1971), together with the facies series locations from Fig. 1*b*. Initial melting of pelite (wet and dry) is taken from Thompson & Tracy (1979, fig. 2). The cross-shaded areas locate the regions where partial melting of pelites and amphibolites may begin. Initial melting of amphibolite (wet and dry) is modelled on the results of Yoder & Tilley (1962, fig. 27). A P–T interval ~100 °C wide is shown as the actual reactions are not well known. The bold line is the *PTt* path for a rock buried at 50 km after crustal shortening, taken from Fig. 3*e* of Part I.

As is discussed in Part I, we expect deeply buried rocks to experience decompression of several kilobars, at nearly constant temperature, early in their exhumation. An example of such a *PT* path (taken from Fig. 3*e* of Part I) is shown in Fig. 2. In interpreting data such as these, then, it is important to determine whether, for example, the pressure ranges shown in

Fig. 2 represent imprecision of the geobarometric technique or represent the results of two precise barometers that were set at different points on *PTt* paths. This problem also extends to the use of mineral assemblages established by discontinuous reactions to define *PT* histories.

2.2. *Simple mineral equilibria and PT points*

There is general agreement among experimentalists as to the relative position of the stability fields of, for example, albite versus jadeite + quartz, or of the Al_2SiO_5 polymorphs, but variance among the absolute positions that are determined emphasizes the influence of experimental procedures, crystalline solutions, cation disorder, kinetic factors and other effects (e.g. Zen, 1969). We assume that further studies will reconcile the experimental differences and consider here only the application of such data to the interpretation of natural mineral occurrences.

At first glance, the *PT* ranges appropriate to the facies series seem to be defined by such observations as the type of Al_2SiO_5 polymorph present in the rock, or the presence of jadeite + quartz rather than albite. However, as the presence of such minerals outside their stability fields is what allows us to infer metamorphic conditions, we clearly must consider kinetic effects during the passage of rocks along a *PTt* path.

For many metamorphic reactions it may be reasonable to assume that rates of change of temperature and pressure are so slow that the reactions remain close to equilibrium. Where this is not a good assumption, however, the *PT* conditions recorded by mineral assemblages will lag behind the actual conditions experienced by the rock along its *PTt* path. This would result in what has been termed 'apparent *PTt* paths' (P. Koons, personal communication) or 'metamorphic jet-lag' (J. Dixon, personal communication). Because, in general, we expect rocks to spend a considerable fraction of their *PT* history within a few tens of degrees of their maximum temperature (Part I, section 5; section 2.3, Fig. 5, this paper) this effect is likely to be least important near the peak of metamorphism, but as the example below shows, kinetic factors may cause appreciable complications.

Fig. 3 shows an Al_2SiO_5 triple point that lies between the experimentally determined triple points of Richardson *et al.* (1969) and Holdaway (1971); no particular significance ought to be attached to this point, as it is chosen for clarity only. Two *PTt* paths calculated in Part I are shown in this figure, as are schematic rates (slow to fast) of transformation between the polymorphs; these are purely illustrative, as data on the kinetics of these transformations are not available.

The shallower uplift path (FGH) passes exactly through the triple point and the deeper path passes well to the high-*T* side of it. Immediately after thickening, both samples are in the kyanite stability field; the shallower sample passes through the triple point at G and enters the stability field of andalusite. At the triple point no reaction will occur in the absence of chemical potential gradients and, depending on the kinetics, there may be anywhere between total and negligible transformation of kyanite to andalusite on the portion GH of the path. Thus this path would result in a rock containing at most kyanite and andalusite, and possibly just andalusite, even though it passed through the triple point.

The second sample passes into the stability field of sillimanite at B; the rate of conversion of kyanite to sillimanite would increase along BC and then decrease along CD. Similarly, it passes into the andalusite stability field at D and conversion of sillimanite to andalusite occurs at an increasing rate along DE and at a decreasing rate thereafter. In each case, again depending on the kinetics, the transformations between polymorphs may be total or partial. If neither transformation were to go to completion the rock would contain the sequence Kya–Sil–And, despite the fact that it did not pass through the triple point. We also note that

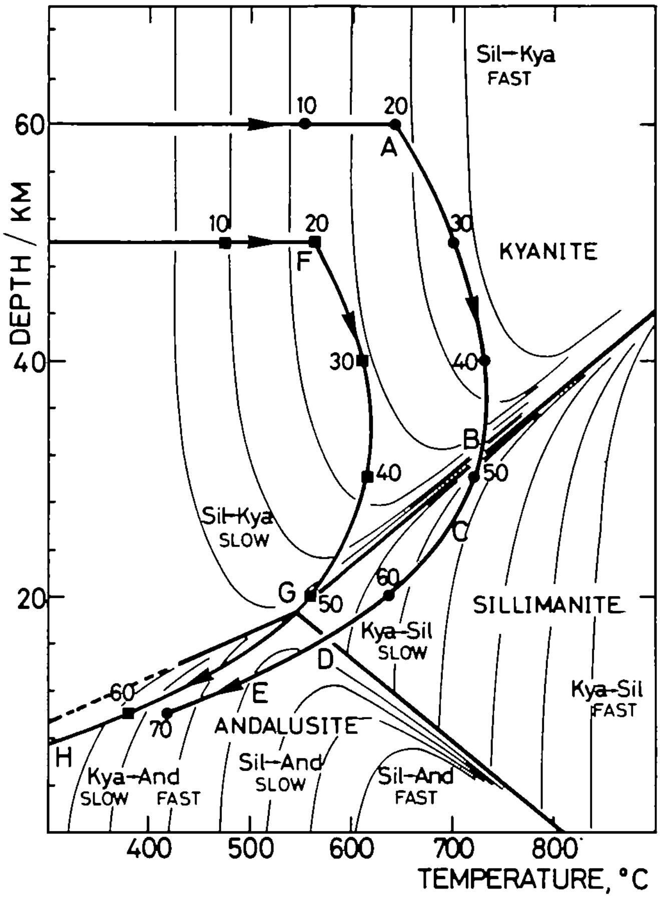


FIG. 3. Depth–temperature plots for two rocks buried at 50 and 60 km immediately after a thrusting event involving the emplacement of a 50 km thrust sheet in a regime governed by Heat Source Distribution II (Part I, subsection 3.2) and a thermal conductivity of 2·25 W m^{-1} K^{-1}; there is a phase of 20 Myr isobaric heating, followed by erosion of 50 km in 50 Myr (Fig. 8*a*, Part I). Solid symbols are placed on the paths at 10 Myr intervals. The Al_2SiO_5 relations shown lie between those of Richardson *et al.* (1969) and Holdaway (1971). The schematic kinetic curves are labelled SLOW to FAST to indicate rates of transformation between polymorphs.

the interpretations of textural relations between two Al_2SiO_5 polymorphs is ambiguous; for example, the *PTt* paths calculated in section 5 of Part I may pass from the kyanite to the sillimanite field either during the nearly-isobaric heating phase or during the nearly-isothermal decompression phase of the path.

2.3. *Fluid controlled reactions*

There is little doubt that dehydration and decarbonation reactions at depth within the crust can release H_2O and CO_2 that will percolate through the overlying rocks. It is commonly argued (e.g. Fyfe *et al.*, 1978; Walther & Orville, 1982) that such percolation will be concentrated into arborescent channels rather than distributed uniformly throughout the overlying rock column. Such localization is frequently inferred in the field from the concentration of retrogressive reactions around fracture zones, but there may also be more widespread fluid motion.

We have argued in Part I that, at least in the deeper crustal levels, the transfer of heat by H_2O or CO_2 is unlikely to be a significant proportion of the total *heat flux* (see also Brady, 1982). However, fluid motion is likely to be an extremely important part of the *metamorphic* history of a rock, for the introduction of a fluid phase may, depending on the chemistry, drive prograde or retrograde reactions that replace one set of minerals by another set that records radically different *PT* conditions. This is a common occurrence during retrograde metamorphism and Ferry (1982) suggests it may also be recorded by prograde reactions. Depending on circumstance, such occurrences may either be fortunate—as continuous reaction is likely to preserve only small parts of *PTt* paths (see below), whereas a phase of fluid motion may result in the recording of two distinct phases of metamorphism—or unfortunate, in that the intersection of a phase of fluid motion with the *PTt* path of the rock can result in the preservation of a fluid-catalyzed or fluid-driven reaction that bears no obvious relation to the peak metamorphic conditions.

2.4. *Distinct metamorphic parageneses, chemical zoning and PTt paths*

If the nucleation and growth kinetics of heterogeneous metamorphic reactions are sufficiently rapid, then mineral assemblages will always be close to equilibrium on a *PTt* path, even if the mineral compositions lag behind the equilibrium ones due to slow diffusion. We have discussed kinetic factors briefly in section 2.2 and we will assume here that apparent *PTt* paths are close to real ones while the reactions that record them are occurring. The classical explanation that reconciles near-equilibrium prograde reactions with the scarcity of retrograde reactions and the preservation of metamorphic minerals at the surface, is that fluids generated in devolatilization reactions during prograde metamorphism are able to leave their source area, and thus cannot drive retrograde reactions later (e.g. Schuiling, 1963). If this explanation holds, then the form of the *PTt* paths calculated in Part I suggests that, in many cases, the peak metamorphic conditions preserved may be close to the maximum temperature experienced by the rock.

