Geoscience Canada

The Thermal Background to Metamorphism - II: Simple Two-Dimensional Conductive Models

C. M.R. Fowler and E. G. Nisbet

Volume 9, Number 4, December 1982

URI: https://id.erudit.org/iderudit/geocan9_4art03

See table of contents

Publisher(s)

The Geological Association of Canada

ISSN

0315-0941 (print)
1911-4850 (digital)

Explore this journal

Cite this article

Fowler, C. M. & Nisbet, E. G. (1982). The Thermal Background to Metamorphism - II: Simple Two-Dimensional Conductive Models. *Geoscience Canada*, *9*(4), 208–214.

Article abstract

In a previous article simple one dimensional models of metamorphic conditions were examined. In this article more realistic two-dimensional models are set up, to show the metamorphic effects of burial, intrusion and thrusting. Models in which erosion takes place demonstrate the development of a metamorphic faciès series and underline the importance of understanding regional stratigraphy in interpreting metamorphism.

The Thermal Background to Metamorphism - II Simple Two-Dimensional Conductive Models.

C.M.R. Fowler and E.G. Nisbet
Department of Geological Sciences
University of Saskatchewan
Saskatoon, Saskatchewan S7N 0W0

Summary
In a previous article simple one dimensional models of metamorphic conditions were examined. In this article more realistic two-dimensional models are set up, to show the metamorphic effects of burial, intrusion and thrusting. Models in which erosion takes place demonstrate the development of a metamorphic facies series and underline the importance of understanding regional stratigraphy in interpreting metamorphism.

Introduction
In the first of these discussions (Nisbet and Fowler, 1982) of the thermal background to metamorphism, we discussed simple one-dimensional problems, most of which were mathematically accessible to senior geology undergraduates. However one-dimensional models do not give an easy intuitive feel of the thermal constraints on metamorphism. In this article several two-dimensional problems are analysed. These are mathematically complex, but the results are much more easily grasped by the student, especially as most geologists are accustomed to thinking about strata in two dimensions. It is hoped that a study of these two-dimensional models will give some idea of the rates of thermal equilibration, and that they will give a qualitative understanding of the movement of heat. The intention of the models is to show how the various different types of metamorphic facies series come to be put together. All of the models discussed are conductive. In nature many metamorphic rocks show evidence for transport of heat by fluids, but that problem is much more complex and not discussed here. We also make no allowance for heat produced or consumed by exothermic or endothermic metamorphic reactions. In some cases this will considerably change our models, but a full discussion of this is out of the scope of the present paper.

Two-dimensional models
Metamorphism may be caused by a wide variety of thermal events; rocks have very different thermal responses to different tectonic events. The models are chosen to demonstrate a variety of possible metamorphic environments; they are by no means exhaustive or even representative of the possible range existing in the earth. Initially, we shall consider cases in which no erosion or deposition takes place, then we shall introduce erosion, since no metamorphic rock comes to be exposed on the surface without undergoing either erosion or tectonic accident. It can be readily seen from the discussion of the geotherm in the first article that erosion immediately has important consequences on the nature of the metamorphism itself, and on the metamorphic facies series eventually created.

A. Non-erosional models
We take three models: a model of burial metamorphism; a model of intrusion in which we consider both basic and granitic intrusions, and a model of overthrusting.

Burial metamorphism. We have constructed a model of a "typical" burial terrain: this consists of a granitic country rock in which a rectangular trough of sediment has been deposited. Beneath both granite and trough is a gneissic continental crust, overlying the mantle. Figure 1a simply shows the standard physical parameters of the model. The model includes a 40 km thick crust; the conductivity throughout is 0.006 cals cm^{-1} $°C^{-1}$, initial temperature gradient is the equilibrium gradient in the granite, gneiss and mantle while the sedimentary trough is at 100° C throughout (this is an arbitrary value and only significant in the very early stages of equilibration of the model), and heat generation in the granite is 6.0 h.g.u., 2.0 h.g.u. in the sediment, 1.0 h.g.u. in the gneiss and 0.1 h.g.u. in the mantle. The model is analogous to a rapidly filled sedimentary basin surrounded by granitic crust; it is perhaps similar to a sedimentary trough formed on a continent above a subduction zone, or to an Archean greenstone belt filled with thick sediment and set in a granitic terrain.

Figure 1b shows how the model has evolved after 20 Ma. The sedimentary trough very rapidly equilibrates towards the temperature of the surrounding country-rock; after 1 Ma the base of the middle of the trough is at about 250° C and after 20 Ma it is at about 270° C, close to equilibrium, with a shallow-level gradient of about 19° C/km. The temperature in the sedimentary trough is thus very strongly influenced by the heat production in the surrounding granite. If the country rock had been mafic material rather than granite the equilibrium gradient would have been lower, as the heat generation of mafic rocks is lower than that of granites (see previous article).

