PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A
MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

# Pressure--Temperature--Time (P--T--t) Histories of Orogenic Belts [and Discussion]

A. B. Thompson, J. R. Ridley and R. Mason

*Phil. Trans. R. Soc. Lond. A* 1987 **321**, 27-45
doi: 10.1098/rsta.1987.0003

*Phil. Trans. R. Soc. Lond.* A **321**, 27–45 (1987) 
*Printed in Great Britain*

# Pressure–temperature–time (*P–T–t*) histories of orogenic belts

By A. B. Thompson and J. R. Ridley†

*Department für Erdwissenschaften, E.T.H. Zentrum CH-8092, Zürich, Switzerland*

Thermal modelling shows that a cycle of crustal thickening and erosion reproduces many of the characteristics of medium-pressure metamorphic terranes. In contrast, the structural and metamorphic features of high-pressure terranes suggest rapid exhumation, possibly tectonically as fault-bounded blocks. Low-pressure metamorphism requires an augmented heat supply. Such terranes are characterized by granite–gneiss domes, and evidence of crustal extension, and hence may be the result of the mechanically likely orogenic sequence of early thickening followed by extension. Whether earlier isograd sequences are extended, condensed, or reset depends upon the relative rates of deformation and thermal relaxation, and when the deformation occurs relative to the thermal peak of metamorphism. Detailed determinations of relations between deformation events and metamorphism is made difficult by the contrast between continuous metamorphic evolution and short time-span deformation events. Combined microstructural and geochronological studies, together with a consideration of the distribution of isograds will give most information on complex, polymetamorphic histories, and allow distinction between regional and local features, especially those due to differential uplift.

## 1. Introduction

Modelling of the thermal development of the crust during and after crustal thickening has reinforced the traditional view that regional metamorphism and orogeny are related. A number of studies of the thermal budget of regional metamorphism have attempted to delineate the major factors that control the development of pressure ($P$) and temperature ($T$) during orogenesis (Oxburgh & Turcotte 1974; Bird *et al.* 1975; England & Richardson 1977; England & Thompson 1984).

There remains a poorer understanding of smaller scale relations between deformational events and metamorphism within an orogenic belt. Microtextural development, or the distribution of isograds in an area, will be controlled by deformation patterns on a scale smaller than considered in the above models. How do likely patterns and rates of deformation during orogenesis affect the thermal structure of an orogenic belt, and what features of the evolving thermal structure are preserved in the isograd pattern observed at the earth's surface?

Comparison of $P$–$T$ conditions within orogenic belts with possible $P$–$T$–$t$ (time) paths calculated for thermal models of regional metamorphism (Thompson and England 1984), suggest that the different facies series of metamorphism (Miyashiro 1961) may indeed be diagnostic of specific tectonic régimes (Miyashiro 1973).

The purpose of this contribution is to suggest systematics among regional structural and metamorphic features of orogenic belts. From a consideration of theoretical aspects of

† Present address: Department of Geology, University of Zimbabwe, P.O. Box MP167, Harare, Zimbabwe.

metamorphic development, suggestions are made of how combined studies of $P$–$T$–$t$ paths, microstructure and field observations of the metamorphic isograds and ages may be used in the evaluation of orogenic histories.

## 2. Modelling of metamorphism and deformation histories during orogenesis

The thermal modelling of orogenic terranes considered by England & Thompson (1984) is for a simplified deformation history in which there is a single phase of deformation giving rise to an amount of crustal thickening. The deformation was considered to be instantaneous. The isostatic adjustment mechanism was simple continuous erosion, which removed only the thickened amount, and hence large volumes of material that remained beneath the surface could be remetamorphosed during (and certainly would only be exposed by) a later event. Despite these simplifications, the single-event models present features that have application to more complex deformation histories (see figure 1).

1. For slow rates of uplift (less than 1 mm $a^{-1}$), the buried rocks experience heating, because of the greater initial contribution of radioactive heating compared with cooling due to advection (England & Richardson 1977).

2. Because of this, the maximum temperature ($T_{max}$) achieved on the $P$–$T$–$t$ path will be at a depth shallower than the amount of burial (figure 1*b*).

3. If geobarometric pressures are recorded at the depth corresponding to $T_{max}$, then samples buried to different depths will record a difference in pressure less than that equivalent to their distance apart following burial (a pressure telescoping effect, see figure 1*b*).

4. As rocks spend 30–50% of the metamorphic cycle within 50 K of $T_{max}$, this $P$–$T$ interval is the most favourable one for the equilibration of mineral assemblages, although there may be some modification during cooling.

5. Metamorphic conditions during orogenesis are continually evolving, thus $P$ and $T$ are continually changing.

These generalities apply to a single-stage event of thrusting of hot material over cold, or the homogeneous thickening of a radioactively layered crust. Different features result from more complex models, e.g. a repeated or continuous (rather than instantaneous) underthrusting of cold material (Rubie 1984; Bradbury & Nolen-Hoeksema 1985; Gillet *et al.* 1985) or two stages of thickening by thrusting (Chamberlain & England 1985). In the two-stage model, a much higher value of $T_{max}$ can be produced (although at later times and at greater depth) than in single stage models (figure 2).

Whatever the complexity of the models, real deformation histories are always more so. Any tectonic history proposed for an ancient metamorphic terrane must be consistent with there being an impetus for erosion to give exposure of once deeply buried rocks (England & Richardson 1977) or with a mechanism whereby crustal blocks are brought to the surface tectonically. In many cases, the tectonic processes themselves provide this mechanism (see below) through the activation of faults and shear zones.

Examples of the forms of possible complexities are found by considering processes in presently active belts. In the Himalayas, crustal shortening by underthrusting is currently taking place, while rocks metamorphosed at an earlier stage of the same orogeny are already exposed at the surface (see, for example, Molnar 1984). (A single orogeny is taken here to encompass all of the effects in one terrane of one set of plate motions.) Seismic observations in the Himalayas



PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

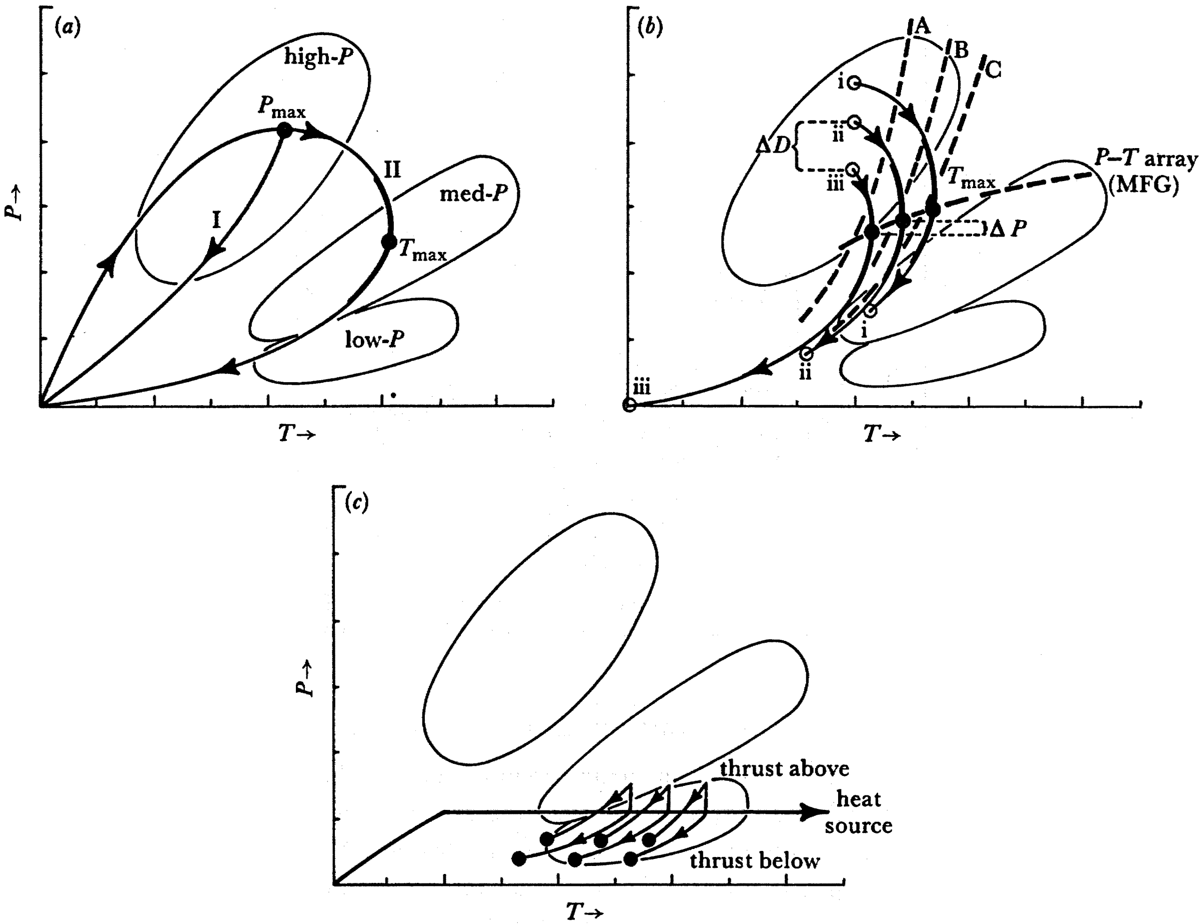



FIGURE 1. Schematic pressure temperature (*P–T*) diagrams showing the relative location of generalized facies series (taken from England & Thompson 1984, p. 903, figure 1) and some simplified single stage *P–T* histories. (*a*) Rapid burial followed by immediate and rapid uplift (path I) preserves high-*P*, low-*T* assemblages. Slower uplift (path II) would cause elimination of the high-*P* facies series and conditions in the vicinity of $T_{max}$ (medium-*P* facies series) would be recorded (see also Oxburgh & Turcotte 1974). (*b*) A single thickening event transposes rocks to three depths, i, ii and iii, in the high-*P* facies series. Slow uplift causes the paths to pass through a lower-*P* facies series where pressure telescoping occurs if geobarometric pressures each close at $T_{max}$ ($\Delta P < \Delta D$). The *P–T* array or metamorphic field geotherm (MFG) shows likely recorded values of *P* and *T* close to $T_{max}$. Different assemblages would be observed through crossing of reactions A, B, C. Only sample iii would reach the surface after erosion has removed the thickened amount. (*c*) Shallow burial and an augmented heat supply (e.g. massive magma invasion) could induce progressive prograde metamorphism. Paths resulting from placing a thrust above or below a specific burial depth are shown.

(Molnar & Deng 1984) and modelling of the evolution of the belt (Tapponier & Molnar 1976; Tapponier *et al.* 1982) suggest that uplift and erosion in the belt is, in part, a function of whether shortening can be taken up by lateral movements or not, i.e. uplift is spasmodic rather than continuous. A fuller description of the evolution of the belt is given by Burg (this symposium.) A similarly prolonged history of convergence giving rise to multiple phases of deformation is seen in the Alps, where a major phase of deformation (backfolding) immediately postdates the peak of metamorphism (Huber *et al.* 1981).

An assessment of the effect of complex deformation histories on the metamorphic evolution of a terrane requires some consideration of the relative rates of large-scale deformation and thermal relaxation.

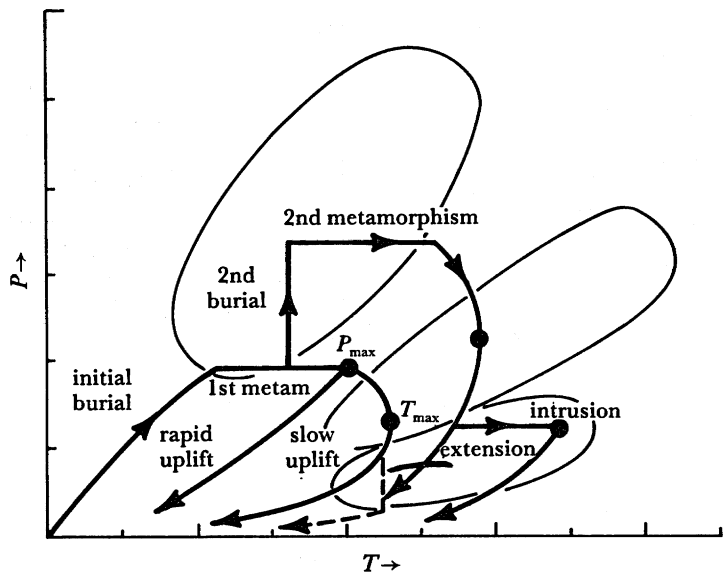



FIGURE 2. Cartoons illustrating how successive tectonic processes can perturb $P$–$T$ histories. Solid circles mark the peak metamorphic conditions for the subsequent paths.

## 3. Rates of regional deformation and rates of thermal relaxation

Any deformation will perturb the thermal structure of the crust. The amount of perturbation will depend on the rate of deformation relative to the rate of thermal relaxation. The time required for a thermal perturbation to diffuse a certain distance can be calculated from the characteristic thermal length $l^2 = \alpha\kappa t$, where $\alpha$ is a factor depending upon geometry, $\kappa$ is thermal diffusivity *ca.* $10^{-6}$ m² s⁻¹ and $t$ is time. Thus for $\alpha = 1$, the times for a perturbation to diffuse 1, 5 and 10 km are about $3 \times 10^4$, $8 \times 10^5$ and $3 \times 10^6$ a.

We are primarily interested in the vertical component of regional deformation as this describes transport of buried rocks to the surface. It we define the rate of deformation in the vertical direction in terms of the initial burial depth, $l_0$, and uplift velocity, $u$, the time needed to remove the perturbation by erosion or by extensional thinning, is $l_0/u$. This timescale can be compared with the conductive relaxation time scale discussed above ($l^2/\kappa$), and the ratio defines a Peclet number $ul^2/l_0\kappa$. Peclet numbers for erosion-controlled exhumation of tectonically thickened crust, range from between 0.05 and 2.6 for the single-stage models considered by England & Thompson (1984, p. 913). For such low values of Peclet numbers, temperature perturbations caused by the motion of the rock diffuse away quickly compared with the time scale of erosion to the surface. Conversely, fast uplift (large Peclet numbers; such as during extensional thinning of lithosphere (England 1983; Thompson & England 1984, p. 944) causes the rock to decompress nearly isothermally. The conditions for isothermal decompression in this case have been assessed by Jarvis & Mackenzie (1980).

Sleep (1979) calculates the rate at which large-scale folds should form in order that isotherms remain almost parallel to the folded surface, by using the characteristic thermal length ($\lambda^2 = 4\pi^2\kappa t$) for a sinewave. For $\lambda = 11$ and 21 km, and $\kappa = 6 \times 10^{-7}$ m² s⁻¹, this gives a maximum duration of folding of $1.6 \times 10^5$ and $5.9 \times 10^5$ a, respectively; and implies average velocities of folding of 4.3 and 3.0 cm a⁻¹, respectively.

Rates and time spans of deformation in orogenic belts are, however, poorly known (see Pfiffner & Ramsay 1982). Crustal thickening and thinning integrated over a whole orogenic belt will be related to rates of plate movements. These will be long-lasting movements: velocities of 1 cm $a^{-1}$ correspond to movements of 100 km over $10^7$ a. The deformation resulting from such plate motions will, however, be unevenly distributed. Individual phases of deformation within one part of an orogenic terrane can be of much shorter duration (see, for example, the age data and shortening distances for thrusts, summarized by Bradbury 1985, p. 132).

Pfiffner & Ramsay (1982) have estimated time spans of individual phases of deformation. They suggest that most phases last $1 \times 10^6$–$5 \times 10^6$ a, with many shorter, down to $1 \times 10^5$ a. These figures imply that deformation can produce significant perturbations of the thermal structure of an orogenic belt on a scale length shorter than the crustal thickness.

### 3.1. *Pre-, syn- and post-metamorphic deformation*

Rapid deformation may perturb the thermal structure of the crust, but whether this is subsequently preserved as deformed isograds depends on whether this deformation pre-, syn- or post-dates the metamorphic peak.

Any deformation post-dating the metamorphic peak will deform the already 'crystallized' isograd pattern, whatever the rate of deformation. Deformation pre-dating the peak of metamorphism may deform the then-existing isograd pattern, but this pattern will subsequently be overprinted during prograde evolution. It may, however, be possible to trace an early event by mineral growth–deformation patterns (Zwart 1962; Bell & Rubenach 1983).

At the peak of metamorphism, deformation will disturb the thermal structure so that, as a generalization, antiforms or uplifted blocks are cooled and synforms or downthrown blocks heated up (Sleep 1979). Isograds may become folded around antiforms and truncate synforms (Fisher 1980). As Chamberlain (1986) has demonstrated, this can result in extremely complex isograd patterns when two stages of noncoaxial folding occur.

Syn- and post-peak metamorphic deformation may give rise to different microtextural imprints. If the cooling of the upthrown block is the result of the deformation itself and juxtaposition is against cooler rocks, microtextures will show features indicating progressive deformation during progressive cooling (e.g. paracrystalline microboudinage, Misch 1969). If the deformation took place after cooling, the microtextures will indicates a distinct phase of retrogression.

### 3.2. *Relative timing of deformation and metamorphism*

Deformation on the scale considered here can affect the timing, and hence the nature of the peak of metamorphism. Figure 3 compares model *P–T–t* paths for two tectonic histories that are initially identical. One history is simply that of thickening by thrusting and subsequent erosion, the other involves, in addition, extension of the overthickened pile 30 Ma after thrusting. The modelling is for the case of extension by movement along a low-angle shear zone cutting the whole crust (figure 3*a*). In this second model, the peak of metamorphism for the rocks below the shear zone occurs immediately at the start of extensional deformation for a wide range of rates of thinning.

This is an example of a tectonic event that would be reflected by a specific deformation–porphyroblast relation (Zwart 1962). As the metamorphic peak occurs at the onset of extensional deformation, porphyroblast growth would be immediately pre-deformational, or continuing through the first increments of deformation (figure 4).

