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Thermal Gradients and Regional Metamorphism in Overthrust Terrains with Special Reference to the Eastern Alps

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With 8 figures in the text

Abstract

About 65 mybp overthrusting took place in the Eastern Alps and a northward-moving complex of thrust sheets up to 15 km thick, was piled upon the European basement, trapping beneath it a mélange of predominantly pelagic, Mesozoic sediments (Peripheral Schieferhülle). Parts of this mélange show evidence of an early blueschist facies metamorphism, and the whole mélange experienced later greenschist to amphibolite facies metamorphism. The results of an analytical thermal model show that the overthrusting produces a major perturbation in the regional conductive thermal gradient. For a few million years after thrusting temperatures beneath the thrust are very low and if the thrust complex is thick enough to give sufficiently high pressures, blueschist metamorphism could occur at that time beneath it. Alternatively the blueschist event could have occurred elsewhere and the blueschists could have been emplaced as part of the mélange. The later metamorphism requires temperatures $\sim 500°$ C at 15 km depth, 25–30 m.y. after thrusting. Even when the maximum reasonable effects of radioactivity both within the thrust complex and the basement are taken into account, the depression of the thermal gradient associated with the overthrusting is still so great that it does not appear possible that the required temperatures of metamorphism can be attained in the time available without an additional contribution of heat from the upper mantle.

INTRODUCTION

There is no other young orogenic belt on Earth which has been studied so long and in such detail as the Alps, and it is clear that they offer an unusually good opportunity for understanding the complex interplay of thermal and mechanical processes which lead to the distribution of strain patterns and metamorphic facies which characterize the axial zones of many orogenic belts.

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The problems which confront any attempt to establish even a semi-quantitative model are formidable. Perhaps the major difficulty arises from the demonstrable invalidity in the case of the Alps, of the assumption made in the case of many other metamorphic terrains, that temperature during metamorphism may be regarded as some kind of relatively simple monotonic function of depth. This assumption may well be invalid in many metamorphic regions, but in the Alps it is demonstrably so.

In the steady-state situation, in the absence of significant tectonic movements temperature normally increases steadily with depth. The main changes in slope of the geothermal gradient are associated with variation in thermal conductivity satisfying the relationship

$$q = k\beta,$$

where q is the upwards heat flux to the surface, and in a steady state situation must remain constant, whatever the variation in k and β, the thermal conductivity and thermal gradient respectively. The values measured for the thermal conductivity of rocks mostly fall in the range 3×10^{-3} to 9×10^{-3} cal/cm sec°C. Thus even in a static situation the thermal gradient may vary with depth by a factor of two or more on passing from one lithology to another.

If the effects of radioactive heat production within the rocks are taken into

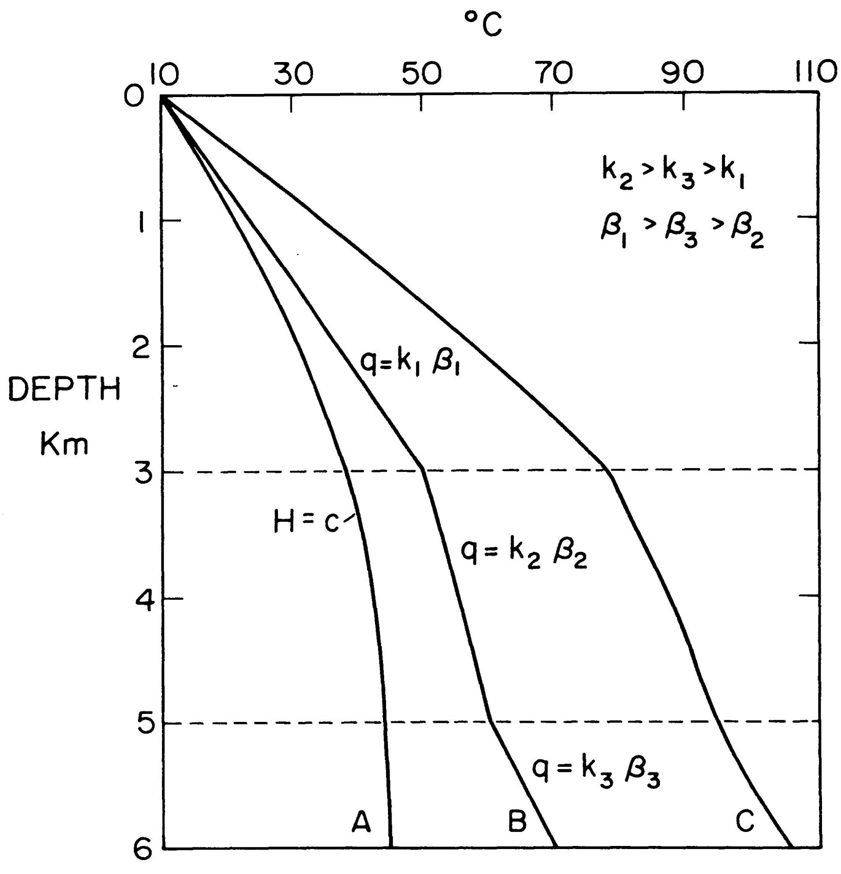

Fig. 1. Variation of thermal gradient (β) with thermal conductivity (k) in a three layer medium; for curve A there is no heat flux from below 6 km and internal heat production (H) is uniform through the 6 km; for curve B all heat is conducted from below 6 km and the heat fluxes at all depths are constant. Curve C is the sum of A and B. Variations in thermal conductivity may occur through variation in mineral assemblage, grain size or rock fabric. See text for discussion.

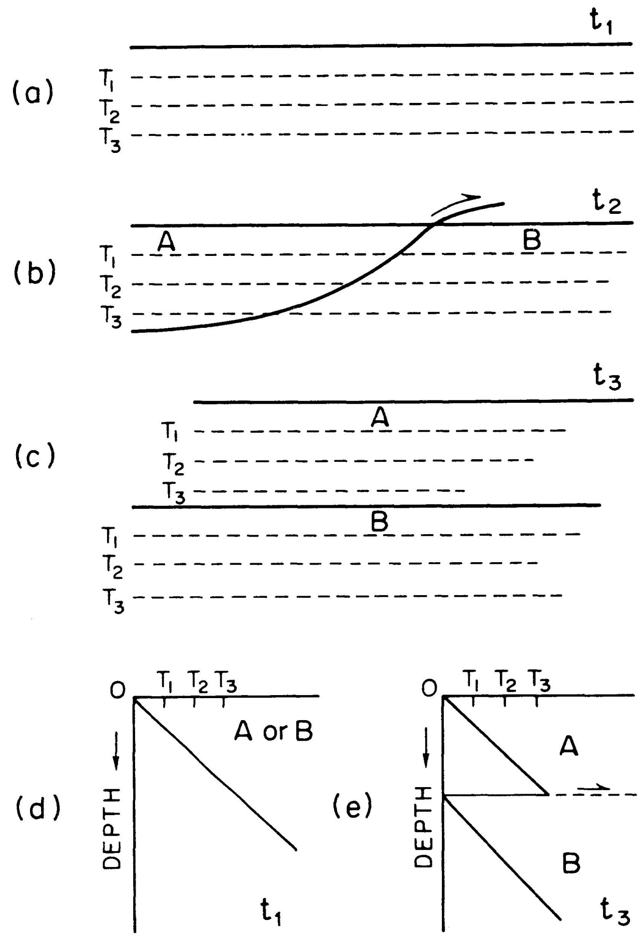

Fig. 2. The thrusting model: (a) crustal profile of presumed initial (t_1) condition; T = temperature $T_1 < T_2 < T_3$ with the associated thermal gradient shown in (d); (b) incipient thrust; (c) situation immediately after thrusting, giving rise to thermal profile (e). For discussion see text.