Fig. 4*a* shows schematically two *PTt* paths in relation to four dehydration reactions. If all fluid released by these reactions escapes from the rock, then the deeper path, XYZ, will record its last prograde reaction at X, before reaching its maximum temperature, T_{max}, whereas the shallower rock, WYZ, would record its last reaction at W—after reaching T_{max}, and at lower pressure.

The exact relations of the peak metamorphic condition, defined by discontinuous reactions, to the highest temperature reached by the rock depends on the details of the *PTt* path and the reactions that may occur. In general, though, we should expect the former to be within a few tens of degrees of the latter. In either case, an influx of H_2O after the rocks had reached Y or Z would result in retrograde reactions under conditions markedly different from the 'peak' metamorphic conditions of X or W (see above). Adequate evaluation of the *PT* conditions of peak metamorphism and of a 'retromorphic' overprint (Albarede, 1976), or from a comparison of evolved fluid inclusions with peak metamorphic conditions (Hollister *et al.*, 1979) can provide two distinct points on a *PTt* path, although the absolute times of two such episodes are rarely known.

An additional source of information on *PTt* paths from discontinuous reactions are inclusions of minerals from earlier parageneses within zoned porphyroblasts that have grown during metamorphism (Rosenfeld, 1970; Thompson *et al.*, 1977). In all cases, if it proved possible to date the individual parageneses, this would give a valuable additional constraint on the metamorphic history.

The chemical zonation in a mineral can also give indications as to the sign of pressure and temperature change during its growth. For a continuous reaction, in which a crystalline

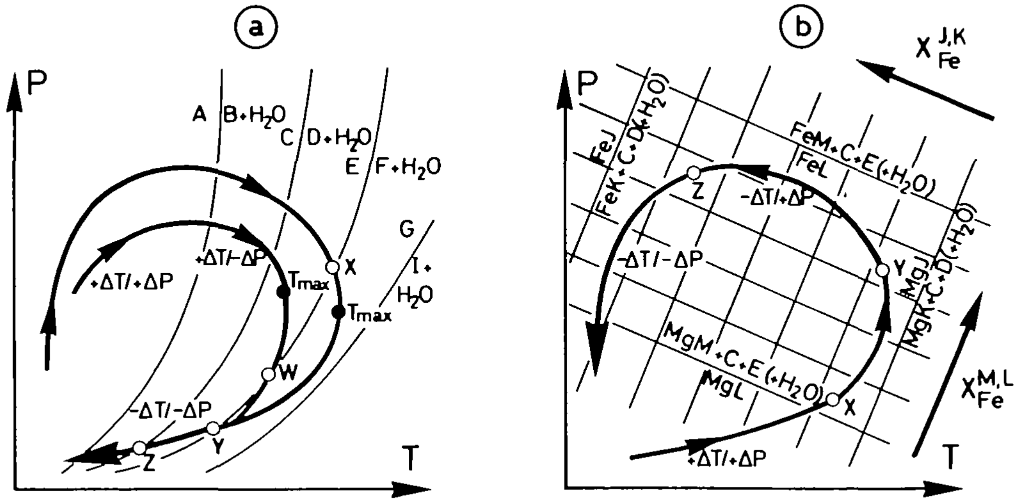


FIG. 4. Schematic P–T diagrams and PTt paths as shown relative to (a) successive dehydration reactions, to illustrate the relationship of T_{max} to the last dehydration reaction; (b) two cross-cutting sets of X_{Fe} compositional isopleths for continuous reactions. The path shown will result in different local reactions in the same specimens, for example, one may release H_2O and the other consume it along a given PTt path segment. In both figures the sense of changing $\Delta T/\Delta P$ is indicated by arrows.

solution reacts with other minerals in the paragenesis to yield other crystalline solutions, the compositional changes are in the same sense for both reactant and product (Fig. 4b). For example, cordierite (Crd) will react to form garnet (Gar) and Al_2SiO_5 + quartz with increasing temperature; both Crd and Gar will increase in Fe content at the expense of Mg, and growth zonation is often recorded in garnet.

Because the compositional isopleths for different continuous reactions often have quite different PT orientations, continuous chemical zoning in different sub-assemblages within the same sample may afford information on distinct portions of the PTt path. This may be seen with the aid of the schematic (Fe–Mg) compositional isopleths for the two continuous reactions

$$J \rightarrow K + C + D\ (+H_2O)$$

$$L \rightarrow M + C + E\ (+H_2O)$$

in Fig. 4b. The path segment XY would show little compositional decrease in X_{Mg} for minerals J and K, but large decreases in X_{Mg} for minerals L and M. The path segment YZ would show large decrease in X_{Mg} for J and K and a small increase in X_{Mg} for L and M. Fig. 4b also schematically illustrates that along segment XYZ the reaction of L to M + C + E is releasing H_2O while K + C + D are hydrating to form J. (Note that, although none of the PTt paths considered in these papers exhibits the form of that in Fig. 4b, portions of such a path may be appropriate for, say, rocks buried close to a thrust.)

2.5. *Diffusional zonation and PTt paths*

Many workers have discussed the widespread and continuous chemical zonation of garnets and a few other minerals (plagioclase, amphibole and pyroxene) in metamorphic rocks. Where it can be shown that such zonation is diffusional and not growth zonation (see above) the potential exists for determining cooling rates from the preserved diffusional profiles. Lasaga *et al.* (1977) give expressions for the interdiffusion of two ions between two adjacent minerals in a regime where the temperature falls at a constant rate. They are able to match well the observed variation in Mg content in a pair of adjacent garnet and cordierite crystals in a rock from central Massachusetts and to place fairly tight bounds on the quantity: (coefficient of interdiffusion of Fe and Mg in the garnet divided by the cooling rate experienced by the rock). Lasaga (1983) has enlarged on this work and expresses the hope that diffusional profiles may be eventually used in 'geospeedometry'—the estimation of cooling rates. The uncertainty in this process lies in the difficulty of determining the coefficients of diffusivity in natural systems accurately enough to tie down cooling rates to better than an order of magnitude (see Lasaga *et al.*, 1977; Lasaga, 1983). In this section we wish only to discuss in general terms the influence of the *PTt* paths calculated in Part I on diffusional processes. For this purpose it is adequate to use the concept of a blocking temperature, below which chemical diffusion becomes so slow that it may be neglected. We use the analysis of Dodson (1973, 1979), who gives a simple expression for the closure temperature of a chemical system consisting of a solid phase in contact with an infinite reservoir. Although his analysis is primarily devoted to the closing temperatures of isotopic clocks, it is a good approximation to the case in which a pair of adjacent silicate minerals exchange cations, and the coefficient of interdiffusivity of the ions is very much lower in one mineral than in the other; (see Dodson 1973; Lasaga *et al.*, 1977; Lasaga, 1983).

Dodson (1973, 1979) gives an expression for the closing temperature of such a system, where the coefficient of diffusivity, D, is given by

$$D = D_0 \exp\left[-E/RT\right] \tag{1}$$

and E is the activation energy for the diffusion process, R is the gas constant and T is the absolute temperature. For a constant rate of temperature change ($\dot{T}$) the closure temperature (T_c) of a grain of radius a is given by

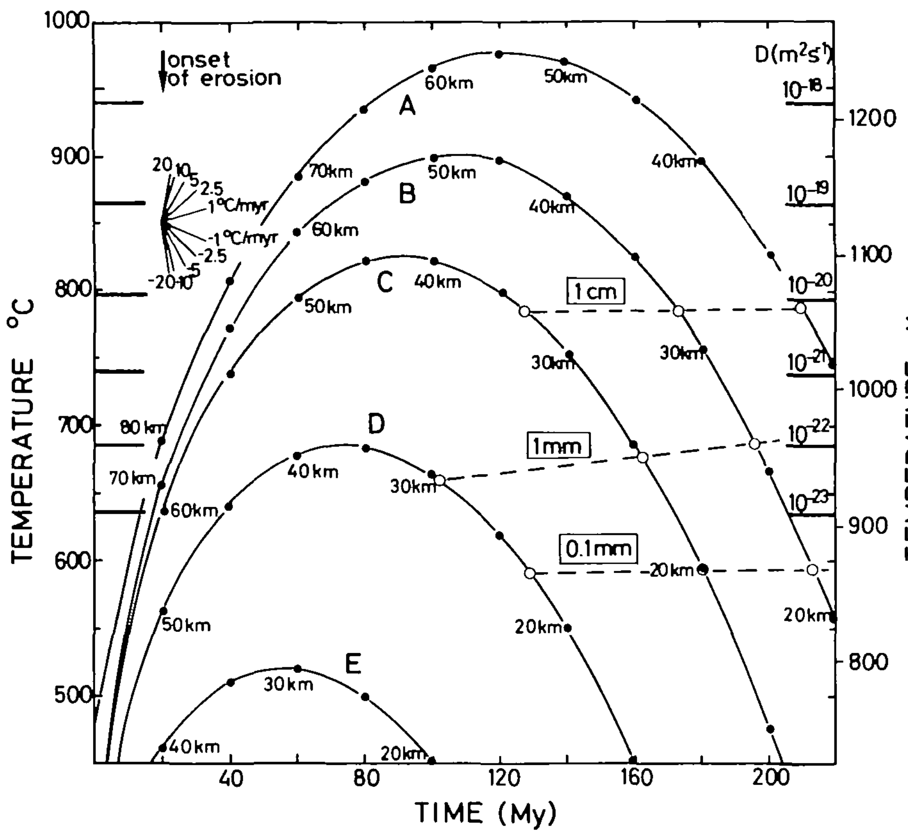
$$\frac{E}{RT_c} + \ln\left[\frac{E}{RT_c^2}\right] = \ln\left[\frac{-AD_0}{a^2\dot{T}}\right] \tag{2}$$

where A is a constant, depending on the geometry of the diffusing grain, that is of order 10.