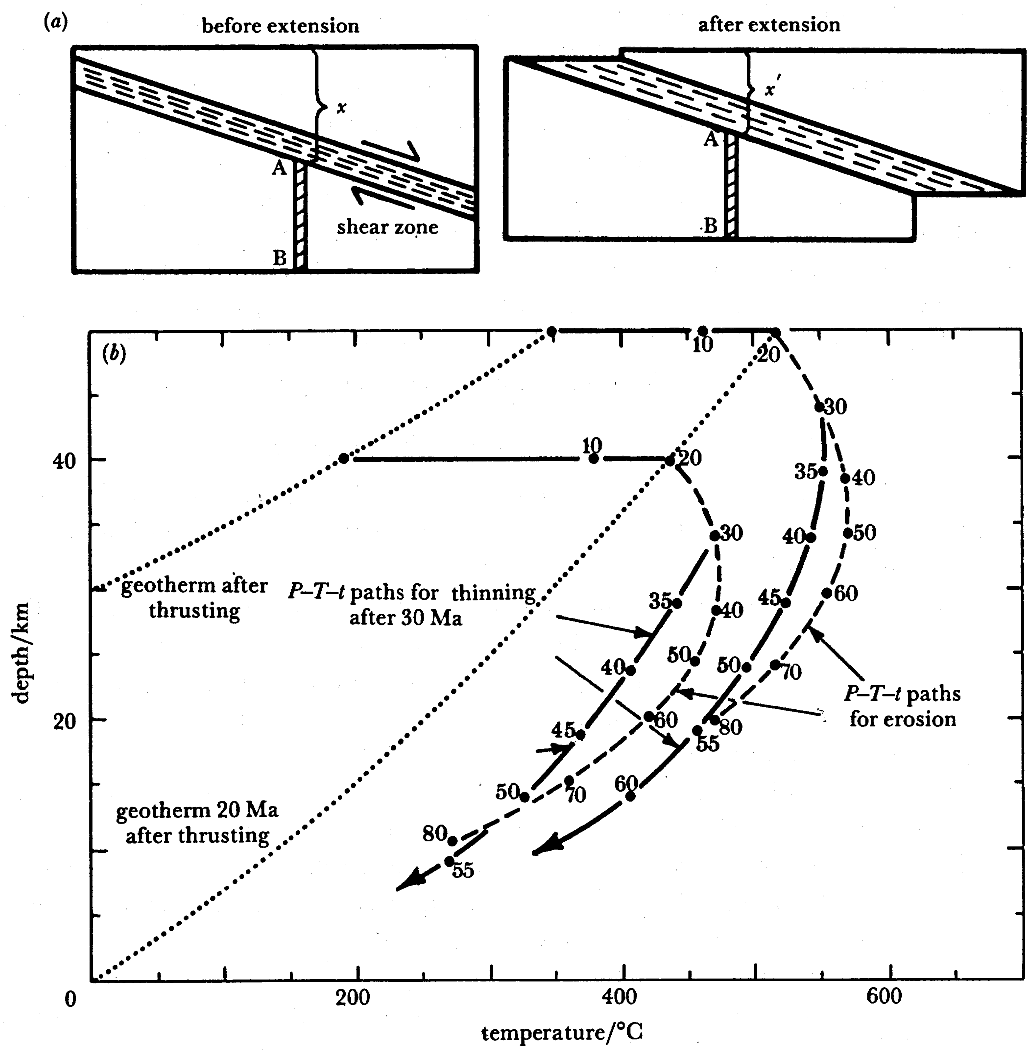



FIGURE 3. Model $P$–$T$–$t$ paths illustrating the influence of a phase of extensional deformation on metamorphic development of overthickened crust. ($a$) shows schematically the single-polarity mode for extension modelled here (after Wernicke 1985). The metamorphic development of a column A–B is traced as it becomes tectonically unroofed. The broken lines in ($b$) show a $P$–$T$–$t$ path resulting from a history of overthrusting and subsequent erosion. The solid lines of ($b$) show resulting $P$–$T$–$t$ paths for the situation with a 5 km thick shear zone at 25–30 km depth at the onset of extension. Rates of movement on the shear zone are such as to give a rate of unroofing of 1 km $Ma^{-1}$. Modelling of the case without extension follows the overthrusting model of England & Thompson (1984) with; a 30 km thick overthrust unit, mantle heat flow 29 mW $m^{-2}$, maximum radiogenic heat production 1.25 μW $m^{-3}$, exponentially decaying erosion with a time constant of 60 Ma, and moderate rock conductivity and heat capacity.

Figure 4 also shows other possible deformation–porphyroblast–growth relations resulting from different tectonic events. In general, specific temporal relationships between phases of mineral growth and deformation may be the result of one or more of the following:

(i) deformation promoting recrystallization and/or access of fluids (Brodie & Rutter 1985);

(ii) shear heating (Reitan 1968; Graham & England 1976);

(iii) tectonic burial and subsequent heating through thermal relaxation (Chamberlain & England 1985); and

(iv) juxtaposition of warm rocks against cooler, hence determining the timing of the metamorphic peak (as considered above).

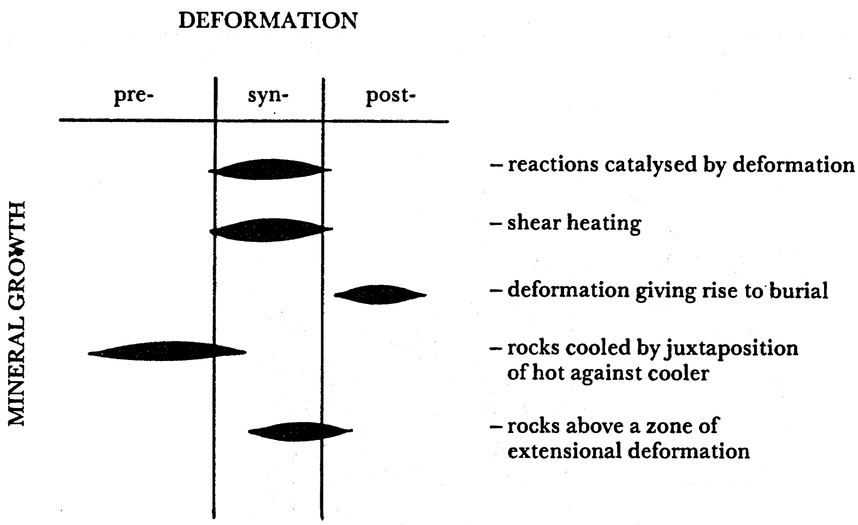



Figure 4. Patterns of the timing between mineral-growth and deformation to be expected from various geometries of deformation and from different causal relationships between deformation and metamorphism. Polyphase metamorphism and deformation will clearly render these simplified relationships ambiguous.

## 4. The determination of *P–T–t* paths for individual rock samples

Examples of segments of *P–T* histories from the literature are presented in figure 5. Those obtained by continuous reaction geobarometry–geothermometry (Spear & Selverstone 1983) rarely record temperature intervals of more than about 50 °C. A similar temperature interval is recorded by zonal arrangements of solid inclusions in garnets (Thompson *et al.* 1977). Longer retrograde and uplift path segments are preserved where a distinct retrogressive event (recorded by mineral assemblages or fluid inclusions) has not completely overprinted the peak metamorphic assemblages. It is expected that long segments of paths indicating concomitant burial and cooling could also be preserved. The studies emphasize that geobarometric–thermometric data obtained from individual samples, by themselves provide quite limited information on overall *P–T* histories.

