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

High Pressure-Temperature Point on an Archaean Geotherm, Implied Magma Genesis by Crustal Anatexis, and Consequences for Garnet-Pyroxene Thermometry and Barometry [and Discussion]

M. J. O'Hara, G. Yarwood, A. B. Thompson, G. C. Brown and Janet V. Watson

Phil. Trans. R. Soc. Lond. A 1978 **288**, 441-456
doi: 10.1098/rsta.1978.0027

Phil. Trans. R. Soc. Lond. A. **288,** 441–456 (1978) [441]
Printed in Great Britain

High pressure–temperature point on an Archaean geotherm, implied magma genesis by crustal anatexis, and consequences for garnet–pyroxene thermometry and barometry

By M. J. O'Hara and G. Yarwood
Grant Institute of Geology, Edinburgh University, West Mains Road,
*Edinburgh EH*9 3*JW, U.K.*

The presence of metasediments indicates derivation of at least part of the 2.8 Ga old Scourie gneiss complex from a wet supracrustal series. The dry, large ion lithophile element depleted character of the present complex indicates an episode of anatexis and removal of granitic magmas during the evolution of the gneisses. This paper concentrates on the evidence for very high temperatures of metamorphism, the nature of the processes of anatexis and magma ascent and their implications for heat transfer in the Archaean crust; finally it reviews the credibility of some garnet–pyroxene thermometers and barometers in the light of the external constraints placed upon possible solutions by the crustal setting of these granulite facies gneisses.

Introduction

The 2.8 Ga old Lewisian rocks which outcrop around Scourie, Sutherland (NW Scotland) are predominantly quartzo-feldspathic layered granulite facies gneisses with smaller volumes of interlayered spinel-hornblende peridotites and pyroxenites, plagioclase–garnet–pyroxenites and some metasediments.

Some general aspects of the mineralogy of the Scourie gneisses have been surveyed and a tentative model suggested for the changes of pressure and temperature during the evolution of this complex, in relation to the conditions necessary for production and preservation of granulite facies rocks in general (O'Hara 1977).

In this paper the critical evidence is summarized for high temperatures and pressures (*ca.* 1150 ± 100 °C; 15 ± 3 kbar†) in these crustal rocks during their metamorphism. These establish a particularly hot and deep seated point among available estimates of Archaean crustal geotherms (Watson 1978).

These high temperatures make it inevitable that substantial partial melting would have occurred in any water-bearing supracrustal series which was subjected to such conditions. The presence of occasional metasediments in the complex (Bowdidge 1969; Barooah 1970; O'Hara, in preparation) indicates that part at least of the complex began as a supracrustal series. The general and very marked depletion (O'Hara 1960) of all the large ion lithophile elements (e.g. K, Rb, Ba and even Na) and also of water, all of which would be concentrated in any granitic partial melt fraction, suggests that an episode of anatexis has indeed occurred in the Scourie gneisses. The crystallization products of the majority of the melts which are inferred to have formed are not to be seen at the structural levels now exposed in the Scourie district. This paper considers the character and temperature of the magmas produced, the

---

† 1 kbar = 10^8 Pa.

influence of their movement on heat transfer in Archaean continental crust and on the consequent form of Archaean geotherms.

The Scourie gneisses exhibit a long history of progressive adjustment of mineral compositions and textures in response to falling temperatures and pressures in the succeeding 1.8 Ga during which time they were excavated and eventually exposed at the pre-Torridonian land surface.

The distribution of elements between coexisting minerals has ceased changing in response to falling pressures and temperatures at a different temperature for each mineral pair concerned. The effective closure temperatures (see Dodson 1976) are expected to vary with the mineral pair involved, the particular elements which are being exchanged and, perhaps, the crystallographic sites on which the exchange is occurring.

These problems are general to metamorphic complexes in which the recovery of the rocks from depth to the surface has been a slow process, unless the temperature of metamorphism was close to or below the closure or blocking temperatures for element interchanges between the particular minerals which are to be used as geothermometers or geobarometers.

Nevertheless, so far as the assumption will ever be justified in slowly cooled gneiss complexes, there are some rocks from Scourie in which the coexisting garnets, augites and hypersthenes may be deduced on textural evidence to have equilibrated. The temperatures and pressures at which equilibration occurred were those of conditions encountered during the history of recovery of these rocks from their peak of metamorphic conditions. Those pressure and temperature conditions can be closely constrained by available experimental data and by other geological considerations.

Many methods have been suggested for obtaining temperatures and pressures of equilibration of coexisting pyroxenes and garnets. These have been applied mainly to mineral assemblages of upper mantle derivation, where any temperature greater than *ca.* 400 °C, and any pressure greater than *ca.* 10 kbar is inherently acceptable. These same methods have been applied to the pyroxenes and garnets from Scourie, and the estimated pressures and temperatures are compared with the ranges of values constrained by the geological data.

These mineral assemblages from Scourie are proposed as a useful test case for future models relating compositions of coexisting garnets, pyroxenes, spinels (and feldspars) to their conditions of equilibration.

EVIDENCE FOR HIGH PRESSURES OF METAMORPHISM IN THE SCOURIE COMPLEX

Rocks described by O'Hara (1961) as 'transition gneisses' or ariegites, are found at the interface between pyroxenites and the relatively iron-rich garnet–augite gneisses. A re-examination of the textures of some of these indicates that the earliest recognizable mineralogy consisted of large (*ca.* 1 cm), relatively magnesian garnets set in a clinopyroxene matrix. At some stage the magnesian garnet became unstable and reacted partially or completely, thus

$$\text{Mg-rich garnet} \longrightarrow \text{less Mg-rich garnet} + \text{hypersthene} + \text{trace augite} + \text{calcic plagioclase} + \text{aluminous spinel} \qquad (1)$$

The reaction products form broad, well crystallized rims around the residual garnets, and some amphibole formed at the same time.

Subsequently, garnet regrew along every grain boundary between the original reaction products, but on a smaller scale (*ca.* 20–50 μm).

From these features, well seen in sample 10038, from Geodh Eanruig, Scourie More, it is inferred that the metamorphism during which the Mg-rich garnet crystallized occurred to the high pressure side of reaction (1) and that the recovery path of these rocks crossed the reaction boundary towards the low pressure side, and then returned towards the high pressure side while temperatures were high enough to permit significant reactions to occur, the regrowth of the garnet occurring at a significantly lower temperature than that at which the original breakdown occurred.

Material consisting of both the garnet and its reaction products was hand-picked from the coarsely crushed rocks, and held overnight at pressures between 10 and 16 kbar at 1150 °C in graphite crucibles. Garnet was largely replaced by its breakdown products at 10 kbar, but appeared to have increased in amount at 12 kbar and grown in abundance in runs at 14 and 16 kbar. At the temperatures of metamorphism deduced below, pressures at least as high as 12 kbar are, therefore, required to account for the original stability of the magnesian garnet–clinopyroxene gneisses. Experiments relevant to the temperature estimates have been conducted initially at a standard pressure of 15 kbar. At temperatures lower than 1150 °C, pressures lower than 12 kbar will suffice to stabilize the magnesian garnets; hence 12 kbar represents the extreme upper limit of acceptable pressure estimates for the conditions of equilibration of the garnets and pyroxenes produced by the reversal of reaction (1) in sample 10038.

