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# The metamorphism of the Dalradian rocks of western Ireland and its relation to tectonic setting

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[Plates 1 and 2]

The metamorphic evolution of Dalradian rocks exposed in the NW Mayo, Ox Mountains and Connemara inliers of western Ireland is reviewed, and new data and revised calculations are presented. There is evidence at a single locality for an early episode of moderately high-pressure metamorphism with the production of crossite–epidote schists. Subsequent Barrovian-style metamorphism overprinted this almost everywhere in NW Mayo and resulted in further heating to produce chlorite–biotite and garnet zones with the garnet isograd near 400–450 °C, and a staurolite–kyanite zone for which conditions were $P = 8 \pm 2$ kbar (1 bar $= 10^5$ Pa) and $T = 620 \pm 30$ °C.

In Connemara, rare staurolite–kyanite schists are now inferred to have formed at lower pressures than those in the Barrovian zones because ilmenite is stable rather than rutile. The metamorphic zones now mapped result from regional scale heating by synorogenic intrusions exposed to the south, and took place at pressures of 4–6 kbar. There is, however, strong evidence to suggest progressive uplift during this heating phase, especially in the south.

It is concluded that the thermal evolution of complex areas such as these can usefully be broken down into a series of modules, each corresponding to metamorphism in a distinct tectonic setting. The sequence identified in the Irish Dalradian is: (i) recrystallization during burial in a low heat flow setting; (ii) thermal relaxation and possibly uplift; and, in Connemara only, (iii) further heating and uplift, probably in the roots of a volcanic arc.

## Introduction

Our knowledge of the relations between regional metamorphism and tectonics comes in part from the gross disposition of different types of metamorphic belt, and in part from the detailed study of the metamorphic evolution of individual belts. Detailed petrological studies of the type presented here are only worthwhile when the stratigraphy and structure of the region are already well known, and in this respect the Irish Dalradian rocks are comparatively well served. More remarkably, the metamorphism of what is demonstrably the same stratigraphic sequence ranges within 50 km from low-temperature, transitional blueschist–greenschist facies assemblages typical of many circum-Pacific 'high-pressure' metamorphic belts, to low-pressure anatectic migmatites (Long *et al.* 1983; see also figure 1). For this reason the Dalradian rocks of Ireland provide a unique opportunity to study the relation between different basic types of metamorphism.

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Metasediments of the Dalradian Supergroup make up much of the southern Highlands of Scotland and can be correlated into Ireland (Harris & Pitcher 1975). They were deposited in Eocambrian to Cambrian times, and a late Precambrian tillite provides one of the most distinctive stratigraphic markers. In Ireland (figure 1), the most extensive Dalradian outcrops are in Donegal, NW Mayo, the Ox Mountains and Connemara. Whereas the other inliers are separated from one another principally by areas of unmetamorphosed Upper Palaeozoic sediments, the Connemara inlier is separated from the others by the low-grade Lower Palaeozoic volcanic and metasedimentary rocks of the south Mayo trough and the ophiolitic mélange of Clew Bay (Ryan *et al.* 1983). It is also unique in lying to the south of the Highland boundary line (Max & Riddihough 1975) and palaeomagnetic results are consistent with its having been rotated relative to the rest of the Irish Dalradian after the peak of metamorphism (D. J. Robertson, personal communication; Morris & Tanner 1977).