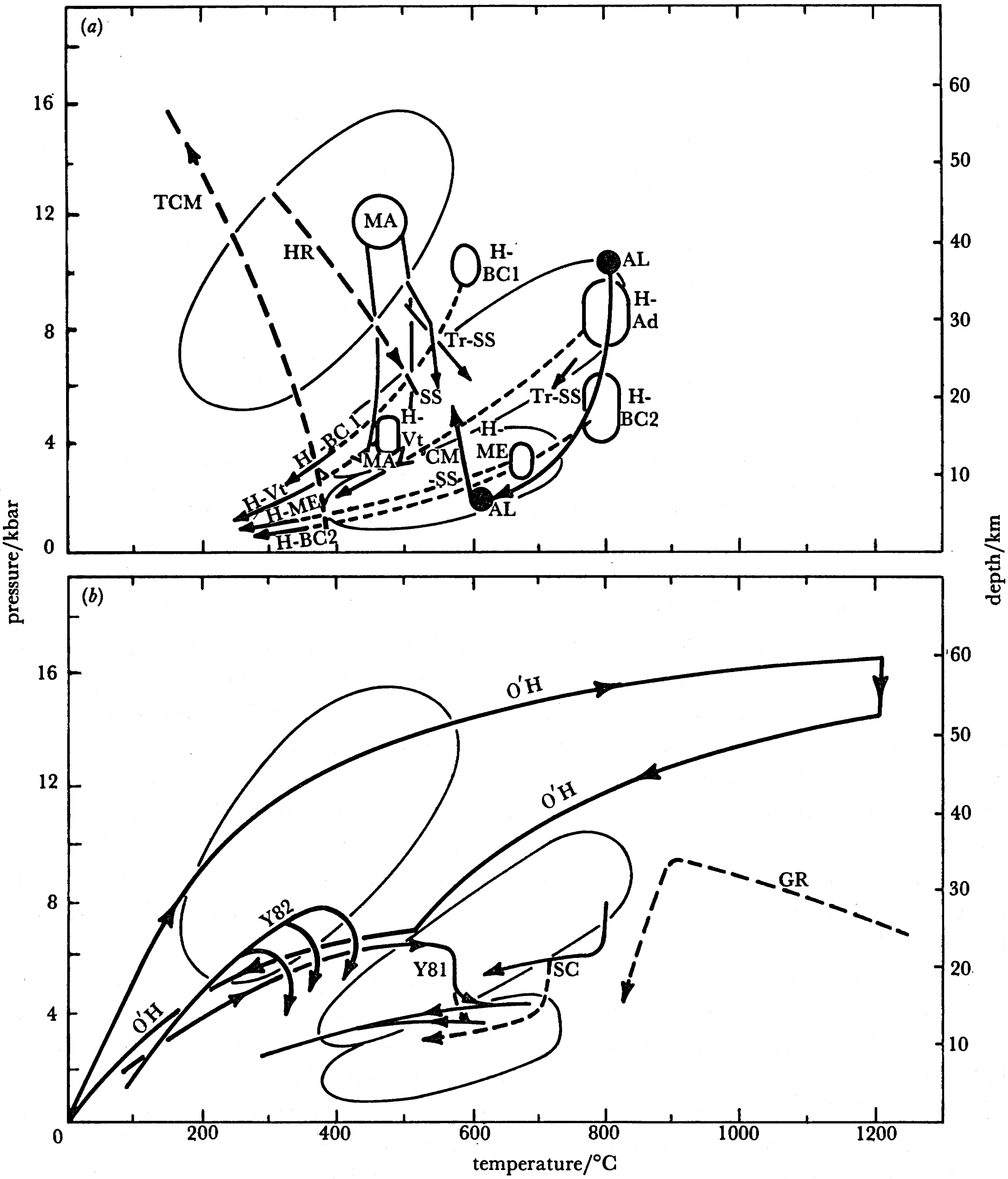
Because the closure temperatures for commonly used isotopic systems appear to be reasonably well constrained (see Dodson 1979; Cliff 1985), the ages obtained for various minerals in a given sample provide important additional information on *P–T* histories by giving an indication of time intervals and by extending the length of the deduced *P–T* path. Data from the Scottish Dalradian have been used by Dempster (1985) to demonstrate that uplift rates may have changed by an order of magnitude during exhumation. The results also constrain the duration of post-metamorphic deformation events and intervening periods of orogenic inactivity. The relative duration of prograde deformational and metamorphic histories are unfortunately not open to such detailed geochronological studies, as isotopic ages are reset up to and after the peak of metamorphism.

Even though microtextural development depends upon a number of complexly interacting kinetic factors (Ridley & Thompson 1986), the shape, orientation and distribution of solid inclusions in porphyroblasts provide the only means of deciphering prograde deformational and metamorphic histories (Bell & Rubenach 1983). Microstructural information is always incomplete. Much information is lost during textural evolution, for instance, as the matrix undergoes grain coarsening during the long time interval close to $T_{max}$ (Ridley & Thompson





PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES






PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

FIGURE 5. (*a*) Examples of deduced *P–T* paths obtained mostly by continuous reaction mineral barometry–thermometry (where $\rho = 2.8$ gm cm$^{-1}$). (CM, Crawford & Mark 1982; HR, Holland & Richardson 1979; MA, Maresch & Abraham 1981; SS, Spear & Selverstone 1983; TCM, Trzcienski *et al.* 1984; Tr, Tracy *et al.* 1976). The paths from Albarede (1976, AL) were determined from peak and retrograde assemblages (two points) and those from Hollister *et al.* (1979; H-) by comparing late fluid inclusions with peak assemblages. The Hollister data are from the Adirondacks (H-Ad), British Columbia (H-BC1, H-BC2), Maine (H-ME) and Vermont (H-Vt).

(*b*) Multistage paths deduced from continuous mineral zoning, different metamorphic assemblages in associated rocks (O'H, O'Hara 1977; GR, Griffin 1971); regional uplift histories (Y-82, Yardley 1982; SC, Schenk 1984) and late contact effects superimposed on regional metamorphism (Y–81, Yardley *et al.* 1981).

1986). Kinetic considerations further suggest that small changes in temperature (less than 10–20 K or pressure (less than 1 kbar†) may remain unrecorded.

Growth discontinuities in garnet have been used (Thompson *et al.* 1977) to suggest that completely different recorded metamorphic histories could be the result of one multistage deformation event and not separate orogenies.

## 5. Data available from metamorphic terranes

Information about tectonic events within an orogenic terrane is available from the geometry of isograds (Chinner 1966; Fox 1975), or from the distributions of recorded pressures, temperatures and ages at the earth's surface. These reflect variations in temperature during the metamorphic history, variations in the amount of uplift (P. H. Thompson 1976) or deformation. Two-dimensional modelling (e.g. Fowler & Nisbet 1982; and schematically figures 8 and 9 of Thompson & England 1984) shows the ranges of $P$, $T$ and age expected at the surface for simple tectonic histories.

### 5.1. *Isograds, isotherms and tectonic events*

Isograds are rarely parallel over great distances, and divergences of isograd spacings are common, for example, in the Lepontine Alps (Wenk 1970) and the Scottish Dalradian (Chinner 1966, 1978). Because of the likely rates of heat transfer and deformation, and the effects of continued heating after deformation (see §3.1), these variations imply deformation at or after the stage of metamorphism recorded by the isograd sequence. It has been argued above that, unless deformation is rapid, heat flow will be at such a rate that isotherms will transect fold and thrust structures, and will only in exceptional circumstances follow such structures. Whether transiently deformed isotherms are recorded in the isograd pattern depends on the timing of deformation and metamorphism.

Lithosphere thinning can condense isograd spacings and thickening extend them (Thompson 1981), although rethickening may give rise to renewed heating and hence overprinting of the isograd pattern.

If the normal orogenic sequence is one of crustal thickening followed by extension of the over-thickened pile, the 'norm' will be for local areas of condensed isograd sequences in orogenic belts. They have been recorded from a number of terranes, and in a variety of orientations; subvertical, condensed-normal and overturned.

An inverted and locally condensed metamorphic sequence is seen along the main Central Thrust of the Himalayas (Bouchez & Pecher 1981; Sinha-Roy 1981). Zones of steeply dipping condensed isograds are found along the Insubric line of the Alps and along the Highland Boundary fault of the Scottish Dalradian (Milnes 1978; Harte & Hudson 1979). A major zone of right-way-up condensed isograds is seen in the Ladakh Himalayas, partly following the Indus–Tsangpo suture (Honegger *et al.* 1982; K. Honegger & E. Herren, unpublished results). Several lines of evidence suggest that these zones are, related to post-peak-metamorphic zones of deformation.

There has been argument over the relative ages of metamorphism and thrusting along the Main Central Thrust of the Himalayas. The inverted isograd sequence has been ascribed to

† 1 bar = $10^5$ Pa.

either heat flow down from a hot overthrust slab (Le Fort 1975; Molnar *et al.* 1983; Jaupart & Provost 1985), or the tectonic disruption of an earlier normal metamorphic zonal sequence. The situation is clearer along the Moine Thrust in Scotland where a similar metamorphic geometry is seen. Here, late deformation under low-grade conditions has disrupted an earlier isograd pattern. Heating of the underlying rocks after thrusting reached only to low temperatures (Johnson *et al.* 1985; Harte *et al.* 1984).

The steep zone along the Highland Boundary fault has been related to $D_4$ (post-peak-metamorphism) folding (Chinner 1978; Watkins 1985). Along the zone of condensed isograds in the Ladakh Himalayas, there is a major phase of deformation related to a low-grade overprinting of, in part, staurolite- and sillimanite-grade rocks; although it is not clear whether the deformation, in juxtaposing warm rocks against cold, caused the decrease in temperatures.

In the Dalradian example, apparent temperature gradients from the biotite to kyanite isograds ($T \approx 250$–$300$ °C) are 50–60 K km$^{-1}$ over 5 km. Such steep gradients could only exist for short times synkinematically, if they were the result of overthrusting of hot rock over cold. Steep inverted gradients can last up to 1 Ma if there is a significant shear-heating contribution (Graham & England 1976), although for these to be preserved the whole pile would need to be cooled shortly after deformation to quench in the inverted metamorphic zonation. Moreover, although shear heating may be an important mechanism in achieving upper-amphibolite facies conditions, it has been shown by Poirier *et al.* (1979, p. 451) that it has little effect at these temperatures and higher.

There is evidence from some of the zones of condensed isograds that, in addition to having been formed late in the metamorphic history, they are zones of extensional deformation. Stretching fabrics in quartz-rich tectonites are found in the vicinity of the Himalayan Main Central Thrust (Bouchez & Pecher 1981) and along the right-way-up condensed sequence in the Ladakh Himalaya.