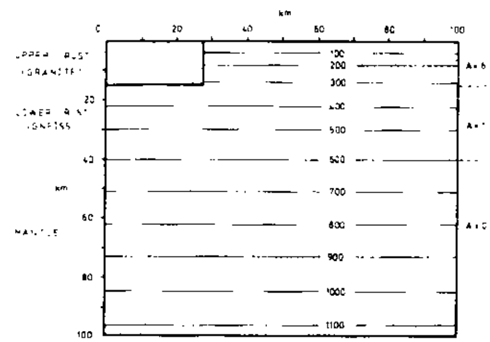


Figure 1a *Dimensions and physical parameters of the two-dimensional burial and intrusion models. As the problem is symmetrical, only half of the model is shown. Thickness of the sediment or intrusion (stippled region) = 15 km, half-width = 27.5 km. Surrounding terrain is initially at its equilibrium thermal gradient. Deep heat contribution* q_m *= 0.5 h.f.u. A = radioactive heat generation in h.g.u. For the burial model, initial temperature of the sediments is taken arbitrarily as 100° C and A = 2 h.g.u. In the basic intrusion, initial temperature = 1100° C and A = 1 h.g.u. In the granitic intrusion, initial temperature = 700° C and A = 10 h.g.u.*
Note 1 h.f.u. = 41.8mWm^{-2}
= 10^{-6} cal $cm^{-2}sec^{-1}$
1 h.g.u. = 0.418×10^{-3}mWm^{-3}
= 10^{-13}cal cm^{-3} sec^{-1}

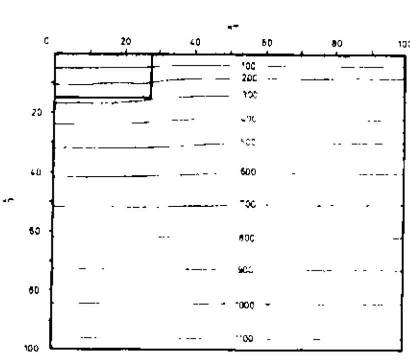


Figure 1b *Burial model after 20 Ma.*

0315-0941-82-04/208-214 $01.00/0

Hence the temperature in the trough at 15 km would be 50° C to 100° C lower. The model thus clearly demonstrates the important control that the surrounding and underlying rocks exert on the P/T path of burial metamorphism. McKenzie (1981) presents a much more rigorous and realistic model of burial heating and oil maturation in sedimentary basins.

Contact and Intrusion Metamorphism. A second class of model may be constructed as the inverse of the first. In this family of models a large igneous body is intruded into a country rock, and then crystallizes and cools at high level (in contrast to intrusion at deep level which can be crudely modelled by increasing the deep heat contribution gm - see discussion of this in previous article). Two different types of intrusion may be considered - a basic intrusion which contains only small amounts of radioactive elements, and a granitic intrusion rich in these elements.

Basic Intrusion. The main parameters of this model are shown in Fig. 1a. A large basic intrusion at 1100° C, 55 km across and 15 km deep is emplaced to a high-level in a country rock of sediments and low-grade metamorphic rocks. The latent heat of crystallization, 100 cals/gm, is spread over a 1 Ma cooling interval. The situations after 1 and 20 Ma are shown in Fig. 2a, b. Clearly contact metamorphism is an important transient phenomenon, but what is also interesting is the relatively small effect the intrusion has some distance away from the contact at high levels. If intrusion is to be a major cause of *regional*, as opposed to *local* metamorphism, the intrusions must form a high proportion of the total rock pile.

Granitic Intrusion. With a granitic intrusion the problem is rather different; the initial temperature of the igneous body is lower, but the internal heat generation is greater. Thus there is less contact metamorphism but a greater long-term effect around the intrusion, where the eventual temperature is higher than in the previous case, for bodies of the same size.

The physical parameters of this model are shown in Fig. 1a, while Figs. 2c, d, e, and f show the evolution of the model after 1 Ma, 2 Ma, 5 Ma, and 20 Ma. In this case the same latent heat as before is spread over a period of 2 Ma. The top of the pluton cools rapidly, but the base of the pluton remains hot, partly because of the internal heat generation. This high basal temperature leads to a general raising of the temperature below and around the base of the pluton, to the extent that eventually some local partial melting may take place in the country rock at depth below the intrusion; in other words, the intrusion itself can sometimes cause smaller further intrusion. If, for instance, the original intrusion were mantle-derived, it might cause later crustal anatexis. This would cause a geologist to find some very confusing field relations - at the top of the pluton "old" intrusive relationships and low $^{87}Sr/^{86}Sr$ initial ratios would be seen; at the base of the pluton and below, "young" partial melting textures and high $^{87}Sr/^{86}Sr$ ratios would be present; it would be very difficult indeed to unravel the time history of the rock.

During the period immediately after intrusion, hydrothermal convection cells are set up by the hot body, especially if the intrusion is in a relatively wet country rock. These cells dominate the heat transfer process, so the simple conductive models considered here should only be regarded as rough guides to the real pattern of metamorphism. Convective heat transfer will tend to speed up the cooling, as heat is moved more quickly. Furthermore, it will tend to concentrate the metamorphic effects nearer the source of heat, as this is where convection is most active. In granites, internal heat generation may prolong the action of convective cells. The presence of water also has profound effects on the mineralogical course of the metamorphism. Nevertheless, the simple conductive model remains a useful guide to the general understanding of metamorphism around a pluton.

