Zeitschrift: Eclogae Geologicae Helvetiae

Herausgeber: Schweizerische Geologische Gesellschaft

Band: 73 (1980)

Heft: 2: Symposium alpine geotraverses with special emphasis on the Basel-Chiasso profile : Lausanne, 4-5 October 1979

Artikel: Heat flow and the metamorphic evolution of the Eastern Alps

Autor: Oxburgh, E. Ronald / England, Philip C.

DOI: https://doi.org/10.5169/seals-164961

Eclogae geol. Helv.	Vol. 73/2	Pages 379-398	7 figures in the text and 1 table	Basle, July 1980

Heat flow and the metamorphic evolution of the Eastern Alps[1])

By E. RONALD OXBURGH[2]) and PHILIP C. ENGLAND[3])

ABSTRACT

The interpretation of heat flow observations within young mountainous terrains such as the Eastern Alps is difficult because corrections must be applied which are comparable with the magnitude of the measurements themselves. These corrections must take account, among other factors, of rugged topography, anisotropic thermal conductivity, regional uplift history, and recent local erosion. The more reliable heat flow values within the axial zone of the Eastern Alps range from about 75 to 90 mW/m². In the Molasse basin to the north thermal gradients measured in oil wells suggest that the heat flow there is unlikely to be less than 100 mW/m².

In reconstructing the thermal history of the region, information can be derived from the study of metamorphic mineral parageneses, isotopic and chemical geothermometers, radiometric age determinations and the present distribution of heat-producing elements. These, when interpreted along with theoretical considerations, suggest that in the axial zone of the Eastern Alps, the interface between basement and cover was at a temperature of about 550 ± 50 °C, 30 my ago. The depth of the interface at that time is less well constrained at 7.5 ± 2 kbar (16-29 km).

It is possible to combine these constraints with those given by the surface heat flow measurements to give approximate estimates of the thermal budget of the Eastern Alpine metamorphism. Because of the uncertainties involved in this, there may be discrepancies between predicted and observed heat flow values: the heat flow might be expected to be higher by as much as 30%. It is also unexpected that the axial zone heat flow should be lower than that in the Molasse basin. The heat flow observations are possibly systematically perturbed by water flows which are orographically controlled, and which depress heat flow values in the axial zone by lateral convective heat transfer in the upper crust.

ZUSAMMENFASSUNG

Die Interpretation von Wärmeflussdaten aus jungen Orogenen wie den Ostalpen wird dadurch erschwert, dass die Korrekturen der gemessenen Werte oft in derselben Grösse liegen wie die Messwerte selbst. Die Korrekturen werden angebracht, um thermische Effekte von Gebirgsrelief, Anisotropie der Wärmeleitfähigkeit, Hebungsgeschichte, rezente, lokale Erosion usw. zu berücksichtigen. Verlässliche Wärmeflusswerte innerhalb der Axial-Zone der Ostalpen liegen im Bereich von 75 bis 90 mW/m². Im nördlich vorgelagerten Molasse-Becken wurden in Erdöl-Explorationsbohrungen geothermische Gradienten beobachtet, nach welchen der Wärmefluss dort kaum niedriger als 100 mW/m² sein kann.

Die notwendigen Ausgangsdaten für die Rekonstruktion der thermischen Geschichte der Ostalpen werden durch metamorphe Mineralparagenesen, Isotopen- und chemischen Geothermometer, absolute Altersbestimmungen sowie durch die heutige Verteilung der radiogenen Wärmequellen geliefert. Diese Daten sowie entsprechende theoretische Überlegungen legen es nahe, dass die Grenzfläche Kristallin/

[1]) Cambridge Earth Science Series: Contribution N° 1.

[2]) Department of Earth Sciences, Downing place, Cambridge, CB2 3EW (England).

[3]) Department of Earth Sciences, Bullard Laboratories, Madingley Rise, Cambridge, CB3 OEZ (England).

Schieferhülle vor 30 Mio Jahren in den Ostalpen auf einer Temperatur von 550 ± 50 °C lag. Die damalige Tiefenlage kann weniger genau angegeben werden (16–29 km entsprechend 7.5 ± 2 kbar).

Die Kombination dieser Randbedingungen mit den Wärmeflussdaten ermöglicht, das thermische Budget der alpinen Metamorphose in den Ostalpen abzuschätzen. Wegen den erwähnten Unsicherheiten können Unterschiede zwischen berechnetem und beobachtetem Wärmefluss auftreten: Die Modellrechnung führt zu bis 30% höhere Wärmeflusswerte. Es ist auch unwahrscheinlich, dass der Wärmefluss in der Axial-Zone niedriger ist als im Molasse-Becken. Die Wärmefluss-Beobachtungen sind möglicherweise durch Bewegungen des Gebirgswassers systematisch gestört, welche den Wärmefluss in der Axial-Zone infolge von lateralem, konvektivem Wärmetransport in der oberen Kruste erniedrigen.

1. Introduction

This paper is written in response to an invitation to review and synthesize the information available on the metamorphic, thermal and geochronological evolution of the Eastern Alps. Information on these topics is so abundant that under normal circumstances such a paper would be very long. It is fortunate, therefore, that there are several recent reviews available on separate aspects of the complete problem, and this paper will be devoted largely to the interpretation of the combined geophysical and geological data in terms of the thermal evolution of a metamorphic belt.

This combined approach is necessary if we are to gain an understanding of the pressure-temperature-time (*PTt*) paths experienced by the rocks of the belt and to arrive at some estimate of the relative importance and magnitudes of the various processes which contributed to the thermal budget of the Alpine metamorphism.

2. Heat flow measurements

As will be discussed later, the thermal development of a metamorphic belt is governed largely by two factors: by the length of time which elapses between the main episode of internal movements (e.g. thrusting, etc.) and the onset of rapid uplift; and by the amount of heat which flows through it during this time. In the Alps, measurement of present day heat flow gives us an estimate of the latter factor, but because of the erosion over the last 20–30 Ma it is an obscured estimate (see below).

Heat flow determinations are difficult to make in active orogenic areas and for that reason reliable observations are rather few. Čermák & Rybach (1979) have collated all information available up to 1978 for the Alpine area. The majority of measurements were made in shallow lakes with essentially the same techniques as are used in oceanic measurements. The unreliability of data collected in this way has been emphasized by England (1978*b*) and Haenel (1979); the magnitudes of the corrections which have to be carried out for regional erosion, the local scouring of valley bottoms during the last glaciation, and for the effects of refraction of heat are uncertain, and large by comparison with the quantity being measured.

In contrast, measurements made in deep holes or tunnels are usually more reliable as the tunnels often penetrate around 1 km underground (and so are less prone to the near-surface perturbations which bedevil lake measurements) as well as sampling heat flow through a greater volume of rock. More reliance is therefore placed on the relatively few tunnel measurements than on the more numerous, and generally higher, determinations in Alpine and Peri-Alpine lakes.

Probably the most reliable heat flow value published for the Eastern Alps is that calculated by CLARK (1961) for the Tauern railway tunnel. This value of 1.8 μcal cm^{-2} sec^{-1} (75 mW m^{-2}) is in agreement with unpublished values measured by the Oxford Heat Flow Group in the eastern half of the Tauern window during the construction of road and hydroelectric tunnels.

Immediately to the north and south of the Alpine chain in the Peri-Alpine Molasse and Po Basin depressions, there are no published reliable measurements. It is of importance, however, to note the temperature information obtained during the drilling operations carried out for oil exploration in the Molasse Basin.

KUNZ (1975) shows that down to depths of several kilometers in the Molasse zone mean thermal gradients in the range 30–50 °C/km occur; although there are important and unquantified sources of error associated with measurements of this kind, the errors are likely to have lowered the apparent values of thermal gradient rather than raised them. This leads to the conclusion that the vertical thermal gradient in the Molasse zone may be 1.5–2.5 times as high as that measured in the axial zone.

Without a good knowledge of the values of thermal conductivity which are typical for each area it is not possible to make a proper comparison of the rates of heat flow. A large number of measurements of conductivity have, however, been carried out on rocks from the axial zone (CLARK 1961; ENGLAND 1976; Oxford Heat Flow Group, unpublished data) and are relatively high (mostly 3–3.5 Wm^{-1} K^{-1}) as would be expected in metamorphic rocks which have very little pore space and are commonly rich in quartz or carbonate. Given these values and plausible estimates for the conductivity range of the quartzose Molasse sediments, it is highly unlikely that the heat flow in the Molasse zone is lower than that in the axial zone and it could well be 50% higher.

3. Regional metamorphic conditions in the Eastern Alps

The metamorphic mineral assemblages observed in the Alps, and their age relations, are usually interpreted as resulting from a phase of crustal compression and thickening, resultant upon continental collision of Late Cretaceous or early Tertiary age (e.g. ERNST 1973; OXBURGH 1972; OXBURGH & TURCOTTE 1974; BICKLE et al. 1975; HAWKESWORTH et al. 1975).