### 5.2. *Differential uplift of metamorphic terranes*

Differential uplift, not directly related to crustal extension, may also be a cause of deformed isograds in orogenic terranes. Relative uplift of blocks much smaller than the dimensions of the whole orogenic belt has been demonstrated and quantified by isotopic dating studies, both in presently active and in older belts. Examples have been recorded from the Scottish Dalradian (Dempster 1985; Bradbury 1985), the Lepontine Alps (Bradbury & Nolen-Hoeksema 1985), the Tauern Window (Droop 1981), the Northwestern Himalayas (Zeitler 1985), and the New England Appalachians (Sleep 1979; Chamberlain 1986). Discontinuities in $P$–$T$ values deduced from mineral barometry–thermometry have been observed at structural breaks between tectonic units (Baker 1985; Bradbury & Nolen-Hoeksema 1985), uplift having given rise to distinct thermal highs at the present level of exposure. The distribution of similar isotopic closure ages delineates the dimensions of tectonic blocks. These are generally 20–100 km, although the configuration of blocks may change during orogenesis as faults and shear zones lock or become activated.

### 5.3. *Old and recent metamorphic terranes*

Tertiary orogenic belts, including the Alps and the Himalayas, are still underlain by considerably over-thickened crust. They are undergoing rapid erosion, and the distribution of metamorphic grades and facies series within them is transient. It is in these Tertiary orogenic belts that it has been found possible to correlate individual tectonic episodes with metamorphic events, e.g. periods of rapid cooling (Bradbury & Nolen-Hoeksema 1985) or areas of rapid uplift (Wagner *et al.* 1977).

The recognition of similar events in older metamorphic terranes requires a consideration of what will be the long-term thermal effects of the types of processes observable in the presently active belts. The thermal effects of repeated or continuous underthrusting (as considered by Rubie 1984; Bradbury & Nolen-Hoeksema 1985; Chamberlain & England 1985) will be most important in currently or recently active belts. Figure 6 illustrates a schematic tectonic history involving differential uplift that will show a metamorphic imprint while the differential uplift is taking place, but in the long term will not appear as a metamorphic anomaly.

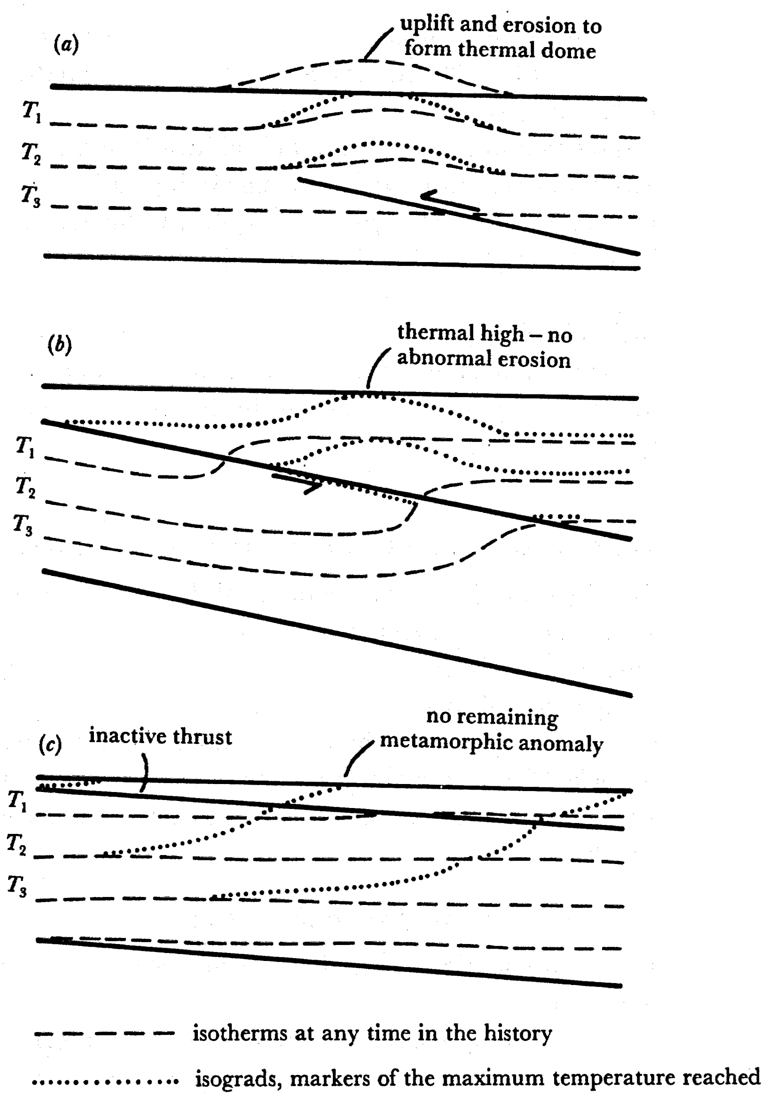



FIGURE 6. Thermal development of the crust during a schematic tectonic history giving rise to contrasting transient and long-term metamorphic imprints. Local uplift over a décollement (*a*) is followed by erosion to form a thermal dome. Regional overthrusting and crustal thickening (*b*) causes a thermal high but without abnormal erosion. After prolonged erosion (*c*), there is no evidence of the thermal dome remaining as a metamorphic anomaly.



PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

The Basin and Range Province, western U.S.A., illustrates aspects of this problem. At present, metamorphic core complexes are undergoing erosion as they form the footwalls of major normal fault systems (Armstrong 1982; Eaton 1982; Wernicke 1985). If the present tectonic activity were to cease, there would be no long-term regional tectonic anomaly to give rise to regional uplift and erosion. The high areas would remain high and the basins remain as basins. One possible older equivalent of the Basin and Range province is the Hercynian of Central Europe (P. H. Thompson & Bard 1982). Here there are blocks of a mixture of low- and medium-pressure (including granulites with eclogite and peridotite inclusions) terranes surrounded by contemporaneous and younger sedimentary basins.

There has been argument over whether the thermal and tectonic controls of metamorphism during the Archaean and Proterozoic were the same as during the Phanerozoic (Thompson *et al.* 1984; Richter 1985). Granulite facies terranes in stable Archaean cratons represent the end product of exhumation of lower crustal material. They appear to be underlain by continental crust of 'normal' thickness (30–40 km), and indicate $P$–$T$ conditions of 7–10 kbar at 700–850 °C (Perkins & Newton, 1981).

The presence of considerable volumes of igneous material in such terranes has been used as an argument that igneous activity was an important factor in increasing the heat supply and in giving rise to uplift and erosion (O'Hara 1977; Wells 1980). If, however, the erosion is the result of tectonic thickening of the crust, this implies that similar processes of tectonic thickening occurred during the Archaean as in more recent orogenic activity (England & Bickle 1984) although Thompson & England (1984) suggest that the $P$–$T$ conditions recorded by granulites are unlikely to be exhumed by erosion following a single crustal thickening event. It is therefore suggested that differential tectonic uplift was important in the emplacement of these terranes to high levels in the crust.

## 6. Metamorphic facies series and tectonic environments

Since the classification by Miyashiro (1961) of metamorphic facies series, a number of authors have attempted to show correlations between tectonic régimes and facies series and rock associations. Miyashiro's classification is followed here, although with the recognition that most orogenic belts are inadequately described as belonging to one facies series.

The following sections summarize some of the deformational, microstructural and metamorphic characteristics of rocks of each metamorphic facies series, to suggest how they may relate to different tectonic environments. Emphasis is placed on the features of recently or currently active orogenic belts.

### 6.1. *Medium–pressure facies series*

The major metamorphic features predicted for a single-cycle thickening and erosion orogenic history have been listed above (§2, and see also Thompson & England 1984). Figure 5 shows that the metamorphic histories of many medium-pressure terranes conform to the predicted pattern. Most importantly, there is often an increase in temperature during the initial stages of uplift.

Many medium-pressure terranes cover large areas, and show an intact, although deformed, sequence of isograd zones, for example the Dalradian of Scotland (Chinner 1966, 1978) and New England (J. B. Thompson *et al.* 1968; Dixon & Lundgren 1968). In the medium- to higher-grade areas of these terranes, the peak of metamorphism (as marked by porphyroblast growth) post-dates one or two phases of penetrative high-strain deformation (Harte & Johnson 1969; Harte *et al.* 1984). In both the Dalradian (Harris *et al.* 1976) and the Alps (Bradbury & Nolen-Hoeksema 1985) there is evidence from the deformation style that the early deformation took place at lower metamorphic grades. It is tempting to relate this deformation to an earlier phase of tectonic thickening.

In many medium-pressure terranes there are phases of deformation of moderate or high total strains post-dating the peak of metamorphism. It is these phases of deformation that appear to be responsible for the presently observed spacing and three-dimensional pattern of isograds. In the Lepontine Alps, such a phase of deformation post-dates porphyroblast growth, but does not appear to be associated with any retrograde reactions (Chadwick 1968; Klaper 1982). This deformation may have caused local redistribution of heat, but was not of such intensity as to cause renewed burial and overall heating of the tectonic pile.

Seismic observations of continental collision zones, for example in the Himalayas (Molnar & Deng 1984), suggests that the take-up of shortening in continental collision will rarely be by simple thickening. Also, thickening may be followed by, or be interspersed with, episodes of lateral movement.