account these relationships are slightly modified. For a uniform distribution of radioactivity through the depth of interest and a uniform conductivity, the upwards heat flux must diminish with depth. In Fig. 1 curve A shows a thermal gradient resulting solely from uniform radioactive heating within the six kilometer depth shown. Curve B shows the gradient which results only from transfer to the surface by conduction of heat received from below. Curve C shows the result of combining curves A and B as could happen in nature. In this example the radioactivity within the upper 6 km, and heat conducted from below that depth make roughly equal contributions to the surface heat flux. The figure shows the effects of several different layers with different values of thermal conductivity. Thus even in a static and steady-state situation variations in the distribution of radioactivity with depth and in the thermal conductivity may bring about significant variations in the geothermal gradient.

So far we have been concerned solely with heat transfer by conduction. In tectonic environments such as the Alps, however, there are also thermal effects associated with tectonic movements. These tend to become important when the rate of movement of rock masses is large by comparison with the rate of transfer of heat by conduction. This may be readily appreciated with reference to Fig. 2. Consider a steady state situation as shown in (a): isotherms T_1, T_2, T_3 are horizontal and increase in value with depth as shown at time t_1. At time t_2 a thrust fault develops; the rocks, A, to the left are *rapidly* transported over the material, B, to the right. The situation immediately after thrusting at t_3, is shown in (c); the rocks of unit A along with their distribution of isotherms now rest upon unit B. The geothermal gradients before and after thrusting are shown in (d) and (e) respectively. At t_3 the geothermal gradient has a saw-tooth form. In the absence of further tectonic activity, the sawtooth will gradually decay and temperatures will change progressively towards an equilibrium gradient identical to that before the thrusting occurred.

The principles behind this process are clear, but the question is whether it is ever likely to be important in nature. We are concerned here with a kind of solid-state thermal convection. In our example we have allowed heat to be rapidly transferred by the physical movement of material and not by thermal conduction, and the physical reality of the model depends upon the quantitative comparison of the effects of these two different physical processes. Such a comparison may conveniently be made by means of the dimensionless physical parameter known as the Peclet number, P.

$$P = \frac{u\,l}{\kappa},$$

where u is the velocity of transport of hot material (in this case the velocity of overthrusting), l is a characteristic length over which thermal conduction is operative (in this case, the thickness of the thrust sheet) and κ is the thermal diffusivity of the rocks.

$$\kappa = \frac{k}{\rho\, c_p},$$

where k is the thermal conductivity of rocks; ρ the density, and c_p their specific heat. In cases where $P > 1$ it may be readily shown that as long as the movement persists convective heat transport controls the temperature distribution and conductive effects are of secondary importance. We therefore evaluate P for the case of a thrust sheet 10 km thick and $u = 10^{-7}$ cm/sec (i.e. 3 cm/yr; a value based on the assumption that regional thrusting processes are related to plate tectonics and may therefore be expected to be characterized by similar rates of movement). A reasonable value for κ is 10^{-2} cm^2/sec. Using these values

$$P = \frac{3.3 \times 10^{-7} \times 10^{6}}{10^{-2}} = 33\,.$$

The requirement that $P > 1$ is thus easily satisfied, and it becomes clear that in regions where large scale overthrusting has occurred it is essential to take into account the perturbation of the geothermal gradient brought about by the thrusting process.

In this paper we attempt to relate the thermal requirements of the regional metamorphic events which have occurred in the Tertiary in the Eastern Alps to temperature distributions which are implied by the large scale tectonic movements and the measured radioactivites of the rocks.

THERMAL RELAXATION IN THRUST SHEETS

We approach this problem initially by considering a fairly simple theoretical model for which an exact solution is provided. We do not here deal with the mathematical derivation of this solution (see TURCOTTE and OXBURGH, in prep.) but the initial assumptions are clearly stated and we consider the effects of varying the input parameters and boundary conditions.

The first model analyzed is that sketched in Fig. 2(e). A thrust sheet is assumed to have been very rapidly emplaced in a single phase of movement. The Peclet number is assumed to be high and the duration of the overthrusing process brief; the temperature distribution immediately after thrusting is taken to have perfect "saw-tooth" form. We consider the subsequent variation of temperature with time in the thrust sheet and the rocks which it overlies. We assume that before thrusting the thermal gradients in both the thrust sheet (hereafter called the slab) and the underlying country rock (hereafter called the basement) are the same. Both zones are considered to be laterally much more extensive than they are thick (i.e. the solution is one dimensional) and to have equal, and uniform thermal diffusivity, $\kappa = 10^{-2}$ cm^2/sec^{-1}.

No post-thrusting, convective process (eg. magmatic intrusion, movement of fluids) is considered at this stage. We also take no account of internal heating by radioactivity and no account of surface erosion (e.g. see CLARK and JAEGER, 1969) or of frictional heating on the thrust fault. The consequences of these factors are discussed later.

a) The Interface Zone

As soon as thrusting is complete conductive heat transfer brings about the gradual decay of the saw-tooth isotherm distribution produced by thrusting. Equilibrium is reached once more when there is a continuous and uniform thermal gradient through both units. We begin by considering temperatures along the interface between them. For convenience and generality we assume the initial upper surface temperature of both units to be zero and we designate the temperature at the base of the slab, T. The act of thrusting brings the base of the slab at temperature T into contact with the zero temperature surface