The lines of evidence for this point of view are discussed extensively by these authors; in this and the next section we confine ourselves to a brief description of the metamorphic history and outline the metamorphic and geochronological evidence on which it is based. Figure 1 shows a sketch geological map of the Eastern Alps, and Figure 2 gives the tectonic stratigraphy of the region.

During the Mesozoic there was epicontinental sedimentation over the southern margin of the European continental region; this margin contained some of the units which are now exposed in the Pennine zone of the Alpine Chain and which in the Eastern Alps consisted of a complex of Hercynian orthogneisses and paragneisses (the Zentralgneis).

Towards the end of the Mesozoic the closure of the ocean to the south of the European land mass culminated, in the Eastern Alps, with the emplacement of at least 10 km thickness of southerly derived Palaeozoic or older schists and gneisses (the Altkristallin) and its sedimentary cover on to the European basement. Much of the strain associated with this overthrusting was accommodated by a zone of predominantly Mesozoic sediments and volcanic rocks sandwiched between the gneissic units and there appears to have been relatively little internal disruption of the Altkristallin during emplacement. The highly strained zone is now exposed as

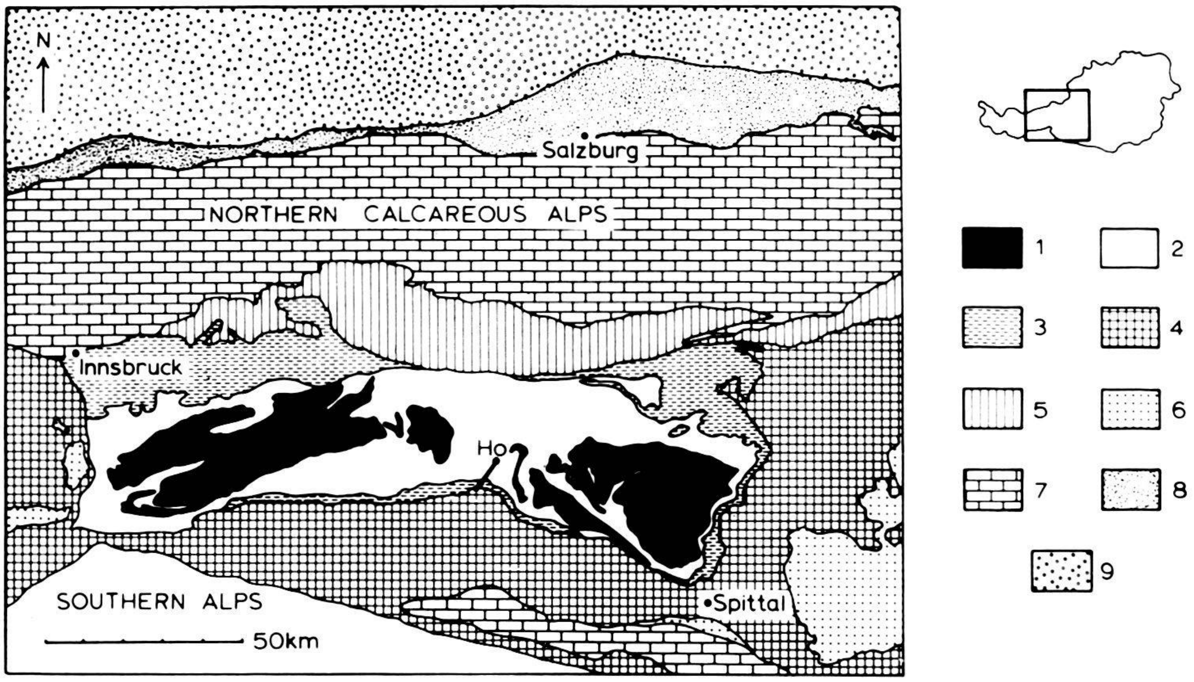

Fig. 1. Sketch map of the Eastern Alps. 1 = Penninic Basement complex; 2 = Peripheral Schieferhülle; 3 = Unterostalpin; 4 = Altkristallin; 5 = Grauwackenzone; 6 and 7 = Palaeozoic and Mesozoic cover of the Altkristallin sheet; 8 = Flysch and Helvetic Zones; 9 = Molasse.

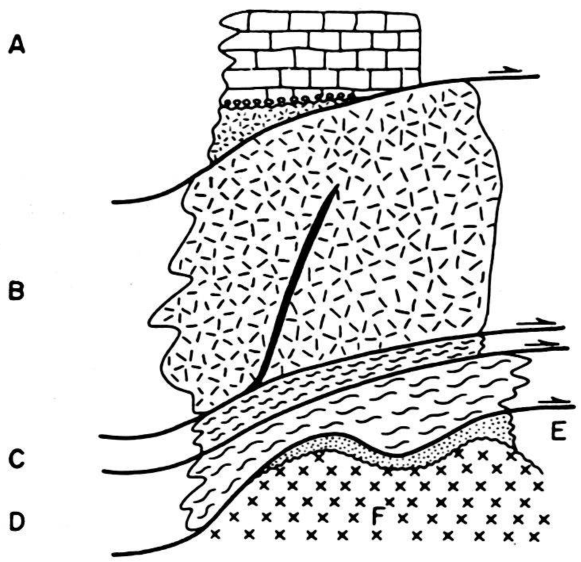

Fig. 2. Sketch of the main tectonic units of the central Eastern Alps (after OXBURGH & TURCOTTE 1974). A = Oberostalpin-sediments with local metamorphics beneath thrust over B = Altkristallin sheet, thrust over C = Unterostalpin, thrust over D = Peripheral Schieferhülle melange, thrust over F and E = the autochthonous European basement with its locally preserved sedimentary cover.

the metasedimentary and metavolcanic rocks of the Peripheral Schieferhülle and the Unterostalpin thrust complex, and it is largely these rocks which provide the evidence of Alpine metamorphic conditions.

It seems likely (Section 4) that the Austroalpine nappes were emplaced by 60–65 my b.p., yet relatively little sedimentation occurred in the Molasse Basin until the Oligocene (e.g. OXBURGH 1968); indeed this observation seems to hold for all the peri-Alpine sedimentary basins (e.g. GUILLAUME & GUILLAUME 1980).

Rapid erosion, initiated in early Oligocene times, has removed much of the allochthonous cover from the area and exposed, in the Tauern window, Pennine basement metamorphosed to amphibolite facies. It is the aim of this review to examine the metamorphic and geochronological observations of this history in an attempt to determine the *PTt* paths followed by rocks in such an orogenic belt and, by linking these observations with geophysical constraints (principally heat flow), to cast light on the thermal requirements for the observed metamorphism – and hence, perhaps, on the underlying tectonic processes.

Major reviews of metamorphism in the Alps have been published by FREY et al. (1974) and ERNST (1973), and these document extensively the main Tertiary (Alpine) metamorphism.

Since the publication of these reviews the most important development is the clear recognition of an earlier and widespread high pressure and low temperature metamorphic event (often referred to as Eo-Alpine) which widely affected the Mesozoic metasediments but is largely obliterated by the later greenschist-amphibolite facies event (FRY 1973; MILLER 1974; HOLLAND 1977, 1979). This high pressure event is not well dated in the Eastern Alps but in the Western Alps a similar event is described in the Zermatt/Saas Fee area and constrained to the interval 60–90 my b.p. (FREY et al. 1979; HUNZIKER 1974).

It is not clear whether the ultra-high pressure eclogite blocks preserved in the south of the Tauern window are related to this event or are much older (HOLLAND 1979; ENGLAND & HOLLAND 1979).

In the Pennine zone of the Eastern Alps, the evidence for the Tertiary (Alpine) metamorphism is almost ubiquitous; this ranges in grade from high amphibolite facies at the deepest exposed levels to greenschist facies at the higher levels. More precise estimates of temperature and pressure conditions attained at particular places within this sequence have been derived by BICKLE (1973), HOLLAND (1977) and DROOP (1979). These estimates are summarized in Table 1.

Studies of oxygen-isotope fractionations between cogenetic mineral pairs are also able to provide information about temperatures reached during the thermal history of a region. Extensive work of this kind has been carried out in the Western Tauern (FRIEDRICHSEN et al. 1973; HOERNES & FRIEDRICHSEN 1974, 1978). They show an E–W trending temperature maximum with a value slightly over 600 °C; the maximum corresponds approximately to the zone of present-day highest topography and is flanked by zones of lower recorded temperatures to north and south falling to values between 400 and 500 °C on the northern and southern margins of the Tauern window.