### 6.2. *High-pressure facies series*

The classical view of high-pressure metamorphism is that it is related to metamorphism in a subduction zone (Miyashiro 1973). Thermal modelling studies, however, have shown that the relevant $P$–$T$ conditions can also be produced in continental collision zones (Oxburgh & Turcotte 1974; England & Richardson 1977; England & Thompson 1984), although evidence of this is likely to be destroyed due to heating of the pile during uplift (see Laird & Albee 1981). This observation has led to the idea that the presence of high-pressure (as opposed to medium-pressure) rocks at the surface is related to features of the uplift history (Draper & Bone 1981; Rubie 1984).

In contrast to medium-pressure terranes, high-strain deformation and the metamorphic peak often appear to be synchronous in blueschists and eclogites (Gosso 1977; Ridley 1984; Bell & Brothers 1985). High-strain deformation may also occur during uplift and be associated with a distinct phase of metamorphic recrystallization (Gosso 1977).

Zonal sequences show various forms and degrees of preservation in blueschist terranes. The classical 'circum-Pacific' blueschist terranes show inverted zonal sequences. In New Caledonia, the sequence appears right-way-up (Brothers 1974) and condensed in the sense that between the bottom and top of the exposed pile (a few hundred metres of section) the assemblages indicate differences in metamorphic pressures greater than 2 kbar (Bell & Brothers 1985).

Where high-pressure metamorphism is seen as an early stage of regional metamorphism in a continental collision zone (Alps, Eastern Mediterranean) any original zonal sequences have been disrupted.

The presence of coesite relics in garnet inclusions in the Dora Maira massif of the Alps (Chopin 1984) not only shows that uplift has taken place from greater than 90 km depth, but also suggests that uplift was much faster than would have been caused by erosional unroofing (Gillet *et al.* 1984). Block faulting and tectonic uplift of blocks within a large-scale fault zone may have been responsible for part of this uplift.



PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

### 6.3. *Lower-pressure facies series*

The frequent occurrence of intruded granites and remobilized basement-gneisses as domes in these terranes has led to the concept of regional-scale contact metamorphism, for example, in Maine (Ferry 1980). The raised heat flow has generally been assigned to magmatic injection from the mantle into the lower crust.

England & Thompson (1984, 1986) have pointed out that in cases where there is no lower-crustal magma addition, the geothermal gradients during And–Sil–facies series metamorphism will be such that the lower continental crust will have passed above the likely solidus. This implies that low-pressure metamorphism and remobilization of the lower crust would be expected to occur together. Convective movement of magmas would locally accentuate high geothermal gradients.

Work in the Basin and Range Province has suggested that the primary cause of high geothermal gradients is crustal extension rather than magma injection (see, for example, Lachenbruch & Sass 1978). Because it is unlikely, in high-grade metamorphic rocks, to be able to distinguish the prograde path, raising of the base of the lithosphere during thinning would leave a similar thermal signature as intrusion of magmas into the base of the crust.

The metamorphic core complexes of the western U.S.A. form a contemporary low-pressure terrane (Armstrong 1982). The Hercynian metamorphism of the Pyrenees is an older example (Soula 1982). Here, high geothermal gradients and diapiric intrusion appear related. The intrusion has given rise to a great spatial variability in deformation style and deformation–porphyroblast relations (Soula 1982). The structures above a diapir are consistent with extensional tectonics.

Lithosphere thinning, such as presently taking place in western U.S.A., will not produce a tectonic anomaly that would induce deep erosion. If thinning is the cause of low-pressure metamorphism, one must invoke special circumstances to explain the uplift. In the Pyrenees, low-pressure metamorphism is Hercynian, while uplift is related to Alpine compressional deformation (Lamouroux *et al.* 1980). Elsewhere, uplift may be of crustal blocks smaller than the size of the whole orogenic belt. If this is the case possible uplift mechanisms include diapir-induced buoyancy, buoyant underplating by subducted continental crust, uplift of blocks between major faults or of blocks attached to adjacent zones of thickened crust. None of these mechanisms would give rise to regional uplift.

## 7. Summary

Considerations of rates of heat flow within the crust indicate how isotherms evolve in response to tectonic events, and how isograd distribution will relate to local and regional structures. Important controls are the relative rates of deformation and thermal relaxation, and whether the deformation predates, is synchronous with, or postdates the metamorphic peak.

Only tectonic events close to or after the metamorphic peak, result in deformed isograds; at the peak, in areas where deformation induces local cooling; after the peak isograd sequences may be folded and expanded or condensed by crustal thickening or thinning. As repeated thickening will generally give rise to renewed heating (Chamberlain & England 1985), the presence of condensed isograd sequences in metamorphic terranes is suggested to be the norm (see also Harte & Dempster, this symposium).

Information on orogenic development during the cooling and uplift stage of metamorphism, obtained by combined geochronological, metamorphic and structural studies, suggests that uplift in a mountain belt is generally very unevenly distributed.

The only indications of deformation events before the peak of metamorphism, and of how these may have transiently affected the thermal structure of the orogenic belt, will come from combined microtextural and structural studies. Microtextural studies may also be important in distinguishing between deformation events at the peak of metamorphism and those significantly post-dating it.

Parameters such as the size and distribution of successive porphyroblasts, and the size of grains in a rock matrix are functions, through reaction kinetics, of heating rates, deformation, and fluid involvement. At present, the quantitative nature of the relations is poorly understood. Useful information may, however, be obtainable by measuring the variation of any parameter across a metamorphic terrane (see Ridley 1986).

Contrasting microtextural histories, and contrasting patterns of isograd distributions in terranes showing the various facies of metamorphism, suggest indeed that the facies series are diagnostic of specific tectonic régimes. Only metamorphism in the kyanite–sillimanite facies series can be adequately modelled as the thermal result of simply crustal thickening and subsequent erosion. Both lower-pressure and higher-pressure metamorphism require a different or more complex history. It is suggested that crustal extension is important in low-pressure metamorphic belts. The presence of igneous rocks in such an environment may be the result of extension rather than the cause of the high heat flow. There may be various causes of high-pressure metamorphism, but the consistent pattern of an end to metamorphic recrystallization at the end of a major deformation event suggests that the rocks are prevented from further heating up. Tectonic uplift as fault-bounded blocks is considered to be important.

We thank M. Casey, D. C. Rubie, N. Mancktelow and B. W. D. Yardley for discussion and P. C. England for review; L. Dietrich for typing and B. Buehlmann for drafting. J. R. R. thanks the Royal Society for support from a European Programme Research Fellowship.

## References

Albarede, F. 1976 Thermal models of post-tectonic decompression as exemplified by the Haut-Allier granulites (Massif Central, France). *Bull. Soc. géol. Fr.* **17**, 1023–1032.
Armstrong, R. L. 1982 Cordilleran Metamorphic core complexes from Arizona to Southern Canada. *A. Rev. Earth planet. Sci.* **10**, 129–154.
Baker, A. J. 1985 Pressures and temperatures of metamorphism in the eastern Dalradian. *J. geol. Soc. Lond.* **142**, 137–148.
Bell, T. H. & Rubenach, M. J. 1983 Sequential porphyroblast growth and crenulation cleavage development during progressive deformation. *Tectonophysics* **92**, 171–194.
Bell, T. H. & Brothers, R. N. 1985 Development of *P–T* prograde and *P*-retrograde isogradic surfaces during blueschist to eclogite regional deformation/metamorphism in New Caledonia, as indicated by progressively developed porphyroblast microstructures. *J. metamorph. Geol.* **3**, 59–78.
Bird, P., Toksoz, M. N., & Sleep, N. H. 1975 Thermal and mechanical models of continent–continent collision zones. *J. geophys. Res.* **80**, 4405–4416.
Bouchez, J. L. & Pecher, A. 1981 The Himalayan Main Central thrust pile and its quartz-rich tectonites in central Nepal. *Tectonophysics* **78**, 23–50.
Bradbury, H. J. 1985 The Caledonian metamorphic core; an Alpine model. *J. geol. Soc. Lond.* **142**, 129–136.
Bradbury, H. J. & Nolen-Hoeksema, R. C. 1985 The Lepontine Alps as an evolving metamorphic core complex during A-type subduction: evidence from heat-flow, mineral cooling ages, and tectonic modeling. *Tectonics* **4**, 187–211.

Brodie, K. H. & Rutter, E. H. 1985 On the relationship between deformation and metamorphism with special reference to the behaviour of basic rocks. In *Metamorphic reactions* (ed. A. B. Thompson & D. C. Rubie), pp. 138–179. New York: Springer.

Brothers, R. N. 1974 High-pressure schists in northern New Caledonia. *Contr. Miner. Petr.* **46**, 216–229.

Chadwick, B. 1968 Deformation and metamorphism in the Lukmanier region, central Switzerland. *Bull. geol. Soc. Am.* **79**, 1123–1150.

Chamberlain, C. P. 1986 Evidence for the repeated folding of isotherms during regional metamorphism. *J. Petr.* **27**, 63–89.

Chamberlain, C. P. & England, P. C. 1985 The Acadian thermal history of the Merrimack Synclinorium in New Hampshire. *J. Geol.* **93**, 593–602.

Chinner, G. A. 1966 The distribution of pressure and temperature during Dalradian metamorphism. *J. geol. Soc. Lond.* **122**, 159–186.

Chinner, G. A. 1978 Metamorphic zones and fault displacement in the Scottish Highlands. *Geol. Mag.* **115**, 37–45.