Evidence for very high temperatures of metamorphism in the Scourie complex

(i) *Ilmenite-magnetite intergrowths*

Intergrowths of a type previously described (O'Hara 1961) are ubiquitous in the gneisses. No independent ilmenite has been seen. Assuming development of these intergrowths by oxidation of an originally homogeneous magnetite–ülvospinel solid solution, the composition of the original spinel would have been *ca.* $Mt_{60}Usp_{40}$. At an oxygen fugacity between those of the quartz–fayalite–magnetite and magnetite–haematite equilibria, a temperature greater than *ca.* 800 °C is required to stabilize this solid solution (Buddington & Linsley 1964).

(ii) *Augite ferro-pigeonite intergrowths*

These have been found in a quartz–ferrohypersthene–augite–magnetite gneiss (O'Hara, in preparation). Lamellae 0.1 mm broad of a pyroxene which is presumed to have been ferro-pigeonite originally, of bulk composition *ca.* $Ca_{16}Mg_{27}Fe_{57}$, are observed parallel to {001} of an augite host whose bulk composition was *ca.* $Ca_{37}Mg_{26}Fe_{37}$. Both of these pyroxenes have exsolved further small bodies of the other pyroxene ($\sim 4 \times 10$ μm) also elongated parallel to {001}. The augite component has also exsolved abundant {100} lamellae of hypersthene. The original 0.1 mm broad ferro-pigeonite lamellae have inverted to ferrohypersthene, with abundant lamellae of augite arranged parallel to {100} of the composite crystal.

Two coexisting clinopyroxenes with compositions similar to those of the augite and ferro-pigeonite from the Scourie rock, have been reported to be stable at 1040 °C at atmospheric pressure (Ross, Huebner & Dowty 1973). The effect of an increase of pressure is probably to reduce the mutual solubility of these two pyroxenes at a given temperature, and hence to raise the temperature required to stabilize this pyroxene pair.

The 0.1 mm thick lamellae of pigeonite are considered to have formed within the stability field of pigeonite of this composition, and at a temperature greater than *ca.* 1050 °C.

(iii) *Augite–hypersthene–plagioclase–ore intergrowths*

Lamellar intergrowths of hypersthene, plagioclase and iron ore arranged parallel to {100} of the host augite were reported from some of the garnetiferous basic gneisses (O'Hara 1961).

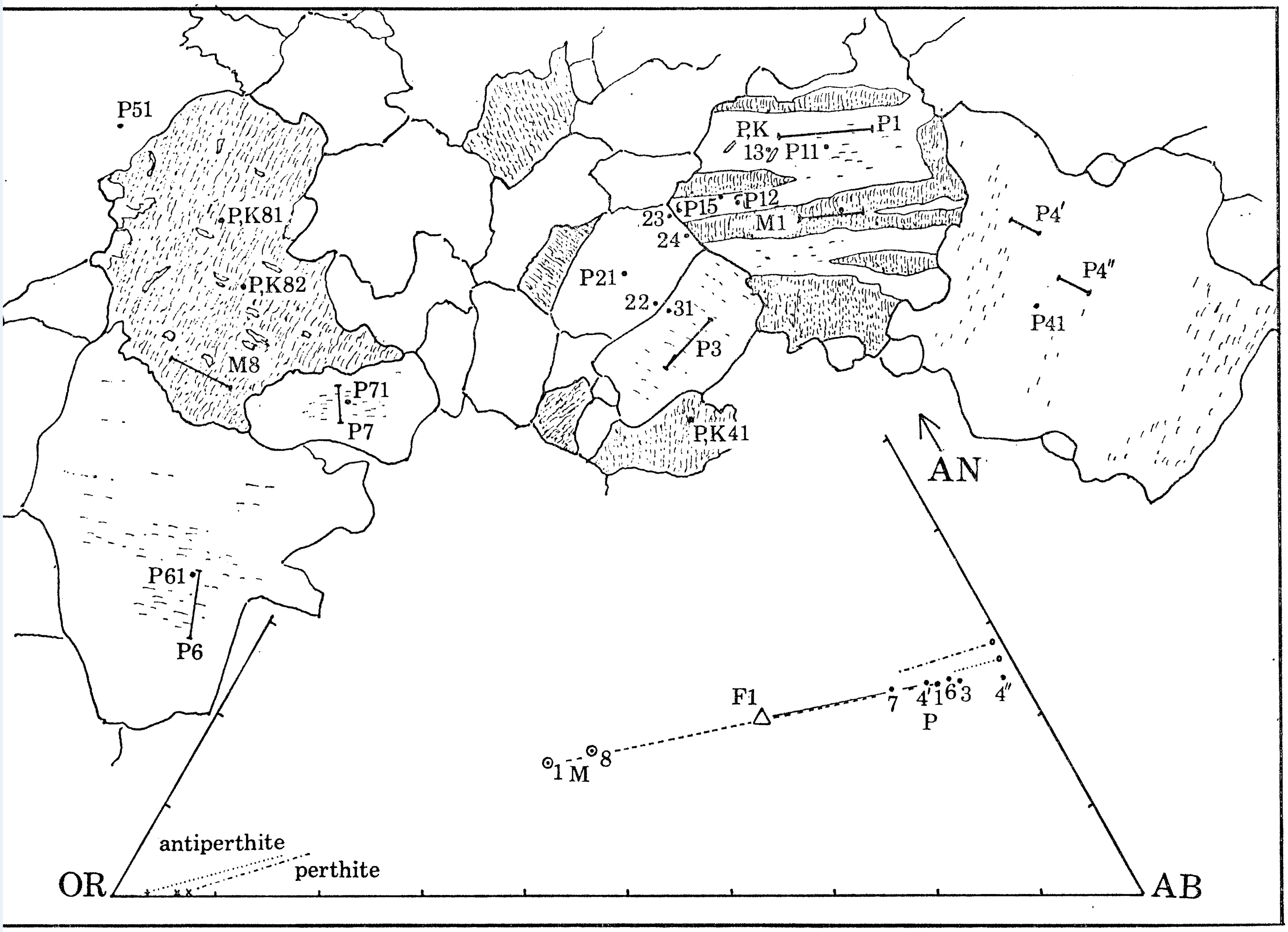

FIGURE 1. Sketch of the texture of the feldspars in part of sample H75123, a granulite facies pegmatite from the west side of the inlet immediately west of Pairc a Cladaich, on the north side of Scourie Bay. The field of view is *ca.* 0.6 cm wide. Lettered and numbered points and traverses show the sites of electron microprobe analyses of plagioclases (P), orthoclases (K) and mesoperthites (M). Results of these analyses are displayed in the inset projection into orthoclase (OR)–albite (AB)–anorthite (AN). The composition of the original feldspar, F1, (which composed the complex intergrowth shown slightly right of centre) has been obtained by integration of traverse analyses M1, P1. A solid tie line F1–P4′ links the postulated highest temperature assemblage, and a broken line links the components at the first stage exsolution (M1, M8 to P1, P6, P7). Temperatures of equilibration of these two assemblages are discussed in the text. A dot–dash line links the coexisting plagioclase and orthoclase in the mesoperthites, e.g. P, K 81, P, K, 82, and P, K 10 (point just above traverse M1) while the dotted line links coexisting plagioclase and orthoclase in the antiperthites, e.g. P, K 13. These latter tie-lines link the present compositions of the actual feldspars at the microscopic level and represent relationships in two independent second stages of exsolution. Temperatures of equilibration below 500 °C are calculated for these assemblages. Study at still higher magnifications will doubtless show a further history of exsolution proceeding to still lower temperatures.