Estimates of temperature may also be made from a study of the compositions of coexisting calcite–dolomite pairs (e.g. BICKLE & POWELL 1977). There are several

Table: *Eastern Alpine metamorphic conditions.*

Locality	Level	Estimate Pressure (kbar)	Temperature (°C)	Source
Alpine metamorphism (35 M.y.b.p.)				
S.E. Tauern Window	Basement/cover interface	7 ± 1.0	570 ± 20	Droop (1979)
	2 km below interface	7 ± 1.5	590 ± 20	Droop (1979)
S. central Tauern Window	Basement/cover interface	6.5 ± 1.0	520 ± 20	Holland (1977)
	3 km below top of basement	10 ± 2	550	Holland (1977)
Western Tauern Window	Basement		600	Friedrichsen et al (1973) Hoernes and Friedrichsen (1974, 1978)
Eo-Alpine event				
S. central Tauern Window	Basement cover interface		350°C	Holland (1977)

difficulties with this kind of temperature indicator, some of which apply equally to the others mentioned earlier. The most serious is that one member of the pair may continue to undergo chemical exchange with its environment until late in the cooling phase of metamorphism thus invalidating the assumption of equilibrium between the pair; alternatively both may exchange but differentially with the same consequence. In ideal circumstances temperature indications such as this may provide minimum estimates of the maximum temperature reached by the system insofar as they presumably record the temperature of the system at some time between the thermal maximum and the present. Estimates of metamorphic conditions discussed in this section are given in Table 1.

4. Geochronological information

In principle, geochronological data may be used to determine the time elapsed since a particular volume of rock cooled through a particular temperature interval or, in some situations, the time of growth of a particular mineral phase. As such, these observations may provide very powerful constraints on the thermal history of an area. In practice their usefulness is limited by uncertainties in the temperature dependence of the diffusion characteristics of the various radiogenic daughter products through different silicate structures.

Recent reviews of the geochronological work in the Eastern Alps have been presented by RAITH et al. (1978), SASSI et al. (1978), FREY et al. (1974), BORSI et al. (1978) and HAWKESWORTH (1976).

The time of emplacement of the Altkristallin upon the Pennine zone is of considerable importance for the understanding of the thermal history; it is, however, rather poorly constrained and insofar as the Altkristallin metamorphic thrust complex might have been emplaced as a series of discrete thrust units, these movements need not necessarily have occurred simultaneously along the Alpine chain. An age of emplacement of about 65 my b.p. is consistent with most stratigraphic and geochronological observations, although emplacement as early as 80 my b.p. is possible (HAWKESWORTH 1976). It is important to note that this span of ages represents uncertainty in the *timing* of overthrusting and does not give an estimate of its *duration*.

Throughout the Eastern Alps, the Pennine basement orthogneisses give whole rock *Rb/Sr* isochron ages in the range 245–280 my suggesting that in the late Permian there was a major phase of post-tectonic granodioritic intrusive igneous activity, perhaps with the general scale and character of that of the Sierra Nevada batholith of the western USA. The metasedimentary Mesozoic series which rest upon this basement (Peripheral Schieferhülle and Unterostalpin) are locally autochthonous but largely parautochthonous or allochthonous; they all show cooling ages which lie in the range 13–35 my. The youngest ages are found in the west and the oldest in the east (RAITH et al. 1978).

As shown later such ages are likely to correspond to or slightly post-date the onset of major uplift and erosion in the eastern Tauern window. Their early Oligocene values are in good agreement with the stratigraphic evidence quoted in Section 3.

The massive allochthonous basement thrust complex formed by the Altkristallin of the Upper East Alpine nappe, overlies the metasedimentary series just described and contributed to their conditions of metamorphism by tectonic burial. As might be expected the lower parts of the Altkristallin are somewhat affected by the elevation of temperature in the underlying metasediments and micas show partial or complete loss of radiogenic argon for a thousand or so meters above the base.

The Altkristallin itself shows a polymetamorphic early history which is variable from place to place; there is evidence of regional metamorphism at about 300 my which is locally overprinted by a later event between 80 and 100 my (BREWER 1969; BORSI et al. 1978).

Tertiary igneous activity is relatively unimportant within the Eastern Alps. None is known in the deepest tectonic units (i.e. the Pennine Zone) but the Altkristallin is penetrated in its southern parts by members of the so-called Peri-Adriatic tonalite suite of shallow intrusions. These give cooling ages of 35–40 my.

5. The processes controlling thermal evolution of the Eastern Alps

We turn now to the question of how these different kinds of temperature, time, pressure and heat flow information may be used together to constrain models for the thermal evolution of this or any other metamorphic belt. It must at the outset be confessed that even in a metamorphic belt as well studied as the Alps it is still impossible to achieve this end fully. The reasons will become apparent.

We begin by considering the pressure-temperature-time (PTt) history of a small volume of rock which starting at the earth's surface is slowly buried, reaches a maximum depth of burial, and then is slowly brought to the surface one more (Fig. 3*a*); if the process of burial is sufficiently slow (e.g. $\ll 1$ mm/yr) the rock will be close to thermal equilibrium throughout the burial.

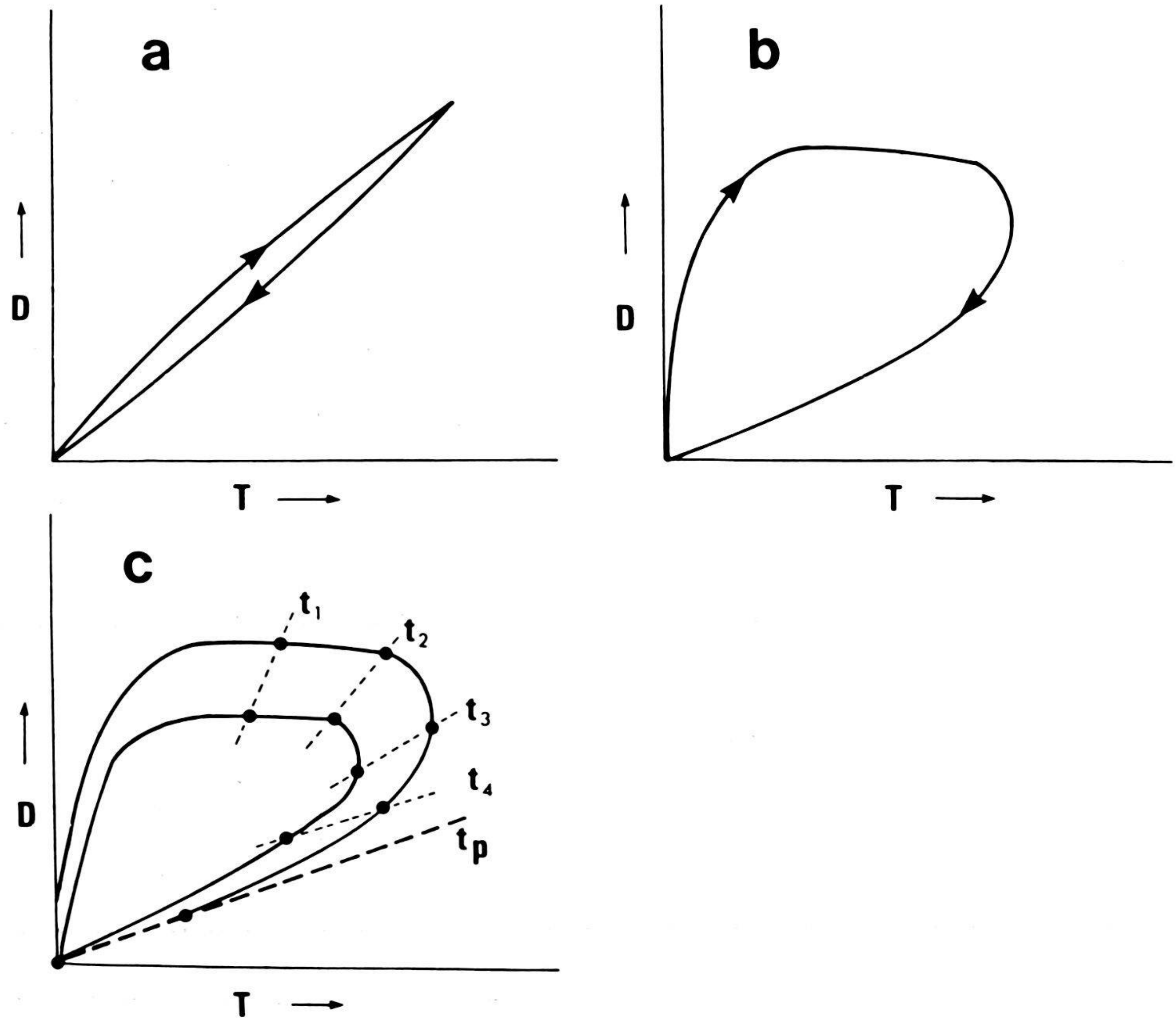

Fig. 3. Sketch PTt paths for imaginary burial and erosional histories. *a*) Rock buried and exhumed very slowly so that its temperature remains close to equilibrium at all parts of the path. *b*) Rock buried rapidly, perhaps by overthrusting, with a period of warming accompanied by little erosion (therefore little decompression), followed by a period of more rapid erosion at a uniform rate. *c*) As 3*b* except that the trajectories for two rocks separated in depth by a fixed distance are shown. The dashed lines join P/T points on the two paths, at times t_1, t_2, t_3, t_4, in the past, and thus indicate approximate geothermal gradients in the region of the sample rocks. The heavy dashed line indicates the present day geothermal gradient.