Chopin, C. 1984 Coesite and pure pyrope in high-grade blueschists of the Western Alps: A first record and some consequences. *Contr. Miner. Petr.* **86**, 67–118.

Cliff, R. A. 1985 Isotopic dating in metamorphic belts. *J. geol. Soc. Lond.* **142**, 97–110.

Crawford, M. L. & Mark, L. E. 1982 Evidence from metamorphic rocks for overthrusting, Pennsylvania Piedmont, U.S.A. *Can. Mineral.* **20**, 330–347.

Dempster, T. J. 1985 Uplift patterns and orogenic evolution in the Scottish Dalradian. *J. geol. Soc. Lond.* **142**, 111–128.

Dixon, H. R. & Lundgren, L. W. 1968 Structure of Eastern Connecticut. In *Studies of Appalachian geology: northern and maritime* (ed. E.-An. Zen), pp. 219–229. New York: Wiley Interscience.

Dodson, M. H. 1979 Theory of cooling ages. In *Lectures in isotope geology* (ed. E. Jaeger & J. C. Hunziker), pp. 194–202. Berlin: Springer Verlag.

Draper, G. & Bone, R. 1981 Denudation rates, thermal evolution and preservation of blueschist terrains. *J. Geol.* **89**, 601–613.

Droop, G. T. R. 1981 Alpine metamorphism of pelitic schists in the south-east Tauern Window, Austria. *Schweiz. miner. petrogr. Mitt.* **61**, 237–274.

Eaton, G. P. 1982 The Basin and Range Province: Origin and tectonic significance. *A. Rev. Earth planet. Sci.* **10**, 409–440.

England, P. C. 1983 Constraints on the extension of continental lithosphere. *J. geophys. Res.* **88**, 1145–1152.

England, P. C. & Bickle, M. 1984 Continental thermal and tectonic regimes during the Archaen. *J. Geol.* **92**, 353–367.

England, P. C. & Richardson, S. W. 1977 The influence of erosion upon mineral facies of rocks from different metamorphic environments. *J. geol. Soc. Lond.* **134**, 201–213.

England, P. C. & Thompson, A. B. 1984 Pressure-temperature-time paths of regional metamorphism I. Heat transfer during the evolution of regions of thickened continental crust. *J. Petr.* **25**, 894–928.

England, P. C. & Thompson, A. B. 1986 Some thermal and tectonic models for crustal melting in continental collision zones. In *Collision tectonics* (*Geol. Soc. Lond., Spec. Publ.* no. 19) (ed. M. P. Coward & A. C. Ries), pp. 83–94.

Ferry, J. M. 1980 A case study of the amount and distribution of heat and fluid during metamorphism. *Contr. Miner. Petr.* **71**, 373–385.

Fisher, G. W. 1980 Isograd migration in response to heat transfer during Acadian folding, eastern Vermont. *Abstr. Geol. Soc. Am.* **12**, 426.

Fowler, C. M. R. & Nisbet, E. G. 1982 The thermal background to metamorphism: simple two-dimensional conductive models. *Geosci. Can.* **9**, 208–214.

Fox, J. S. 1975 Three-dimensional isograds from the Lukmanier Pass, Switzerland, and their tectonic significance. *Geol. Mag.* **112**, 547–564.

Gillet, P. Davy, P. Ballevre, M. & Choukroune, P. 1985 Thermomechanical evolution of a collision zone: the example of the Western Alps. *Terra Cognita* **5**, 399–404.

Gillet, P. Ingrin, J. & Chopin, C. 1984 Coesite in subducted continental crust; P–T history deduced from an elastic model. *Earth planet. Sci. Lett.* **70**, 426–436.

Gosso, G. 1977 Metamorphic evolution and fold history in the eclogitic micaschists of the upper Gressoney Valley (Sesia-Lanzo zone, western Alps. *Rc. Soc. ital. Miner. Petr.* **33**, 389–407.

Graham C. M. & England, P. C. 1976 Thermal regimes and regional metamorphism in the vicinity of overthrust faults: an example of shear heating and inverted metamorphic zonation from Southern California. *Earth planet. Sci. Lett.* **31**, 142–152.

Griffin, W. L. 1971 Genesis of coronas in anorthosites of the upper Jotun Nappe, Indre Sogn, Norway. *J. Petr.* **12**, 219–243.

Harris, A. L., Bradbury, H. J. & MacGonigal, M. H. 1976 The evolution and transport of the Tay Nappe, Scot. *J. Geol.* **12**, 103–113.

Harte, B., Booth, J. G., Dempster, T. J., Fettes, D. J., Mendum, J. R. & Watts, D. 1984 Aspects of the post-depositional evolution of Dalradian and Highland Border Complex in the southern Highlands of Scotland. *Trans. R. Soc. Edinb.* **75**, 151–163.

Harte, B. & Hudson, F. C. 1979 Pelite facies series and the temperatures and pressures of Dalradian metamorphism in E. Scotland. In *The Caledonides of the British Isles – Reviewed* (*Geol. Soc. Lond., spec. Publ.* no. 8) (ed. A. L. Harris, C. H. Holland & B. E. Leake), pp. 323–337.

Harte, B. & Johnson, M. R. W. 1969 Metamorphic history of Dalradian rocks in Glens Clova, Esk and Lethnot, Angus, Scotland. *Scott. J. Geol.* **5**, 54–80.

Holland, J. B. & Richardson, S. W. 1979 Amphibole zonation in metabasites as a guide to the evolution of metamorphic conditions. *Contr. Miner. Petr.* **70**, 143–148.

Hollister, L. S., Burruss, R. C., Henry, D. L. & Hendel, E. M. 1979 Physical conditions during uplift of metamorphic terrains, as recorded by fluid inclusions. *Bull. Mineral.* **102**, 555–561.

Honegger, K., Dietrich. V., Frank, W., Gansser, A., Thoeni, M & Trommsdorff, V. 1982 Magmatism and metamorphism in the Ladakh Himalayas (the Indus-Tsangpo Suture Zone). *Earth planet. Sci. Lett.* **60**, 253–292.

Huber, M., Ramsay, J. & Simpson, C. 1981 Deformation in the Maggia and Antigorio nappes, Lepontine Alps. *Eclog. geol. Helv.* **73**, 593–606.

Jarvis, G. T. & McKenzie, D. P. 1980 Sedimentary basin formation with finite extension rates. *Earth planet. Sci. Letts.* **48**, 42–52.

Jaupart, C. & Provost, A. 1985 Heat focussing, granite genesis and inverted metamorphic gradients in continental collision zones. *Earth planet. Sci. Lett.* **73**, 385–397.

Johnson, M. R. W., Kelly, S. P., Oliver, G. J. H. & Winter, D. A. 1985 Thermal effects and timing of thrusting in the Moine Thrust zone. *J. geol. Soc. Lond.* **142**, 863–874.

Klaper, E. M. 1982 Deformation und Metamorphose in der noerdlichen Maggia Zone. *Schweiz. miner. petrogr. Mitt.* **62**, 47–76.

Lachenbruch, A. & Sass, J. H. 1978 Models of an extending lithosphere and heat flow in the Basin and Range Province. *Mem. geol. Soc. Am.* **152**, 209–250.

Laird, J. & Albee, A. L. 1981 High pressure metamorphism in mafic schist from northern Vermont. *Am. J. Sci.* **281**, 97–126.

Lamouroux, C., Soula, J. C., Deramond, J. & Debat, P. 1980 Shear zones in the granodiorite massifs of the Central Pyrenees and the behaviour of these massifs during Alpine orogenesis. *J. struct. Geol.* **2**, 49–53.

Le Fort, P. 1975 Himalayas: The collided range. Present knowledge of the continental arc. *Am. J. Sci.* (Rodgers vol.) **275**-A, 1–44.

Maresch, W. V. & Abraham, K. 1981 Petrography, mineralogy, and metamorphic evolution of an eclogite from the island of Margarita, Venezuela. *J. Petr.* **22**, 337–362.

Milnes, A. G. 1978 Structural zones and continental collision, central Alps. *Tectonophysics* **78**, 369–392.

Misch, P. 1969 Paracrystalline microboudinage of zoned grains and other criteria for synkinematic growth of metamorphic minerals. *Am. J. Sci.* **267**, 43–63.

Miyashiro, A. 1961 Evolution of metamorphic belts. *J. Petr.* **2**, 277–311.

Miyashiro, A. 1973 *Metamorphism and metamorphic belts.* (492 pages.) New York: Halsted Press, Wiley and Sons.

Molnar, P. 1984 Structure and tectonics of the Himalaya: Constraints and implications of geophysical data. *A. Rev. Earth planet. Sci.* **12**, 489–518.

Molnar, P., Chen, W. P. & Padovani, E. 1983 Calculated temperatures in overthrust terrains and possible combinations of heat sources responsible for the Tertiary granites in the Greater Himalaya. *J. geophys. Res.* **88**, 6415–6429.

Molnar, P. & Deng, Q. 1984 Faulting associated with large earthquakes and the average rate of deformation in central and eastern Asia. *J. geophys. Res.* **89**, 6203–6228.