Experimental studies on one of the garnetiferous basic gneisses (H75116) indicate that all plagioclase, hypersthene and iron ores can be dissolved in clinopyroxene at a temperature of 1100 °C at 15 kbar, but complete solution is not achieved at 900 or 1000 °C.

Sample H75116, in common with many of the more garnetiferous basic rocks from Scourie, contains all its plagioclase as narrow rims surrounding the garnets, with most of the hypersthene forming an outer rim surrounding the plagioclase, and the ore minerals (see (i) above) restricted to the granular augite matrix.

This texture, like the lamellar intergrowths previously described, is interpreted as due to exsolution of excess jadeite, calcium tschermak's component, hypersthene and iron–titanium spinel from an original high temperature pyroxene during the slow cooling of the complex.

(iv) *Orthoclase–plagioclase intergrowths*

Intimate regular, crystallographically orientated intergrowths of plagioclase (*ca.* An_{27}) and orthoclase (*ca.* Or_{95}) occur in the leucocratic gneisses associated with the basic and ultrabasic masses (O'Hara 1960, 1965). The mesoperthites are interpreted as exsolution intergrowths produced from a homogenous feldspar phase (*ca.* $Ca_{15}Na_{37}K_{48}$) and are comparable in many respects with those described by Smith (1974) from the high temperature, but low pressure, granulite facies gneisses of Rogaland, Norway. The mesoperthite coexists in some cases with independent antiperthite plagioclase (*ca.* $Ca_{23}Na_{69}K_8$).

Rarely, mesoperhite and antiperthite are themselves seen in orientated intergrowth (figure 1, grain 1). This also is interpreted as an exsolution or breakdown texture produced by the recrystallization of a phase (*ca.* $Ca_{19}Na_{54}K_{27}$) stable at a still higher temperature. This still higher temperature phase apparently coexisted with a relatively potassium-rich plagioclase (figure 1, grain 3 or 4) of composition *ca.* $Ca_{24}Na_{67-70}K_{9-6}$. Such highly calcic alkali feldspar compositions have not previously been reported, even from dry high temperature syenites. The effect of pressure is to widen the solvus between coexisting feldspars, hence raising the temperatures required to stabilize the original feldspars still further.

Application of the two feldspar geothermometer advocated by Stormer (1975) to the highest temperature feldspar assemblage suggests temperatures of *ca.* 1270 °C at 15 kbar pressure. Its application to the mesoperthite–antiperthite assemblage suggests temperatures of *ca.* 1000 °C at 15 kbar.

Using the revised two feldspar thermometer of Powell & Powell (1977) these temperatures become respectively *ca.* 650 and *ca.* 670 °C. These are inverted with respect to their textural relations. This inversion of the textural order of the parageneses probably results from the invalid nature of the assumptions that X_{Ca} in the alkali feldspar and X_K in the plagioclase are both small. Revision of the relationship to allow for the presence of significant amounts of these components has yielded estimated temperatures of *ca.* 1145 and *ca.* 850 °C respectively at 15 kbar for the higher temperature feldspar parageneses (R. Powell & M. Powell, personal communication).

(v) *Granulite facies pegmatites and other dry partial melts*

The leucocratic gneisses which are the source of the high temperature feldspar parageneses described in (iv) above are interlayered with the ultrabasic and garnetiferous basic gneiss complexes. Many of these are concordant layers with granulite facies mineralogy and are composed of abundant quartz and orthoclase–mesoperthite (*ca.* $Ab_{50}An_{25}Or_{25}$). These rocks have a good foliation, picked out by the lenticles of blue quartz, and have not been melted since the peak of the metamorphism.

In places, however, these leucocratic gneisses can be seen to cut across the foliation of the basic gneisses, but retain an internal mineral foliation parallel to the regional gneissic foliation (Bowdidge 1969). This indicates that melts of these compositions existed before the end of the metamorphic reconstruction.

Elsewhere these layers can be seen passing into somewhat coarser, unfoliated pegmatitic (*ca.* 1 cm grain size) modifications which retain, in part, the blue quartzes and dark feldspars of their granulite facies mineralogy. Such rocks, of which sample H75123 (figure 1) is an example, appear to have been melted, and then to have crystallized, in a nearly dry state after the climax of metamorphism. The scarcity of water during subsequent cooling is inferred from the preservation in the rocks of orthoclase, and of blue quartzes, which are among the first features to be destroyed in granulite facies rocks to which water has gained access during their cooling.

Experiments on the sample H75123 in the absence of water at 15 kbar indicate that partial melting begins about 1200 °C and is well advanced at 1300 °C. If water had been a significant component of the natural melts, the solidus temperatures would have been depressed below those at which the exceptionally calcic mesoperthites might have crystallized.

Summarizing, then, the combination of field, mineralogical and experimental evidence for the leucocratic gneisses indicates attainment of a temperature between 1050 and *ca.* 1250 °C near the climax of metamorphism. Evidence from exsolution structures in garnetiferous basic gneisses and metamorphosed ironstones support the estimate of original temperatures in excess of the range 800–1000 °C. Pressures were at least as great as 12 kbar, but there is as yet no evidence to support the stability of garnet–peridotite in this province. This implies pressures of less than *ca.* 18 kbar (O'Hara, Richardson & Wilson 1971). The pressure of metamorphism is, therefore, set at 15 ± 3 kbar, corresponding to depths of *ca.* 50 ± 10 km.

MAGMATIC AND THERMAL IMPLICATIONS OF THE HIGH METAMORPHIC TEMPERATURES

The argument which indicates that the present granulite facies complex at Scourie is a residuum of partial melting has been outlined in the Introduction. Partial melting, with the production of water-rich granitic–granodioritic melts would have begun at temperatures close to 650 °C, and would have continued with every incremental increase in temperature thereafter, irrespective of whether the successive partial melts (*a*) were aggregated *in situ* (equilibrium partial melting); (*b*) drawn off and erupted directly towards the surface (fractional partial melting); or (*c*) drawn off and aggregated in a magma chamber before commencing their progress towards the surface.

The amount of water likely to be held in the rocks before the onset of melting is unlikely to have exceeded 3 % by weight. The first liquids to form might have contained more than 10 % by weight of water. Fractional partial melting according to models (*b*) or (*c*) would, therefore, have rapidly depleted the residual complex in water. Only small amounts of melt need have formed at higher temperatures in models (*b*) or (*c*) after production of a relatively large initial amount of water-rich liquid at the onset of melting.

Given the nature of the residuum, and the probable peak temperature of anatexis (metamorphism) it is possible to predict the character of the last liquids extracted with some certainty.

The majority of the Scourie gneiss complex is composed of quartz-rich andesine–augite gneisses with primary hypersthene and titanium-rich magnetite. The last liquids produced must, therefore, have been hot (*ca.* 1150 ± 100 °C) magmas just saturated with quartz, andesine, augite, hypersthene and titanian magnetite at *ca.* 15 kbar. Such liquids would have contained relatively large amounts of normative quartz, and would have had relatively high ratios of plagioclase (probably oligoclase) to potassium feldspar in the C.I.P.W. norm. Their compositions might have resembled those of granodiorites or possibly quartz diorites.

If the model of anatexis suggested in (*a*) above is correct (equilibrium melting), these characteristics would have to apply to the *whole* of the magma which eventually separated and ascended into the upper parts of the crust. Furthermore, the concentrations of large ion lithophile (l.i.l.) trace elements such as Rb, Ba in the bulk magma would have to be those appropriate to partition between that liquid and the residual andesine crystals of the Scourie gneisses.

The relevant distribution coefficients are not large (Drake & Weill 1975). The present concentrations of such elements in the quartzo-feldspathic gneisses are extremely low, hence their concentrations in the last liquids to be separated from the gneisses must, in model (*a*), also have been low.

If anatexis is supposed to have proceeded along the lines of model (*a*), four conclusions follow. The original bulk chemistry of the complex undergoing partial melting would have had to have been depleted in l.i.l. elements. The magmas produced would have been granodiorite or quartz-diorite of low total water content (well below saturation with water vapour), low l.i.l. element concentration and particularly high temperature.

If, however, anatexis is supposed to have proceeded along the lines of model (*b*) by fractional melting, these conclusions would apply only to the last small batches of liquid extracted. The bulk of the liquids produced might have had the l.i.l. element concentrations, water contents and temperatures more commonly associated with the granitic rocks, and the original pre-anatexis complex need not have shown l.i.l. element depletion. Nevertheless, in this model the late arrival at high structural levels of at least small amounts of dry, hot, l.i.l. element depleted granodiorite or quartz–diorite is predicted. If anatexis is supposed to have proceeded along the lines of model (*c*), this last constraint would be removed.

Model (*c*), one of fractional anatexis with aggregation of the melt fractions elsewhere prior to the final emplacement, fractional crystallization and consolidation of the aggregated magmas, best describes the character of granitic intrusive complexes at higher crustal levels.

In all these models, the low initial isotope ratios of strontium and lead which are reported from the Scourie complex would characterize the anatectic magmas produced.

The effects of anatexis and magma ascent upon the heat flux towards the surface are to some extent model dependent. Assuming model (*c*) and postulating that aggregation of the partial melts occurred at a relatively high level in the crust, the effect would be to make the metamorphic region become an exceptionally good channel for the transfer of heat and heat producing elements towards the surface once temperatures in the metasedimentary pile had risen above *ca.* 650 °C. Locally the geotherm would be distorted, having a steep (large $\mathrm{d}T/\mathrm{d}P$) limb, rising from the surface conditions to the solidus temperatures of the crystallizing granodioritic–granitic magmas at some relatively shallow depth, followed by a less steep limb (perhaps not very different from the adiabatic gradient for the ascent of the magmas) extending back to a point on the steady state geotherm.

The marked inflexion in the geotherm established in an Archaean metamorphic belt would have been set at some depth *less than* that at which the simple adiabat for the ascending granodioritic magmas intersects the wet solidus for such materials, because the forced adiabatic crystallization of the magmas as they intersected the wet liquidus at low pressures would cause an increase in temperature of the systems (Tuttle & Bowen 1958) and hence an intersection with the wet solidus at some still lower pressure. As the deep seated region undergoing anatexis heated up and lost magma, so the shallow seated limb of the geotherm would become steeper and steeper, but always terminating at some point on the wet granite–granodiorite solidi. At greater depths, the gradient ($\mathrm{d}T/\mathrm{d}P$) would be less than that of the steady state geotherm until the seat of the anatexis had been passed.

This model is consistent with the general evidence on Archaean geotherms summarized by Watson (1978, this volume, fig. 1). It integrates the evidence from granite–greenstone belts with that from granulite facies terrains and would obviate the apparent need to form these two types of province in regions of fundamentally different heat flow in the Archaean crust. It would suggest that the data points for pressures and temperatures at depths of 15–30 km in the Archaean crust, conspicuously sparse in Watson's (1978) fig. 1, will be derived from terrains which have seen the passage, if not the emplacement of large volumes of acid magmas.

Some of this distortion of the hypothetical steady state geotherm in uniform materials would necessarily be long-lived because of the transfer of heat producing elements into the upper crust with the magmas. Assuming rates of uplift and erosion in the order of 0.02 mm/a and a mean depth of granite–granodiorite emplacement of 7 km (*ca.* 2 kbar confining pressure), the temperature distortion produced by the transfer of K, U, Th into the upper crust would have lasted for more than 350 Ma.

Establishment of the points on the steady state crustal geotherm is a more difficult matter. However, a deep-seated crustal complex undergoing *fractional* anatexis will have lost not only a large amount of latent heat of fusion but also the majority of its heat-producing elements early in the cycle of heating. It is, therefore, unlikely to have attained temperatures very much higher than those appropriate to the steady state geotherm near the base of the continental crust at the relevant time in Earth history.

CALIBRATION OF GARNET–TWO PYROXENES THERMOMETERS AND BAROMETERS

A model of the cooling and recovery history of the Scourie complex has been proposed (O'Hara 1977) which accommodates the requirements (imposed by transition gneiss samples such as 10038, mentioned above) that the cooling path should pass to the low pressure side of the breakdown curve for the more magnesian garnets (but not for that of more iron-rich garnets), and then pass back towards the high pressure side. During this cooling history it is proposed that the gneisses cooled from *ca.* 1150 ± 100 to *ca.* 450 ± 50 °C while buried at depths between 45 and 25 km, during a period of about 400 Ma.

As the complex cooled, solid solutions stable at the higher temperatures exsolved and in part recrystallized. Adjacent minerals continued to exchange elements as the temperature fell, and to order their internal structures, each individual process closing over separate relatively narrow temperature intervals set at absolute values of temperature which will depend upon the elements and the minerals involved, and the cooling rate (Dodson 1976).

Garnet–pyroxene barometry has in the past depended in principle upon first estimating the temperature from the distribution of calcium between coexisting orthopyroxene and clinopyroxene, and then the pressure at that particular temperature from the alumina content of orthopyroxene coexisting with garnet (O'Hara 1967; Boyd 1973; Wood & Banno 1973). The method has been applied to rapidly quenched garnet–lherzolite assemblages brought up very rapidly in kimberlite diatremes, where it may be assumed that few problems of closure exist, but the only geological constraints on the possible answers are that they should lie at temperatures below the beginning of melting of the garnet–lherzolites and at pressures greater than *ca.* 10 kbar.

In the case of the garnet–two pyroxene assemblages of the Scourie, and other, granulite facies gneisses the closure problems may be severe, although the geological constraints on the possible answers are tight enough to provide a test of the success or otherwise of different methods of estimating temperature and pressure. It is certain that the thermometers and barometers will not record the peak conditions of metamorphism of the complex unless this lay below their respective closure temperatures. There is, moreover, no sure basis for assuming that they will together refer to an actual point or points on the real cooling and unloading path followed by the complex. This problem is general to deep seated metamorphic complexes. There are, however, at least two parageneses among the Scourie gneisses where it may be assumed that the coexisting garnets and pyroxenes are as close to mutual equilibration as is ever likely to be observed in such rocks.

In the garnet-rich basic gneiss, H75116, large (1 cm) garnets of uniform, unzoned composition totally enclose a few small grains of augite and hypersthene which are also unzoned, and in mutual contact.

In the transition gneiss 10038, garnet of uniform composition has reformed as narrow rims between plagioclase, hypersthene, spinel and amphibole (also of uniform composition) and around occasional large augite grains whose outer portions have undoubtedly recrystallized.

Electron microprobe analyses of these minerals are shown in table 1.

Table 2 shows the results of applying a variety of thermometers and barometers to these garnet–two pyroxene assemblages. The plausible answers are constrained by the following considerations.

The compositions of coexisting pyroxenes of sample H75116 in the body of the rock differ little from those quoted in table 1. These generally form 0.1–1.0 mm grains. Experiments in which the powder of this rock was held in graphite capsules at 900 °C overnight at a pressure of 15 kbar yielded a product in which there was a significant decrease in the amount of hypersthene, and a decrease in the Ca/(Ca + Mg + Fe) ratio of the clinopyroxene. Irrespective of all considerations of kinetics, and attainment of equilibrium, this result is taken to indicate that the pyroxenes attained their present compositions at a temperature significantly lower than 900 °C. The coexisting fine-grained pyroxenes quoted from sample 10038 are considered to have equilibrated at a still lower temperature, following the previously mentioned model (O'Hara 1977) (although the larger augites and hypersthenes present in this sample resemble those quoted from H75116).

The pressures of equilibration of the garnets and orthopyroxenes are taken to lie in the range 8.5 ± 3.5 kbar in accordance with the model previously suggested (O'Hara 1977) and controlled in part by the experimental data for sample 10038 described above.

Temperature estimates in table 2, rows 1 and 12 are based upon an ideal solution model for the coexisting pyroxenes, and calibrated against the Davis & Boyd (1966) determination of solvus limits in the diopside–enstatite system.

However, Powell (1978) has demonstrated that an ideal solution model is not applicable to the two pyroxenes; temperature estimates in rows 4, 13, 14 and 15 are based on his revised solution model, and are calibrated against the corrected form of the diopside–enstatite solvus (Nehru & Wyllie 1974; Howells & O'Hara 1975; Mori & Green 1975, 1976; Lindsley & Dixon 1976) with due allowance for the pressure effects noted by Howells & O'Hara (1975).

Temperature estimates in row 5 are based upon the Ca–Mg exchange reaction between orthopyroxene and garnet (Powell 1978) calibrated principally against the experimental data of Akella (1976).

Table 1. Analyses of coexisting pyroxenes, garnets and spinels suitable for tests of geothermometers and geobarometers

	SiO_2	TiO_2	Al_2O_3	Cr_2O_3	FeO	MnO	MgO	CaO	Na_2O
H75116: relatively large inclusions in large garnets									
Opx$_{11}$	52.51	0.06	2.69	n.d.	21.09	0.21	23.29	0.59	0.02
Cpx$_{11b}$	50.37	0.76	5.21	n.d.	7.89	0.06	12.43	21.65	0.99
Gt$_{11}$	39.10	0.08	21.98	n.d.	24.91	0.76	7.87	6.35	0.04
Opx$_{12}$	51.40	0.13	3.56	n.d.	22.15	0.20	21.94	0.37	0.01
Cpx$_{12}$	50.50	0.62	5.50	n.d.	8.40	0.05	11.89	21.72	1.03
Gt$_{12}$	39.30	0.04	22.01	n.d.	24.44	0.76	8.02	6.42	0.03
Opx$_1$	52.04	0.13	2.97	n.d.	21.51	0.19	22.99	0.48	0.02
Cpx$_1$	49.70	0.58	5.56	n.d.	10.16	0.15	12.18	20.21	0.99
Gt$_{A1}$	39.13	0.05	21.95	n.d.	24.82	0.76	7.89	6.30	0.04
10038: intimately intergrown garnet, two pyroxenes, spinel (+plagioclase, amphibole)									
Opx$_{1,2}$	53.14	0.08	2.70	0.02	17.46	0.11	26.17	0.26	0.03
Cpx$_{2,3}$	49.45	0.84	6.53	0.05	6.44	0.06	12.46	22.42	0.85
Gt$_{4,5}$	39.43	0.01	22.37	0.05	22.95	0.59	9.03	6.13	0.01
Sp$_2$	0.02	nil	59.73	2.08	24.32	0.07	12.93	0.07	nil
Cpx$_2$	49.90	0.82	6.26	0.05	6.15	0.04	12.79	22.49	0.81
Gt$_1$	40.80	nil	22.27	n.d.	22.48	0.49	8.58	6.75	0.10
Opx$_3$	52.16	0.09	4.79	0.01	17.15	0.10	24.73	0.26	0.04
Cpx$_5$	50.37	0.71	6.08	0.03	5.39	0.02	13.02	22.58	0.76
Gt$_6$	39.87	0.01	22.28	0.04	23.20	0.47	9.00	5.68	nil
Sp$_5$	nil	nil	59.20	2.58	24.58	0.05	12.30	0.06	nil
Opx$_6$	52.20	0.09	4.27	n.d.	17.82	0.12	24.97	0.31	0.02
Cpx$_8$	49.07	0.81	6.48	n.d.	6.48	0.05	12.43	22.28	1.00
Gt$_8$	39.83	0.01	22.46	nil	22.24	0.58	9.06	6.80	nil
Sp$_7$	0.03	nil	59.98	1.68	25.27	0.08	12.68	0.06	nil

H75116. Opx$_1$–Cpx$_1$ are compiled from 150 μm traverse analyses (to integrate fine scale exsolution lamellae). Gt$_{A1}$ is average of 4 analyses in the enclosing garnet, including 2 points away from inclusions.

10038. Coexisting spinel analyses grouped with pyroxenes and garnets. Spinels also contain 0.30 % NiO. The coexisting plagioclases are *ca.* An$_{74}$; Cpx$_2$–Gt$_3$ are immediately adjacent; Opx$_{1,2}$ and Gt$_{4,5}$ are adjacent. All grouped analyses are within a maximum of 100 μm of each other.

These last two groups of temperature estimates are more credible than the first in view of the solution of orthopyroxene and alumina into clinopyroxene which occurs in *ca.* 17 h at 900 °C, 15 kbar in sample H75116, and the probability that the garnet–two pyroxene assemblages quoted from 10038 have equilibrated at still lower temperatures than those in H75116. Temperature estimates based on the Raheim & Green (1974) garnet–clinopyroxene thermometer fall into the credible range also. Results based on Saxena's (1976) two pyroxene thermometer (row 16) appear to be too low.

Pressure estimates in rows 12–15 are based upon the rather small effect of pressure upon the

Table 2. Temperature (°C) and pressure (kbar) estimates for equilibration of some Scourie garnet–pyroxene parageneses

	H75116 inclusions within large garnets			10038 garnet growths between breakdown products of earlier garnets		
†	Opx$_{11}$–Cpx$_{11}$ –Gt$_{11}$	Opx$_1$–Cpx$_1$ Gt$_{A1}$	Opx$_{12}$–Cpx$_{12}$ –Gt$_{12}$	Opx$_{1,2}$ Cpx$_{2,3}$ Gt$_{4,5}$	Opx$_3$ Cpx$_5$ Gt$_6$	Opx$_6$ Cpx$_8$ Gt$_8$
1	828	888	805	811	824	789
2	691 ± 24	733 ± 25	727 ± 26	668 ± 24‡	661 ± 24	628 ± 23
3	775 ± 18.0	842 ± 26.1	808 ± 17.5	698 ± 13.1	693 ± 13.3	642 ± 11.1
4	792 ± 10	767 ± 10	699 ± 4	653 ± 2	660 ± 2	655 ± 2
5	726 ± 31	686 ± 28	652 ± 26	625 ± 25	630 ± 25	635 ± 25
6 (*a*)	0.8	0.5	**1.7**	0.8	**4.3**	**3.5**
(*b*)	3.2	3.6	**0.9**	4.1	**4.4**	**2.8**
7 (*a*)	**6.8**	**6.7**	**9.3**	**7.0**	**11.7**	**11.1**
(*b*)	**4.5**	**3.7**	**8.6**	**3.7**	**11.8**	**10.5**
8 (*a*)	7.6	9.3	3.7	8.8	3.8	3.0
(*b*)	9.9	12.2	4.4	12.1	3.7	3.6
9 (*a*)	593 **4.0**	725 **2.6**	600 **5.5**	566 **3.7**	547 **8.6**	522 **7.9**
(*b*)	611 **1.9**	737 **0.4**	607 **4.7**	602 **1.0**	545 **8.7**	527 **7.3**
10 (*a*)	498 **15.5**	686 **11.7**	499 **16.6**	529 **12.1**	477 **18.9**	454 **18.7**
(*b*)	513 **13.7**	698 **9.2**	502 **16.1**	528 **12.3**	475 **19.2**	457 **18.2**
11 (*a*)	614 **1.5**	751 3.6	615 **3.7**	610 **0.1**	565 **5.9**	528 **7.0**
(*b*)	638 1.4	763 6.6	623 **2.9**	636 3.8	563 **6.1**	532 **6.5**
12	26.9	29.8	17.3	29.6§	32.7	33.2
13	841 26.1	926 46.9	695 5.1	651 6.0	660 8.4	658 13.1
14	397 **27.9**	864 31.6	608 **4.6**	v. high	v. high	591 2.2
15	799 18.8	815 25.1	708 15.9	655 12.9	633 13.1	657 11.5
16	298	292	237	272	no convergence	

All pressures in bold type are negative.

†1. Two-pyroxene temperature (Wood & Banno 1973, equation 27).

2. Clinopyroxene–garnet temperature (Raheim & Green 1974) at 8.5 ± 3.5 kbar.

3. Simultaneous solution, temperature from Raheim & Green (1974), pressure from Powell (1978, equations 22–26).

4. Two-pyroxene temperature (Powell 1978, equation 22) at 8.5 ± 3.5 kbar.

5. Orthopyroxene–garnet temperature (Powell 1978, equation 26) at 8.5 ± 3.5 kbar.

6. Orthopyroxene–garnet pressure (Wood & Banno 1973, equation 17), temperature as in (1).

7. As (6), but with revised ΔH, ΔS values of Wood (1974).

8. Orthopyroxene–garnet pressure (Wood 1974, equation 12), temperature as in (1).

(*a*) $X_{\mathrm{Al}}^{\mathrm{M1}} = \frac{1}{2}\mathrm{Al}$;
(*b*) $X_{\mathrm{Al}}^{\mathrm{M1}} = \frac{1}{2}(\mathrm{Al}+\mathrm{Na}-\mathrm{Fe}^{3+}-\mathrm{Cr})$ (1+ and 3+ charges balanced over sites, only Al and Si in 4-coordination)

9. As (6), but temperature from (2), simultaneous solutions for *P*, *T*.

10. As (7), but temperature from (2), simultaneous solutions for *P*, *T*.

11. As (8), but temperature from (2), simultaneous solutions for *P*, *T*.

(*a*) $X_{\mathrm{Al}}^{\mathrm{M1}} = \frac{1}{2}\mathrm{Al}$
(*b*) $X_{\mathrm{Al}}^{\mathrm{M1}} = \frac{1}{2}(\mathrm{Al}+\mathrm{Na}-\mathrm{Fe}^{3+}-\mathrm{Cr})$

12. Clinopyroxene–garnet pressure (Raheim & Green 1974), temperature as in (1).

13. Clinopyroxene–garnet pressure (Raheim & Green 1974), temperature from (4), simultaneous solutions for *P* and *T*.

14. Clinopyroxene–garnet pressure (Raheim & Green 1974), temperature from (5), simultaneous solutions for *P* and *T*.

15. Orthopyroxene–garnet pressure (Powell 1978, equation 26), temperature from (4), simultaneous solutions for *P* and *T*.

16. Orthopyroxene–clinopyroxene temperature (Saxena 1976).

‡ For cpx$_2$–gt$_1$ (table 1), 667 ± 24 °C.

§ For cpx$_2$–gt$_1$ (table 1), 29.3 kbar at 810 °C.

Fe–Mg and Ca–Mg exchange reactions between pyroxenes and garnets. The calculated pressures are therefore critically dependent upon the precision of the temperature estimated from one of the thermometers. Few of the results are credible.

Pressure estimates in rows 6–11 are based upon the changing solubility of alumina in orthopyroxene coexisting with garnet using ideal solution models, in some cases with empirical corrections for Fe/Mg ratio. The relatively large ΔV for the relevant reactions should make this a promising geobarometer, but Powell (1978) demonstrates that the results are liable to great uncertainties, not least because of the considerable uncertainty which attends the distribution of aluminium between the octahedral and tetrahedral sites in orthopyroxenes. This is brought out by the (*a*) and (*b*) results in table 2, rows 6–11.

There are, however, serious doubts about the uniqueness of the results for alumina solubility in orthopyroxene against which this geobarometer has been calibrated (Howells, Begg & O'Hara 1975; Howells 1975; Howells & O'Hara 1978, this volume).

The occurrence of some apparently credible results in row 8 can give no confidence. The empirical correction for iron content which is built into the relation used has been calibrated against pyroxene compositions grown from seeded glass starting materials, which is not the technique most likely to yield equilibrium.

If allowance is made for the real uncertainties attaching to the various estimates (see Powell 1978) all approaches listed in table 2 could encompass at least part of the pressure–temperature region within which the recrystallization of the minerals is constrained to have occurred.

One cannot therefore distinguish between the two alternatives, lack of precision in the geothermometers and geobarometers or lack of equilibrium due to closure problems between the coexisting minerals listed in table 1, either of which might account for the wide spread of estimates shown in table 2. At this stage in the development of these techniques it must be accepted that the accuracy of the estimates of equilibration temperatures and pressures in samples, such as the garnet–lherzolite nodules from kimberlites, may be poor.

Conclusions

The Archaean granulite facies gneisses at Scourie retain evidence of a slow, complex cooling history at relatively high temperatures and pressures. Presence of relics of metasediments in the complex suggests that during the approach to the highest temperatures and pressures, dehydration and depletion of the complex in large ion lithophile elements occurred, probably by a process of fractional anatexis with aggregation of the melt fractions before final emplacement of the granodioritic magmas at higher crustal levels. Relatively rapid convective heat transfer due to the movement of these magmas would have distorted temperature distributions in this segment of the Archaean crust for a long period of time. General geological considerations and experimental data for the rocks themselves permit the possible conditions of equilibration of garnet–two pyroxene assemblages in some of the rocks to be closely constrained. The calibrations and assumptions of various geothermometers and geobarometers of possible application to upper mantle rocks may be tested by their success in predicting the desired conditions from the reported mineral compositions. At present the accuracy and precision of these methods is less than satisfactory with regard to pressure estimates.

References (O'Hara & Yarwood)

Akella, J. 1976 *Am. Miner.* **61**, 589–598.
Barooah, B. C. 1970 *Scott. J. Geol.* **6**, 221–225.
Bowdidge, C. R. 1969 Ph.D. thesis, University of Edinburgh.
Boyd, F. R. 1973 *Geochim. cosmochim. Acta* **37**, 2533–2546.
Buddington, A. F. & Lindsley, D. H. 1964 *J. Petrol.* **5**, 310–357.
Davis, B. T. C. & Boyd, F. R. 1966 *J. geophys. Res.* **71**, 3567–3576.
Dodson, M. H. 1976 *Nature, Lond.* **259**, 551–553.
Drake, M. J. & Weill, D. F. 1975 *Geochim. cosmochim. Acta* **39**, 689–712.
Howells, S. 1975 *Ext. Abstracts, Int. Conf. Geothermom. Geobarom.*, Pennsylvania State University.
Howells, S., Begg, C. & O'Hara, M. J. 1975 *Phys. Chem. Earth* **9**, 895–902.
Howells, S. & O'Hara, M. J. 1975 *Nature, Lond.* **254**, 406–408.
Howells, S. & O'Hara, M. J. 1978 *Phil. Trans. R. Soc. Lond.* A **288**, 471–486 (this volume).
Lindsley, D. H. & Dixon, S. A. 1976 *Am. J. Sci.* **276**, 1285–1301.
Mori, T. & Green, D. H. 1975 *Earth planet. Sci. Lett.* **26**, 277–286.
Mori, T. & Green, D. H. 1976 *Am. Miner.* **61**, 616–625.
Nehru, C. E. & Wyllie, P. J. 1974 *Contr. Miner. Petr.* **48**, 221–228.
O'Hara, M. J. 1960 Ph.D. Thesis, University of Cambridge.
O'Hara, M. J. 1961 *J. Petrol.* **2**, 248–276.
O'Hara, M. J. 1965. *Geol. Mag.* **102**, 296–314.
O'Hara, M. J. 1967 In *Ultramafic and related rocks* (ed. P. J. Wyllie), pp. 393–403. New York: Wiley.
O'Hara, M. J. 1977 *J. geol. Soc. Lond.* (In the press.)
O'Hara, M. J., Richardson, S. W. & Wilson, G. 1971 *Contr. Miner. Petr.* **32**, 48–68.
Powell, R. 1978 *Phil. Trans. R. Soc. Lond.* A **288**, 457–469 (this volume).
Powell, R. & Powell, M. 1977 *Mineral. Mag.* (In the press.)
Raheim, A. & Green, D. H. 1974 *Contr. Miner. Petr.* **48**, 179–203.
Ross, M., Huebner, J. S. & Dowty, E. 1973 *Am. Miner.* **58**, 619–635.
Saxena, S. K. 1976 *Am. Miner.* **61**, 643–652.
Smith, J. V. 1974 *Feldspar Minerals* **2**, Berlin: Springer.
Stormer, J. C. 1975 *Am. Miner.* **60**, 667–674.
Tuttle, O. F. & Bowen, N. L. 1958 *Mem. Geol. Soc. Am.* **74**, 153 pages.
Watson, J. V. 1978 *Phil. Trans. R. Soc. Lond.* A **288**, 431–440 (this volume).
Wood, B. J. 1974 *Contr. Miner. Petr.* **46**, 1–15.
Wood, B. J. & Banno, S. 1973 *Contr. Miner. Petr.* **42**, 109–124.

Discussion

A. B. Thompson (*E.T.H., Zurich, Switzerland*). In their discussions on granulite facies metamorphism both Professor Watson and Professor O'Hara referred to 'dry' conditions. This concept no doubt needs further clarification in view of the confusion that may arise regarding the possible presence or absence of fluid in rocks metamorphosed to this grade. I am thinking in particular of those who wish to model the physical characteristics of the lower crust and upper mantle. Very few of the rocks considered to characterize the ill-defined granulite facies are without substantial quantities of hydrated minerals, notably phlogopite-rich micas and some amphiboles. Even the absence of hydrated minerals does not necessarily imply the high temperatures and pressures usually inferred for assemblages assigned to the granulite facies. If the $a\mathrm{H_2O}$ is low enough (presumably by dilution of the equilibrium fluid by some other abundant species, such as CO_2) then anhydrous assemblages may replace their hydrated counterparts several hundreds of degrees below the conventional limits of the 'granulite facies'. In fact large gradients in $\mu\mathrm{H_2O}$ may be extremely important in initial melting processes in the lower Crust and upper Mantle, where the increasing volume of melt effectively desiccates the enclosing rock. Obviously it is the value of $a\mathrm{H_2O}$ that determines whether or not a hydrate will form and the amount of H_2O then determines the modal amount of hydrate. It is certainly true that increasing temperature can produce the same results as decreasing $a\mathrm{H_2O}$, or vice versa, but this obviously does not mean that we can ignore one of these variables. There have been no systematic studies directed toward evaluating the value of $a\mathrm{H_2O}$ during 'dry' granulite facies metamorphism and until this is done we cannot adequately describe those physical properties that are dependent upon water concentration rather than on the amount of water.

It is appropriate to make some remarks about further problems in the application of cation exchange geothermometers and geobarometers in view of the extremely high pressures and temperatures deduced by Professor O'Hara for some of the Scourie dyke rocks. First, many of the commonly used 'calibrations' for ortho–clinopyroxene equilibria are incorrectly formulated as they attempt to fit a miscibility gap between two phases of different structure in terms of a solvus between phases of the same structure. Secondly, most minerals exhibit cation ordering with cooling (not necessarily slow even in geological time). In the case of coexisting isostructural phases, ordering will displace the 'solvus' to lower mutual solubility at constant temperature and pressure or conversely to higher temperatures at constant composition and pressure. Therefore, any attempt to use data from isostructural exsolved phases (e.g. haematite–ilmenite or alkali feldspars) could give erroneous estimates of temperature if ordering occurs during cooling.

Recent studies of ours on metagabbros from the Adirondacks in New York State, U.S.A., reveal similar mineralogy and textures to those shown by Professor O'Hara illustrating metamorphism of the Scourie dykes. Microprobe studies of adjacent minerals reveals complex compositional gradients presumably induced by diffusion at elevated temperature and pressure. If only occasional microprobe analyses are made without reference to precise location in the thin section and the currently available exchange thermometers and barometers are used, then great variation in estimated pressures (± 5 kbar) and temperatures (± 400 °C) can result. These are problems quite apart from the quality of the data and the exact solution models used in the calibration.

M. J. O'Hara and G. Yarwood. We share the prejudice that partial melting of wet crustal rocks at depth (where the initial melts would contain up to 18 % water in equilibrium with a water-rich vapour), followed by removal of these melts, is the most probable method by which granulite facies rocks became desiccated (and depleted in incompatible elements). We envisage the possibility of small amounts of an $\mathrm{H_2O}$-poor vapour remaining in the complex at Scourie during and after the peak of metamorphism and clearly there was sufficient water to produce substantial amounts of amphibole and a little mica, but only in the most magnesian rocks. These minerals appear to be 'primary', or at least to have formed very early in the cooling history.

Several features of the Scourie complex suggest, however, that the activity of water, and its absolute amount, were low during the later stages of metamorphism and the cooling history:

(*a*) The Scourian pegmatites would presumably have concentrated water in solution. On final consolidation this water would have been released and made available for alteration of the adjacent granulite facies gneisses. Effects which can be ascribed to this are of minor importance suggesting that the water contents of the pegmatitic melts were low, and hence that water was not available to enter these melts (the pegmatites form a very small proportion of the complex). Low activity of water in the vapour phase is a necessary consequence of this deduction.

(*b*) Retrograde alteration of the granulite facies gneisses near Scourie is largely confined to the formation of rims of amphibole and biotite around augite and hypersthene respectively in the acid and intermediate gneisses, followed by replacement of the interior of the grains by an aggregate of amphibole and quartz. The small grain size of the products and the observation that the earliest, marginal alteration affects pyroxene grains whose external shapes and bulk compositions had already been determined by re-equilibration events which may not have terminated until temperatures below 800 °C, combine to imply that amphibole in acid gneisses only becomes stable in equilibrium with the vapour phase at low temperatures. Because amphibole was stable in the more magnesian rocks, apparently at a much higher temperature ($>$ 1000 °C?) it is possible that much of the amphibole formation in the acid gneisses occurred in response to later diffusion of water into the complex, rather than because water had been present throughout as a minor component of the vapour phase. This would be in accord with the patchy nature of the preservation of fresh granulite facies material. Further studies will be necessary to establish in detail the manner in which the availability and activity of water varied in the Scourie complex during the metamorphism and subsequent cooling history.

The question of water availability does not affect the reading of the mineralogical evidence in favour of very high metamorphic temperatures. Stability of pigeonite (*ca.* $Ca_{16}Mg_{27}Fe_{57}$), the high temperatures indicated by two-pyroxene parageneses of the ultrabasic rocks, and the remarkable feldspar parageneses all point in the same direction. Experiments at 15 and 20 kbar in dry or 'damp' systems have reproduced the feldspar parageneses corresponding to the first stage of exsolution of the Scourie feldspars between 1000 and 1250 °C, but have so far failed to stabilize the mesoperthite of composition *ca.* $Ab_{50}An_{25}Or_{25}$.

We are in full agreement with Dr Thompson concerning errors in formulations and calibrations of geothermometers, and the question of cation ordering, which is part of the closure problem, concerns us greatly. As our paper attempts to show we are not yet in a position to distinguish the relative importance of the two factors in producing the wide range of implausible answers displayed in our table 2.

A principal purpose of the study of the Scourie gneisses was to evaluate the effects of local variations in mineral compositions upon the temperatures and pressures calculated for a single rock. It would appear that mineral grains are more homogeneous in composition in the Scourie rocks, perhaps reflecting a slower cooling rate (e.g. the ilmenite–magnetite thermometer appears to have closed at a much lower temperature at Scourie) than affected the Adirondack metagabbros. Nevertheless, there are small variations in grain compositions adjacent to their boundaries which appear (unpublished data) to imply continuing equilibration to somewhat lower temperatures on a scale of *ca.* 10 μm. The mineral compositions shown in table 2, H75116, and used elsewhere in our comments on these rocks (e.g. O'Hara (1977) *Geol. Soc. Lond. J.*, in the press) are representative of many point analyses taken over the interiors of relatively large grains, and the two groups of three columns illustrate that the variation of the apparent temperature and pressure of equilibration resulting from variation in mineral composition is somewhat smaller than that resulting from the use of different thermometers and barometers. The resultant total range of solutions shows a variation of ± 345 °C, ± 37.4 kbar in the one sample (H75116). This range does not, in our opinion, even include the actual temperature of the granulite facies metamorphism at Scourie. Irrespective of the more conspicuous composition gradients which may be frozen in near grain boundaries in faster cooled complexes, the minerals of the Scourie gneisses yield solutions for pressure and temperature which imply very severe closure problems. When we understand these better, there will be a wealth of information about cooling and uplift rates to be derived from these data.

G. C. Brown (*Department of Geophysics, Liverpool University*). I should like to sound a note of caution over the application of crustal fusion hypotheses to the genesis of Archaean granites. Former views that crustal fusion would occur as extrapolated geothermal gradients near 50 °C/km intersected experimental curves were based on the premise that such gradients existed in the crust *before* any melting occurred. I think we should realize that the maximum P–T conditions deduced from mineral geothermometers and geobarometers are the product of increased crustal temperatures *due* to magmatism. Furthermore, even these maximum conditions are turning out to reflect gradients of less than 50 °C/km for Archaean crust, e.g. 35–40 °C/km from Professor Watson's data and 27 °C/km from Professor O'Hara's data. These data suggest the application of crustal fusion models to the genesis of Archaean granites are on a less sound basis than when first suggested by Fyfe (*Phil. Trans. R. Soc. Lond.* A. **273**, 457–461 (1973)).

Janet V. Watson. I agree with the point made by Dr Brown. As I hope my text makes clear, I think that the available evidence for high geothermal gradients in Archaean times relates mainly to periods of granite emplacement. The gradients of interim periods remain to be established.