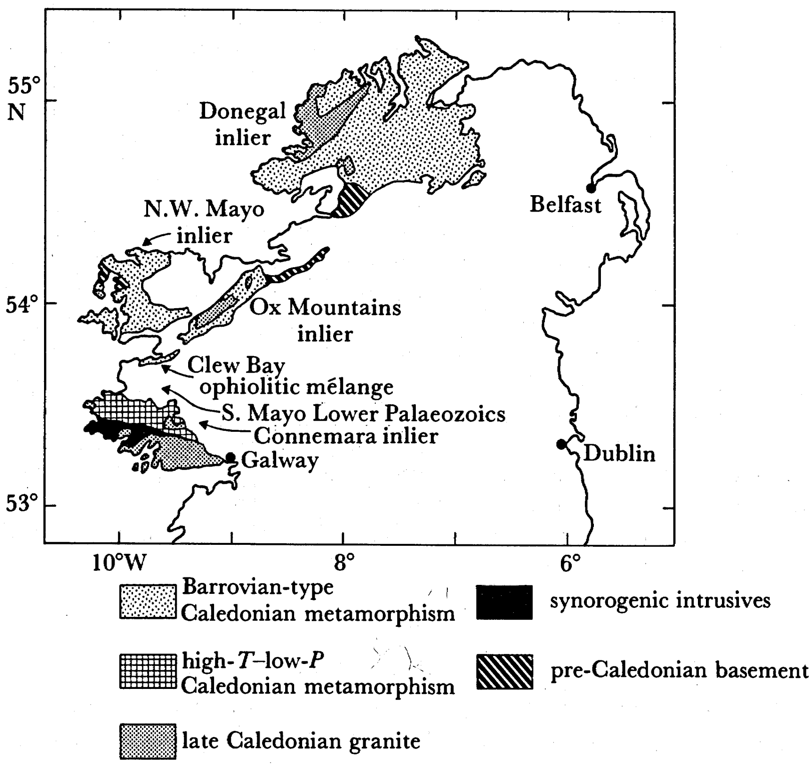


Figure 1. Map to show the distribution of inliers of Dalradian and older metamorphic rocks in Ireland. The location of the low-grade, Lower Palaeozoic inlier in S. Mayo is also indicated.

In this paper we synthesize published and unpublished data on the pressures and temperatures of metamorphism in the Dalradian of western Ireland, with particular emphasis on the NW Mayo and Connemara inliers, and attempt to deduce the changes in $P$–$T$ conditions with time and in relation to the development of structures. The resulting metamorphic and structural models provide important constraints on the Grampian orogeny during which most of the metamorphism took place.

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## Metamorphism in NW Mayo

Despite a number of recent investigations, some uncertainty remains about the structure and stratigraphy of this inlier. Trendall & Elwell (1963) produced the first metamorphic map of the area; subsequently it was shown by Sutton & Max (1969) that the migmatitic rocks of the western part of the inlier are, in fact, reworked basement gneisses. There are fundamental differences between alternative interpretations of the region (see, for example, Phillips *et al.* 1969; Crow *et al.* 1971; Max 1972; Kennedy 1980), but since the controversy primarily surrounds stratigraphic correlation within the Caledonian rocks and the relative importance of different deformation phases it is not of fundamental significance for the work described here.

The metamorphism in NW Mayo is predominantly of Barrovian type, and although the scarcity of pelitic lithologies and poor inland outcrop causes considerable uncertainty, a zonal map has been drawn up in figure 2. The northern part of the inlier displays the highest metamorphic grades with sporadic occurrences of kyanite and staurolite in pelites; an isolated