O'Hara, M. J. 1977 Thermal history of excavation of Archaean gneisses from the base of the continental crust. *J. geol. Soc. Lond.* **134**, 185–200.

Oxburgh, E. R. & Turcotte, D. L. 1974 Thermal gradients and regional metamorphism in overthrust terrains with special reference to the Eastern Alps. *Schweiz. miner. petrogr. Mitt.* **54**, 641–662.

Perkins, D. & Newton, R. C. 1981 Charnocktite geobarometers based on coexisting garnet–pyroxene–plagioclase–quartz. *Nature, Lond.* **292**, 144–146.

Pfiffner, O. A. & Ramsay, J. G. 1982 Constraints on geological strain rates: Arguments from finite strain states of naturally deformed rocks. *J. geophys. Res.* **87**, 311–321.

Poirier, J. P., Bouchez, J. L. & Jonas, J. J. 1979 A dynamic model for aseismic ductile shear zones. *Earth planet. Sci. Lett.* **43**, 441–453.

Reitan, P. H. 1968 Frictional heat during metamorphism quantitative evaluation of concentration of heat generation in time. *Lithos* **1**, 151–163.

Richter, F. M. 1985 Models for the Archean thermal regime. *Earth planet. Sci. Lett.* **73**, 350–362.

Ridley, J. 1984 The significance of deformation associated with blueschist facies metamorphism on the Aegean island of Syros. In *The geological evolution of the eastern Mediterranean* (*Geol. Soc. Lond., spec. Publ.* no. 17) (ed. N. E. Dixon & A. H. F. Robertson), pp. 545–551.

Ridley, J. 1986 Modelling of the relations between reaction enthalpy and the buffering of reaction progress in metamorphism. *Mineralog. Mag.* (In the press.)

Ridley, J. & Thompson, A. B. 1986 The role of mineral kinetics in the development of metamorphic microstructures. In *Fluid-rock interactions during metamorphism.* (ed. J. V. Walther & B. J. Wood), pp. 154–193. Berlin: Springer Verlag.

Rubie, D. C. 1984 A thermal-tectonic model for high-pressure metamorphism and deformation in the Sesia Zone, Western Alps. *J. Geol.* **92**, 21–36.

Schenk, V. 1984 Petrology of felsic granulites, metapelites, metabasics, ultramafics, and metacarbonates from Southern Calabria (Italy): prograde metamorphism, uplift and cooling of a former lower crust. *J. Petr.* **25**, 255–298.

Sinha-Roy, S. 1981 Metamorphic facies and inverted metamorphic sequences of the Eastern Himalayas crystalline rocks. In *Metamorphic tectonites of the Himalaya*, (ed. P. S. Saklani), pp. 279–302. New Delhi: Today and Tomorrow's Printers.

Sleep, N. H. 1979 A thermal constraint on the duration of folding with reference to Acadian geology, New England, U.S.A. *J. Geol.* **87**, 583–589.

Soula, J. C. 1982 Characteristics and mode of emplacement of gneiss domes and plutonic domes in central-eastern Pyrenees. *J. struct. Geol.* **4**, 313–342.

Spear, F. S. & Selverstone, J. 1983 Qualitative P–T paths from zoned minerals: Theory and tectonic applications. *Contr. Miner. Petr.* **83**, 348–357.

Tapponier, P. & Molnar, P. 1976 Slip-time field theory and large-scale continental tectonics. *Nature, Lond.* **264**, 319–324.

Tapponier, P., Peltzer, G., Le Dain, A. Y. Armigo, R. & Cobbold, P. 1982 Propagating extrusion tectonics in Asia, new insights from simple experiments with plasticine. *Geology* **10**, 611–616.

Thompson, A. B. 1981 The pressure-temperature (P,T) plane viewed by geophysicists and petrologists. *Terra Cognita* (spec. issue), pp. 11–20.

Thompson, A. B. & England, P. C. 1984 Pressure-temperature-time paths of regional metamorphism. II. Their inference and interpretation using mineral assemblages in metamorphic rocks. *J. Petr.* **25**, 929–955.

Thompson, A. B., Richter, F. M., Ahrendt, H., Bickle, M. J., Burke, K. C., Dymek, R. F., Frisch, W., Gee, R. D., Kroener, A., O'Nions, R. K., Oxburgh, E. R. & Weber, K. 1984 The long term evolution of the crust and mantle. In *Patterns of change in Earth evolution.* (ed. H. Holland & A. F. Trendall), pp. 389–406. Berlin: Springer Verlag.

Thompson, A. B., Tracy, R. J., Lyttle, P. T. & Thompson, J. B. 1977 Prograde reaction histories deduced from compositional zonation and mineral inclusions in garnets from the Gassetts Schist, Vermont. *Am. J. Sci.* **277**, 1152–1167.

Thompson, J. B., Robinson, P., Clifford, T. N. & Trask, N. J. 1968 Nappes and gneiss Domes in West-central New England. In *Studies of Appalachian geology: northern and maritime*, (ed. E.-An. Zen), pp. 203–217. New York: Wiley Interscience.

Thompson, P. H. 1976 Isograd patterns and pressure-temperature distributions during regional metamorphism. *Contr. Miner. Petr.* **57**, 277–295.

Thompson, P. H. & Bard, J.-P. 1982 Isograds and mineral assemblages in the eastern axial zones, Montagne Noire (France): implications for temperature gradients and P–T history. *Can. J. Earth Sci.* **19**, 129–143.

Tracy, R. J., Robinson, P. & Thompson, A. B. 1976 Garnet composition and zoning in the determination of temperature and pressure of metamorphism, Central Massachusetts. *Am. Miner.* **61**, 762–775.

Trzcienski, W. E., Carmichael, D. M. & Helmstaedt, H. 1984 Zoned sodic amphibole: petrologic indicator of changing pressure and temperature during tectonism in the Bathurst area, New Brunswick, Canada. *Contr. Miner. Petr.* **85**, 311–320.

Wagner, G. A., Reimer, G. M. & Jaeger, E. 1977 Cooling ages derived by apatite fission-track, mica Rb–Sr and K–Ar dating: The uplift and cooling history of the Central Alps. *Memorie. Ist. geol. miner. Univ. Padova* **30**, 3–27.

Watkins, K. P. 1985 Geothermometry and geobarometry of inverted metamorphic zones in the W. Central Scottish Dalradian. *J. geol. Soc. Lond.* **142**, 157–165.

Wells, P. R. A. 1980 Thermal models for the magmatic accretion and subsequent metamorphism of continental crust. *Earth planet. Sci. Lett.* **46**, 253–265.

Wenk, E. 1970 Zur Regionalmetamorphose und Ultrametamorphose im Lepontin. *Fortschr. Miner.* **47**, 34–51.

Wernicke, B. 1985 Uniform-sense normal simple shear of the continental lithosphere. *Can. J. Earth Sci.* **22**, 108–125.

Yardley, B. W. D. 1982 The early metamorphic history of the Haast Schists and related rocks of New Zealand. *Contr. Miner. Petr.* **81**, 317–327.

Yardley, B. W. D., Shepherd, T. J. & Barber, J. P. 1981 Fluid inclusion studies of high-grade rocks from Connemara, Ireland. In *Migmatites, melting and metamorphism*, (ed. M. P. Atherton & C. D. Gribble), pp. Liverpool: Shiva.

Zeitler, P. K. 1985 Cooling history of the NW Himalaya, Pakistan. *Tectonics* **4**, 127–151.

Zwart, H. J. 1962 On the determination of polymetamorphic mineral associations, and its application to the Bosost area (Coastal Pyrenees). *Geol. Rdsch.* **52**, 38–65.

*Discussion*

R. Mason (*Department of Geological Sciences, University College London, Gower Street, London*). In the reconstruction of $P$–$T$–$t$ paths from relict mineral assemblages, both those surviving as inclusions in later crystals and those in cores of zoned grains surviving more general overprinting, what allowances are made for the overstepping of equilibrium reaction conditions that must occur if the reactions are to proceed, and is there a possibility that some minerals grow metastably?

A. B. Thompson. Allowance for the overstepping of equilibrium reaction conditions along $P$–$T$–$t$ paths can only be made by presuming that the amount of overstepping is proportional to the enthalpy or entropy change of the reaction. There is always the possibility that some minerals grow metastably which indicates that inferences about $P$–$T$–$t$ paths from mineral assemblages can only be made by considering several mineralogical indicators in a range of lithologies that might be expected to have undergone similar $P$–$T$–$t$ histories. These problems are considered in some detail by Loomis (1983), Ridley & Thompson (1986) and Walther & Wood (1984).

*References*

Loomis, T. P. 1983 Compositional zoning of crystals: a record of growth and reaction history. In *Kinetics and equilibrium in mineral reactions* (ed. S. K. Saxena), pp. 1–61. New York: Springer-Verlag.

Ridley, J. & Thompson, A. B. 1986 The role of mineral kinetics in the development of metamorphic microstructures. In *Fluid–rock interactions during metamorphism* (ed. J. V. Walther & B. J. Wood), pp. 154–193. Berlin: Springer-Verlag.

Walther, J. V. & Wood, B. J. 1984 Rate and mechanism in prograde metamorphism. *Contr. Miner. Petr.* **88**, 246–259.



PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES