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Meditations on Metamorphism

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Dead flies cause the ointment of the apothecary to send forth a stinking savour: so doth a little folly him that is in reputation for wisdom and honour—Ecclesiastes, Chapter 10, verse 1.

I chose this quotation as what I thought to be an appropriate commentary on my talk today. An occasion like this is I think a useful opportunity to indulge oneself in a little folly. Some of you may think after I have done with speaking that it has been more than that, and that the "stinking savour" has truly offended, nevertheless I have over the last few months had these meditations on metamorphism and I do intend to use the President's privelege of talking on an occasion such as this when I am safeguarded by long established tradition from an open discussion at the end.

Of all the definitions of the metamorphic process, I think that by Turner and Verhoogen (1961, p. 450) is the most complete and most elegant:

"Metamorphism is the mineralogical and structural adjustment of solid rocks to physical or chemical conditions which have been imposed at depths below the surface zones of weathering and cementation and which differ from the conditions under which the rocks in question originated".

The definition, as understood by metamorphic petrologists, clearly excludes superficial processes such as weathering and diagenesis. So far we may have general agreement. The problem comes when we try to subdivide metamorphism. The classical subdivision into CONTACT and REGIONAL is fraught with difficulties because often it is based on erroneous criteria. How many of us would agree with Winkler's (1967, p. 2) statement "Contact metamorphic rocks lack schistosity" and how many of us would say that the presence of a hornfels texture proves contact metamorphism? Hornfels derives from the German *horn* or hoof of animals, and *fels* meaning rock, hence the definition of Winkler (1967, p. 227) "a non-schistose rock, splintery on impact. The edges of thin rock chips occasionally are transculent like a horn. ... Hornfelses are typically produced by contact metamorphism of clays, fine-grained greywackes, etc., and occasionally by regional metamorphism". Please note the last phrase, *occasionally* *by regional metamorphism*. I believe that Winkler was justified in adding this phrase because I believe that there is no definite indication of the type of metamorphism simply based on the rock texture and mineralogy alone. To prove contact metamorphism, it is necessary to prove beyond a shadow of doubt that the source of heat was from a hot magmatic body. Unless a metamorphic aureole can be mapped around the pluton in which the metamorphic grade can be shown to increase towards the contact, then one has not definitely proven contact metamorphism. And what are the extents of these aureoles? Harker (1939) mentions examples of the Shap granite in Westmorland, England, as 1,200 yards, but says that the larger granite masses of Cornwall, England, have aureoles up to two or three miles in breadth. But one cannot generalise on this because there are many factors such as size of pluton, temperature of magma and most important of all ambient regional temperature of the country rocks into which the magma intruded.

If most of us would readily accept hornfels as the criterion *par excellence* for contact metamorphism, then schist and gneiss are undoubtedly taken to indicate regional metamorphism. To digress somewhat at this time, I am often asked to differentiate in unequivocal terms *schist* and *gneiss*. Wenk (1963) has given the following somewhat droll definitions which nevertheless are useful in the field (The definitions are in translation):

> Schist: "when hit with a hammer, rocks having a schistose fabric (i.e. schists) split perfectly parallel to 's' into plates, 1–10 mm in thickness, or parallel to the lineation into pencil-like columns". Whereas Gneiss: "splits parallel to 's' generally along mica or hornblende layers, into plates and angular blocks, a few cm to tens of cm in thickness, or parallel to B into cylindrical bodies (i.e. pencil gneisses)".

Now let us see if these generalizations regarding schistosity and hornfels texture are valid.

To enable us to do this we have to relate metamorphism to a pressure-temperature diagram, such as the one I have drawn here (fig. 1). The first point that needs explaining, but is never done by any author, is the *pressure-depth of burial* equivalence shown on the left and right of this diagram. The metamorphic pressure $\mathbf{P}$ is determined essentially by load pressure due to the superincumbent rock column, and is designated $\mathbf{P}_l$. For common reactions of progressive dehydration of minerals and decarbonation, the fluid pressure $\mathbf{P}_f$ is assumed $= \mathbf{P}_l$. For assemblages containing hydrous minerals the further assumption is made that $\mathbf{P}_{OH_2O} = \mathbf{P}_f = \mathbf{P}_l$, and for carbonate rocks $\mathbf{P}_{OCO_2} = \mathbf{P}_f = \mathbf{P}_l$. But you say: does not tectonic pressure play a role. My reply would be *generally no*. An increment of tectonic overpressure would not be expected above $\mathbf{P}_l$ in any significant amount except in strong rocks at low temperatures. The problem in deciphering metamorphism is not a lack of pressure (we have more than enough by assuming only $\mathbf{P}_l$) but in a lack of heat. Significant amount of tectonic overpressure would provide a source of embarrassment for the metamorphic petrologist. I will return to this point. Assuming then that metamorphic reactions occur only under simple load pressure $\mathbf{P}_l$, let us convert this to a depth of burial.