If uplift and erosion of the overburden took place similarly slowly and (improbably) there had been no change in the thermal properties of the rocks during the burial, the upward trajectory should be identical to that of burial except in the opposite sense. The non-coinidence of the two trajectories in Figure 3*a* implies that although both burial and uplift were slow, both were slightly too rapid for equilibrium to be maintained.

Although it may happen that in certain slowly subsiding sedimentary basins rocks follow *PTt* trajectories such as just described, this probably happens rarely, if ever, in metamorphic belts; a trajectory such as shown in Figure 3*b* is probably more common and is certainly more appropriate for an overthrust terrain such as the Alps. If a thick pile of thrust sheets or nappes is emplaced at rates of centimeters per year (i.e. the characteristic velocity of plate tectonic movements) any rock which is buried beneath the pile experiences a very rapid increase in pressure but there is little time for conductive heating. Having reached its burial depth the rock will there undergo steady conductive warming (the horizontal portion of the curve, Fig. 3*b*) until such time as it reaches equilibrium with its environment or until the warming is terminated by removal of the overlying cover rocks by uplift and erosion. It will be seen from Figure 3*b* that the decreasing pressure limb of the trajectory comprises two parts, a heating part and a cooling part. *Even though uplift and erosion has begun, the rock continues to warm* because heat is being supplied to it from the conductive thermal relaxation of the rock pile, and from radiogenic heat sources, and because it is still sufficiently deeply buried to be unaffected by the erosion surface which is moving down through the rock pile towards it. In this situation the "thermal maximum" is attained when sufficient of the overburden is removed for the effect of upward cooling to be felt and the temperature begins to fall. Note that the rock does not experience the conditions of maximum temperature and maximum pressure at the same time (ENGLAND & RICHARDSON 1977).

We now consider the trajectories of a pair of rocks, one situated several kilometers immediately beneath the other. We assume that this spatial relationship is maintained during burial and return to the surface. A pair of trajectories must now be constructed having the general form of those shown in Figure 3*c*, the inner loop corresponding to the upper of the two rocks. If particular times (t_1, t_2 ...) are now identified on the trajectories it is clear that a line joining the same time point on the different trajectories corresponds to a portion of the actual geothermal gradient for that time. The dashed lines for $t_{1,2,3,4}$ therefore show the change with time of the geothermal gradient over the depth interval separating the two rocks. Finally if the upper member of the pair is exposed at the surface today, the last member of the family of geothermal gradients (heavy dashed line, t_p) represents the present geothermal gradient. The lower member of the pair is, of course, still buried. It should be noted that all the segments of gradient have for simplicity been drawn as straight lines; in reality all would have some curvature.

It is clear from the foregoing discussion that the type of metamorphic history experienced by a rock will be dependent on the relations between at least three timescales: the duration of the burial process, the duration of the uplift process, and the decay time of the thermal disturbance introduced by overthrusting (or indeed by any form of crustal thickening).

The time taken for tectonic burial in the case of the Alpine overthrusting event is hard to estimate; it is probable, however, that overthrusting occurred at velocities typical of plate tectonic movements ($\gtrsim 3$ cm/yr) and thus that the *duration* of the ~ 100 km) of overthrusting was less than 3 my.

If the time required for burial is short by comparison with the time required for the decay of the thermal disturbance produced by overthrusting, we are justified in considering the burial to be rapid in the sense of the discussion of the initial burial portions of the *PTt* paths of Figures 3*b* and 3*c*.

Oxburgh & Turcotte (1974, Fig. 3) discuss the kind saw-toothed thermal profile which might be expected to result from the rapid overthrusting which has just been outlined. They show that the time taken for the temperature at the base of a thrust sheet to progress 1/e of the way to equilibrium (that is: from 0.5 to 0.68 of the equilibrium temperature) is approximately

$$t_{1/e} = \frac{2d^2}{\kappa}, \tag{1}$$

where d is the thrust sheet thickness and κ its thermal diffusivity. A typical thermal diffusivity for crustal rocks is 30 km^2 my^{-1} and a likely range of thrust sheet thickness in the Eastern Alpine case is 15–30 km; thus the time constant for conductive relaxation is in the range of 15 to 60 my. These figures are large by comparison with the 3 my required for the emplacement of the thrust pile and we are justified in assuming that burial is rapid.

The approximately isobaric warming portions of the *PTt* paths in Figures 3*b* and 3*c* reflect that interval between the thrust sheet emplacement (ca. 65 my b.p.) and the onset of significant erosion (ca. 35 my b.p.). A temperature rise of some hundreds of degrees Celsius is expected during this interval (Oxburgh & Turcotte 1974; Bickle et al. 1975; England 1978*a*) as its duration is comparable with the thermal relaxation times just discussed. The magnitude of this rise depends on such variables as the thrust sheet thickness and the amount of heat flowing through the basement beneath it. We shall discuss later how the observed metamorphic grade, with the heat flow and geochronological controls, may be used to give rough estimates of the magnitudes of these variables.

The final portion of the *PTt* paths considered in Figures 3*b* and 3*c* is the decompression phase which has two parts separated by the thermal maximum: the first portion is either isothermal or with some warming, and passes into the second which is characterized by cooling with decompression as the rocks approach the surface.

For the majority of geological phenomena which can be observed and measured at the surface today it is the shape of this final decompressional phase in *PTt* trajectory that is important. It is clear that if uplift and erosion were very slow the trajectory would have the shape shown in Figure 3*a* and if they were virtually instantaneous rocks would reach the surface without change from the maximum temperature reached at depth. At any instant the thermal profile depends on the relative importance of two competing processes. A crude comparison of the rates may be made by use of the Peclet number (Pe)

$$Pe = \frac{d^2/\kappa}{d/u} = \frac{ud}{\kappa}, \tag{2}$$

where d is a characteristic distance (e.g. depth of burial), u is the rate of erosion, and κ is the thermal diffusivity. The upper of term in the left hand expression may be compared with equation (1) and is a characteristic time for the conductive transfer of heat; the lower is the characteristic time for a buried rock to reach the surface. If Pe is much greater than one the effects of conductive cooling may be ignored and if it is much smaller than one, they dominate and erosional effects are subordinate.

Unfortunately, in the case of the Eastern Alps, neither simplification can be made: we know that the erosional timespan is about 30 Ma, and that the depth of erosion is between 15 and 30 km for the axial part of the belt. This gives a Peclet number of between 0.25 and 1 indicating that the transport of heat by upwards movement of the rocks is of equivalent importance to that of conduction.

This conclusion holds for many orogenic belts and its general implications are discussed more fully by ENGLAND & RICHARDSON (1977) and RICHARDSON & ENGLAND (1979).

Interpretation of geochronological and metamorphic observations

We now consider how the ideas discussed above, and those outlined by ENGLAND & RICHARDSON (1977) and RICHARDSON & ENGLAND (1979) may provide a conceptual framework for the interpretation and synthesis of the geological observations in the Eastern Alps. This is done with the help of one-dimensional finite difference calculations on a hypothetical situation having some similarity to the Alps; the details of this kind of numerical study are given by ENGLAND & RICHARDSON (1977). Trajectories are calculated for three points lying above each other at a vertical separation of 5 km, the highest of these points being initially on the land surface. They are taken to have been buried rapidly by an overthrusting process to depths of 30, 35 and 40 km, respectively. The burial occurred at 65 my b.p. and at 35 my b.p. uplift and erosion began. The resulting *PTt* trajectories are shown in Figure 4 and the values of the other input parameters are given in the legend. Provided that there is a relatively long period between burial and the onset of uplift, i.e. some tens of millions of years, the later thermal history is insensitive to the details of the burial process and the earlier parts of the trajectories are not shown.

The first point to emphasise is that *every individual rock particle in the metamorphic belt has its own, and unique, PTt path.* The possibility that there are significant differences in *PTt* paths must be taken into account when information derived from rocks in different areas is used to establish the thermal history of a region. The spacing of the time marks along individual trajectories shows how the rate of cooling of any particular rock increases as it approaches the surface. The uppermost rock cooled 300 °C in the last 10 my, 150 °C in the previous 10 my and about half that in the previous 10 my. During this last 10 my the trajectories of all rocks are similar because of rapid cooling to the surface. The present day geotherm is shown by the series of stars; the deepest trajectory (which is omitted below 20 km for clarity) would terminate at the 10 km point on the present geotherm.

With the help of Figure 5 which shows parts of the 30 (Tr_{30}) and 35 km (Tr_{35}) trajectories from Figure 4, it is possible to examine some of the consequences of the differences in cooling trajectories. The rock following Tr_{30} cools through 500 °C at about 26 my b.p. while that following Tr_{35} does so at about 16 my b.p. There would therefore be an apparent difference of 11 my between radiometric ages based on a mineral such as amphibole, which had a blocking temperature in the area of 500 °C. As the rocks approach the surface the effect of their trajectory difference becomes less; there is an apparent age difference of 8 my for systems with a blocking temperature around 375 °C and about 6 my for 255 °C.