At depth the long-term metamorphic imprint of the granite body is greater

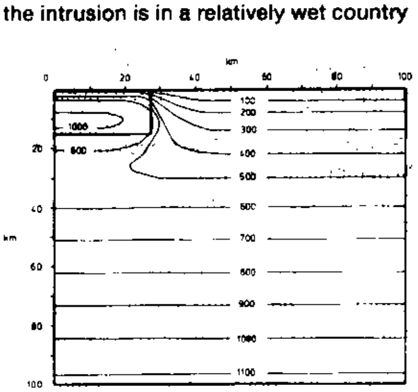
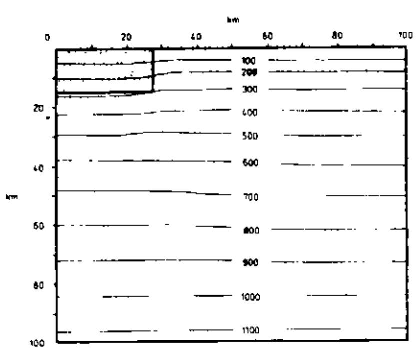
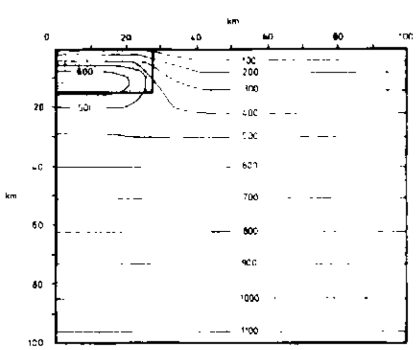


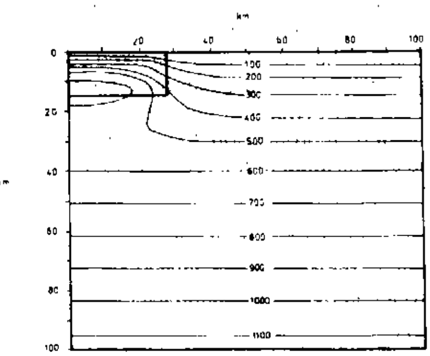
Figure 2a *Basic intrusion after 1 Ma.*



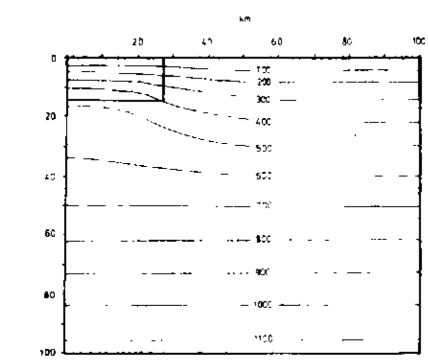
2b *Basic intrusion after 20 Ma.*



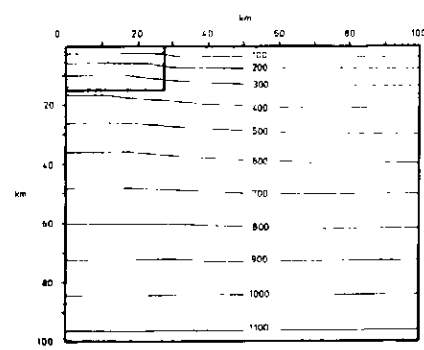
2c *Granitic intrusion after 1 Ma.*



2d *Granitic intrusion after 2 Ma.*



2e *Granitic intrusion after 5 Ma.*



2f *Granitic intrusion after 20 Ma.*

than that of the basaltic body. Whereas basaltic intrusion would have to form a significant proportion of the total rock pile to have a "regional" effect, slightly fewer granitic intrusions would be needed to produce a regional metamorphism in which the entire rock suite is significantly raised above its equilibrium temperature.

One further point may be made at this stage, though it applies generally to all regional metamorphism. As the rock pile heats up in response to some tectonic event and begins to melt at depth, the melt which forms will tend to be rich in those elements (K, Th, U) which produce the heat. Over time this process will effectively "scour" the deep crust of heat-producing elements and lead to a concentration of these elements in shallow-level intrusions (recognized as "late" or post-tectonic) and pegmatites, and eventually in sediments and seawater, after erosion. The net effect is a marked concentration of heat production in the top of the crust. Whatever the initial distribution of heat production, this will lead to the stabilization of the rock pile to a non-melting equilibrium.

Thrust Models. A wholly different type of metamorphism may be envisaged - that produced by overthrusting events (e.g. Niggli, 1970). Fig. 3a illustrates an example, with a large overthrust slice of granite-gneiss material emplaced over mafic rock. Real parallels include subduction zones or an area such as the Eastern Alps (Bickle *et al.*, 1975) where a thick overthrust crystalline block has produced metamorphism below it. In our simple model thrusting is instantaneous. Perhaps the most interesting feature of this model is that one thrusting "event" necessarily leads to two very distinct metamorphic "events". Immediately after thrusting (Fig. 3b) the hot base of the overthrust pile thermally reequilibrates with the cool underthrust rocks beneath. These latter rocks are very likely to have been rich in volatiles (e.g. spilites in a subducted oceanic slab) and thus very rapid retrograde metamorphism will take place in the upper slab, coupled with equally rapid prograde high pressure metamorphism in the lower slab. At the deeper end of the overthrust wedge local partial melting may take place if large amounts of volatiles move from the lower block into the hot lower crust of the upper block. This may produce granitic intrusions at high level.