Consider a cylindrical column of crust, of area 1 cm^2. The pressure $\mathbf{P}_l$ at depth $\mathbf{h}$ is $\mathbf{h} \times \mathbf{d} \times \mathbf{g}$ dynes.

If we take granodiorite and mica schist as average crustal rocks, then an average crustal density of $\mathbf{d} = 2.7$ gm/cc is obtained. The pressure then $= \mathbf{h} \times 2.7 \times \mathbf{g}$ dynes; $\mathbf{g}$ is normally taken as 981 cms/sec^2.

So at a depth of 1 cm the pressure is 2.7×981 dynes
So at a depth of 1 metre the pressure is $2.7 \times 981 \times 10^2$
So at a depth of 1 kilometer the pressure is $2.7 \times 981 \times 10^5$ dynes.
1 bar of pressure $= 10^6$ dynes. 1 kilobar $= 10^9$ dynes.

$$\text{Hence the pressure at 1 kilometer depth} = \frac{2.7 \times 981 \times 10^5}{10^9} \text{ bars}$$
$$= 0.26487 \text{ kilobars.}$$

Or let us remember that the pressure at 10 km = 2.65 kilobars or roughly 30 km depth = 8 kilobars of load pressure for a rock density of 2.7 gm/cc. I have used this conversion to illustrate different kinds of metamorphism which can be encountered within the earth's crust. (fig. 1)

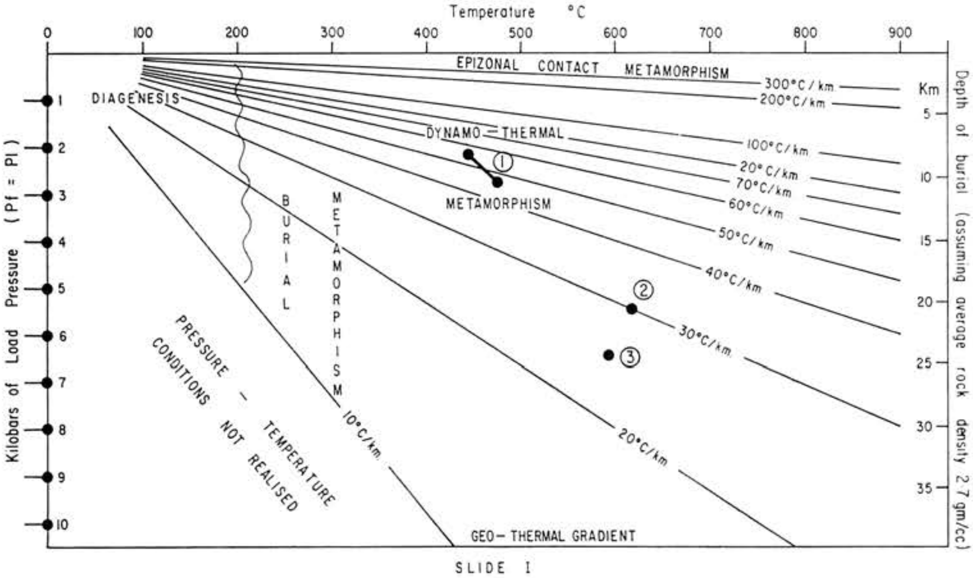

Fig. 1. Diagram to show various geothermal gradients which may occur in metamorphism. The fluid pressure during metamorphism P_f is assumed equal to the hydrostatic or load pressure P_l. Points 1, 2, and 3 are taken from fig. 7.

The metamorphic type is best described in terms of temperature and pressure and if all the rocks in one orogenic belt have been correctly interpreted, their conditions of pressure and temperature will be expected to lie along one of the geothermal gradients or along a limited range of neighbouring gradients, for which I have shown the round number examples. Thus we may say that one orogen was characterised by a low geothermal gradient, e.g. generally 20° C/km (this is now referred to as the high pressure facies series, but this is an unsatisfactory term because it implies that all the rocks were deeply buried; indeed, as you see they were under a range of pressures corresponding to the range of burial depths). I propose we call such a facies the low geothermal gradient facies series. This is generally what is called the Barrovian Facies series (after Barrow, 1893), well displayed in the Dalradian of the Scottish Highlands.

Then we may have an orogen characterised by a very high geothermal gradient around 70° C/km. This is commonly referred to as the low pressure or the Abukuma facies series (Miyashiro, 1961). Common thermal gradients around deep "granite" contacts have been shown by Jaeger (1959) to be as follows: For a granitic tabular vertical sheet 2000 m thick intruded at 800°C with a crystallization range 800° to 600°C and intruded at a depth at which the ambient regional rock temperature was 100°C, the thermal gradients around the intrusion have been shown to be within the range 150° to 300°C/km. These are the gradients which characterise thermal or contact metamorphism.