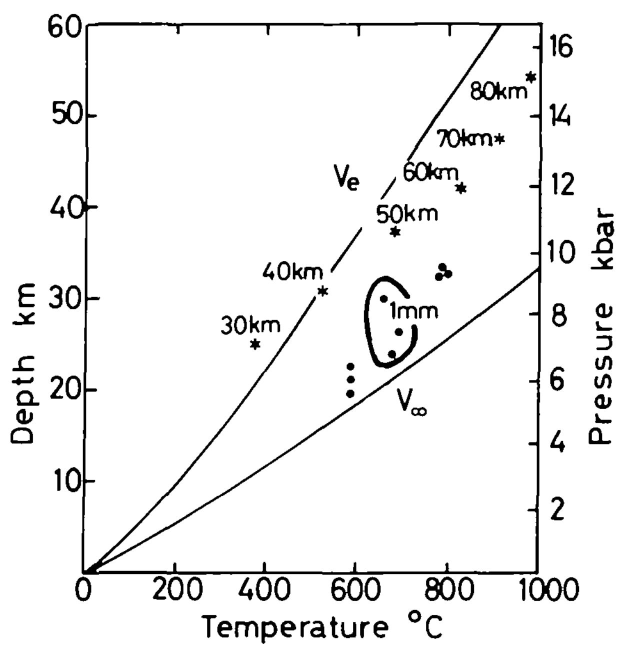
Fig. 5 shows the temperature–time paths followed by rocks buried in the crustal thickening episode shown in Fig. 8b of Part I. The crust is thickened by a single thrust sheet 50 km thick, there is a period of 20 Myr without erosion, then the 50 km of overburden is removed in the next 200 Myr. Paths are shown for rocks buried at 80(A), 70(B), 60(C), 50(D) and 40(E) km in the thickened pile—respectively 30, 20 and 10 km below the thrust, at the thrust, and 10 km above the thrust.

FIG. 5*a*. Temperature–time curves for rocks buried in a regime governed by Heat Source Distribution II and a conductivity of 2·25 W m^{-1} K^{-1} that is subjected to a 50 km thrust at time zero, to 20 Myr of isobaric heating, then to erosion of 50 km in the interval 20 to 220 Myr (Fig. 8*b*, Part I). Paths are shown for rocks that were buried at depths of 80 km (A), 70 km (B), 60 km (C), 50 km (D) and 40 km (E) immediately after thrusting. Also shown (on the right hand axis) are the coefficient of interdiffusivity of Fe and Mg in garnet, given by equation (1), using the parameters given by Freer (1981), and the closing temperatures calculated from equation (2) for grains of radii 1 cm, 1 mm and 0·1 mm, using the cooling rates calculated from these curves, and the same diffusivity data. Solid circles are at 20 Myr intervals on the curves and the depths during exhumation are marked every 40 Myr.

FIG. 5*b*. Solid circles show closing temperatures (equation 2) from Fig. 5*a* for different grain sizes plotted against the pressures at the time of closing; the solid line encloses the conditions for a = 1 mm; the upper group of three points are for a = 1 cm and the lower for a = 0·1 mm. Also shown are the undisturbed geotherm before thrusting (V_c; equation 1, Part I) and the steady state geotherm that would be obtained after thrusting, if erosion were not to occur (V_∞). The asterisks with depths beside them denote the *PT* conditions at which the rocks initially buried to those depths reached their maximum temperatures on the uplift paths of Fig. 5*a*.


onset of erosion
D(m^2s^{-1})
A
B
C
D
E
60km
50km
70km
40km
30km
20km
80 km
1 cm
1mm
0.1mm
1°C/myr
-1°C/myr
TEMPERATURE °C
TEMPERATURE K
TIME (My)

(*a*)


Ve
V∞
80km
70km
60km
50km
40km
30km
1mm
Depth km
Pressure kbar
Temperature °C

(*b*)

The first point to note about these paths is that the rocks spend a large proportion of the total metamorphic cycle within 50 to 100 °C of the peak metamorphic temperatures that they reach. Path A spends 130 Myr, out of the 220 Myr, within 100 °C of its peak temperature of 975 °C, and 95 Myr within 50 °C of this; the corresponding intervals are 120 (85) Myr for Path B, 115 (80) Myr for Path C, 105 (75) Myr for Path D and 100 (70) Myr for Path E.

This form of *Tt* path is common to most of the cases illustrated in Part I; the exceptions are rocks *overlying* thrust planes (Figs. 3 and 6–8, Part I), which cool rapidly immediately after the event and may not subsequently attain temperatures higher than their initial ones. Rocks buried below the thrust surface or in the other classes of model (Figs. 4 and 5, Part I) all exhibit the kind of paths shown in Fig. 5*a*, and the discussion below probably applies to many rocks exposed in deeply eroded terrains. Also illustrated in Fig. 5*a* are coefficients of diffusivity calculated using the data given by Freer (1981) for the interdiffusion of Fe and Mg in garnet; namely, $6{\cdot}1 \times 10^{-4}$ m^2 s^{-1} for D_0 and 344 kJ $mole^{-1}$ for E in equation (1). The bars on the temperature axes of Fig. 5*a* show that for this choice of parameters, the coefficient of interdiffusivity changes by five orders of magnitude between 940 °C and 640 °C. This means that, for example, the rock following Path C spends 60 Myr above the 10^{-20} m^2 s^{-1} isopleth, then passes from the 10^{-20} m^2 s^{-1} isopleth to the 10^{-22} m^2 s^{-1} isopleth in the next 40 Myr.

These remarks apply equally to other thermally activated processes that yield *PT* estimates of metamorphic conditions, and it might at first sight seem that (as was discussed in the previous section) the form of *PTt* paths tends to favour the preservation of mineral assemblages and/or compositions that are close to peak metamorphic conditions. This is not necessarily so, however.

We might suppose that a geobarometer is calibrated on the basis of the distribution of cations between garnet and some other coexisting minerals with which it is in contact, and, as above, regard the other minerals as having much higher diffusivity than the garnet, so that the rate of diffusion of the ions in the system is governed by their diffusion in the garnet. The choice of garnet as the least diffusive is, of course, arbitrary; in a real case one would apply these arguments to whatever mineral had in fact the lowest diffusivity. The data discussed by Freer *et al.* (1982) suggest that the diffusion of calcium in pyroxenes may be the process that governs the closing of geobarometers in the situations that we describe here; however, they also show that the diffusivity of calcium in silicates is so low that it is hard to determine accurately. We have chosen the data quoted above for the interdiffusion of Fe and Mg in garnet (Freer, 1981) as geologically reasonable values in order to illustrate a general principle; the remarks that follow ought not to be taken to apply exactly to any real choice of cations, geobarometer or geothermometer. The reader is referred to Lasaga (1983) for a discussion of cation diffusion in less simplified mineral systems following more simplified thermal histories.

Open circles joined by dashed lines illustrate closure temperatures calculated from equation (2) for these data. Equation (2) gives the temperature at which diffusion in the garnet effectively ceases for a given grain size and cooling rate, and thus defines the position in *PT* space where a geobarometry point is determined on a particular path. As there is a relatively small range in cooling rates in Fig. 5*a*, grains of a given radius close at nearly the same temperature on each temperature–time path and the principal differences in closing temperature arise from differences in the assumed grain size. (The relevant 'grain size' in this context may not be the size of individual grains in the rock, but of subgrains with boundaries along which the diffusing species may move much faster than they can through the crystal lattice.) In all the cases illustrated, grains that pass appreciably above their closure temperatures, subsequently close at temperatures considerably below the maximum, and do so within a moderate pressure interval (less than about 3 kb) because of the rapid change of diffusivity with time on the cooling portion of the paths. If the closure temperature for a chosen grain size should happen to lie close to the peak metamorphic temperature, then the mineral compositions will indicate near peak temperature but in this case they may be set over a large range of pressures (perhaps 6 to 10 kb).