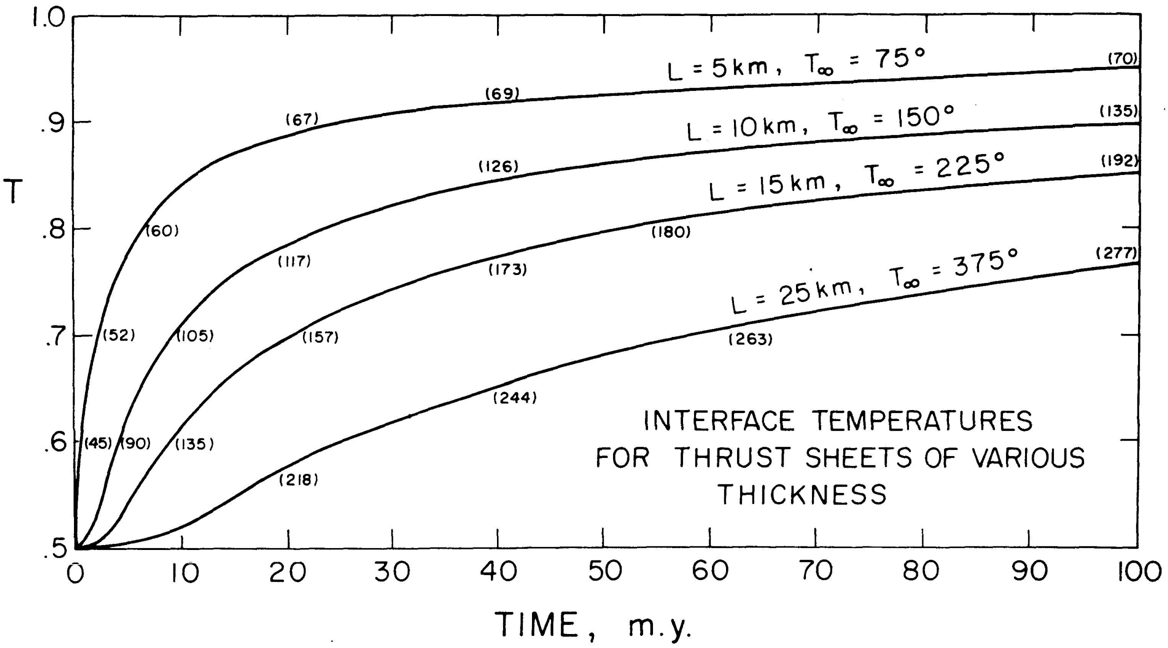

Fig. 3. Temperature as a function of time on the interface between a thrust slab and its basement for slabs 5, 10, 15 and 25 km thick. Times are times elapsed after thrusting. T is the initial temperature and also the equilibrium temperature at the base of the thrust. Shown in parentheses are actual temperatures in °C for an initial thermal gradient through both slab and basement of 15°/km; the corresponding equilibrium temperatures are shown as T_∞. For discussion see text.

of the basement. The moment contact is made the temperature on the interface becomes 0.5 T. The subsequent variation in interface temperature is shown in Fig. 3 for the first 100 m.y. after thrusting.

Curves are plotted showing the interface temperatures under thrust sheets of different thicknesses. In all cases the final equilibrium temperature must be T, the initial temperature at the base of the thrust sheet. The thinner the thrust sheet, the more rapidly does conductive equilibration occur. It is apparent that for thrust sheets ten or more kilometers thick, periods of tens of millions of years elapse before there is a close approach to equilibrium. These curves may be applied to any value for the initial thermal gradient, β, simply by selecting an appropriate value of T for the slab thickness curve used. If the thickness of the slab is L, then $\beta = T/L$. As an example, temperatures in degrees C are shown against each curve on the assumption that $\beta = 15°\ km^{-1}$.

b) Temperatures Within the Slab

In order to investigate temperature distributions above the interface it is convenient to select a single value for the thrust sheet thickness and examine that situation in some detail. We therefore consider temperature distributions within and (in the next section) below a thrust sheet 15 km thick. This is a

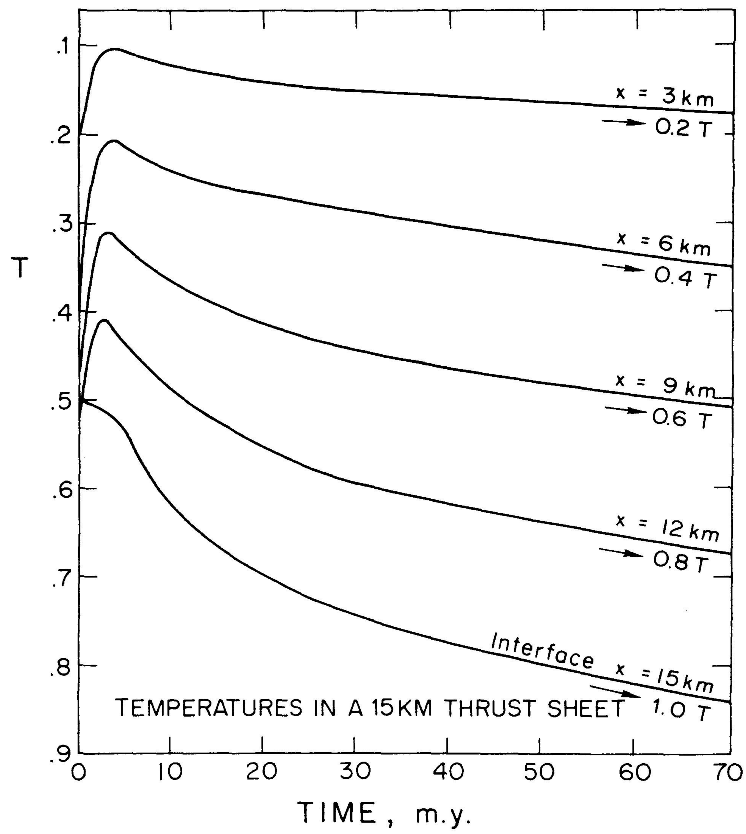

Fig. 4. Temperature as a function of time and depth (x) within a 15 km thick thrust sheet. The equilibrium value for T is shown on each depth curve. T and time as in Fig. 3.

value which is thought to be an upper limit for the central part of the Eastern Alps (OXBURGH et al. 1971). Temperature as a function of time at various depths in such a sheet is given in Fig. 4. After thrusting there is at all depths an initial cooling to a temperature close to one half the initial temperature at that depth. The temperature minimum at each depth occurs progressively later upwards through the slab, but even at 3 km it occurs only 4 m.y. after emplacement and shortly after that, steady temperature increases occur at all depths. The "sawtooth" initial temperature distribution and associated negative thermal gradients are damped out in less than a million years and by 3.3. m.y. the gradient through the slab is nearly linear and has about half the value of the initial gradient. The curves are carried out to 70 m.y., the presumed maximum duration of cooling since the main overriding of the Upper (and Middle) East Alpine Nappes (see below).

c) Temperatures Within the Basement

The temperatures in the basement are shown in Fig. 5. There is initially a very large heat flux downwards from the slab into the basement in response to the steep negative thermal gradient across the interface shortly after thrusting; this allows a rapid initial basement temperature rise but after about 25 m.y. warming occurs much more slowly.