These considerations have a bearing on the interpretation of the fission-track age studies of WAGNER et al. (1977). They describe a systematic decrease in fission track age with topographic elevation at a number of sites in the Central Alps. They interpret this variation in terms of differences in times of cooling through the 120 °C isotherm, the higher samples giving the older ages because they cooled earlier. They show that in a number of areas there is a nonlinear relationship between age and topographic elevation and use the differences in line slope along with other geochro-

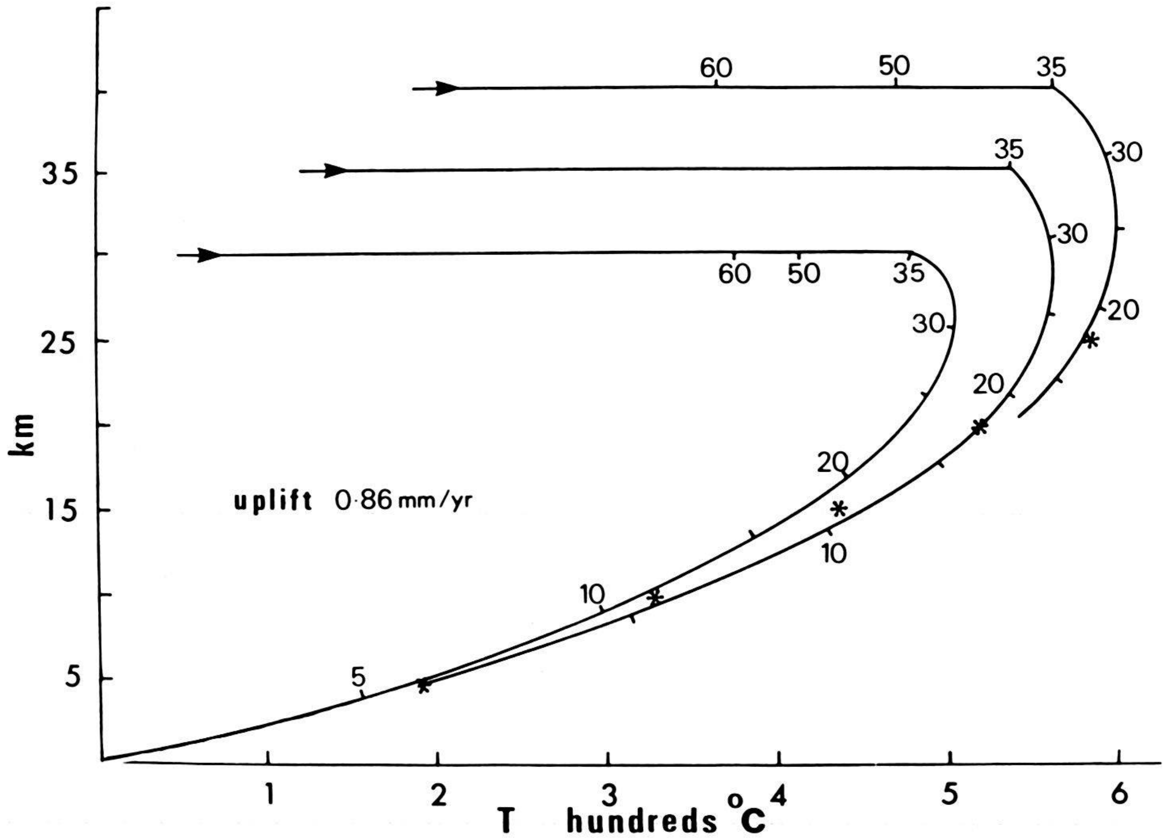

Fig. 4. *PTt* paths for three rocks in a model which resembles the East Alpine metamorphic episode. The rock has thermal conductivity 2.5 Wm^{-1} K^{-1} and thermal diffusivity 10^{-6} m^2 s^{-1} throughout. Radiogenic heat production is distributed as $A(Z) = A_0 e^{-Z/D}$ in the rock, where A_0 is 3 μW m^{-3} and D is 8 km, temperatures are initially in equilibrium with this and with a constant heat flux of 25 mW m^{-2} 65 my b.p.; a thrust sheet 30 km thick, of identical thermal properties is emplaced on the basement. At 35 my b.p. erosion commences at 0.86 mm/yr, which exposes the top of the basement by the present day. The lines show the *PTt* paths followed by rocks at the top the basement and 5 and 10 km below it. Tick marks on each path are every 10 my, starting 60 my b.p. Note the slight negative geothermal gradient implied for 60 my b.p. (5 my after thrusting) with temperatures around 350 °C extending at least 10 km below the thrust (pressures of 8–11 kbar). Stars indicate temperatures at the present day.

nological observations to calculate variations in rate of uplift and erosion. Uplift rates derived in this way are broadly consistent with those based on geodetic observations and they seem to show both recent increases and decreases in uplift rate in different areas. The specimens used by WAGNER et al. must have followed slightly different *PTt* trajectories and examination of Figure 5 shows that if rocks following Tr_{30} and Tr_{35} were today exposed at the surface the apparent rate of uplift calculated in this way would have increased from about 0.5 mm/yr 15–25 my ago to a rate very close to the true rate of 0.86 mm/yr measured on low temperature systems such as fission tracks. Thus apparent nonlinearity of uplift may result from a comparison of specimens following different trajectories during uplift of a constant rate. The effect shown in Figure 5 is much smaller than that recognised by WAGNER et al. (e.g. changes from 0.3 to 1.1 mm/yr over six million years) but it would not be difficult to construct models in which the effect was much larger than we have shown, and it is interesting to note that several of their cooling curves (WAGNER et al. 1977, Fig. 4) show an acceleration in apparent erosion rate over the last 20 my. On the other hand, a trajectory given by a constant rate of uplift cannot explain rates which seem to decrease with time. Furthermore, the systematic patterns of

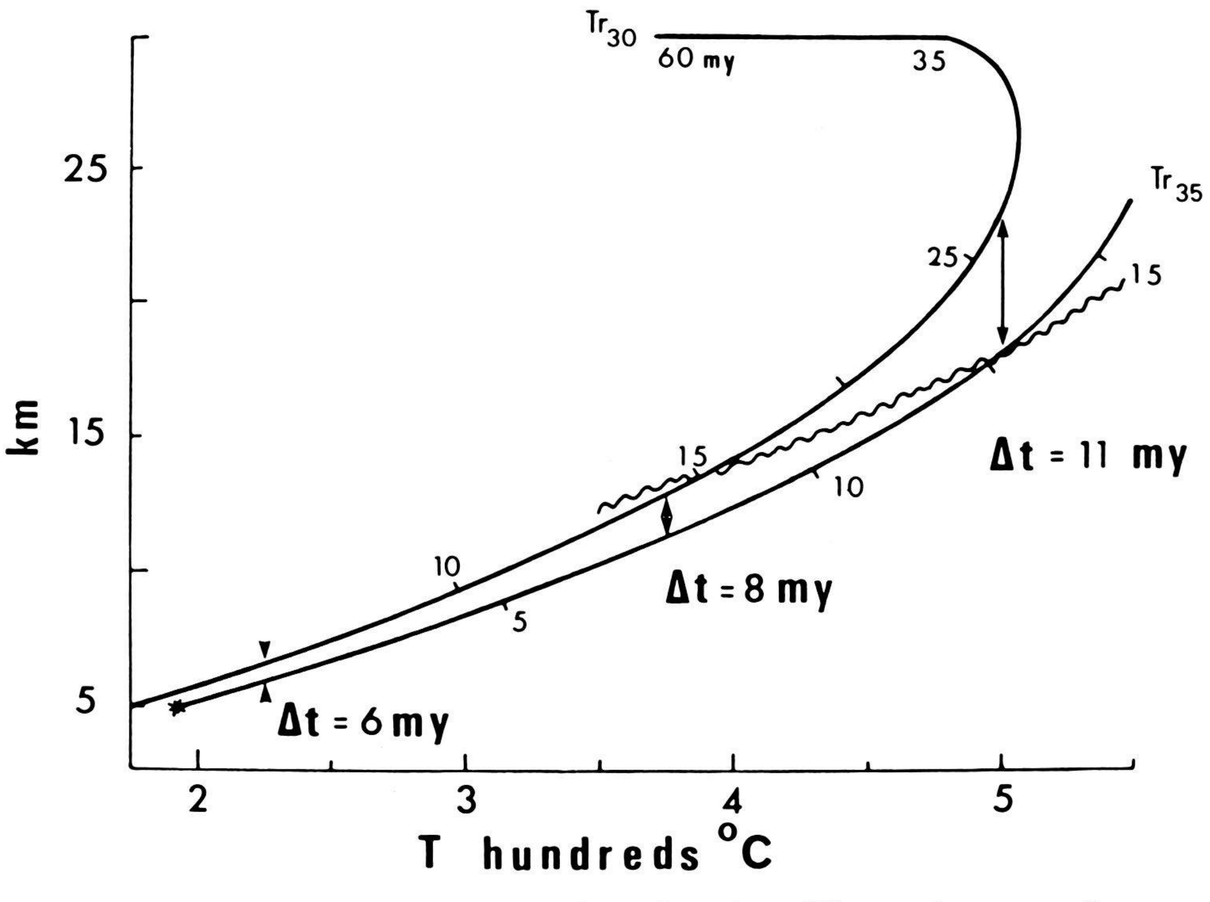

Fig. 5. Expanded portion of two paths from Figure 4, to show differences between cooling ages, recorded by the two sample rocks. The differences, Δt, are illustrated for isotopic systems with closing temperatures of 500 °C, 375 °C and 225 °C. The wavy line joins *P/T* points at 15 my b.p., illustrating that a phase of deformation at this time would predate a phase of mineral growth at, say, 450 °C in the upper rock, and post date it in the lower one.

regional variation recognized by WAGNER et al., strongly support their interpretation of true regional variations in uplift rate.

As mentioned earlier the elevation of temperature in the lower part of the Altkristallin during the Alpine metamorphism was sufficient to bring about partial or complete resetting of the various isotopic systems. A zone of partial resetting for any particular system overlies a zone of complete resetting immediately above the Pennine zone. In terms of Figure 5 this means that the trajectories for successively deeper levels have involved higher maximum temperatures, and successively longer periods above the blocking temperature for diffusion of the appropriate species. As argued by HAWKESWORTH (1976) for the southeast Tauern, the highest - topographically and numerically - ages which are completely reset, should approximate to the time of the thermal maximum at that depth. In fact they will slightly post-date it. For the southeast Tauern they are close to 35 my.

A second and quite different effect of trajectory difference is also shown in Figure 5. A phase of deformation occurring at 15 my b.p. would clearly affect the upper part of the rock pile at lower temperatures than the deeper part. We note, however, that a late phase of mineral growth occurring say at about 450 °C, would

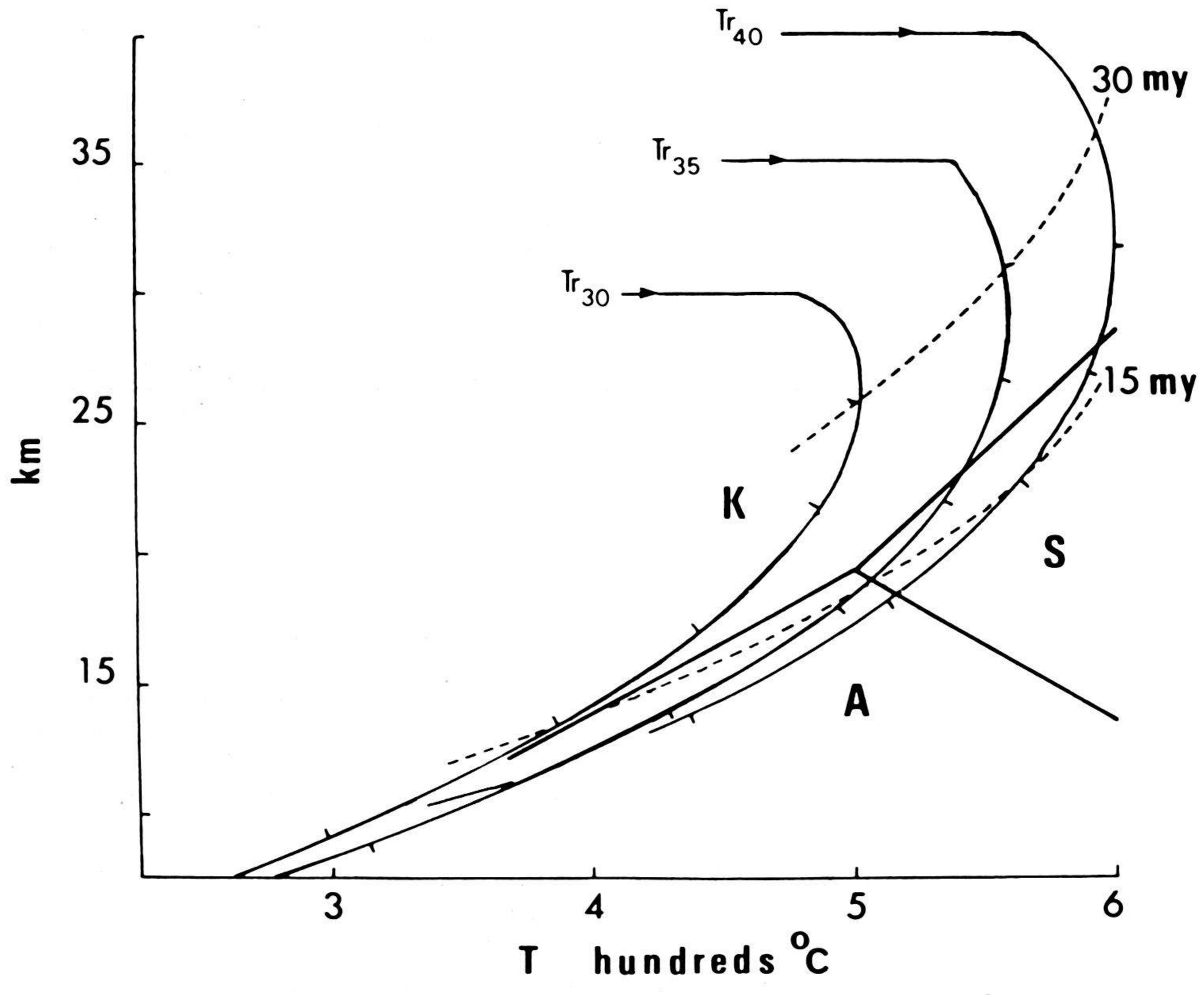

Fig. 6. As Figure 4, showing a portion of the alumino-silicate phase diagram and the 30 and 15 my geotherms.

be pre-tectonic for Tr_{30} (i.e. the shallower part) and post-tectonic for Tr_{35}, perhaps making the tectonothermal history of the area difficult to interpret.

Figure 6 is once more based on Figure 4 but shows all three trajectories. Parts of the 30 my b.p. and 15 my b.p. geotherms are shown and their marked curvature is apparent, a steepening of the gradient occurring in the uppermost zone where the effects of erosion are felt most strongly. The figure also illustrates very clearly how the thermal maxima for each trajectory do not occur at the maximum depth of burial, and how the maxima occur progressively later with increasing depth, e.g. 31 my b.p. for Tr_{30} and 20 my b.p. for Tr_{40} (England & Richardson 1977; Richardson & England 1979). We now consider the 'metamorphic geotherm' derived by Bickle & Powell (1977). In an area of exceptional relief and exposure in the central Tauern Peripheral Schieferhülle they showed that the maximum temperatures given by calcite-dolomite geothermometry increased downwards over a vertical distance of several kilometers. They interpret this increase with depth as a direct determination of the thermal gradient at the thermal maximum of metamorphism. Figure 6 shows that the line determined by Bickle & Powell must be one joining thermal maxima at different depths; it must be polychronous and give a *steeper gradient* than was actually present.

A portion of the Al_2SiO_5 phase diagram is also shown, largely because it contains univariant boundaries with both positive and negative slopes. It is of interest to compare the growth histories of phases developed in these two different cases. In the case of positive slopes the phase boundary and geotherm are sub-parallel to each other and the same reaction may take place almost simultaneously over a considerable depth interval. On the other hand in the case of reactions occurring at phase boundaries with steep or negative slope (e.g. sillimanite to andalusite in Figure 6 on Tr_{35} and Tr_{40}) they are expected to be strongly diachronous and have large, if in many cases undetectable, age differences over relatively short distances.

The foregoing discussion gives an idea of the considerations to be taken into account in using pressure/temperature data to define a geotherm. It should be clear that the geotherm changes significantly with time and that P/T information which is available is likely to refer to different depth ranges at different times.

Interpretation of heat flow observations

These considerations, however, do give a better framework for understanding present day heat flow observations - in particular by employing metamorphic and geochronological data to determine the erosional perturbation to heat flow in a region. Conversely it may be possible to use present day heat flow measurements to place some constraints on the conditions of metamorphism, as the following example suggests.

The effect of erosion is to enhance the surface heat flow by bringing hot rocks closer to the surface; because of the time and length scales involved (see above) a surface heat flow determination in the Alps will be influenced by all the post-Oligocene erosional history of the area, although it will be most sensitive to recent variations in erosion rates.

For example, if the uplift interpretation of WAGNER et al. (1977) is correct, the present day conductive heat flow in the Simplon area should be significantly higher than in surrounding areas, because of the high rate of uplift since 20 my b.p. The measurements of CLARK & NIBLETT (1956) suggest that this is indeed the case. They record a value of 92 mW m^{-2} for the Simplon area where WAGNER et al. estimate recent erosion rates of about 1 mm/yr. In contrast the Gotthard area appears to have been uplifted at about half that rate and has a recorded heat flow of about 67 mW m^{-2} (CLARK & NIBLETT 1956). Although the true errors on the heat flow values are large the difference between them is probably significant.