Another important feature of the curves of Fig. 5*a*, is that the rocks follow similar uplift paths once they are past their peak metamorphic conditions—albeit at different times. For example, paths A, B and C all pass through 30 km depth at temperatures within a few degrees of 750 °C, despite their radically different initial and peak metamorphic conditions; however the times of passing 30 km are very different, being 220, 170 and 140 Myr, respectively. As the closing temperature is relatively insensitive to cooling rate, this means that rocks from a given terrain that are used as geobarometers of the kind described here will not, in general, preserve a record of their peak metamorphic conditions but will preserve a record whose range is more affected by variations in the parameter D_0/a^2 (equation 2) than by variations in peak metamorphic conditions. This is illustrated in Fig. 5*b* where the closure *PT* conditions for the three grain sizes in Fig. 5*a* are compared with the steady state geotherm for the unthickened crust (V_e), that for the thickened crust (V_∞) and the *PT* points at which the rocks experienced their maximum temperatures. The closure conditions occupy a smaller range of *PT* space than do the maximum-temperature *PT* points, and they lie in considerably higher *T*/*P* conditions.

As shown in Fig. 5*b*, we should expect such data to yield a small pressure or temperature field (within the limits associated with variation of D_0, E and a^2 for a given system) even if a suite of rocks had experienced peak metamorphic conditions that range in pressure and temperature by several kilobars and several hundred degrees. It presumably does not need emphasizing that a *PT* point determined with a geobarometer and a geothermometer that closed at different temperatures would not represent any point of the *PT* path that the rock followed.

The exception to these considerations occurs when the relevant minerals grow below (and never pass above) the closure temperature for diffusion, and in this case it is perfectly possible to have shallowly-buried rocks recording higher pressures than more deeply buried rocks. For example, the compositions of 1 cm grains grown during prograde metamorphism on path D (in Fig. 5*a*) could reflect pressures in the depth range 50 to 35 km, while the same-sized minerals on paths C to A would record burial depths of around 30 km.

Our remarks have been necessarily simplified in this section and it is difficult to make specific points about systems in which coefficients of diffusion are very strong functions of temperature, and when diffusion mechanisms, let alone activation energies may be unknown in the region of interest. Where zonation by diffusion is involved, we expect that, because of the very strong temperature-dependence of diffusivity, such zonation would only preserve information on a relatively narrow temperature range (around 100 °C for the cases shown in Fig. 5; see also Lasaga *et al.*, 1977; Lasaga, 1983), and that a similarly narrow range of experienced temperature may separate minerals exhibiting growth-induced zonation during prograde metamorphism from unzoned minerals from which diffusion has removed such zonation. It may in principle be possible to determine diffusion data to define closure temperatures for geobarometers in the same way as is done for isotopic clocks; if this is done with sufficient reliability, one could then use geobarometers to give simultaneous *P* and *T* determinations.

Although it may appear that the action of temperature-dependent diffusion along the kinds of *PTt* paths expected in orogenic terrains involves unwelcome complexity, this very complexity offers the hope of determining *PTt* paths from rocks in a way which simpler reactions cannot do (e.g. Harrison & McDougall, 1982; Lasaga, 1983).

2.6. *Metamorphism and deformation*

Because of the direct relation of rock displacements to temperature perturbation, the establishment of relationships between the style and intensity of deformation and the metamorphic development of a terrain is clearly desirable. Regional mapping of phases of metamorphic growth relative to a series of deformational episodes is capable of determining such a chronology (e.g. Ramsay, 1963; Rast, 1965) but, as with other indicators discussed in this section, it is important to understand how the nature of the *PTt* paths followed by rocks may affect the information we extract from them.

Oxburgh & England (1980, p. 392) emphasized that a short phase of deformation (which may be regarded as happening instantaneously on these time scales) will, because of the diachronous nature of the metamorphism, occur at different stages of the thermal evolution of rocks buried to different depths in the crust. Fig. 6 illustrates *PTt* paths for three rocks initially buried at 40, 50 and 60 km. The thermal development is governed by a 35 km thrust and by the other parameters described for Fig. 3*e*, Part I. A phase of deformation (*d*) occurs

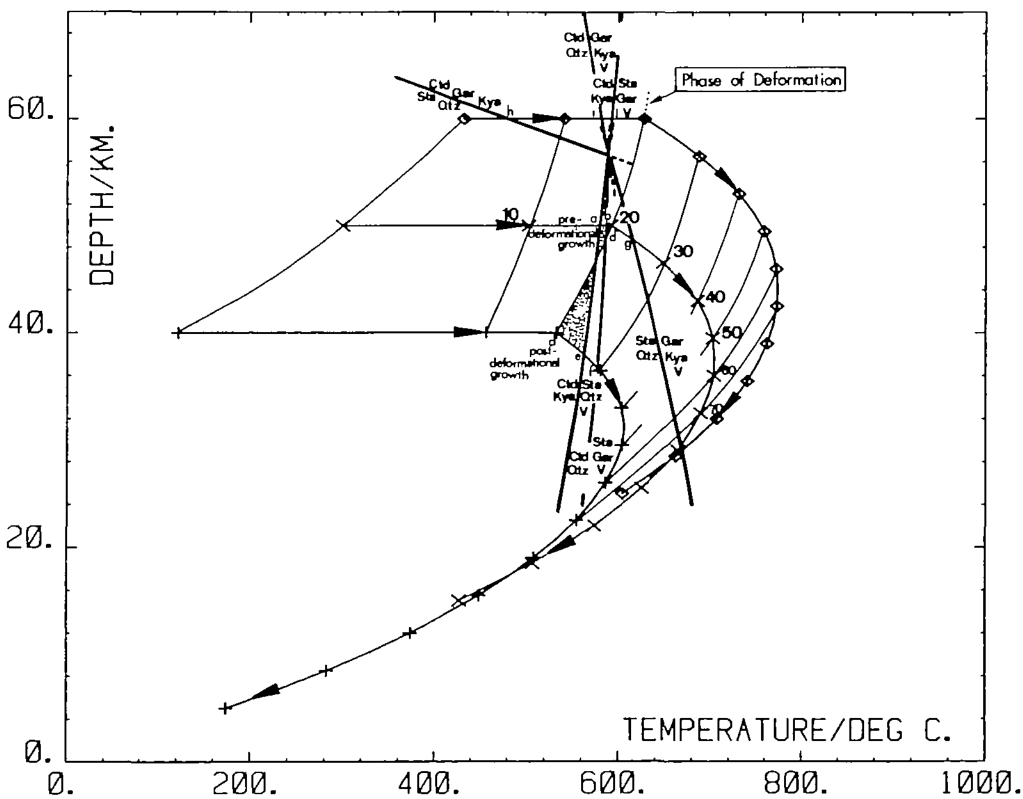


FIG. 6. Three *PTt* paths for isobaric heating and uplift of samples initially buried at depths of 60, 50 and 40 km (from Fig. 3*e*, Part I). Isochronous points are connected by lines. The phase of deformation, *d*, is assumed to be very short compared with the duration of the *PTt* paths. The experimentally determined curves from Ganguly (1972, p. 342) indicate that reactions can be pre-deformational for one sample (50 km) and post-deformational for another (40 km).

20 Myr after thickening, coincident with the beginning of uplift. Also shown are experimentally determined equilibria (Ganguly, 1972, fig. 3) for some common porphyroblast minerals in pelitic rocks. It will be seen that growth of staurolite (Sta) and quartz (Qtz) from chloritoid (Ctd) and kyanite will be pre-deformational for the 50 km sample at *a* and post-deformational for the 40 km sample at *e*. Likewise, the growth of garnet (Gar) and Sta from Ctd + Qtz will be pre-deformational for the 50 km sample at *b* and post-deformational for the 40 km sample at *f*. The 60 km sample will show a completely different mineral reaction history from these, and only the 50 km sample will exhibit the prograde growth of Gar + Kya from Sta + Qtz (at *g*).

3. *PTt* PATHS IN CONTINENTAL COLLISION ZONES

Section 2 briefly discusses the means by which *PTt* data may be obtained, and some of the pitfalls involved in interpreting them. In this section we discuss the forms of *PTt* paths that would be expected during collision zone metamorphism under differing conditions of heat supply and tectonic setting; as in Part I we discuss only the large scale features of such paths, recognizing that smaller scale variations cannot be interpreted except on an *ad hoc* basis. It is unlikely that any one section of a *PTt* path obtained from an area would provide unequivocal evidence of a tectonic setting but several such paths observed regionally, in conjunction with structural, textural and geochronological studies can provide valuable constraints on processes in metamorphic belts.

3.1. PTt paths followed during continental collision

An event of continental collision, by thickening the continental crust establishes a distribution of heat supply that would, in steady state, support geothermal gradients very much higher than those in the undisturbed crust (Part I, section 5 and Appendix B). Immediately after this collision, though, the thermal gradients are considerably less than the undisturbed gradients, and the difference between starting and equilibrium conditions may be expressed as a perturbation of several hundreds of degrees throughout the lithosphere (Part I, Fig. B1).

The extent to which this perturbation decays is governed by the time that is available before exhumation terminates the heating history of individual rocks. As discussed in Part I, the time scales for diffusion of heat and for the erosional process are comparable, and this means that steady state geothermal gradients are not achieved, although thermal gradients appreciably higher than the undisturbed ones are to be expected.