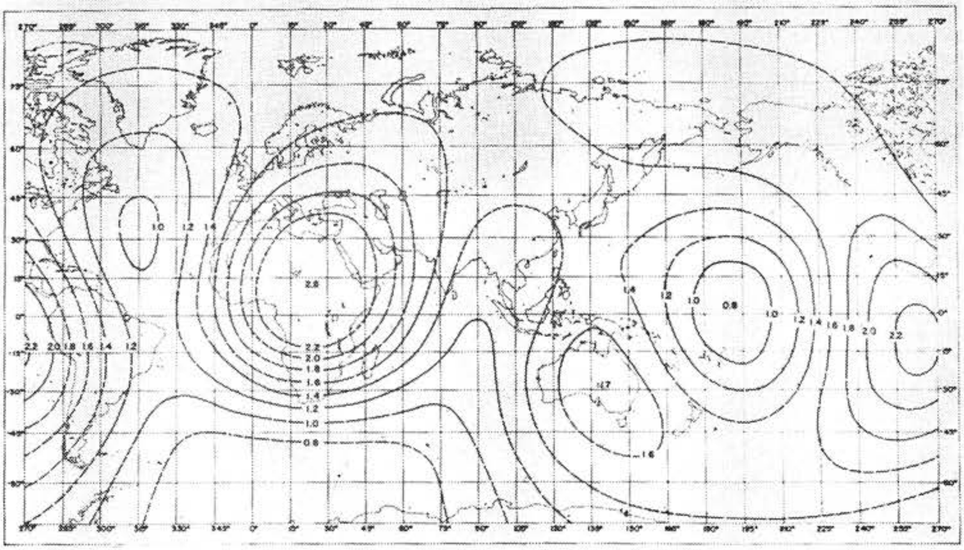

Fig. 2. Heat flow pattern of the surface of the earth taken from Lee and Uyeda (1965, fig. 44). Contour lines are in millicalories/cm²/sec and are dashed over regions where no data exist.

Let us now see what order of geothermal gradients can be found in the crust today. Lee and Uyeda (1965, fig. 44) have given a diagrammatic summary of 987 present day heat flow values collected from a worldwide distribution (fig. 2). However this major paper summarises about 2000 observations. Heat flow is measured in millicalories per cm^2 per sec. The authors have shown that variations greater than 0.2μ calories/cm^2/sec. are significant. At the 95% confidence level, the world's mean heat flow on land is $1.5 \pm 10\%$ millicalories/cm^2/sec. and the average does not differ significantly from that over the oceans. The surface heat flow can be related to geothermal gradient by the equation

$$\mathbf{q} = \mathbf{K}\ (\mathbf{dT/dZ})$$

where $\mathbf{q}$ = surface heat flow in μ cal/cm^2/sec
K is the thermal conductivity of the rocks in cal/cm/sec
dT/dZ is the thermal gradient.

Thus the thermal gradient

$$\mathbf{dT/dZ} = \mathbf{q/K} \text{ in } ^\circ\text{C per cm}$$
$$= \mathbf{q/K} \times 10^5 \text{ in } ^\circ\text{C per km.}$$

The thermal conductivities of rocks can be measured. The conductivity decreases with temperature and increases with pressure. Hence Beck (1965) concluded that the effects of rising temperature and pressure in burial tend to cancel each other out. The thermal conductivity depends on pore fluids, orientation of crystals, and mineralogy. Limestones are of average 4.7 millicalories/cm/sec. Granite has been computed and measured by Beck (1965) as 8.0 . Shales about 6. Feldspar, muscovite and sericite about 5.5. Norite and peridotite were listed by Clark (1966) as around 6.

So in our heat flow map of the world, if we were to take the high value over the African continent of 2.6 millicalories/cm²/sec and assume an average thermal conductivity of 6×10^{-3} cal/cm/sec,

$$\text{the thermal gradient } ^\circ\text{C/km} = \frac{2.6\times10^{-6}}{6.0\times10^{-3}}\times10^{5} = \frac{2.6}{6.0}\times10^{2} = 43^\circ\text{C/km}.$$

and a low value over the Pacific of 0.8 would give a thermal gradient of

$$\frac{8.0}{6.0}\times10^{2} = 13^\circ\text{C/km}.$$

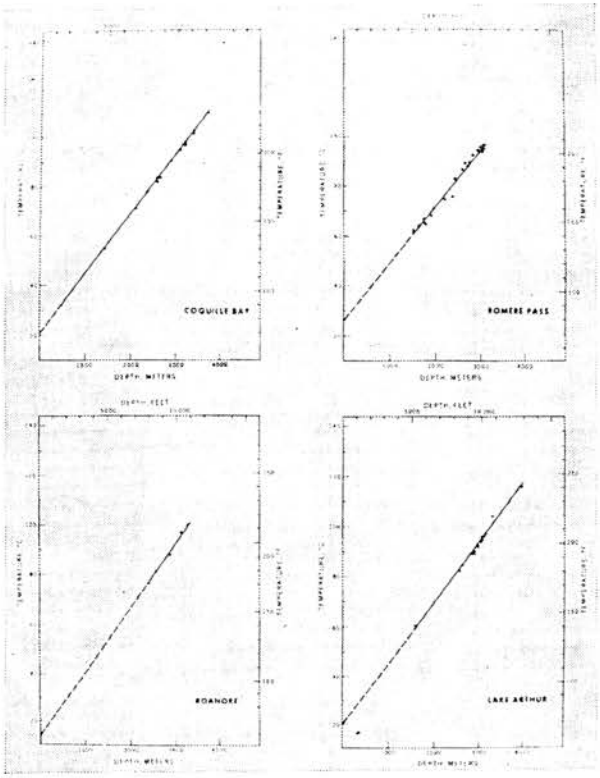

Fig. 3. Some temperature—depth plots of gas fields in southern Louisiana, U.S.A. from Jaml, Dickey, and Tryggvason (1969, fig. 2). Reproduced by permission of American Association of Petroleum Geologists.

And let us now take the specific example of the Gulf Coast of Louisiana, U.S.A.

Jaml, Dickey and Tryggvason (1969) have shown from 123 gas producing fields along the gulf coast of Louisiana, from direct measurement of temperatures in wells drilled to exceed 10,000 feet (3048 m), that temperature gradients in this region can be calculated to range from 18° to 36°C/km and that most average around 23°C/km.

Figure 3 shows the kind of data on which this study is based. Note the good linear correlation of temperature with depth. Of this increase of temperature with depth, it is well for geologists to recall the early observation of Robert Boyle the physicist in 1671 (quoted by Bullard, 1965) that "beer does not freeze in Moscow in cellars that were not above 12 or 14 feet deep". It is not recorded how many beer cellars he sampled before making this generalization, but his reputation as a physicist was at stake and we can be sure he made a careful study. He interpreted the increase of temperature with depth as being due to sources of heat within the earth.

In the Louisiana study, which must have been less exciting that Boyle's study, temperatures were measured in a bottom hole pressure 'bomb' from a few weeks to several years after the drilling had stopped so that thermal conditions would have returned to normal after being upset by drilling. The authors took the average shale density as 2.4 gm/cc at a depth of 3000 m and a calculated thermal conductivity of 5.0×10^{-3} cal/cm/sec. All their results were plotted and contoured by a computer and their map (fig. 4) shows the heat flow distribution pattern over the Gulf Coast area. They computed an average geothermal gradient over the whole region of 23.7°C/km based on the average heat flow of 1.18 millicalories/cm²/sec.

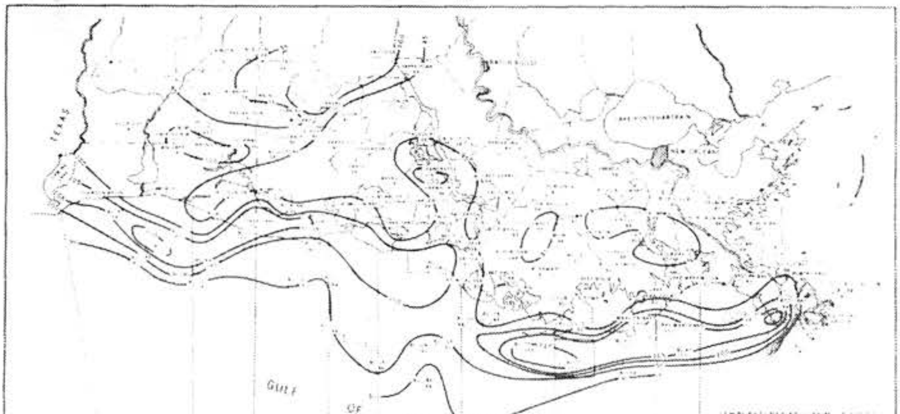

Fig. 4. Temperatures in South Louisiana at a depth of 3043 metres. Contour interval is 5°C. From Jaml, Dickey, and Tryggvason (1969, fig. 5). Reproduced by permission of American Association of Petroleum Geologists.

If the hot region shown immediately offshore, which coincides with the deepest known part of the sedimentary basin, is extrapolated to the lower levels of the basin to a depth of 15 km, the temperature indicated would be 400°C using a safe average gradient of 24°C/km (pressure equivalence = 3.8 kb); then the Tertiary rocks of the Gulf Coast which are buried at this depth are probably undergoing metamorphism high in the greenschist facies and well on their way to amphibolite facies of dynamothermal metamorphism at this very day without any help from orogenesis.

From the compilation of heat flow measurements by Lee and Uyeda (1965) we can therefore confidently assume that normal geothermal gradients today present in the earth's crust range from about 10° to about 40°C/km and from the work on the Gulf Coast, we see that gradients of 24°C/km are normal for sedimentary basins.

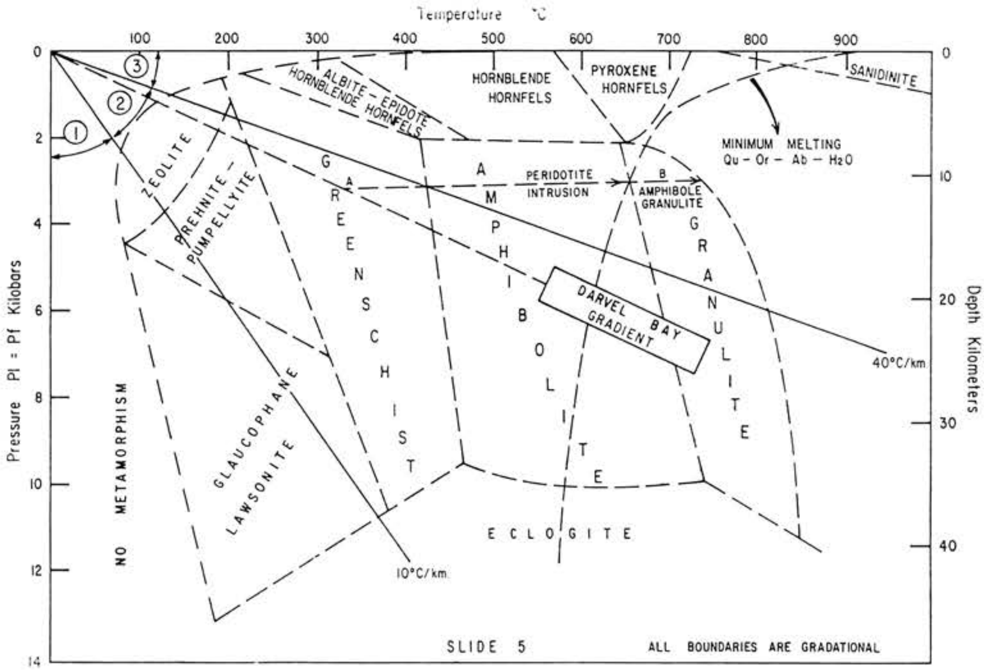

Fig. 5. Pressure—temperature scheme of the metamorphic facies after Turner (1968). Regime 1 = regions of abnormally low geothermal gradient; 2 = regime of normal gradient; 3 = regime of exceptionally high geothermal gradient. See text for details of the Darvel Bay gradient.

Let us see now what this means in terms of the metamorphic facies. The next slide (fig. 5) shows a recent summary of the pressure/temperature scheme of the metamorphic facies as outlined by Turner (1968). The whole classification is based on the assumption, which generally seems to be valid, that the pressure acting on rocks during metamorphism is actually load pressure and that tectonic overpressure, if any, is insignificant. I have subdivided the diagram into three regimes: 1 = terranes of abnormally low geothermal gradient. 2 = terranes of normal geothermal gradient; everything within this field may be considered as normal dynamothermal metamorphism, and can be explained by the principle of uniformitarianism accepting the present day range of geothermal gradients extrapolated backwards in time to apply to Phanerozoic orogenic systems. Regime 3 is a terrane of abnormally high geothermal gradient, and includes what Miyashiro (1961) has called the Abukuma facies series. Regime 3 characteristically includes the hornfels facies series resulting from the high thermal gradients around igneous putons. Holland and St. John Lambert (1969) have equated the greenschist facies with widespread schistosity, similar and isoclinal folding; the amphibolite facies with similar and isoclinal folding initially, then flow folding and diapiric gneiss domes in the upper grades; and the granulite facies with simple structural styles. The hornfels facies is characterised by hornfelsic textures or crudely developed schistosity transitional to the greenschist and amphibolite facies styles with increasing depth of burial.

I have shown on this diagram (fig. 5) the probable geothermal gradient for the Darvel Bay area of Sabah, which has been described by Hutchison and Dhonau (1969, 1971). The highest grade of regional metamorphism now exposed is greenschist facies, which would indicate that the rocks now exposed at the surface were buried during orogenesis at a depth of 10 km or so. In the islands such as Tabawan and Silumpat, the grade of metamorphism has been shown to increase from greenschist facies to hornblende granulite facies towards a syntectonic peridotite intrusion. The geothermal gradient no longer follows the regional gradient near the peridotite body, but with constant pressure (equivalent to 10 km depth) the temperature increases towards a maximum of around 700°C towards the peridotite intrusion. This is an example of a contact aureole around an alpine-type peridotite intrusion, but note that the high temperature peridotite cannot cause any of the metamorphic facies commonly associated with contact metamorphism, and since the intrusion is syntectonic and deep seated, it takes the metamorphism through a sequence of dynamothermal metamorphic facies —through amphibolite to hornblende granulite facies. All these rocks within the aureole of the peridotite are well foliated and the hornblende well lineated as a result of the continuing pressure of 3kb. Hence if the intrusion is into a crustal level which already is undergoing conditions of dynamothermal metamorphism, the hornfels facies will not be possible and the aureole will be characterised by an increase of dynamothermal facies.

Hence I would like to introduce a new classification for metamorphism, as follows:

A metamorphic aureole, characterised by the hornfels facies should be referred to as a *contact aureole*.

Contact metamorphism should be reserved for high level metamorphism related to intrusions into rocks which would otherwise be unmetamorphosed without the intrusion.

Thermal aureole should be used to refer to an increase of dynamothermal metamorphism towards a deep seated pluton. Away from the influence of the pluton, the rocks would have to be shown to be undergoing regional dynamothermal metamorphism. The aureole would be characterised by the dynamothermal facies series—greenschist and amphibolite. The definition of thermal aureole is consistent with the usage of *thermal dome* after Wenk (1962) to indicate localised regions within an orogenic terrain which had a locally higher thermal gradient. Thermal domes look like thermal aureoles on outcrop maps but lack the central pluton; both are caused by a higher thermal gradient than normal, the aureole by a hot pluton, the dome by a higher heat flow from the underlying mantle.

My next diagram (fig. 6) compares the effects of two intrusions into different levels within an orogenic belt. Intrusion 1 emplaced in the mesozone at a depth of 10 km will cause an increase in the dynamothermal metamorphic facies around itself from the regional greenschist to amphibolite facies, so that amphibolite facies metamorphism

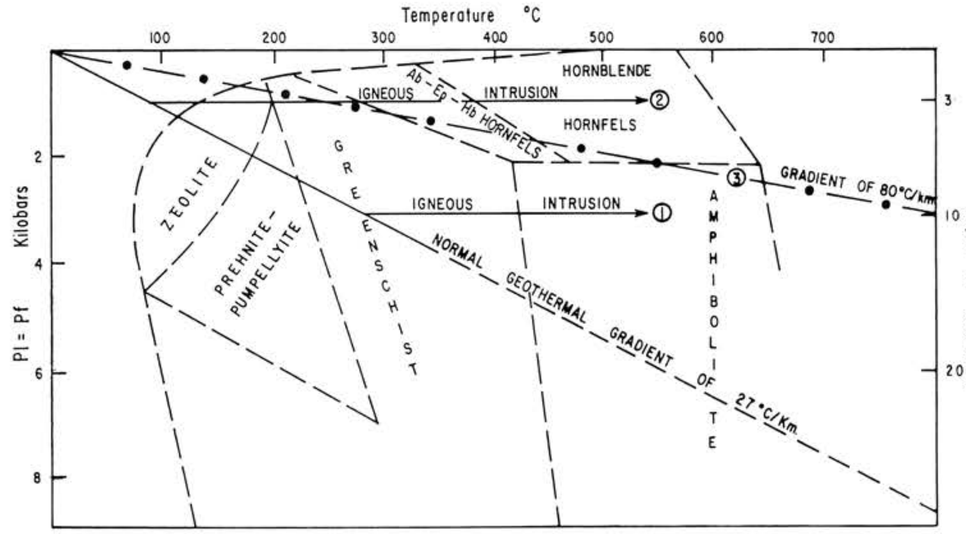

Fig. 6. Comparison of the effects of (1) a granite emplacement into the mesozone; (2) a granite emplacement into the epizone; and (3) a regional high geothermal gradient. See text for discussion.

will take place at a depth of 10 km instead of 18 to 20 km that would be necessary had the pluton not been emplaced. Intrusion 2, on the other hand is emplced into the epizone of the orogenic belt and without it, at a depth of 3 km the rocks are not undergoing any kind of metamorphism. The increase of temperature towards the pluton will cause the formation of hornfels facies rocks. This is *sensu classico* what we know as contact metamorphism, characterised by hornfels, whereas example 1 is also contact metamorphism but the auroele is devoid of hornfels and is characterised by foliated dynamothermal rocks. Hence for 1 I propose we call the aureole around the pluton a *thermal aureole*, and the contact around 2 a *contact aureole*. It should be noted here that the contact metamorphism in Malaya is mainly of type 1, the granite plutons are generally of the mesozone, so that the rocks at the level of the plutons are regionally under greenschist facies; the extra heat from the batholith may locally increase the metamorphism to amphibolite or hornblende hornfels facies (Hutchison, in preparation).

Some regions of the earth's crust are characterised by very high geothermal orogenic gradients such as 3 (fig. 5). The effect of contact metamorphism caused by an intrusion into such a region would clearly be much less marked than in a Barrovian terrane. In such a region the hornfels facies may equally well result from regional as from contact metamorphism. These metamorphic terranes have been called the low pressure or Abukuma facies series. Established examples are to be found in the Ryoke belt of Japan (Miyashiro, 1961), the central Pyrenees of Spain (Zwart, 1962), northern Portugal (Brink, 1960) and the Tasman metamorphic belt of New Zealand (Landis and Coombs, 1967). In all of these, kyanite is absent in pelitic rocks and andalusite goes to sillimanite with increasing temperature.

The deduction of thermal gradients in orogenic belts must be based on experimental laboratory studies of mineral transformation. The next slide (fig. 7) summarises some

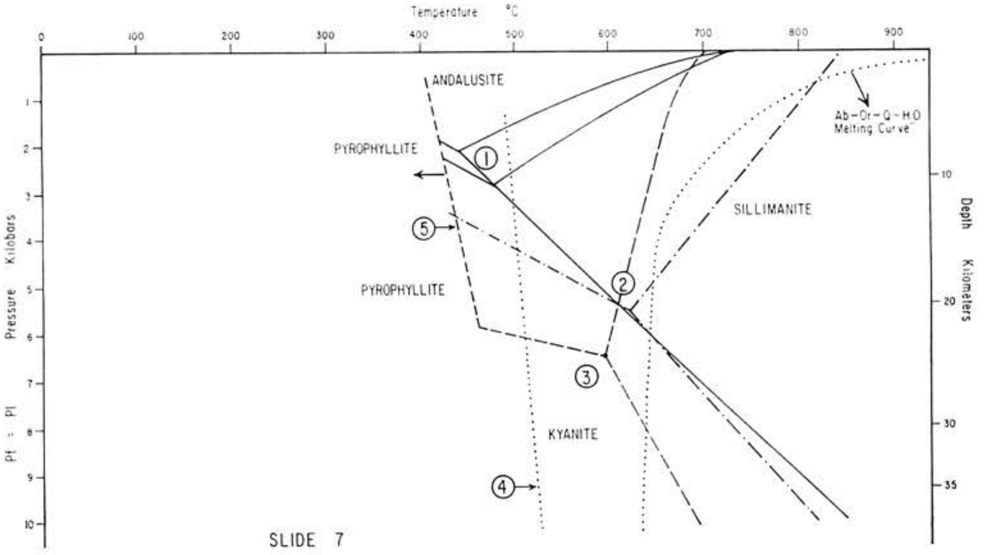

Fig. 7. Pressure—temperature diagram showing the positions of the andalusite-kyanite-sillimanite triple point according to (1) Newton (1966) and Weill (1966); (2) Richardson, Gilbert, and Bell (1969); and (3) Winkler (1967). Lines 4 (after Winkler, 1967) and 5 (after Kerrick, 1968) give the upper limits of the stability field of pyrophyllite.

of the confusion that exists in this field. One of the major mineral groups in metamorphic interpretation is the Al_2SiO_5 polymorphic group andalusite—sillimanite—kyanite. The slide shows the more recent determinations of the stability fields of these three phases as consistent with the work of Newton (1966) and Weill (1966). The position of the triple point at around 470°C and pressure of 2.6 kb (point 1) has been accepted by Turner (1968) and by Fyfe and Mackenzie (1969) as being consistent with the entropy values of the various polymorphs.

This low pressure position of the triple point differs radically from earlier determinations. Point 3 shows that given by Winkler (1967), but the last word has not been said on the positioning of the triple point because Richardson, Gilbert and Bell (1969) gave point 2.

Line 4 gives the subdivision of the stability field of sheet-silicate pyrophyllite on the left and kyanite and andalusite on the right after Winkler (1967), while 5 is the boundary given by Kerrick (1968). The position of 4 can hardly be accepted because it almost eliminates andalusite completely. Indeed 5, which is more acceptable, shows also that the andalusite field is extremely restricted in terms of pressure and temperature combinations.

I included this slide to illustrate to you the imperfection of the experimental foundations of metamorphic petrology. Indeed Fyfe and Mackenzie (1969) have cast doubts on the truly polymorphic relationship of kyanite, sillimanite and andalusite. There may well be slight differences in the chemistry of the three minerals which have not been taken into account by experimenters. These laboratory studies have generally ignored the possibility of the andalusite-sillimanite-kyanite group not being truly polymorphic. Study of chemical variations is urgently required. Also of importance may be the different morphology that sillimanite may exhibit—one as large prisms, the other as fine mats, called "fibrolite". These two contrasting habits could well be expected to vary chemically, but this kind of research has not been done.

Out of all this confusion, however, an increasing number of metamorphic petrologists are beginning to accept point 1 as the approximately correct triple point. Let us return now to our first slide (fig. 1). Assuming the invarient point 1 on this diagram, then terranes of andalusite-sillimanite facies series would require a thermal gradient 60 or more °C/km on the basis of $\mathbf{P}_f = \mathbf{P}_l$. This kind of high geothermal gradient is difficult enough to conceive under the simple model of metamorphism in which the pressure is solely considered to be load pressure. If there were a significant increment of tectonic overpressure, then the gradient may be pushed downwards into the kyanite field. The fact that andalusite—sillimanite regional metamorphism does exist is strong evidence for the general absence of tectonic overpressure in any significant amount. These high geothermal gradients are also characteristic of aureoles around intrusions, but clearly they have existed in regional metamorphic terranes in the past, otherwise kyanite would have been formed. The additional presence of cordierite in these rocks further points to low pressure conditions, for at pressures in excess of the range 5 to 7 kb (a guess based on field and experimental evidence) cordierite becomes unstable and the magnesium of pelites is accommodated exclusively in almandine and biotite (Turner, 1968). In any case there is no doubt that cordierite is characteristic of low pressure conditions (Schreyer and Yoder, 1964).

Andalusite is a common mineral in the Malay Peninsula. Cordierite is fairly abundant (Hutchison, in preparation), and sillimanite is well developed in the Stong Migmatite Complex of Kelantan where it coexists with cordierite (Hutchison, 1969). The only recorded occurrence of kyanite in Malaya has been noted by MacDonald (1968) in the Taku Schist. The inference from the relative absence of kyanite and abundance of cordierite is that the Malayan Orogen was one of high geothermal gradient around 60 or more °C/km, and as we have seen from our previous discussion in regions where there is a high geothermal gradient, it will be difficult to distinguish the effects of contact metamorphism from those of regional metamorphism. Indeed the exercise becomes futile because both effects would give similar mineral paragenesis.

Can we account for such high geothermal gradients? The only places which have high thermal gradients today are in the so called geothermal areas such as New Zealand, Iceland, Italy, the western United States, Japan and Mexico.

McNitt (1965) has generalized by saying that all occur in regions of late Cenozoic volcanic activity and that the heat is probably derived from shallow intrusive bodies. Gradients as high as 800°C/km have been measured but do not persist to depth. However a gradient of 180°C/km has been measured to a depth of 1000 metres. Many of these regions are characterised by numerous and complex fault systems along which numerous hot springs emanate. Turner (1968) speculates that the high geothermal gradients necessary for the formation of the Alpine Schist Zone of New Zealand may have been genetically related to the Alpine Fault. So also the major dislocation of the Median Line is close to the Abukuma low pressure metamorphic belt of Japan.

I should like to end off by suggesting that Malaya has been a region of abnormally high geothermal gradient, and indeed may still be, though no measurements are available from deep bore holes. It is interesting to speculate that the reason for the high thermal gradient of the Malayan Orogen may have resulted from the impressive amounts of volcanic material which have characterised the Malayan Geosyncline from Lower Palaeozoic through Mesozoic times. Further more it is conceivable that the abundant Late Carboniferous to Late Cretaceous granitic activity, with its culmination in the Late Triassic, so heated up the upper tectonic levels of the Malayan Orogen that the high geothermal gradient, though genetically related to the abundant granite plutons, cannot now be related directly to any single granite contact. Similarly the Abukuma terrane of Japan is characterised by a large amount of granitic rocks. "Synkinematic granites are usually abundant in the high-grade parts of the metamorphic terrane" (Miyashiro, 1961, p. 281). Hence the introduction of large amounts of magma may have raised the regional temperature generally but, as in Malaya, the metamorphism is regional in aspect and cannot be related in terms of contact metamorphism to any single pluton. The high flow of heat from the mantle or from the deeper levels of the crust may have been facilitated also by the impressive amount of faulting which characterises the peninsula. Indeed the common occurrence of hot springs along these fault zones suggests that the country continues to the present day to be a region of high geothermal gradient. But it is for the geophysicists to prove or disprove this theory.

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