The speed of the thermal reequilibration after thrusting should be stressed (Fig. 3b) - it is at first very rapid and "inverted" thermal gradients are probably very short-lived except where a major heat source such as shear heating or exothermic metamorphism is present in the upper slab (Graham and England, 1976). The resulting geotherm in the thrust zone and below has a very small dT/dP of the order of a few °C/km.

After initial reequilibration comes a long period (30-50 Ma or more depending on the size of the pile) in which a slow build-up of the geotherm takes place. This is a period of prograde metamorphism throughout the pile with the removal of water to higher levels during recrystallization. Finally partial melting takes place at the base of the pile and the heat production is redistributed until stability is reached.

Thus two metamorphic events may be distinguished: a very early and rapid retrogression in the upper slab and progression beneath, followed by slow progression throughout and finally partial melting and "late" intrusion to high level.

B. Erosional Models; The Development of a Metamorphic Geotherm.

Any metamorphic rock must undergo erosion if it is to become exposed at the surface (or else it must be exposed by tectonic disruption, which can be thought of as equivalent to fast erosion). Thus erosion is essential to produce a metamorphic belt. However it can be seen from the above discussion of the geotherm in the previous article and from the work of England and Richardson (1977) that the process of erosion itself has a fundamental effect on the structure of the geotherm, and the P/T path through which any metamorphic rock passes is to a large extent controlled by erosion. Secondly England and Richardson pointed out that the "metamorphic geotherm" - the geotherm inferred by the Pressure and Temperature measured in metamorphic rocks exposed at the surface - is polychronic and strongly influenced by the erosion path followed by the rocks. We discuss this further below.

Miyashiro (1973) has introduced the concept of a "metamorphic facies series" to represent the series of metamorphic facies observed in a single metamorphic terrain. For the purpose of this paper we shall define a metamorphic facies series as the curve in P/T space obtained when P/T data from metamorphic rocks from a single coherent tectonic/metamorphic terrain are plotted together. In many metamorphic terrains relict minerals from an earlier event may also exist - these would be expected to give a different curve since the P/T conditions of metamorphism and the erosional history of the earlier event would be unlikely to

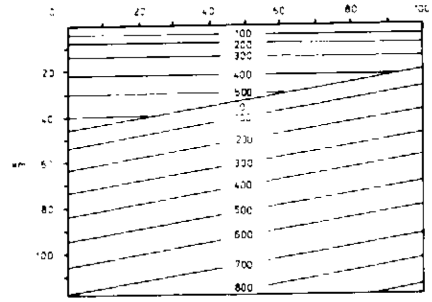
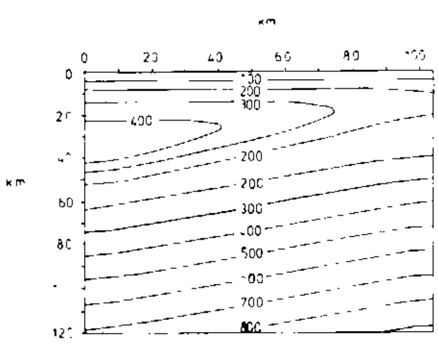
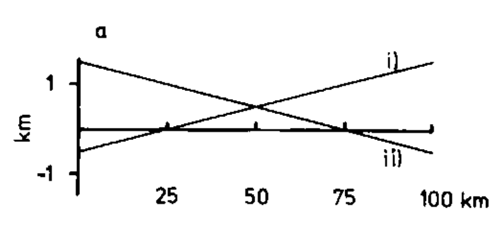


Figure 3a *Overthrust model - initial conditions.*



3b *Overthrust model after 1 Ma.*



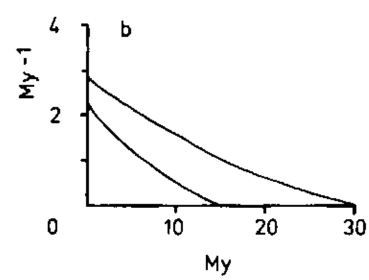


Figure 4 *Erosion functions. At any particular time and place, rate of erosion U = (space function) x (time function) in km/Ma.*

4a *Space function across the burial and intrusion models. Two cases were considered: i) erosion of the country rock and deposition upon the sediments or intrusion; ii) erosion of the sediments or intrusion and deposition on the country rock. In the overthrust model erosion is constant in space across the model.*

4b *Time function. Lower curve: burial and intrusion models. Upper curve: overthrust model.*

be exactly similar. We should like, however, to emphasize the non-unique character of thermal equilibration; a similar end result may be obtained by a variety of different paths.

We shall now examine the various thermal models of the previous section to see how they are influenced by erosion. The erosional model chosen for the burial and intrusion models is as follows: 1) In space across the 200 km model we have considered both strong erosion at the margins and deposition in the centre; and also strong erosion at the centre and deposition on the margins. In both cases

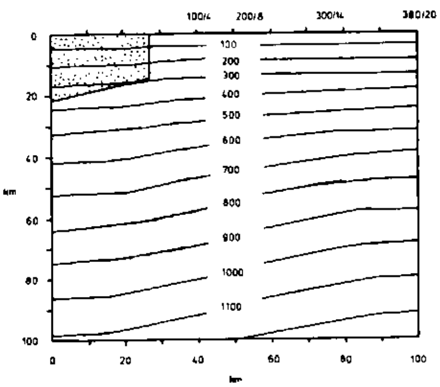


Figure 5 *Burial model after 20 Ma of erosion of country rock and deposition upon sediment trough. Figure shows maximum temperature (° C/depth (km)) attained for rocks now exposed at the surface after 20 Ma. The depth/temperature paths followed by these samples are the same as those shown in Figure 7a.*

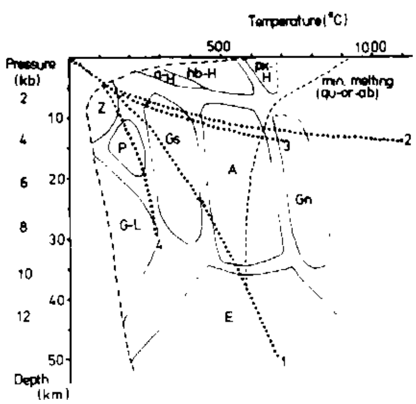


Figure 6 *Pressure/Temperature curves (Metamorphic facies series) obtained from exposed rocks in the models after erosion (see Figs. 5, 7, 8). Dotted lines show facies series. 1) equilibrium series (Figs. 5, 7a, 8-high grade); 2) basalt intrusion after erosion (Fig. 7b); 3) granitic intrusion (Fig. 7c); 4) overthrust model - low grade (Fig. 8). Facies fields after Turner (1968). Z = zeolite, P = prehnite-pumpelleyite, G-L = glaucophane-lawsonite, Gs = greenschist, A = amphibolite, E = eclogite, Gn = granulite, a-, hb-, px-H = albite, hornblende and pyroxene hornfels.*

the erosion rate varies linearly across the model (see Fig. 4a); 2) In time we have chosen an exponential function, erosion rate decaying as $e^{-t/15}-e^{-1}$ for 15 Ma, where t is in Ma. This is shown in Fig. 4b.

In the case of the thrust model we have chosen different erosional constraints; an initial erosion rate of 2.8 km/Ma across the whole model, decaying in time as $e^{-t/30}-e^{-1}$ (t in Ma.)

Burial model with Erosion. As a simplest case we consider the "burial" trough of sediment, and continue to deposit in the middle of the trough at an initial rate of 1.1 km/Ma while eroding at the edge of the model at an initial rate of 3.3 km/Ma; the rate of erosion and deposition falling off with time as described above.

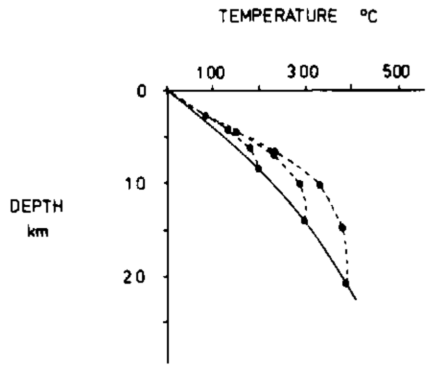
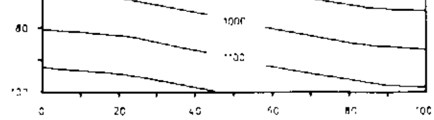
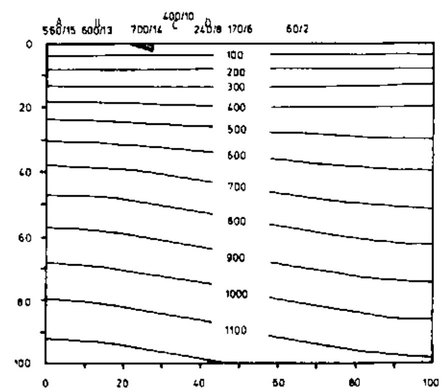


Figure 7a *Temperature-depth paths followed by points finally exposed at the surface, for intrusion and sedimentary trough models with subsequent erosion of country rock and deposition on intrusion/trough. Starting equilibrium gradient - solid line. Circles - temperature every 2 Ma for first 6 Ma of erosion.*

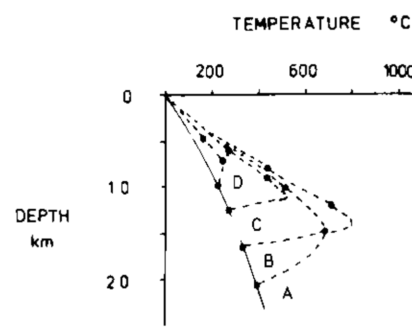


7b *Basalt intrusion model after 20 Ma of erosion of the intrusion and deposition on the country rock. Stipple shows the remnant of the intrusion.*

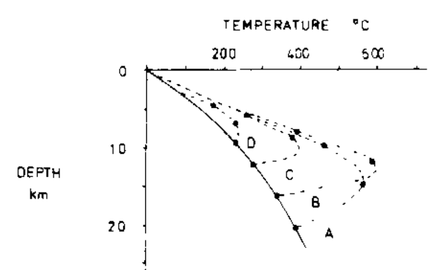
The result is shown in Fig. 5, which reflects the state of the rock pile after 20 Ma. By this stage the burial trough has been further covered by sediment, while in the country rock deep erosion has taken place, to a depth of 15-20 km at the edge of the model. Fig. 5 also shows the maximum temperatures attained during the 20 Ma time period (i.e. 15 Ma of erosion/deposition and 5 Ma of thermal reequilibration) by the rocks which are



7c *Granitic intrusion (see Fig. 2c) after 20 Ma of erosion of the intrusion and deposition on the country rock Stipple shows the remnant of the intrusion. In Fig. 7 numbers above models give the maximum temperature (° C) attained by rocks now exposed at the surface, and the depths (km) at which these temperatures were reached.*



7d *Temperature-depth paths for basalt intrusion of Fig. 7b.*



7e *Temperature-depth paths for granite intrusion of Fig. 7c.*

exposed at the surface after 20 Ma. These are the rocks available to a field geologist, and they are the rocks from which he will set up a metamorphic facies series. The maximum temperature attained is not necessarily preserved by the highest-grade minerals or the temperature of the maximum entropy assemblage but for our purposes it is sufficient to assume that the mineral assemblages studied by our geologist will be those formed when the rock reaches its highest temperature. For a brief discussion of time-temperature-transformation behaviour the reader is referred to Putnis and McConnell (1980). As the rock cools it will equilibrate to a lower temperature and the mineral composition will alter (e.g. by reactions at the rims of the minerals), but reaction kinetics become markedly slower as the rock cools, and thus there will be a good chance of preserving the highest-grade minerals if erosion is fast enough.

In our simple case of burial metamorphism, the trough of buried sediment has no heating effect on the country rock. Thus when the country rock is eroded the P/T curve which can be plotted from the highest-grade minerals in the exposed rocks is simply that of the initial equilibrium thermal gradient in the rock: in this case the "metamorphic geotherm" is identical to the equilibrium geotherm no matter what the rate of erosion is (provided that erosion is fast enough to "quench" the mineral compositions at their highest temperatures). The metamorphic facies series produced by the event is that of a normal equilibrium geotherm in the country rock (Fig. 6). We shall call this Facies series 1.

Intrusion models with Erosion. With the two intrusional models discussed above, things can be rather different. If erosion is as above (the intrusion being buried and the country rock being eroded) little metamorphic effect is seen even from these very large intrusions. With the exception of a localized contact zone (of the order of 5 km across) in both cases, the country rock gives a metamorphic facies series identical to the equilibrium series 1 and the net result is similar to that shown in Fig. 5. A real example of this could be the Great Dyke of Zimbabwe, which has only a restricted contact zone. Fig. 7a shows the depth-temperature paths followed by individual points exposed on the surface after 20 Ma. For a fuller discussion of the significance of these paths, see England and Richardson (1977) and Thompson (1981).

On the other hand, if the intrusion is eroded and deposition takes place on the country rock, very marked effects are seen. This is because deep-seated rocks, close to the intrusion, are now being eroded. The resulting facies series is one of very low dP/dT (facies series 2 for basalt and facies series 3 for granite as shown in Fig. 6). It can be seen from Fig. 7b, c that although the original intrusion has been almost completely eroded, the metamorphic imprint of intrusion and erosion is widespread and lasting. Our geologist would be surprised to discover that the apparently small granite in Fig. 7c had apparently been responsible for such a major metamorphic effect. Figs. 7d and 7e show some depth-temperature paths for points exposed on the surface after 20 my.

Thrusting models with Erosion. Erosion of the thrusting model provides some interesting clues to the genesis of "paired" metamorphic belts. Our model is constructed to show this; however the remarks also apply qualitatively to any overthurst terrain, but with correspondingly lower pressures and temperatures in thinner thrust sheets. Fig. 8 shows the state of the thrusting model after 30 Ma of erosion. In time, erosion is initially very strong and then decays exponentially according to $U = 4.5\ (e^{-1/30} - e^{-1})$ where U = erosion in km/Ma and t = time in Ma (Fig. 4b). In space, erosion is the same across the top of the model; all parts being equally eroded at any particular time.

Initially only the overthrust unit is exposed at the surface. Since the thrusting event has a general cooling effect on the upper block, the highest temperature any part of the upper block experiences is its initial temperature prior to thrusting. Rocks from the upper block will thus probably retain relict minerals from their previous high-grade environment. The extent of the mineral reequilibration is probably very dependent on the availability of volatiles rising from the underthrust sheet into the hot base of the upper sheet. If local partial melting occurs (e.g. in the upper sheet on the extreme left of the model) prograde effects will be seen.

Simultaneously with the metamorphism of the upper block the underthrust block will experience prograde metamorphism. As erosion takes place the deep-level rocks are lifted up towards the surface, and this has a cooling effect which progressively halts and reverses the rise in temperature of the rocks in the underthrust block. Thus, rocks which are initially at depths of, say 10-20 km below the thrust surface, will attain a maximum temperature and then cool again as they are uplifted. The maximum temperatures attained in this underthrust block are also controlled by the time of initiation and the rate of erosion, and also by any shear heating in the upper slab (Graham and England, 1976). Eventually the thrust surface is exposed by the erosion, and the following metamorphic events are seen by a field geologist: 1) An early (prior to thrusting) high grade event in the overthrust block, shown as Facies series 1 in Fig. 6, overprinted by 2) a post-thrusting retrograde event in the remaining part of the overthrust block. In real rocks this event would be controlled by the introduction of large amounts of volatiles from the underthrust rocks. The overthrust block would be rapidly cooled and hydrated. The degree of "overprinting" would depend upon reaction kinetics (exponentially related to temperature) and the availability of volatiles; 3) A high

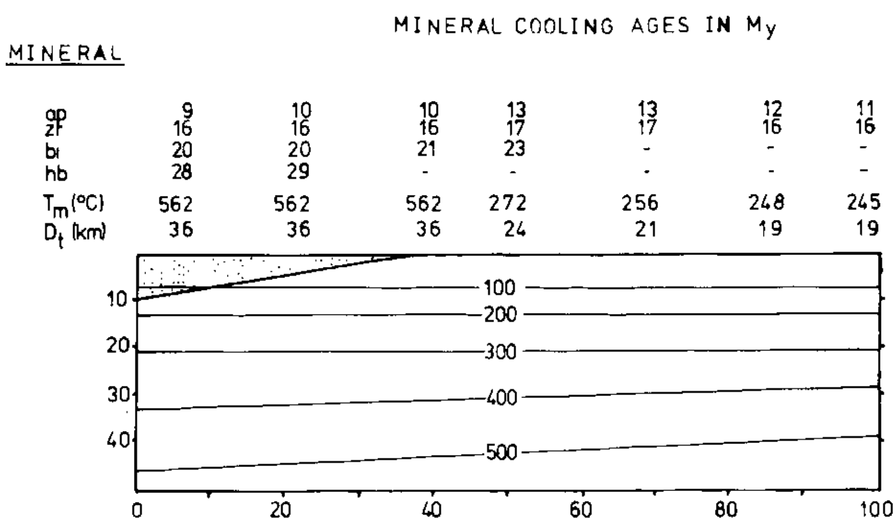


Figure 8 *Overthrust model (see Fig. 3) after 30 Ma of erosion. Stipple - remnants of overthrust. T_m = maximum temperature attained by rock which is now at the surface. D_t = depth at which this temperature was reached. Figure also shows time (in Ma before exposure) of closure of various radiometric systems in the rocks now at the surface. ap = apatite fission track, zr = zircon fission track, bi = biotite K/Ar, hb = hornblende K/Ar.*

pressure, low temperature event in the underthrust rocks, shown as Facies series 4 in Fig. 6.

An overthrust of the dimensions modelled here would thus produce the classic "paired metamorphic belts"; one high grade (Facies series 1) and one low grade (Facies series 4). Much smaller overthrusts (e.g. at high tectonic levels in mountain belts) would be qualitatively similar, but the metamorphic effects might be restricted by kinetic factors or else small or difficult to distinguish.

The time at which the rocks in the overthrust pile attain their maximum temperatures depends on their original depth and on erosion rates; thus radiometric ages would probably show a significant spread.

Radiometric ages

The age of a radiometric system, such as a mineral, generally depends on the way in which the daughter product became sealed into the system as it cooled. In a typical system a "blocking" temperature exists: - this is the temperature below which the system can be thought to be closed (Dodson, 1973). Different minerals, different dating techniques and different rates of cooling all produce different blocking temperatures; thus we have here a very powerful tool for working out the thermal history of a mountain belt.

Harrison *et al.* (1979) and Harrison and Clark (1979) have attempted to use this tool in the interpretation of a pluton. For moderate cooling rates they suggested the following closure temperatures: 94° C for apatite fission tracks, about 175° C for zircon fission tracks, 177-260° C for biotite K/Ar, and 400-550° C for hornblende K/Ar. Sm/Nd and Rb/Sr whole rock ages in many cases probably reflect the original age of the rock. Consider first the basaltic and granitic intrusions illustrated in Fig. 2. Underneath the intrusion, minerals which "close" around 500° C would have their radiometric clocks started about 1-2 Ma after the start of cooling. In contrast, minerals which "close" at around 300° C would not start recording until 20 Ma in the case of the basalt intrusion, which has little internal heat generation. In the case of the granite a mineral with a 300° C blocking temperature would not close at all underneath the intrusion.

The depth/temperature paths shown in Fig. 7 are of great interest, in what is essentially a simple case of the variable erosion of an equilibrium geotherm. Note the great spread in radiometric ages that would be obtained from different mineral 'clocks'. Furthermore, the blocking temperatures are dependent on *rate* of cooling, which is different in each case.

In Fig. 8 we show the effect of using these various dating techniques across our thrust model's 30 Ma erosion surface - the dates shown are the dates that would be measured by a geologist working on this surface. It can be seen that the dates provide a very useful tool indeed for studying the cooling and erosional history of the pile. The model involves rapid erosion of the overthrust pile; in many natural examples there may be profound differences in closure ages of minerals which were initially produced by the same tectonic event.

Implications

Most regional metamorphism is probably connected in some way to subduction, but the immediate controlling process of metamorphism in any particular terrain is likely to be an event such as intrusion or thrusting. To interpret a metamorphic terrain fully it is not sufficient to know the P/T conditions undergone by the rocks: a full interpretation of the thermal history of the rocks also involves studying the erosional and radiometric history of the terrain. Our thermal models demonstrate that a *general* knowledge of the stratigraphy of an area is essential before its metamorphic study can be properly unravelled.

Convective movement of fluid has not been discussed in these models, yet it is a very major factor in heat transport around plutons, and during dehydration of overthrust terrains. As a rough general rule fluid movement speeds up thermal reequilibration, and reduces to some degree the extent of aureoles around intrusions, while at the same time promoting metamorphic reactions and having a major impact on the thermodynamics of reequilibration.

Acknowledgements

We thank J. Martignole, R.S.J. Lambert, A.B. Thompson, P.C. England and Jo Laird for comments, and the Royal Society, London for support. Much of the computing was carried out during tenure of fellowships at E.T.H. Inst. für Krist. und Petrographie and E.T.H. Inst. fur Geophysik, Zürich, Switzerland: our thanks to Professors Mueller, Thompson, Tromsdorff and Hsu. Thanks too, to L. McMonagle.

References

Bickle, M.J., C.J. Hawkesworth, P.C. England and D. Athey, 1975, A preliminary thermal model for regional metamorphism in the Eastern Alps: Earth Planet. Sci. Lett., v 16, p. 13-28.

Dodson, M., 1973, Closure temperatures in cooling geochronological and petrological systems: Contrib. Mineral Petrol., v. 40, p. 259-274.

England, P.C. and S.W. Richardson, 1977, The influence of erosion upon the mineral facies of rocks from different tectonic environments: J. Geol. Soc. Lond., v. 134, p. 201-213.

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

Harrison, T.M., R.L. Armstrong, C.W. Naeser, and J.E. Harahal, 1979, Geochronolgy and thermal history of the Coast Plutonic Complex near Prince Rupert, British Columbia: Can. J. Earth Sci., v. 16, p. 400-410.

Harrison, T.M. and G.K.C. Clarke, 1979, A model of the thermal effects of igneous intrusion and uplift as applied to Quottoon pluton, British Columbia: Can. J. Earth Sci., v. 16, p. 411-420.

McKenzie, D.P., 1981, The variation of temperature with time and hydrocarbon maturation in sedimentary basins formed by extension: Earth Planet. Sci. Lett., v. 55, p. 87-98.

Miyashiro, A., 1973, Metamorphism and metamorphic belts: Allen and Unwin, London, Wiley N.Y., 492 pp.

Nisbet, E.G. and C.M.R. Fowler, 1982, The thermal background to metamorphism-1. Simple one-dimensional models: Geoscience Canada, v. 9, p. 161-164.

Niggli, E., 1970, Alpine metamorphose und Alpine Gebirgsbildung Fortschritte Mineral v. 47, p. 16-26.

Putnis, A. and J.D C. McConnell, 1980, Principles of mineral behaviour, Blackwell Scientific Publications, Oxford, 257 p

Thompson, A B., 1981, The pressure-temperature (P T) plane viewed by geophysicists and petrologists: Terra Cognita, Journal of European Union of Geosciences, vol. 1, p. 11-20

Appendix

The general differential equation governing heat conduction in a moving medium is:

$$\underline{u} \bullet \nabla T + \frac{\delta T}{\delta t} = \frac{K \nabla^2 T}{\rho c} + \frac{A}{\rho c}$$

with T = temperature field
K = conductivity
ρ = density
c = specific heat
A = heat production
$\underline{u}$ = velocity

For a two-dimensional model undergoing uplift and erosion, but no horizontal motion, this equation becomes:

$$\frac{\delta T}{\delta t} = \frac{K}{\rho c}\left(\frac{\delta^2 T}{\delta x^2} + \frac{\delta^2 T}{\delta z^2}\right) + \frac{A}{\rho c} + \frac{u \delta T}{\delta z}$$

For the one-dimensional models examined this equation was solved numerically using the Crank-Nicholson implicit finite-difference approximation (Carslaw and Jaeger, 1959). Solutions for the two-dimensional models were computed using explicit finite-difference approximations and/or the slightly faster alternating-direction implicit (ADI) finite-difference method (Douglas, 1955; Pearceman and Rachford, 1955). The accuracy of the two-dimensional solutions is not much better than 5%, but this is quite sufficient to study the development of the temperature field and determine the type of metamorphic facies series each model can yield.

Carslaw, H.S and Jaeger, 1959, Conduction of heat in solids, Oxford University Press.

Douglas, J., 1955, On the numerical integration of $\frac{\delta^2 u}{\delta x^2} + \frac{\delta^2 u}{\delta y^2} = \frac{\delta u}{\delta t}$ implicit methods, J Soc. Indust Appl Math., v 3, p. 42-65.

Pearceman, D.W. and H H. Rachford, 1955, The numerical solution of parabolic and elliptic differential equations: J Soc. Indust Appl. Math., v 3, p. 28-41

MS received April 5, 1982;
Revised August 1, 1982