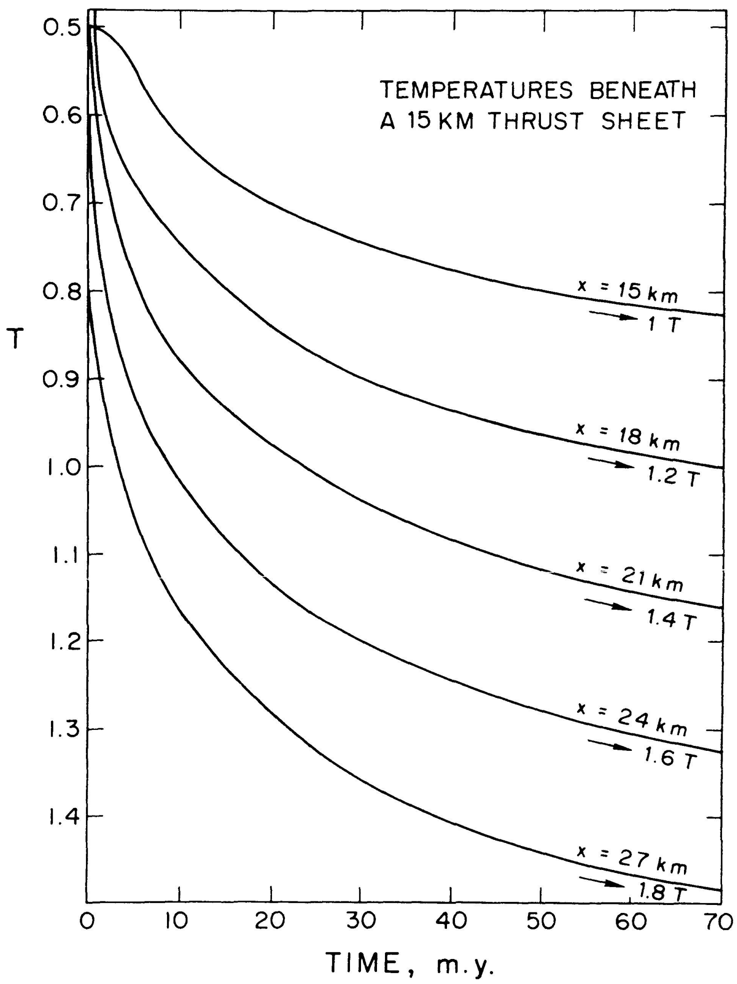

Fig. 5. Temperature as a function of time and depth (x) beneath a 15 km thrust sheet. Symbols as for Fig. 4.

The results for slab, basement and the interface are summarized in Fig. 6 as a series of thermal gradients from 3.3 to 60 m.y. through a 15 km thrust sheet and the basement upon which it rests. It is scarcely apparent at the scale of Fig. 6, but all except the initial and final gradients are slightly sigmoidal.

d) The Effects of Varying Input Parameters

As has been shown the curves can be applied to any desired initial value of the thermal gradient. Although temperature profiles have been presented only for a thrust sheet 15 km thick, the effects of other thicknesses at various times may be roughly estimated by comparison of Fig. 3, which shows interface temperatures as a function of time and sheet thickness, with Fig. 6. The effect of varying the thermal diffusivity is proportionally and inversely to change the time (t) i.e. the product κ t must remain constant so that if in Fig. 3 it was desired to use a thermal diffusivity of 2×10^{-2} cm^2 sec^{-1} rather than 10^{-2} cm^2 sec^{-1}, the shape of all curves would remain unchanged, but the values of all times indicated on the horizontal scale would be halved.

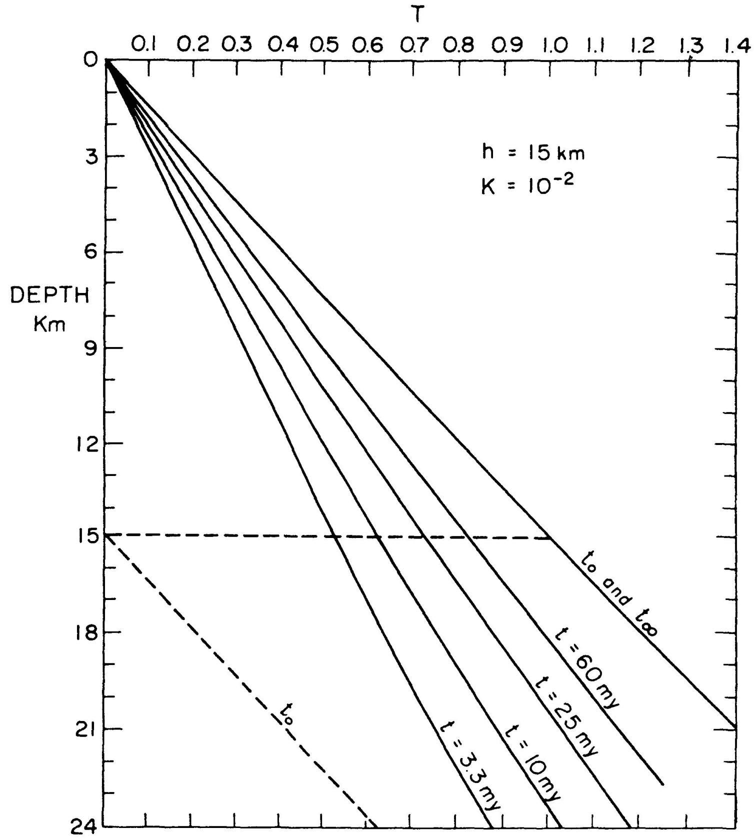

Fig. 6. The conductive thermal relaxation of a 15 km thick thrust sheet with $\kappa = 10^{-2}$ cm^2/sec. The initial (t_0) temperature distribution has sawtooth form (solid and dashed lines) and the equilibrium (t_∞) profile is given by the solid line. Profiles for four intermediate times are shown. See text for discussion; T as defined for Figs. 3, 4 and 5.

REGIONAL METAMORPHISM IN THE EASTERN ALPS

It is not appropriate to review here the geological history of this region and we simply outline those events which have a bearing on the present problem. More extensive discussions of relevant aspects of the regional geology are to be found in Kober, 1955; Clar, 1965; Tollmann, 1963; Exner, 1964; Gwinner, 1971; Oxburgh, 1968a,b; Oxburgh et al., 1971; Ernst, 1973; Frasl and Frank, 1966.

During the Mesozoic, epicontinental sedimentation occurred along the southern margin of an extensive continental region which was continuous northwards with the present day Bohemian Massif and the older basement rocks of Europe. Some time, or at several times, between the latest Cretaceous and the Miocene this region was overridden by allochthonous thrust sheets from a generally southerly direction. This overriding appears to have been associated with the subduction of oceanic lithosphere which formerly lay to the south of the European continental land mass and separated it from the continental region from which the allochthonous thrust sheets were derived (for discussion of the process see Oxburgh, 1972). It seems that when continental collision occurred, a substantial quantity of pelagic sediment was

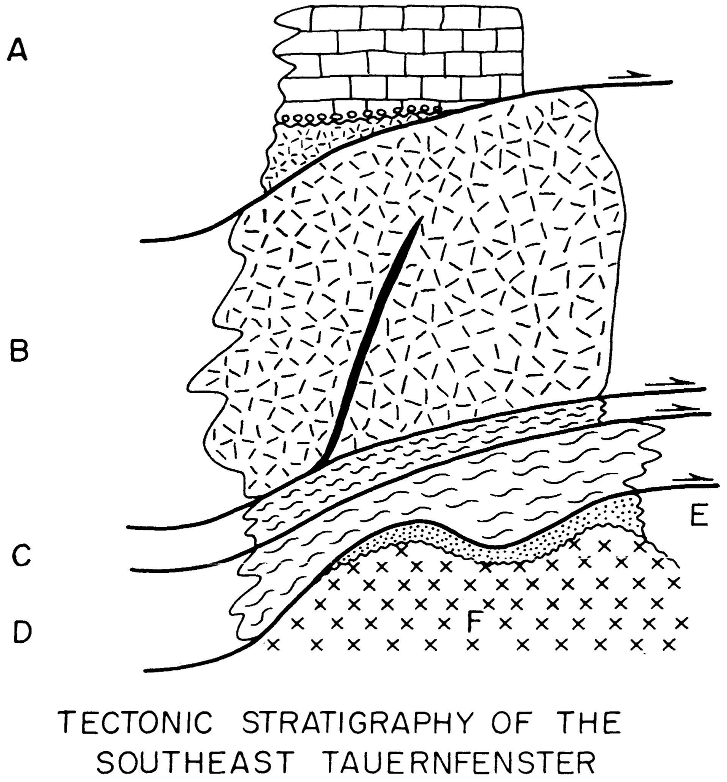

Fig. 7. Sketch of the main tectonic units in the central Eastern Alps. A: Oberostalpin – sediments with local metamorphics beneath thrust over B: the Altkristallin (Mittelostalpin) showing discordant dyke, thrust over C: Unterostalpin thrust over D: Peripheral Schieferhülle mélange thrust over F and E: the autochthonous European basement with its locally preserved sedimentary cover. In this paper the main interest is in explaining the metamorphic history of unit D. Units D, E and F form the so-called Pennine Zone of the Eastern Alps.

dragged northward underneath the thrust sheets and today lies between them and the underlying European continental basement. Within the Tauern Window these allochthonous pelagic rocks are seen as part of the Peripheral Schieferhülle (Fig. 7 D). They overlie a complex of Hercynian orthogneisses and paragneisses (Central Gneiss and Inner Schieferhülle, Fig. 7 F) with its autochthonous and parautochthonous cover of Mesozoic epicontinental sediments (Fig. 7 E).

The main allochthonous units above the Peripheral Schieferhülle have been variously classified and subdivided. The largest element, however, is a 300 km × 200 km body of Palaeozoic or older schists and gneisses (the Altkristallin, Fig. 7 B). The thickness of this unit is variable and is hard to establish but in its southern parts is likely to be about 10 km. The meaning of this "thickness" is itself difficult to interpret because the unit could be composed of a number of slices resting upon each other (for an extreme position see Metz, 1966). There undoubtedly are important shear zones and discontinuities in metamorphic grade within the Altkristallin (e.g. Wright, 1973) but detailed mapping is not yet sufficiently extensive to establish their significance and extent. Beneath the Altkristallin and above the Peripheral Schieferhülle there generally occurs a thinner tectonic unit known as the Unterostalpindecke or, locally, the Matrei Zone (Fig. 7 C). This comprises predominantly Mesozoic sedimentary units which have undergone varying degrees of tectonic dislocation. The Altkristallin is overlain today by patchily distributed sedimentary rocks of late Palaeozoic to Mesozoic age which have undergone little or no regional metamorphism. Although some of these sediments are clearly autochthonous (Mittelostalpin Mesozoicum of TOLLMANN, 1963) the remainder may form part of a separate and formerly extensive, overlying tectonic unit (Oberostalpin of TOLLMANN, Fig. 7A). It was concluded by OXBURGH et al. (1971) that 15 km was a reasonable upper limit to this allochthonous pile resting upon the Peripheral Schieferhülle.

There is relatively little direct evidence on the time of formation of this pile of thrust sheets. However, the retrogressed mineral assemblages and overprinted K/Ar ages in the base of the Altkristallin indicate that at any rate the greater part of the thrust sheet pile was in place before peak metamorphic temperatures were reached (about 40 mybp, see below). The movements must also have begun after the deposition of the youngest sediments involved. These are generally regarded as late Cretaceous (TOLLMANN, 1963) but in the western part of the Eastern Alps early Tertiary sediments are recorded by OBERHAUSER (1968) from the Peripheral Schieferhülle of the Rhaetikon. If the Altkristallin does comprise a number of separate slices, times of overriding could well have varied along the Alpine chain; in addition, individual slices could have undergone northward movement in a number of distinct episodes. Movements of the overlying Oberostalpin were at least in part independent of those of the Altkristallin. The isotopic evidence of BREWER (1970) suggests that the main thrusting may have occurred at about 65 m.y. This is consistent with all the other lines of evidence and we accept it for the purposes of our model.

There is evidence of two main periods of regional metamorphism. The most widely developed event affects all the rocks underlying the Altkristallin to some extent and produces variable amounts of retrogression in the lower parts of the Altkristallin (PREY in EXNER, 1962; WRIGHT, 1973; BICKLE, 1973). The mineral assemblages developed in this metamorphism have been described by many authors (e.g. ANGEL, 1939; EXNER, 1957, 1964; CLIFF et al. 1971; ERNST, 1973). At the base of the Peripheral Schieferhülle in the south eastern Tauern pelitic rocks occur in the amphibolite facies of regional metamorphism and CLIFF et al. (1971) have proposed peak metamorphic temperatures of about 550° C. The grade of metamorphism falls towards the north where greenschist facies mineral assemblages occur. These temperatures are consistent both with those proposed by BICKLE (1973) on the evidence of calcite-dolomite geothermometry, for the region south of the Gross Glockner in the center of the Tauern window, and with the metamorphic temperatures estimated by HOERNES and FRIEDRICHSEN (1974) on the basis of oxygen isotope work in the Central Tauern. There is considerable difficulty in establishing reliable depth/pressure controls for the amphibolite facies metamorphism. In the absence of work on pressure sensitive mineral assemblages, estimates of pressure depend upon geological estimates of the maximum reasonable thickness of the overlying pile of allochthonous thrust sheets. The observations of CLIFF et al. (1971) suggest 14 ± 4 km as the depth of burial of the upper surface of the Central Gneiss. It should be noted that if depths of burial significantly greater than this are advocated it is difficult to accommodate the great volumes of erosional debris which would have to have been deposited in the known peri-Alpine Tertiary basins during the mid and late Tertiary.

Occasionally evidence for an earlier metamorphic event is found. CORNELIUS and CLAR (1939) have described eclogites in the Obere Schieferhülle and FRY (1973) reports the local development of lawsonite schists within the same unit. It is clear that these mineral assemblages belong to a high pressure-low temperature series. In any case, fabric evidence confirms the separate identities of the two metamorphic episodes. The greenschist-amphibolite event is demonstrably later than all the main folding deformations in the Tauern (CLIFF et al. 1971; BICKLE, 1973) whereas the high pressure-low temperature assemblage fabrics are strongly deformed by folding.

There is no modern published study of the mineralogy of these rare east Alpine blueschists and it is not known whether they compare more closely with Sanbagawa or Franciscan terrains (ERNST and SEKI, 1967). It is, however, presumed that pressures greater than 15 km of burial and temperatures lower than 400° C or even 300° C are required for their formation.

In the greater part of the Tauern Window, however, the rocks show evidence of only the later greenschist/amphibolite event, and it follows that either the earlier blueschist event was not universally developed in the Peripheral Schieferhülle or that in many places all trace of it has been obliterated by the later metamorphism.

If the former is true, it is necessary to consider whether the blueschist metamorphism may predate the emplacement of the Schieferhülle mélange. In that case, the blueschist metamorphism could have taken place to the south of the Tauern, and a mélange of blueschists and previously unmetamorphosed, or less metamorphosed, pelagic material could have been emplaced on top of the Pennine basement at the time of overthrusting.

The age of the greenschist/amphibolite metamorphic recrystallization is thought to be about 40 mybp on both stratigraphic and radiometric evidence (CLIFF et al. 1971). The age of the earlier event is unknown.

In concluding this review of pertinent geological information we note that there is no Tertiary igneous activity known within the Tauern window but that the Central Gneiss and parts of the overlying metasedimentary units have been mineralized along steep and late joint surfaces. No ages have yet been determined from these mineral veins but it is probable that they are about 20 m.y. old. NORRIS et al. (1971) have shown that the joint pattern is apparently related to the uplift of the Tauern and this uplift has been dated by CLIFF et al. (1971) at about 20 m.y. Some Eocene intrusives cut the Altkristallin south of the Tauern (CLIFF et al. 1971; WRIGHT, 1973; D. J. WATERS, unpubl. data).

To summarize, we assume that at about 40 mybp the Peripheral Schieferhülle of the Southeast Tauern recrystallized at about 500° C and 15 km depth ($\sim$5 kb). About 25–30 m.y. earlier (at 65 mybp) there had been a major episode of overthrusting which resulted in the burial of the Peripheral Schieferhülle to this depth. Some time earlier than 40 m.y., at least some of those rocks underwent a blueschist facies metamorphic event (say $T < 350°$ C, $P > 5$ kb) but it is not known whether this preceded, accompanied, or followed the overthrusting.

APPLICATION OF THE THEORETICAL MODEL

We now attempt to relate the thermal requirements of the two phases of regional metamorphism to our analysis.

a) The "Early" Blueschist Event

It is clear (Figs. 4, 6) that for a while after the overthrusting, temperatures beneath the overthrust mass are unusually low. Any reasonable thermal gradient in the thrust mass will give temperatures low enough, or even too low, for blueschist formation. The limiting constraint is not temperature but pressure. Experimental evidence suggests that the overburden pressure must be 15 km or more (P total > 5 kb) and this must therefore be the minimum thickness of the thrust complex if blueschists are to be generated beneath it. Such a thickness is at the upper limit of estimates for the East Alpine thrust complex and so it is possible that the East Alpine blueschists formed during or shortly after the main overthrusting. If so the age of this metamorphism would provide the best evidence for the time of the main thrust movements.

Equally, however, as discussed earlier, blueschist metamorphism could have taken place in a trench-subduction zone complex to the south of the present Alpine chain, before the Peripheral Schieferhülle rocks were transported to their present position. In this case the East-Alpine thrust complex could have carried northwards beneath it a mixture of pelagic material some of which was blueschist, but other parts of which could have been much less recrystallized, thus explaining the present day irregular and sparse distribution of the former.

Note that most of the simplifying assumptions made in our model do not affect the main conclusion. Because we are concerned with effects which follow immediately after thrusting, the effects of radioactive heating may be neglected. If there is transient convective heat transfer upwards from beneath the slab by water this will reduce or eliminate the early downwards heat flux from the slab and cause temperatures beneath it to rise more slowly than in our model.

If the thrust complex is not essentially a single unit but is built by the stacking of a series of thinner sheets, the initial thermal gradient through the "slab complex" will have multiple "saw teeth" (see OXBURGH et al. 1971, Fig. VI-1(b)). For the same pre-thrusting thermal gradient, however, the heat content of the "slab complex" will in all cases be smaller than in the case of the single slab which we have analyzed, and the effect of this is again to slow down the temperature rise below the slab.

If we confine our attention to the time of overthrusting and shortly after, there seems to be only one physical process which could modify the conclusions reached so far, namely the generation of frictional heat during the overthrusting process. This process has been analyzed (TURCOTTE and OXBURGH, 1969; OXBURGH and TURCOTTE, 1968) with special reference to heat generation at the upper surface of a descending lithospheric slab in island arc regions.

The rate of heat generation, q, is given by $q = u\,\tau$, where u is the velocity of thrusting and τ is the shear stress. It was shown that for $u = 10$ cm/yr and $\tau = 2.64 \times 10^8$ dynes/cm^2, $q = 2$ μ cal/cm^2 sec. The local temperature rise is governed by the balance between the rate of heat generation and rate of conduction of heat away from the movement zone. It is not known what values would be appropriate for τ beneath the overthrust slab but the value used above is likely to be an upper limit; the true value may be significantly lower (HUBBERT and RUBEY, 1959).

In any case, heat generation continues only as long as the movement lasts and with a value of $u = 10$ cm/yr, movement would be completed in 1 m.y. Thus the total amount of heat generated is rather small by comparison with the heat content of the rock volume of interest. The main question therefore is whether heat is removed from the zone of movement sufficiently slowly for there to be a significant, transient local temperature elevation. We do not analyze this problem in detail here, but if circulating fluids had access to the thrust zone and there was convective transfer of heat, there would be little noticeable effect. In the absence of such circulation, heat transfer would be by conduction alone and the effect might be significant. In this case the rocks in the immediate vicinity of the movement zone would be heated briefly during movement, would cool to the local ambient temperature after movement ceased, and then would experience a temperature rise once more during the main post-thrusting restoration of thermal equilibrium.

b) The 40 mybp Amphibolite Facies Event

We now turn to the later metamorphic event. We test our models according to their ability to generate temperatures of $\sim 500°$ C immediately below a

15 km slab, $\sim$30 m.y. after overthrusting occurred. It is clear from Fig. 6 that simple overthrusting of a 15 km slab with a 15°/km initial thermal gradient across a terrain with a similar initial gradient, will *never* provide the appropriate metamorphic temperatures immediately beneath it by conductive heating. For the moment retaining this simple model we see that suitable temperatures may be reached *either* by increasing the thickness of the slab, *or* by increasing the initial thermal gradient within it, *or* both; i.e. by choosing the initial conditions for the slab in such a way that the pre-thrusting conditions at the depth of its base are close to those required for metamorphism. In the present case this amounts to assuming an initial gradient of about 45° C/km if metamorphic conditions are to be reached in 30 m.y. (i.e. 35 mybp), unless the slab thickness is increased beyond limits which seem geologically reasonable.

In the foregoing discussion we have made the simplifying assumption that the thermal gradient and conductivity (and thus the heat flow) within the thrust mass and the new basement upon which it comes to rest, were the same. In many cases this will not be a reasonable assumption.

The initial temperature (i.e. pre-thrusting temperature) of the base of the thrust slab depends on (1) its thickness, (2) its thermal properties, (3) the amount and distribution of radioactivity within it and (4) the heatflux received from the underlying material. Figure 8 (left hand column) shows a possible pre-thrusting relationship between the slab and its original basement, and the thermal gradient through them (A–A′).

Variables 1, 2 and 3 do not change between the pre- and post-thrusting situations, but 4 may well vary. If, once it has come to rest, the sheet receives either a greater or lesser heat flux from beneath than it received before thrusting, the final equilibrium temperature at its base must vary and is given by

$$T_{b'} = T_b + (q_{b'} - q_b)\frac{h}{k},$$

where T_b and $T_{b'}$ are the pre- and post-thrusting equilibrium temperatures at the base (b) of the slab and q_b and $q_{b'}$ are the pre- and post-thrusting values of the upward flux of heat at b; h is the depth to b, and k is the mean thermal conductivity of the thrust mass. In round terms this means that a value of $\Delta q = q_{b'} - q_b$, of 0.5 μ cal/cm² gives an increase (or decrease, if Δq is negative) of 100° C in the equilibrium temperature at the base of a 15 km thrust sheet.

Thus the final post-thrusting equilibrium temperature profile in the slab and the temperature at its base will depend only upon the properties of the slab, and the heat flux from the new basement upon which it comes to rest. The *rate* at which equilibrium is approached however, will depend both on the properties of the basement and the initial gradient in the slab, and this in turn depends upon the pre-thrusting heat flux from below.

We therefore consider the form of the final equilibrium gradient. The geological model is shown in Fig. 8 (right side). The slab is assumed to comprise 5 km of sedimentary rocks with negligible internal heat production overlying 10 km of schists and gneisses with uniform heat production and A = 6 HGU, a probable limiting upper value for heat production within them.

Below the thrust we assume there to be a mean thickness of 4 km of Peripheral Schieferhülle with a uniform heat production of A = 6 HGU, which should again represent a limiting upper value.

The Schieferhülle rests upon the basement complex which is of pre-dominantly granodioritic composition and it is the heat flux from this which is of great importance. Studies in intrusive complexes elsewhere (LACHENBRUCH,

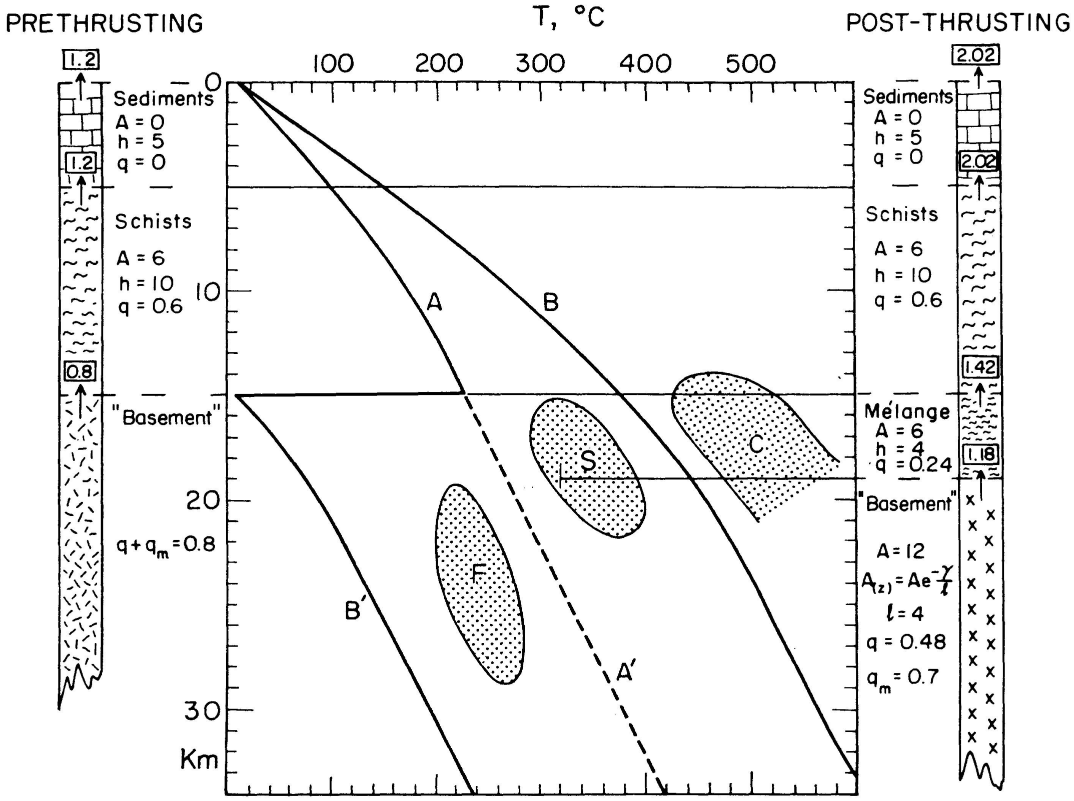

Fig. 8. The relationship between equilibrium thermal gradients, crustal radioactivity, mantle heat-flow and P/T regimes for regional metamorphism. Side columns: A = heat production in HGU (cal/cm³ sec × 10^{-13}), h = thickness of the layer in km, q = local contribution to the total heat flux (i.e. q = Ah × 10^{-2} μcal/cm²sec). Within each column the cumulative heat flow across each interface is given in a rectangle. The left column gives a possible pre-thrusting crustal structure for the source region of the slab and the equilibrium gradient within it (A–A′). After thrusting the crustal structure appears as on the right-hand side, where the top 15 km of the column on the left comes to rest on a melange and basement with the characteristics shown. The gradient immediately after thrusting is (A–B′) and the new final equilibrium gradient is B. For reference, zones F, S and C are the supposed P/T fields for Franciscan metamorphism. Sanbagawa metamorphism and the 40 mybp East Alpine amphibolite facies metamorphism.

1968; BIRCH et al., 1968) have shown that in many such situations heat production and surface heat flow satisfy the relationship

$$q_0 = q_m + A_0 l,$$

where q_0 is the surface heat flow, q_m is a "deep" contribution to the surface heat flux and is apparently uniform within a single heat flow province, A_0 is the heat production of the surface rocks, and l is a constant (with the dimensions of length) characteristic of a particular province. LACHENBRUCH (1968) has shown that in regions where there have been variable amounts of erosion the expression may be generalized

$$q_0 = q_m + A_0 \exp\frac{-x}{l}.$$

This relationship thus implies an exponential fall off in heat production with depth (x). HAWKESWORTH (1974) has shown that such a relationship holds in the Pennine basement complex exposed in the eastern part of the Tauern window and that values of $A_0 = 12$ HGU and $l = 4$ km are appropriate. There is no independent evidence for the value of q_m; provisionally we take a value of 0.7 μ cal/cm^2 sec (ROY et al., 1972).

The resulting post-thrusting thermal gradient when equilibrium has been attained is shown in Fig. 8, curve B. Although the gradient is now somewhat closer to the conditions required for the regional metamorphism, temperatures are still too low at equilibrium and would be significantly lower only 30 m.y. after thrusting (compare Fig. 6). There now remain three ways in which the temperatures may be elevated: (1) increasing the thickness of the thrust sheet, (2) increasing the heat flux from the basement complex, and (3) increasing q_m.

In evaluating these three alternatives we note that 30 m.y. is the time required for the temperature beneath a 15 km thrust sheet to reach 0.75 T (Fig. 6). Thus if a temperature of 500° C is required at the base of the thrust after 30 m.y. this is equivalent to an equilibrium temperature at that depth of $\sim$650° C. This is, of course, only a rough estimate as the model upon which it is based is different from the one we are now considering, in ways already emphasized; nevertheless it should give a reasonable guide. We consider these alternatives therefore in the light of their ability to provide sub-thrust equilibrium temperatures of about 650° C; an equilibrium which is never reached but which must be supposed if high enough temperatures are to be attained in the time available.

Geological objections to increasing the thickness of the thrust sheet have been mentioned earlier. In practice, an unreasonable thickening to more than 35 km would be required, because as the thickness of the sheet increases, so does its time for equilibration (e.g. after 30 m.y. a 25 km sheet has reached only 0.6 T at its base (Fig. 3)). This alternative is therefore regarded as improbable.