No unique interpretation is possible but if, as seems plausible, the difference in the maximum temperatures deduced for the Gotthard and Simplon regions (350 °C and 500 °C; WAGNER et al. 1977, Fig. 4) reflect differences in burial depth at the peak of metamorphism, this supports the suggestion that these erosion rate differences have persisted over most of the uplift history. The heat flow and geochronological observations may be satisfied by models in which at the onset of erosion in each area the heat flow through the level which is *now* the land surface was around 35 mW m^{-2} (CLARK & JÄGER 1969), with erosion of about 15 km in the Gotthard region and about 30 km in the Simplon. This gives an estimate of the amount of heat flowing through the Pennine basement at the peak of the Lepontine metamorphism. Note first that because of the slow thermal relaxation, this may represent less than three-quarters of the equilibrium heat flow (OXBURGH & TURCOTTE 1974) and secondly that it is not the only estimate compatible with present day heat flow.

CLARK & JÄGER give an illuminating discussion of the interaction between the various parameters in this kind of problem and, most importantly, show that parameter estimation in orogenic processes although difficult, is possible.

The metamorphic and geochronological data summarized above give us sufficient constraints to attempt an estimation of the important processes driving the East Alpine metamorphism. The arguments which follow are necessarily brief, but reference is made to sources where some of the more terse statements are given fuller treatment.

The principal parameters affecting the metamorphic conditions attained during the burial history which has been outlined are the thermal conductivity, the depth of burial and the heat supplied by radiogenic sources in the crust and from the upper mantle below the thickened pile (OXBURGH & TURCOTTE 1974; BICKLE et al. 1975; ENGLAND 1978*a*).

For geologically reasonable choices of the thermal conductivity, ranges of the other principal parameters can be found which would yield the Alpine metamorphic conditions after 30 my of burial. The extremes of these ranges are about 15 km of burial with the sum of radiogenic heating in the basement and mantle heat flux amounting to 100–120 mW m^{-2} to 30–35 km of burial and 60–70 mW m^{-2} coming from the basement and upper mantle (OXBURGH & TURCOTTE 1974; BICKLE et al. 1975; ENGLAND 1978*a*).

The depth range encompassed by these estimates corresponds approximately to the pressure uncertainties in Table 1; the much tighter bounds on temperature are used to constrain the other parameters. Thus the well-known imprecision of geobarometers limits us to a finite, but still large, range of the important parameters. At one end of the scale, a heat flux from lower crust and upper mantle of 120 mW m^{-2} must almost necessarily involve lithospheric thinning and the high geothermal gradients which are associated with island arcs or crustal rift zones. At the other end, a contribution of 60 mW m^{-2} is close to the average heat flow for continental crust and would require no thermal perturbation apart from burial to account for the Alpine metamorphism.

These possibilities are further constrained by the present day heat flow; as erosion will enhance the surface heat flow, it is clear that the solutions with lowest burial depths and highest heat flow requirements will violate the observation of around 80 mW m^{-2} for the present-day surface heat flow in the area. In fact, the requirement that the Pennine basement be exposed at the surface today means that even those solutions with 30 km burial depth and thus the lowest present heat flow result in a heat flow of 90 mW m^{-2}, 10 mW m^{-2} higher than observed. This discrepancy is within error, however, and could be removed by making special assumptions about the erosional history (e.g. if most of the erosion had happened by about 10 my b.p.).

Oxburgh & Turcotte (1974) and England (1978*a*) propose transient heat flow regimes, such as might be associated with Benioff zone magmatism, which would product temperatures of 500–600 °C beneath 15 km thrust sheets and which would not violate present day surface heat flow constraints.

Until better geobarometers are available there will be no clear choice between metamorphism driven by enhanced heat flow beneath a thin thrust sheet and metamorphism by deeper burial in an environment of average continental heat flow; however, the P/T data are likely to underestimate burial depths (above; England & Richardson 1977).

Conclusions

We have presented a model of the Tertiary Eastern Alpine metamorphism in terms of the thermal relaxation of a continental crust thickened primarily by overthrusting, and have illustrated the PTt paths which may result from such an event. The interactions of these paths with the observables from which they are deduced (mineral assemblages, cooling ages, etc.) are sometimes complicated, and great care must be taken in interpreting such observations.

These complications are reduced if evidence for a succession of P/T conditions is obtained from single specimens. This kind of approach has been applied by Holland (1977) and Droop (1979) in the Eastern Alps, and an interesting variant has been described by Holland & Richardson (1979) who calculate part of a trajectory by deriving temperatures and pressures for the formation of successive zones in strongly zoned amphibole crystals. Their trajectory begins at about 15 kbar and 200 °C and terminates at about 6 kbar and 550 °C and is defined by an array of ten points over that interval.

It is worth pointing out that in the Alps there is clear and unambiguous evidence for a relatively long time interval between burial of the metamorphic pile and its exhumation by erosion (~30 my). This period of burial plays a key role in the thermal history of this or any other area to which this kind of model may be relevant. There appear to be relatively few orogenic belts for which the timing of the movements is sufficiently well-documented to show whether such a hiatus is a general feature of collision-type orogenic belts. The work of RICHARDSON & ENGLAND (1979) suggests that it may be. They propose that in collision margins limited subduction of continental crust may occur if it undergoes a series of pressure-induced phase changes involving the breakdown of plagioclase and a density increase of about 10%. They show that after burial for some tens of millions of years the subducted crust should have warmed sufficiently to revert to lower density mineral assemblages, giving rise to surface uplift and erosion.

In conclusion we emphasise that the density and reliability of our metamorphic, isotopic and heat flow information is as good, or in some respects better, than that for any comparable orogenic belt in the world. On the other hand, it is still inadequate to define the *PTt* trajectories on which the next generation of metamorphic and thermal interpretations of orogenic belts will depend.

Postscript

In conclusion we return to the apparent pattern of heat flow observed today in the Eastern Alps and in the Molasse zone to the north. Although there may be some question about the precise rates and lateral and temporal variation of rates of recent uplift, there is little doubt that the axial zone is a region of net uplift and that the average rate must lie between 0.5 and 1 mm/yr. Equally the Molasse zone has subsided over the same period but the rate was of the order of 0.2 mm/yr. Today the Molasse thermal gradients are about twice as high as those measured in the axial

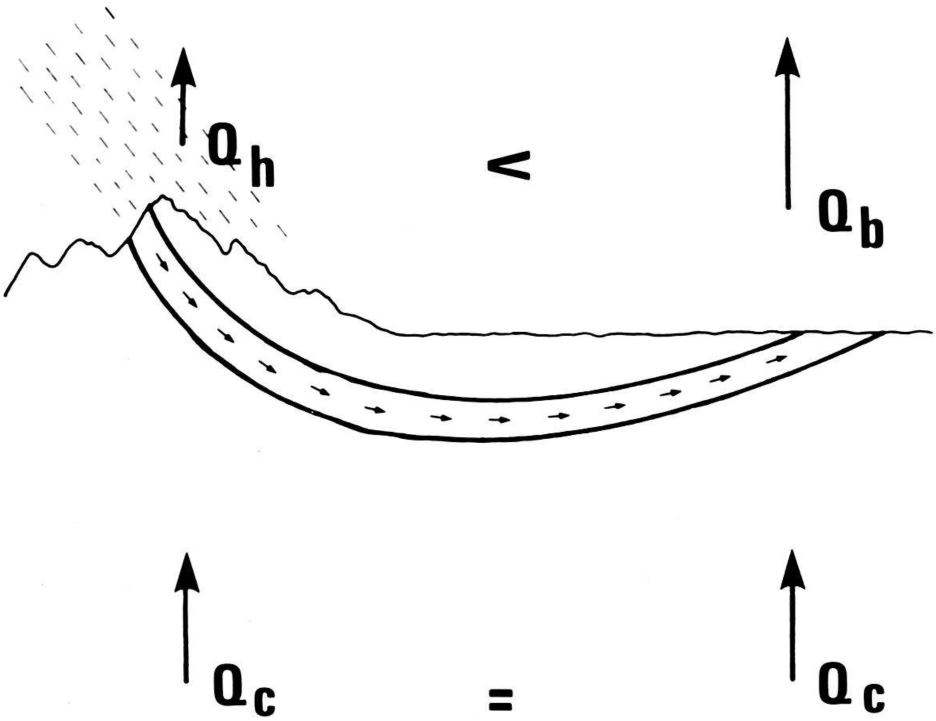

Fig. 7. Cartoon illustrating the possible effect of orographically driven deep water movement on surface heat flow. A uniform flux from below, Q_c, is present beneath both the hills and the plain. Downward movement of water in the hills reduced the surface flux observed there, Q_h, and enhances that in the plain, Q_b.

zone: when allowance is made for probable differences in thermal conductivity the heat flow may be about the same in the two areas or possibly somewhat higher in the Molasse. In view of the pronounced effect of vertical movements on thermal gradients and heat flow discussed earlier, it is very hard to understand why the heat flow in the region of rapid uplift is not significantly higher than in the region of net basement depression.

This is a puzzle that still awaits solution but it is perhaps worth speculating on the remote possibility that the thermal gradient measured in the upper thousand metres or so of the axial zone is not the true conductive thermal gradient but is modified through convective heat transfer by the slow percolation of ground water along joints and fractures. It is simple to show that if the axial zone metamorphic rocks had contained systems of joints or fractures which gave the rocks a bulk porosity comparable to that of clay characteristic flow velocities of as little as a millimeter per year would be sufficient to lower the axial zone heat flow by about 30%. If this flow were orographically driven from the axial zone towards the flanks the effect would be to enhance the heat flow in the lower flanking areas (Fig. 7). It should be emphasised that there is no evidence that such a process operates or that crystalline rocks can have suitable fracture permeabilities down to depths of three or four kilometers as would be required. Until the heat flow paradox is explained, however, it remains a nagging possibility which cannot be completely discarded.

REFERENCES

BICKLE, M.J., & POWELL, R. (1977): *Calcite-dolomite geothermometry for iron-bearing carbonates, the Glockner Area of the Tauern Window, Austria.* - Contrib. Mineral. Petrol., *59*, 281-292.

BICKLE, M.J., HAWKESWORTH, C.J., ENGLAND, P.C., & ATHEY, D. (1975): *A preliminary thermal model for regional metamorphism in the Eastern Alps.* - Earth and planet. Sci. Lett. *26*, 13-28.

BORSI, S., DEL MORO, A., SASSI, F.P., SANFERRARI, A., & ZIRPOLI, G. (1978): *New geopetrologic and radiometric data on the Alpine history of the Austridic continental margin south of the Tauern Window (Eastern Alps).* - Mem. Ist. Geol. Mineral. Univ. Padova *32*, 3-17.

BREWER, M.S. (1969): *Excess radiogenic argon in metamorphic micas from the Eastern Alps, Austria.* - Earth and planet. Sci. Lett. *6*.

ČERMÁK, V., & RYBACH, L. (Ed.) (1979): *Terrestrial Heat Flow in Europe.* - Springer, Heidelberg.

CLARK, S.P., Jr. (1961): *Heat flow from the Austrian Alps.* - Geophys. J. r. astron. Soc. *6*/9, 54-63.

CLARK, S.P., Jr., & JÄGER, E. (1969): *Denudation rate in the Alps from geochronologic and heat flow data.* - Amer. J. Sci. *267*, 1143-1160.

CLARK, S.P., Jr., & NIBLETT, E.R. (1956): *Heat flow in the Swiss Alps.* - Month. Not. r. astron. Soc. [Geophys. Suppl.] *7*.

DROOP, G.T.R. (1979): *Metamorphic studies in the South East Tauern Window, Austria.* - Unpubl. thesis, Univ. Oxford.

ENGLAND, P.C. (1976): *Thermal modelling from heat flow and velocity modelling from seismological data, with special reference to the Eastern Alps.* - Unpubl. thesis, University Oxford.

— (1978*a*): *Some thermal considerations of the Alpine metamorphism - past, present and future.* - Tectonophysics *46*, 21-40.

— (1978*b*): *The effect of erosion on palaeoclimatic and topographic corrections to heat flow.* - Earth and planet. Sci. Lett. *39*, 427-434.

ENGLAND, P.C., & HOLLAND, T.J.B. (1979): *Archimedes and the Tauern eclogites: the role of buoyancy in the preservation of exotic eclogite blocks.* - Earth and planet. Sci. Lett. *44*, 287-294.

ENGLAND, P.C., & RICHARDSON, S.W. (1977): *The influence of erosion upon the mineral facies of rocks from different metamorphic environments.* - J. geol. Soc. (London) *134*, 201-213.

ERNST, W.G. (1973): *Interpretative synthesis of metamorphism in the Alps.* - Bull. geol. Soc. Amer. *84*, 2053-2078.

FREY, M., HUNZIKER, J.C., FRANK, W., BOCQUET, J., DAL PIAZ, G.V., JÄGER, E., & NIGGLI, E. (1974): *Alpine metamorphism of the Alps - a review.* - Schweiz. mineral. petrogr. Mitt. *54*/2-3, 247-290.

FRIEDRICHSEN, H., MULLER, G., & STAHL, W. (1973): *Sauerstoff-Isotopenuntersuchungen an Mineralen eines metamorphen Profils der Hohen Tauern, Österreich.* - Contr. Mineral. Petrol. *40*, 159-168.

FRY, N. (1973): *Lawsonite pseudomorphosed in Tauern greenschist.* - Min. Mag. *39*, 121-122.

GUILLAUME, A., & GUILLAUME, S. (1980): *L'érosion Plio-Quaternaire dans les Alpes - Bilan quantitatif.* - Eclogae geol. Helv. *73*/1, 326-329.

HAENEL, R. (1979): *A critical review of heat flow measurements in sea and lake bottom sediments.* In: ČERMÁK, V., & RYBACH, L. (Ed.): *Terrestrial Heat Flow in Europe.* - Springer, Heidelberg.

HAWKESWORTH, C.J. (1976): *Rb-Sr geochronology in the Eastern Alps.* - Contr. Mineral. Petrol. *54*, 225-244.

HAWKESWORTH, C.J., WATERS, D.J., & BICKLE, M.J. (1975): *Plate tectonics in the Eastern Alps.* - Earth and planet. Sci. Lett. *24*, 405-413.

HOERNES, S., & FRIEDRICHSEN, H. (1974): *Oxygen isotope data of the polymetamorphic area of the northern Ötztal-Stubai Alps (Tirol).* - Schweiz. mineral. petrogr. Mitt. *54*, 769-788.

— (1978): *Oxygen isotope studies of the Austroalpine and Penninic units of the Eastern Alps.* In: CLOSS, H., ROEDER, D., & SCHMIDT, K. (Ed.): *Alps, Apennines, Hellenides.* - Inter-Union Comm. Geodyn. Sci. Rep. *38*, 127-131.

HOLLAND, T.J.B. (1977): *Structural and metamorphic studies of eclogites and associated rocks in the central Tauern Region of the Eastern Alps.* - Unpubl. thesis, Univ. Oxford.

— (1979): *High water activities in the generation of high pressure kyanite eclogites of the Tauern Window, Austria.* - J. Geol. *87*, 1-27.

HOLLAND, T.J.B., & RICHARDSON, S.W. (1979): *Amphibole zonation in metabasites as a guide to the evolution of metamorphic conditions.* - Contr. Mineral. Petrol. *70*, 143-148.

HUNZIKER, T.C. (1974): *Rb-Sr and K-Ar age determination and the Alpine tectonic history of the Western Alps.* - Mem. Ist. Geol. Mineral. Univ. Padova *31*.

KUNZ, B. (1975): *Temperaturmessungen in Erdölbohrungen der Molassezone Oberösterreichs.* - Mitt. Österr. geol. Ges. *68*, 51-58.

MILLER, C. (1974): *On the metamorphism of the eclogites and high-grade blueschists from the Penninic terrane of the Tauern Window, Austria.* - Schweiz. mineral. petrogr. Mitt. *54*/2-3, 371-384.

OXBURGH, E.R. (1968): *An outline of the geology of the Central Eastern Alps.* - Proc. Geol. Assoc. London *76*/1.

— (1972): *Flake tectonics and continental collision.* - Nature *239*, 202-204.

OXBURGH, E.R., & TURCOTTE, D.L. (1974): *Thermal gradients and regional metamorphism in overthrust terrains with special reference to the Eastern Alps.* - Schweiz. mineral. petrogr. Mitt. *54*, 641-662.

RAITH, M., RAASE, P., KREUZER, H., & MULLER, P. (1978): *The age of the Alpidic metamorphism in the Western Tauern Window, Austrian Alps, according to radiometric dating.* In: CLOSS, H., ROEDER, D., & SCHMIDT, K. (Ed.): *Alps, Apennines, Hellenides.* - Inter-Union Comm. Geodyn. Sci. Rep. *38*, 140-148.

RICHARDSON, S.W., & ENGLAND, P.C. (1979): *Metamorphic consequences of crustal eclogite production in overthrust orogenic zones.* - Earth and planet. Sci. Lett. *42*, 183-190.

SASSI, F.P., BORSI, S., DEL MORO, A., SANFERRARI, A., & ZIRPOLI, G. (1978): *Contribution to the geodynamic interpretations in the Eastern Alps.* In: CLOSS, H., ROEDER, D., & SCHMIDT, K. (Ed.): *Alps, Apennines, Hellenides.* - Inter-Union Comm. Geodyn. Sci. Rep. *38*, 154-160.

WAGNER, G.A., REIMER, G.M., & JÄGER, 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.* - Mem. Ist. Geol. Mineral. Univ. Padova *30*, 3-27.