With the exception outlined below, the form of *PTt* paths followed by rocks during continental collision is unlikely to be sensitive to the geometry of the thickening process; the time scale for the decay of the *differences* in temperature between, for example, crustal thickening by a single thrust sheet and by homogeneous strain, is very short compared with the erosional time scale (Part I, section 5 and Appendix B). Consequently the main difference to be expected between the two end-members of the crustal thickening geometry is in the first 20 Myr or so, when there would be a phase of isobaric cooling *within* a thrust sheet that would be absent in the case of homogeneous strain (Figs. 3–5, Part I; and compare *PTt* curve 'Above' in Fig. 7*a* with curves in Fig. 7*b*). Rocks *below* thrust sheets (*PTt* curve 'Below' in Fig. 7*a*), like those in homogeneously thickened crust (Fig. 7*b*), will experience nearly isothermal compression unless they are close enough to a warm thrust sheet to experience a marked warming phase during loading (see, for example, Crawford & Mark, 1982; Spear & Selverstone, 1983), or unless loading is slow compared with the time required for thermal relaxation.

Fig. 7 illustrates some of the kinds of *PTt* paths that would be expected in continental collision zones, based on calculations carried out in Part I. Figs. 7*a* and 7*b* illustrate the cases

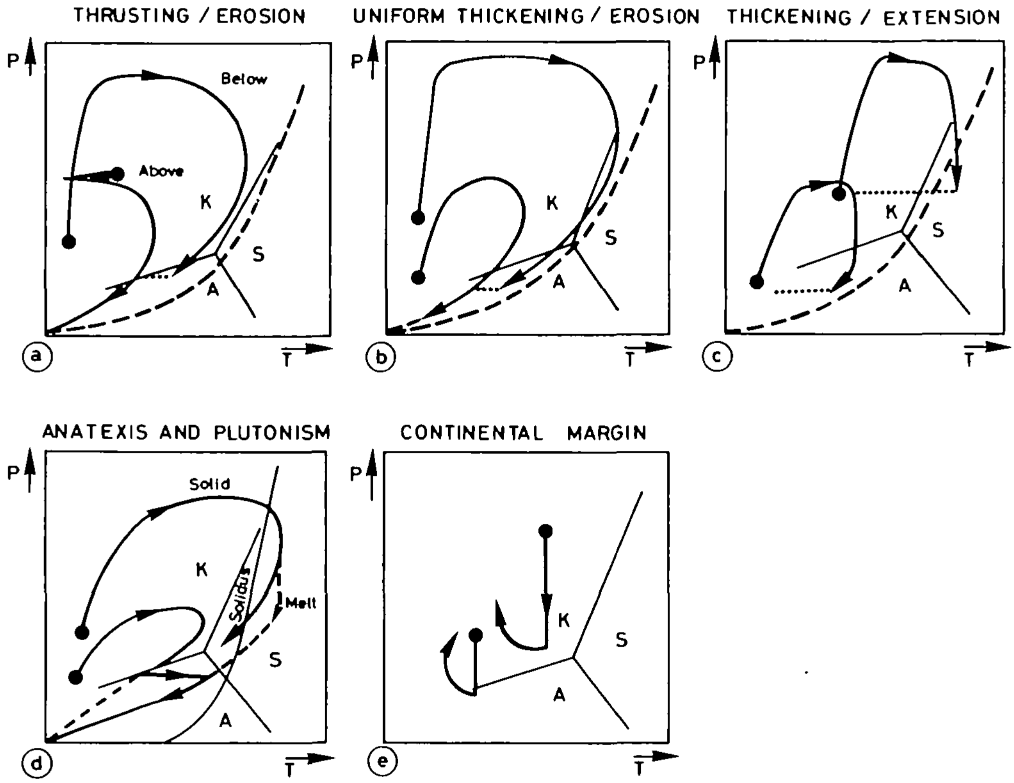


FIG. 7. Schematic PT diagrams showing the forms of PTt paths produced by a variety of tectonic processes, based on the arguments of section 3.1. The paths are limited to those expected during continental collision, and even so, this is not regarded as an exhaustive catalogue. A schematic "dry" solidus is shown in d as a solid line and in a, b and c as dashed lines. The relative stability of Al_2SiO_5 polymorphs is shown by K, A and S.

discussed above, where crustal thickening is by a single large thrust (Fig. 7*a*) or by homogeneous strain (Fig. 7*b*)—the geologically plausible approximation to this case might involve many imbricate thrusts in the brittle layer of the crust, each producing a thermal perturbation with a very short time constant. In these cases, which are discussed in Part I, section 5, the earliest phase of the PTt path (burial) shows strongly increasing pressure and moderate increase of temperature ($+\Delta T/++\Delta P$) within the glaucophane–jadeite facies series. This would be characterized by a transition from zeolite, through prehnite–pumpellyite to glaucophane–lawsonite facies and, if burial were deep enough, to glaucophanitic eclogites. Aragonite could form from calcite and, again at the higher pressures, jadeite + quartz from plagioclase.

As discussed in Part I, sections 5 and 6, certain regimes will permit the preservation of such mineral assemblages because the temperatures never rise out of the Gla–Jad facies series. These conditions are ones in which the initial heat supply in the pre-thickening lithosphere is low (mainly Heat Source Distribution I, Figs. 3*a*, *d*, *g* and 4*a*, *d*, *g* of Part I—corresponding to probable present day conditions beneath old shields: Part I, section 3.2) or in which the initial heat supply is normal, but the thickening process affects the entire lithosphere, so that the post-thickening heat supply to the crust is very low (Fig. 5, Part I). Rocks following paths such as these would experience prograde metamorphism entirely within the Gla–Jad facies series, including a late stage of almost isothermal decompression.

Rocks that experience burial in regimes with thermal budgets regarded as average for the continental lithosphere (Heat Distribution II—Figs. 3*b*, *c*, *d*, 4*b*, *c*, *d* and 6 to 8 of Part I—produces a steady state surface heat flux equal to the continental average) also experience an initial phase of moderate temperature and large pressure increase ($+\Delta T/++\Delta P$), but the preservation of high *P*–low *T* facies rocks from this phase would require either an absence of H_2O (which would otherwise enhance the kinetics or enable rehydration reactions to occur), or an uplift in a much shorter time than is illustrated in Part I—less than about 30 Myr (Oxburgh & Turcotte, 1974; England & Richardson, 1977; Draper & Bone, 1981).

If they are not exhumed rapidly, the rocks then experience a period of heating at depth and temperature rises of several hundred degrees are likely. The earlier part of this phase will be characterized by relatively large increases in temperature and small or no decrease in pressure, while in the later stages, as the rate of temperature increase diminishes, there will be large decreases in pressure accompanying a small increase and then decrease in temperature (Fig. 7*a*, *b*, this paper; Part I, section 5). The temperature increases involved in this phase depend strongly on the heat source distributions (Figs. 3 to 5, Part I) but are relatively insensitive to the total duration of the exhumation event, provided that it is longer than about 50 Myr (Figs. 6 to 8, Part I). However, the longer the delay between the assembly of over-thickened crust and the onset of rapid erosion, the closer the pile will be to its new steady state geotherm; consequently, rocks in slowly evolving belts will start to cool at greater depths than equivalent rocks in more rapidly-eroding terrains (Part I, section 5 and Appendix C). Finally, as the rock approaches the surface, it will experience large decreases in both temperature and pressure.

Most of the examples illustrated for Heat Source Distribution II (Part I, Section 3.2) show *PTt* paths that pass into or close to the kyanite–sillimanite facies series during the stage of near-isothermal decompression. Individual reactions may be recorded that indicate prograde metamorphism (in the sense of $+\Delta T$) but the predominant characteristic of this section of the path is the large change in pressure ($-\Delta P$) while the rocks are within a few tens of degrees C of the maximum temperature they achieve (see Fig. 5). We should expect to preserve small segments of *PT* paths with this characteristic, depending on the reactions possible for a given rock composition. Such paths may be revealed in solid inclusions in porphyroblasts, even if macroscopic assemblages indicative of earlier conditions were replaced.

It is important to recognize that these paths pass at a very high angle through the locus of peak metamorphic temperatures that are experienced by the rocks (England & Richardson, 1977; Part I, Figs. 3–8; Fig. 1, this paper). Such a locus or piezothermic array (section 2), might be indicative of a *spatial* progression towards a region of deeper burial (or greater heat supply; see Figs. 8, 9), but would not in any way indicate the *temporal* progression of metamorphism along the *PTt* paths of the individual rocks.

We have assumed in all the cases examined here that once the crust has returned to its original thickness, no more erosion occurs. It is, however, common to find that old orogenic belts are rejuvenated by later events; often these tectonic events expose rocks that were more deeply buried in the earlier event, without introducing any metamorphic overprint on them; consequently, it is of interest to examine the *PTt* paths of rocks that do not reach the surface in our models.

Each of the models illustrated in Part I finishes the erosional phase of its development with a geotherm that is hotter than the steady state profile associated with its heat supply. The last part of the *PTt* paths experienced by rocks still below the surface at this stage will be one of isobaric cooling. However, this event (depicted by the dotted lines in Figs. 7*a*, *b*) will cover a small range in temperature and will occur at conditions well below the maximum temperatures experienced by the rocks.

England & Thompson (1984*b*), noting the evidence for extensional activity as a late phase in the development of compressional terrains—as has, for example, been documented extensively in the Basin and Range Province of western North America (see Eaton, 1982, for a recent review), in Tibet (Molnar & Tapponnier, 1975, 1978) and in the Andes (Suarez *et al.*, 1983)—illustrate the *PTt* paths that might be followed by rocks subject to such tectonic unroofing. The burial phase is presumably similar to that in the cases discussed so far, but the exhumation phase takes place at a much faster rate than is achieved by erosion; consequently, the deeply buried rocks experience a much greater amount of decompression before they begin to cool appreciably. Indeed, in the cases illustrated by England & Thompson (1984*b*) the most deeply buried crustal rocks experience the entire extensional phase without cooling. At the end of the extension, such rocks may lie at temperatures several hundred degrees above the steady state geotherm, and so experience isobaric cooling from close to the highest temperatures they reach as the last phase of their *PT* history. This is shown schematically in Fig. 7*c*. Evidence of isobaric cooling is not, of itself, diagnostic of magmatic involvement in the metamorphic budget, contrary to the suggestion made by England & Richardson (1977).

England & Thompson (1984*b*) also discuss in more detail than has been possible here, the generation of melts within the continental crust during collision events. Particularly in the cases where the crust is thickened by 35 km more with average surface heat flow (60 mW m^{-2}) or greater (Figs. 3, 4, 6 and 8, Part I), rocks buried in the lower one-third of the crust can achieve amphibolite–granulite conditions and anatexis. It is an important conclusion of that study that crustal anatexis as a result of burial (which has frequently been suggested in the past, e.g. Bott, 1954; Tuttle & Bowen, 1958; Dewey & Burke, 1973) is readily achieved in a system where the crust is doubled in thickness, and fluid is available to induce melting, even in the absence of heat transfer by magma from the mantle.

Once melt has accumulated in sufficient volume to rise through the crust, the calculations described so far do not describe accurately the *PTt* paths followed by rocks in the system. It is still possible, though, to make qualitative statements about them. For the rocks that just pass into the melting field on the uplift path, segregation of melt is less favoured, owing to the high viscosity of first melts (Shaw, 1972) and their low volume fraction, whereas deeper burial and higher temperatures would result in the production of lower viscosity, more voluminous melts; this could permit the segregation and ascent of S-type granites from metasedimentary sources or I-type 'granites' from amphibolite, calc-alkaline magmatic rocks and biotite–plagioclase–quartz gneisses. The distance of ascent of such melts depends very strongly on their viscosity and density contrasts and is hard to predict.

Even if a high proportion of the deeply buried rock were to melt, we should expect the residue to follow a *PTt* path qualitatively similar to those that are illustrated in Part I on the assumption of no melting. We may also reasonably represent the magmatic ascent by a path of nearly isothermal decompression (unlikely to be preserved by its mineralogy). The main departure from the calculated *PTt* paths will be experienced by the overlying rocks intruded by the magma, which will show the isobaric heating characteristic of contact metamorphism. If the intrusive activity is of regional extent, an entire terrain could exhibit such a path. The *PTt* paths for such situations are shown schematically in Fig. 7*d*.

Some *PTt* paths calculated for rocks initially buried deeply in regimes of higher heat-flux or lower conductivity pass within the andalusite–sillimanite field (Figs. 3 and 4, Part I). They all pass into the field on the uplift parts of the paths ($-\Delta T/-\Delta P$), and it seems that metamorphism to And–Sil facies series in a prograde sense ($+\Delta T/\pm\Delta P$) is not likely without additional heat supply from intrusives. As those paths lying closest to the And–Sil field have in any case passed through the likely range for anatexis and are also underlain by paths that would involve crustal melting, we should expect to find a close association of And–Sil metamorphism and intrusive activity in continental collision zones (see Part I, section 6).

We consider one other form of metamorphism that may be of relevance in collision zones, although we know of no evidence for this at present. The telescoping of passive continental margins may be a common feature of the early stages of orogenesis (e.g. Helwig, 1976; Jackson, 1980). If the idealized approach of Part I is followed, we should expect that subsequent thermal metamorphism would overprint any record of *PT* paths followed during the extension and subsidence of continental margins. However, the real geometry of a collision zone might involve the exposure of deeper levels of a passive margin without their being subjected to higher grade metamorphism. The paths followed by rocks in such a tectonic setting show an initial phase of isothermal decompression—as in Fig. 7*c*—followed by a phase of cooling and compression ($-\Delta T/+\Delta P$) as the geotherm relaxes from its elevated position and sediments accumulate on the subsiding margin (Fig. 7*e*).

3.2 Influence of heat supply, depth of burial and time scales in continental collision metamorphism

The thermal development of regions of continental extension has been extensively investigated, and relations exist between the thermal evolution and such measurable quantities as subsidence and surface heat flow which permit fairly precise determination of the thermal history of such regions (McKenzie, 1978; Royden *et al.*, 1980). In contrast, the thermal development of compressional zones remains less approachable. This is partly because of the apparently greater variety of geometries in collision zones, and partly because of the less direct relationship between the conditions governing the thermal development of orogenic belts and the observables.

In this sub-section we discuss the use of *PT* data, heat flow and geochronological studies in determining the heat supply, the depth of burial and the time scales of orogenic development This discussion is mostly a matter of principle, as relatively few studies have been carried out.

Because *PTt* paths are relatively insensitive to the time scale of erosion, provided that it is longer than about 50 Myr (Part I, Figs. 6 to 8), they should not be expected to constrain the duration of an orogenic episode. A relatively small change in the heat supply to the orogen produces changes in maximum metamorphic grade that are comparable with those obtained by changing the erosional time scale by a factor of 4 (Part I, Figs. 3 to 5).

The seminal work of Clark & Jäger (1969) demonstrates the added resolution that is possible if geochronological and heat flow data are added to *PT* data. Bickle *et al.* (1975) and England (1978) showed that knowledge of the present day heat flow permits discrimination between metamorphism at shallow depth in a high *T*–low *P* terrain and burial at moderate depth in a medium *P*–medium *T* terrain for the case of the Tauern metamorphism of the Eastern Alps. Royden & Hodges (1984) have presented a technique for inferring metamorphic conditions from portions of *PT* paths, but this, too, is relatively insensitive to rates unless it is combined with heat flow or geochronology data.

Surface heat flow is a useful constraint in young orogenic belts, while it is still reflecting the thermal perturbations set up during recent deformation. Surface heat flow in older terrains can also be a useful constraint; if it can be demonstrated that the present day heat flow is not augmented by any more recent source it may reasonably be regarded as an estimate of the minimum heat available to drive metamorphism. Cooling data from the closing of isotope or fission track systems are also useful in older belts, provided that the precision on the dates is considerably less than the likely duration of thermal events (say, 30 Myr, e.g. Harrison & McDougall, 1982).

As discussed in section 2.5, if an adequate knowledge of diffusion parameters becomes available, the 'geospeedometry' techniques advocated by Lasaga (1983) may also provide valuable constraints. We feel that the uncertainties involved in relating laboratory determinations of chemical diffusivity to diffusion in natural systems may always be too great;

however we would like this feeling to be proved wrong. The analysis of Dodson (1973) or Lasaga (1983) shows that any cooling rate determination depends directly on the chemical diffusivity assumed (see equation (2) for a simple case). We should thus require to know values of diffusivity to within a factor of 3 for such geospeedometry to yield valuable results. This emphasizes the continuing need for analyses based on the decay of thermal perturbations, for these depend on an estimate of the average thermal diffusivity of the lithosphere, which is currently better constrained.

We should like to emphasize that for such analyses to be useful they must consider the entire geologically plausible parameter range for a given area in order to determine constraints on the thermal budget for metamorphism; it is not sufficient merely to demonstrate that one particular combination of parameters is capable of producing the metamorphism (see discussion in section 5, Part I, and Bickle *et al.*, 1975; England, 1978). It is in the nature of this kind of study that the constraint that is most commonly found is an estimate of the minimum heat supply required to produce the observed metamorphism.

Note that in terrains where uplift is rapid (less than about 10 Myr) the heating due to radiogenic sources will be less, and T_{max} will be achieved close to P_{max}. On such short time scales the importance of metamorphic reactions as heat sources and heat sinks may be comparable to that of the radiogenic heat sources (Part I, section 3.3).

3.2.a. *Facies series and tectonic setting of metamorphism*

As we remarked above, evidence that a rock has passed through a given facies is not evidence that the metamorphism took place under a thermal regime whose steady-state geotherm would cross the same area of *PT* space. Figs. 3 to 8 of Part I show the ranges of conditions in which rocks undergoing regional metamorphism in a continental thickening event are within 50 °C of the maximum temperature that they experience (thick lines on Figs. 3 to 8, Part I). These envelopes have similar slopes to the fields in *PT* space that are occupied by the facies series in these figures. We argue in section 2 that the form of *PTt* paths during this kind of metamorphism may favour the preservation of peak metamorphic mineral assemblages that represent conditions within a few tens of degrees of the true maximum temperatures experienced by the rocks, so this correspondence is not surprising.

These figures also show the steady state geotherm, for the pre-thickening thermal conditions of each model, in relation to the *PTt* paths and to the envelopes defining the area of *PT* space that lies within 50 °C of the peak temperatures on the *PTt* paths. In most cases, except where the entire lithosphere is thickened (Part I, Fig. 5), the steady state geotherms lie within these envelopes and it may be tempting to equate a *PT* array of peak metamorphic conditions with a steady state geotherm. However, a single example will show the error of such an action: in Fig. 3*e* of Part I, the locus of points at which rocks attain their maximum temperatures does in fact lie close to the steady state geotherm for the pre-thickening conditions, but variations in erosional history, for the same original thermal conditions, will remove that coincidence (see Figs. 6 to 8, Part I). The locus at which these same *PTt* paths pass, later in their development, through temperatures 50 °C below their maximum is shown by the thick solid line in Fig. 3*e*, Part I; this line lies close to the steady state geotherm appropriate to Fig. 3*f*, Part I. Whereas the thermal regime represented in Fig, 3*e*, Part I corresponds to the average surface heat flow for continental crust (section 3.2, part I), that in Fig. 3*f* corresponds to the average surface heat flow for young tectonic regions.

This example also shows that *PTt* paths, when available, permit much better discrimination than do individual *PT* points: the rocks initially buried at 60 km in Fig. 3*e*, Part I and at 40 km in Fig. 3*f* both pass through 740 ± 10 °C and 35 ± 2 km while within 50 °C of their peak temperatures, but their *PTt* paths differ radically, and rocks buried to similar depths in the different environments experience maximum temperatures that differ by 200 °C to 300 °C.

Metamorphism in both the andalusite–sillimanite and the glaucophane–jadeite facies series may be expected during continental collision and these facies alone ought not to be regarded as diagnostic of arc or subduction processes. There is extensive regional development of eclogite facies metamorphism in mafic, pelitic and carbonate composition rocks in the Sesia zone of the Italian Alps, in New Caledonia and in Norway (Compagnoni *et al.*, 1977; Black *et al.*, 1982; Bryhni *et al.*, 1977). Such metamorphism might represent burial similar to, but at higher temperatures than, say the Franciscan Gla–Jad metamorphism in California. It is also attainable in continental collision and, if exhumation is sufficiently rapid, it may be preserved without overprinting (this requirement holds equally for the preservation of high *P*–low *T* assemblages generated in subduction zones). Glaucophane–jadeite assemblages as mineral cores or inclusions in higher *T* assemblages (as found by Laird & Albee (1981) in Vermont) are consistent with the *PTt* progression expected during continental collision metamorphism, where the long time scales involved are expected to lead to later overprinting of all early high *P*–low *T* assemblages (Oxburgh & Turcotte, 1974; England & Richardson, 1977; Richardson & England, 1979).

We remarked in section 5 of Part I that Gla–Jad facies mineral assemblages ought commonly to be preserved without overprinting from metamorphism accompanying doubling in thickness of average continental crust, if the entire continental lithosphere is thickened in proportion to the crust (e.g. Part I, Fig. 5). In fact, the typical assemblages from metamorphic belts interpreted as resulting from continental collision indicate metamorphism at much higher temperature than Gla–Jad facies series. The explanation for this that we prefer is that the entire lithosphere is not normally thickened uniformly during continental collision. Whether this is because the lower lithosphere becomes unstable, and drops off (Houseman *et al.*, 1981), or whether some mechanically less well-defined process such as 'flake tectonics' (Oxburgh, 1972) is common, it is difficult to say. The *PTt* paths of Figs. 3, 4 and 6 to 8, Part I, are calculated under the assumption that the lower lithosphere is not appreciably thickened during orogeny. We should expect from these calculations that regional Gla–Jad facies series metamorphism would only be preserved without overprinting in continental collision belts if the original geothermal gradients were very low or exhumation very rapid.

Deciding between different origins for Gla–Jad facies series rocks will clearly involve the use of other indicators—oceanic associations, style of deformation, etc. The adequate characterization of exhumation rates of rocks from different tectonic environments using geochronological or kinetic means (e.g. Wagner *et al.*, 1977; Carlson & Rosenfeld, 1981) may lead to a discrimination between subduction zone and continental collision metamorphism on the basis of time scales. This certainly ought to be possible in the case of subduction zone metamorphism of sediments in the kind of environment envisaged by Ernst (1965), England & Holland (1979) and Cloos (1982) in which slices of sediment may move upward at rates of centimetres per year rather than the millimetre or less per year that characterizes erosion. Such rocks may follow *PT* paths that are controlled by the gross features of subduction zone thermal regimes, and thus show both pressure and temperature decreases ($-\Delta T/-\Delta P$) on the upward paths (Ernst, 1971).

Andalusite–sillimanite facies series metamorphism is commonly associated with island arc terrains in modern belts (Miyashiro, 1961) and the thermal requirement for this metamorphism is often taken to be addition of mantle-derived magma (e.g. Oxburgh & Turcotte, 1971 and Wells (1980), who presents a thermal analysis of an Archaean accretion–differentiation orogen). We have shown in Part I that And–Sil facies series metamorphism in continental collision metamorphism is unlikely without high initial geothermal gradients, and is not attained at all in the prograde sense ($\pm\Delta P$, $+\Delta T$) on any of the purely conductive paths illustrated. However, extensive And–Sil metamorphism is to be expected if the lower crust experiences partial melting—and this is a condition that may readily be obtained under conditions of average mantle heat flow, if the crust is thickened by a factor of 2 or more (Figs. 3, 4, 6–8, Part I; see also England & Thompson, 1984*b* and Brady, 1982). Thus, And–Sil facies series metamorphism, even with the presence of crustally-derived melts, ought not to be taken as evidence of appreciable mantle involvement in the metamorphism. It is certainly diagnostic of higher than average geothermal gradients at the time of metamorphism but such gradients can be produced in widely differing tectonic environments—it would be unwise, for example, to equate the arc vulcanism above subduction zones and the widespread Neogene vulcanism in Tibet (Burke *et al.*, 1974; Kidd, 1975), much of which is many hundreds of kilometres from any Benioff zone. In an old metamorphic belt, without the benefit of constraining seismic data, the distinction between sources for the metamorphic heat must be made on the grounds of the chemical and isotopic composition of igneous rocks in the terrains.

Within the framework outlined in this paper, the association that would appear to be diagnostic of continental-collision metamorphism is metamorphism in or near the kyanite–sillimanite facies series, observed in conjunction with segments of *PTt* paths that record decompression of several kilobars at nearly constant temperature. Metamorphism in the Kya–Sil facies series has been interpreted in the past—mainly in the Alpine chain—in terms of thickening of the continental crust in a regime of normal mantle contribution to heat flow (Clark & Jäger, 1969; Bickle *et al.*, 1975; Richardson & Powell, 1976; England, 1978; Oxburgh & England, 1980; Chen & Molnar, 1981; Brady, 1982), although alternative views are also possible (e.g. Oxburgh & Turcotte, 1974).

3.2.b. Progressive regional metamorphism

In terrains that exhibit spatially progressive metamorphism (increasing grade towards some area of the belt) there is frequently disagreement as to the cause of the increase in grade. Some authors relate this to 'thermal doming'—that is, to proximity to an augmented heat supply—while others relate it to increasing burial depth (see for example, Niggli, 1970; Wenk, 1975; Milnes, 1975 for discussion of the Lepontine thermal 'high' of the Central Alps).

In Figs. 8 and 9 we show the *PTt* evolution of two terrains in which changing metamorphic grade at the final land surface is controlled respectively by variations in burial depth and variations in heat supply. Figs. 8 and 9 are constructed by assembling side-by-side the temperature–depth–time curves for two sets of one dimensional systems. In the first case (Fig. 8) an initial thermal profile governed by Heat Supply II, with conductivity 2·25 W m^{-1} K^{-1} (see Part I, section 2) is overlain by thrusting to give an overburden varying from 50 km to zero km thickness, and erosion of this overburden takes place over the interval 20 to 120 Myr after the thrust sheets are emplaced. This approximates (by neglecting horizontal diffusion of heat) a two dimensional situation in which, for example, a nappe pile of variable thickness is emplaced on a continental basement of constant thermal regime. In the second case (Fig. 9), the depth of overthrusting remains constant at 30 km, while the basal heat supply changes laterally from 10 mW m^{-2} to 40 mW m^{-2}, all other parameters are the same as in Fig. 8. In Figs. 8 and 9 the maximum temperatures experienced by rocks that lie on the final erosion surface, as well as the times and depths at which they were reached, are labelled in the figures for 120 Myr post-thickening (at the end of erosion), as are the times and pressures at which rocks passed 50 °C below their maximum temperature. In the case of metamorphism with variable basal heat supply, peak temperatures are attained at the same times and same depths across the profile; while in the variable burial case the highest tempera-