The second alternative is to increase the heat flow due to radioactivity within the basement. This could be done by increasing either A_0 or l in the exponential expression given earlier. A_0 cannot be significantly increased without exceeding geochemically reasonable limits, but the value used for l, the exponential scale depth, could perhaps be increased. The value of 4 km found by HAWKESWORTH from direct observation is about half that deduced by other workers from indirect means (e.g. ROY et al. 1972). There is no reason to doubt HAWKESWORTH's result, but the distribution of heat producing elements in the basement might itself have been changed by the Alpine metamorphism and might therefore differ today from the premetamorphic distribution. Increasing the value of l to 8 km, increases the basement heat flux to 0.96 μ cal/cm^2 sec and increases the equilibrium temperature below the thrust fault to about 475° C. This alternative is therefore inadequate to explain the observations.

We are left therefore with the possibility of increasing the mantle contribution to the basement heat flow. To achieve an equilibrium subthrust temperature of about 650° the value of q_m must be increased from 0.7 to about 1.8 μ cal/cm^2 sec. This is an unsatisfactory alternative in the sense that there is no obvious way of testing it. The value proposed for q_m is, however, similar to values deduced for other active tectonic areas (ROY et al., 1972). In so far is it seems likely that a subduction zone operated in this area during the early Tertiary (OXBURGH, 1972), it is possible that an increased value of q_m could be associated with dissipative heating along it (e.g. OXBURGH and TURCOTTE, 1968). Thus although this remains very much an *ad hoc* solution to the problem, it appears to be the least unreasonable.

EFFECT OF EROSION AND CONVECTIVE HEAT TRANSFER

It is interesting to consider the nature of the "metamorphic peak" i.e. the maximum temperature conditions of which a record remains in the mineralogy of the rocks. It is clear that thermal equilibration within and below the thrust pile takes place so slowly that at any given reference horizon temperatures would still be rising today if the process had not been interrupted in some way. This interruption seems to have occurred 25–20 m.y. after thrusting – the time of the metamorphic peak, i.e. the time after which the temperatures no longer rose. The most probable cause of the interruption is the onset of erosion at the top of the thrust pile. The effect of a given amount of erosion is simply to bring the constant temperature, cold upper surface of the pile closer to any given reference horizon below, and thus to proportionally reduce the equilibrium temperature for that horizon. Successive increments of erosion progressively reduce that equilibrium temperature until the actual temperature is higher than the equilibrium temperature and cooling begins. This onset of the erosional termination of metamorphism is seen in the Oligocene Molasse sedimentation.

We have made no comment on the possible mechanism of heat transfer associated with the postulated abnormally high value of the heat flow from the mantle. If some of this heat were transferred by a convective rather than a conductive process (OXBURGH and TURCOTTE, 1971), the most plausible means in the present case would be by the movement of hot water. The operation of any such process has the effect of increasing the "virtual conductivity" of the material in the depth interval within which it operates. For any given heat flux across a medium, the effect of increasing the thermal conductivity of the medium is to reduce the thermal gradient required to sustain the heat flux (Fig. 1). Thus it is conceivable that the high temperatures required for Alpine metamorphism could have been attained above a basement within which the thermal gradient was rather low. If at some time after the metamorphic maximum had been reached (say, the onset of uplift) the convective mechanism ceased to operate, heat transfer would occur subsequently only by conduction but along a very low thermal gradient. The consequent conductive heat flow through the basement would remain low until a normal conductive gradient had been reestablished. This observation could have a bearing on the surprisingly low heat flow values sometimes measured in young metamorphic terrains (CLARK and JAEGER, 1969) and conceivably on the low values measured in some young intrusive bodies (LACHENBRUCH, 1968).

If significant convective heat transfer took place within the mélange zone rocks during their regional metamorphism any thermal gradient deduced by internal means (e.g. differences in metamorphic grade at different depths or paleotemperature measurements) could not be used to deduce a heat flow during metamorphism or, by extrapolation to the surface, a depth of burial.

It is worth pointing out that a conceivable alternative cause of the termination of metamorphism might be a change in the mechanism of heat transfer, i.e. if fluids began to circulate over some significant depth interval where they had previously been inactive. Provided the overall heat flux into the system was unchanged, equilibrium temperatures at all depths greater than the top of the convecting zone would be reduced because of an increase in the virtual conductivity of a system.

Equally, however, an increase in the virtual conductivity of a system by the action of moving fluids would be associated with a similar increase in the virtual thermal diffusivity. This would lead to more rapid thermal equilibration within the zone of fluid circulation, and this in turn to more rapid equilibration of the system as a whole. Thus the abnormally high value for q_m proposed above might be somewhat reduced, but would still be high.

CONCLUSIONS

An attempt has been made in this paper to draw attention to certain factors which have previously received relatively little attention but which could have a significant effect on metamorphic temperature regimes in overthrust terrains. For reasons of length, the derivation and calculation of the results presented have been virtually omitted. Solutions to nearly all the problems encountered are to be found in CARSLAW and JAEGER (1959). The analytical treatment of the gradient relaxation after thrusting will be published elsewhere.

Several general inferences may be drawn from the model studies described in this paper. It appears that where regional metamorphism is associated with large scale thrusting, the time required for full thermal equilibration is so great that temperatures in the rocks undergoing metamorphism should continue to rise until the onset of uplift and significant erosion. The maximum grade of metamorphism preserved in a series of rocks buried by and metamorphosed under a thrust sheet would vary both with any regional changes in the thickness of the thrust sheet and with the time of onset of regional uplift. Regions uplifted earlier than others would be expected, other things being equal, to be further from thermal equilibrium and thus at lower temperatures, and at lower metamorphic grade at the time of uplift.

A more rigorous analysis of the various transient models discussed above could most conveniently be carried out by numerical methods and several models have been computed by BICKLE (1973). It is hoped, however, that the present approach may make it possible for those who disagree with the choice of values for any particular parameter used, to readily calculate their own models.

We favor a model therefore in which the thrust mass was emplaced on top of a basement with an abnormally high heat flow of about 2.5 μ cal/cm^2 sec (mélange $q=0.25$; gneiss complex $q=0.48$, $q_m=1.8$ μ cal/cm^2 sec). The high value of q_m was transient and associated with activity on an underlying mantle slip zone.

For a short time after thrusting temperatures below the thrust slab were low enough for blueschist metamorphism but it is uncertain whether the thrust complex was thick enough to provide sufficiently high pressures.

After an initial phase during which there was a downward flux of heat from the sheet into the cold upper part of the basement, temperatures in the basement, the slab and the mélange trapped between them continue to equilibrate at a rate which was largely dependent upon the thickness of the thrust slab. After about 30 m.y. temperatures of about 500° were reached beneath the thrust, sufficient for the amphibolite facies metamorphism of the mélange rocks.

The main conclusions of this paper are:

1. It is possible that blueschists may form under thick overthrust sheets.
2. In the Eastern Alps it is difficult to achieve the metamorphic conditions appropriate for the amphibolite facies event without an anomalously high transient flux of heat from the mantle to the crust. By inference this could also be true for the Central and Western Alps.

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