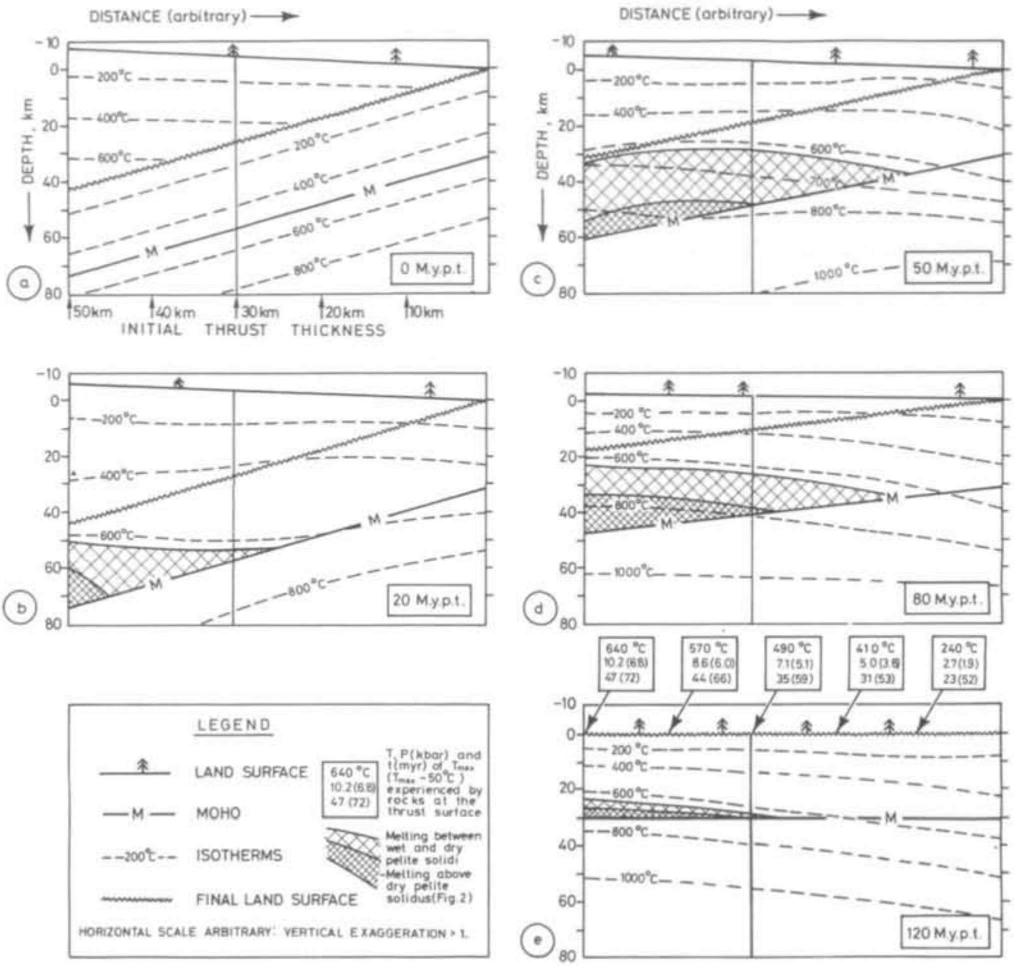


FIG. 8. Cross-section of the thermal evolution of a hypothetical terrain subject to an overthrusting event that emplaces a thrust sheet of varying thickness onto a crust of 30 km original thickness. The thermal regime before thrusting is governed by Heat Source Distribution II (Part I, subsection 3.2) and a thermal conductivity of 2·25 W m^{-1} K^{-1}; thrust thickness decreases from 50 km on the left to zero on the right. There is a 20 Myr period of isobaric heating then the thrust overburden is eroded off in the next 100 Myr (to make the initial thrust surface the final land surface). Vertical exaggeration is arbitrary, but greater than unity, so lateral conduction of heat may be neglected and, while the results are presented as a cross-section through a two dimensional field, they may equally well be applied to any other geometry of overburden, provided the slope of the thrust remains small. Isotherms and extents of partial melting fields are shown at 0 (8*a*), 20 (8*b*), 50 (8*c*), 80 (8*d*) and 120 (8*e*) Myr post thrusting (M.y.p.t.). Boxes above Fig. 8*e* show the temperature (°C), pressure (Kb) and time (M.y.p.t.) of peak metamorphic conditions (maximum temperature, T_{max}), and the pressure and time at which the rocks dropped 50 °C below peak temperature (T_{max} − 50 °C) for the rocks on the land surface at the end of erosion. See legend for further details.

ture metamorphism seen on the final land surface is both youngest and at the greatest pressure. In either case we should expect the simple picture to be perturbed by the upward motion of melts in the higher grade terrains (see above), or by differential erosion (Fowler & Nisbet, 1982).

This offers a criterion for distinguishing between hypotheses for spatially progressive metamorphism. The progressive metamorphism documented by Frey *et al.* (1974) for the Central Alps appears to fall into the category of metamorphism under an increasing load, particularly

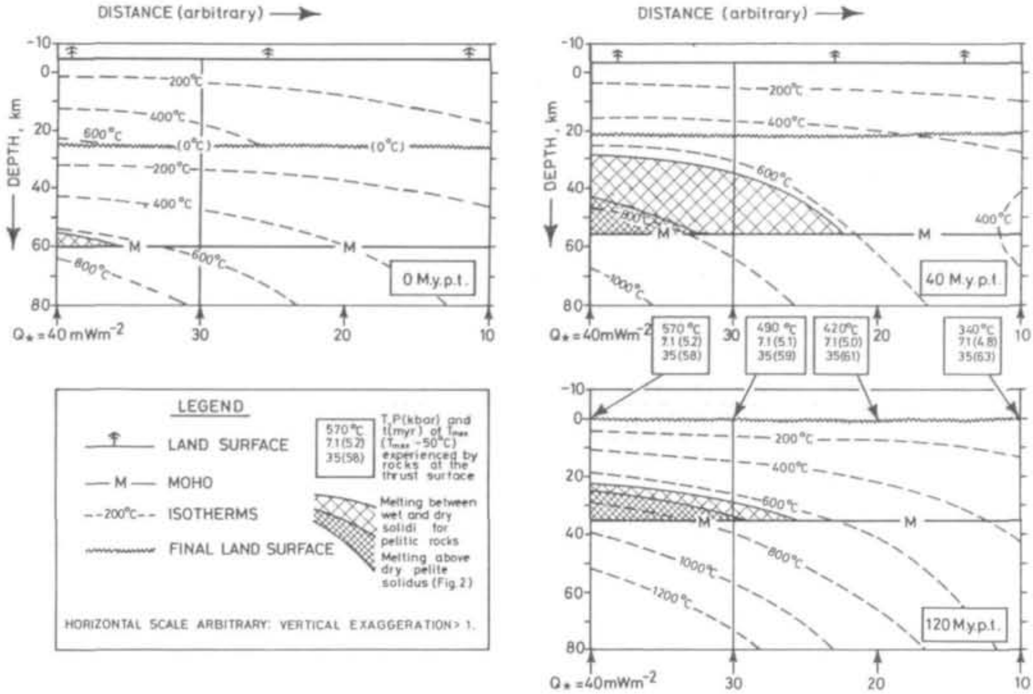


FIG. 9. As Fig. 8, except that the initial thrust thickness is a constant 30 km and it overrides a crust of 35 km; the thermal regime is as in Fig. 8, but the mantle heat flow varies from 40 mW m^{-2} on the left to 10 mW m^{-2} on the right.

when it is considered that assemblages may record conditions at pressures below the maximum temperature (section 2). The regional distribution of radiometric ages, for example in the Lepontine Alps (Wagner *et al.*, 1977) and in the Scottish Highlands (Dewey & Pankhurst, 1970; Borradaile & Hermes, 1980) can also be used to distinguish between the two mechanisms of progressive metamorphism in Figs. 8 and 9.

4. CONCLUSION

The information on metamorphic regimes that most theoretical studies are able to produce is related to large wavelength perturbations to the thermal structure of the continents, that decay on a time scale of tens of millions of years (e.g. Clark & Jäger, 1969; Oxburgh & Turcotte, 1974; Bickle *et al.*, 1975; England & Richardson, 1977; England, 1978; these papers; Royden & Hodges, 1984), and our conclusions like the rest of this paper must be read bearing in mind the points raised in section 4, Part I. In this paper we have discussed some of the many problems involved in using metamorphic petrology data to infer the thermal budgets controlling metamorphism in particular environments. To date there has been relatively little quantitative work of this kind carried out; we have tried not to overemphasize the work that has been done, because it is biased toward the study of Alpine-type belts, which are not necessarily representative of the continental collision process as a whole. They certainly are not as far as tectonic style is concerned—for example, the zone of distributed deformation accompanying the India–Asia collision far exceeds in area the belts of the Alps and the Himalaya—but whether Alpine belts are typical of continental collision zones in the degree of mantle involvement in their thermal budgets will only be known after further data become available.

Despite the somewhat negative tone introduced in the early part of this paper, where the pitfalls between the acquisition of metamorphic petrology data and their interpretation in terms of thermal budgets are discussed, we are convinced of the usefulness of data on *PTt* paths in investigating the thermal development of orogenic belts. In this regard it is encouraging that a large amount of information on *PTt* paths is becoming available from structural, geochronological, mineral zoning, cation and atomic diffusion and other studies (a fraction of which have been referenced above). Although some of the information appears at present to be consistent with the approach adopted in these papers (e.g. Tracy *et al.*, 1976; Thompson *et al.*, 1977; Holland & Richardson, 1979; Spear & Selverstone, 1983; Droop & Bucher-Nurminen, 1984), it is likely that combinations of such studies will reveal aspects of the evolution of metamorphic belts that have not been dealt with in this paper, and will lead to an improved understanding of the thermal evolution of orogenic belts and of the continental lithosphere as a whole.

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