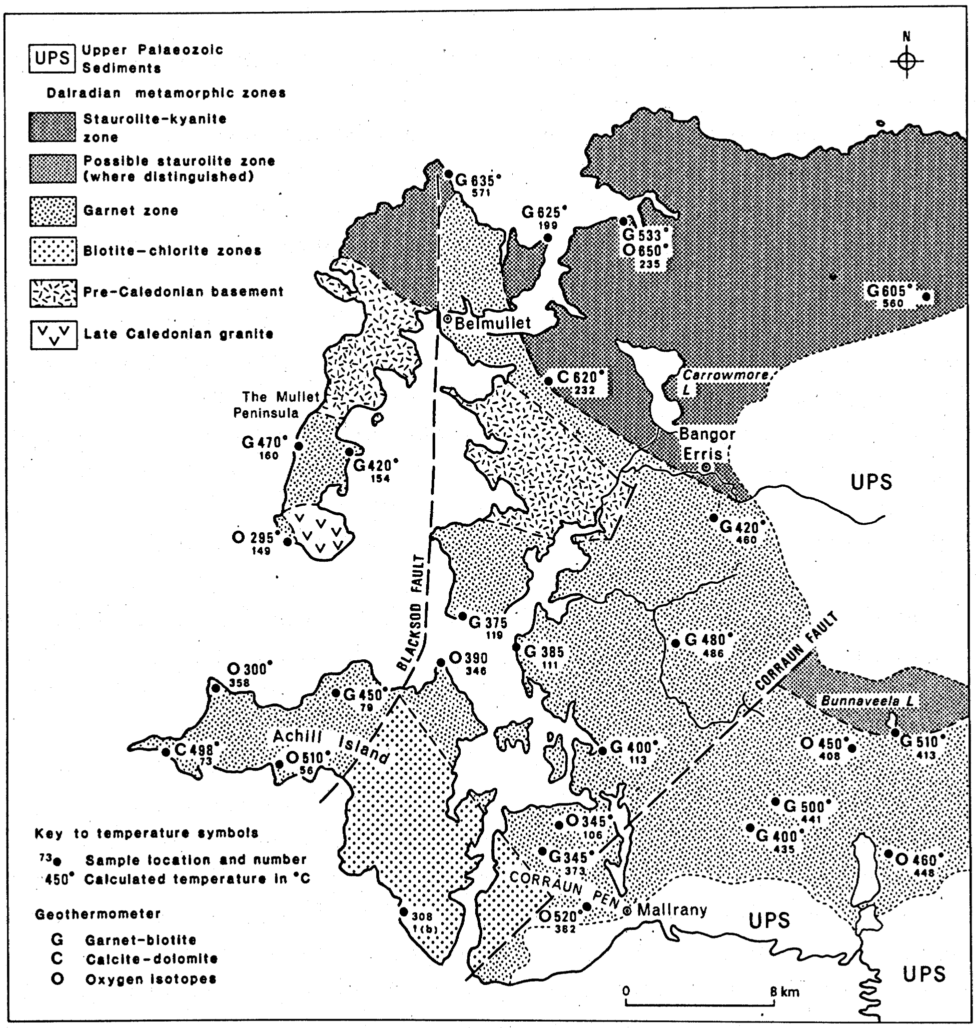


Figure 2. Metamorphic map of the NW Mayo inlier showing zonation based on pelite and semi-pelite mineral assemblages and the locations of samples for which geothermometer temperatures have been determined, together with results.

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occurrence of kyanite and staurolite intergrown in a vein near Bunaveela Lough suggests a local continuation of relatively high-grade rocks. In general it is not possible to subdivide staurolite and kyanite zones, but much of the inlier is of garnet grade, and garnet is found in a wide range of rock compositions. The typical pelite assemblage in the garnet zone is garnet + biotite + muscovite + albite or oligoclase + quartz ± chlorite; however, rare chloritoid schists have been found. In the semi-pelites, K-feldspar is commonly present. Epidote is a common accessory. Lower grade rocks occur in the southern part of Achill Island and adjacent parts of the Corraun peninsula. Here, plagioclase is invariably albite and the white mica is phengite with distinct pale brown pleochroism. Although no separate chlorite zone can be mapped with certainty, biotite is absent from some areas, or present only as inclusions in albite.

Sporadic metabasites show systematic variations in assemblage from actinolite + chlorite + albite + epidote (+ quartz, mica or stilpnomelane) at the lowest grades to hornblende + albite + epidote + chlorite and hornblende + plagioclase + epidote ± garnet with increasing grade. One garnet zone sample contains both albite and oligoclase.

### *Metamorphic textures and metamorphic history*

Even at the lower metamorphic grades, the pelitic and semi-pelitic rocks of this inlier are relatively coarse grained, with feldspar porphyroblasts up to 2 mm across occurring in a finer grained (up to 0.2 mm) matrix. In the garnet zone, garnet sometimes occurs as porphyroblasts up to 3 mm in size and often incipiently altered to chlorite at their margins. In other samples, however, garnet occurs as small euhedral grains, as little as 0.1 mm in diameter, present as inclusions in albite or oligoclase. Sometimes, comparable small garnets occur in the rock matrix also. Porphyroblasts at the highest grades are commonly 2 mm across and occasionally coarser. Phyllosilicates co-existing with staurolite and kyanite may exceed 1 mm in length.

The sequence of mineral growth relative to deformation has been investigated by Kennedy (1969), Max (1973) and Gray (1981). In general, two schistosities at most, S1 and S2, can be identified in individual sections, but a later S3 fabric associated with retrogression is sometimes present. Recent workers have assumed that the dominant fabric is normally S2. The metamorphic peak in the staurolite–kyanite zone clearly occurred after the D2 deformation (MP2 in the terminology of Sturt & Harris 1961). Garnet growth in these rocks is early (MP1–MS2). At lower grades there is an extensive region of plagioclase porphyroblast schists. These are well developed on the mainland N.E. of Achill Island (Kinrovar schists), but also occur throughout much of Achill Island itself and sporadically elsewhere (Trendall & Elwell 1963; Kennedy 1969). The plagioclase is albite over much of the southern part of this area, but oligoclase in the Kinrovar schists. Included fabrics in the plagioclase porphyroblasts usually indicate pre- to syn-tectonic growth with respect to the dominant foliation, and this has previously been taken to mean that growth was MP1 to MS2. However, albite porphyroblast schists from Achill Island are texturally and mineralogically identical to those of the SW Scottish Highlands, that have recently been shown by Watkins (1983) to be of retrograde origin. It is therefore possible that this is the case in NW Mayo also, and that an S3 fabric in fact dominates in the feldspar porphyroblast schists. On the other hand, Kennedy (1969) considered that there had been both an early phase of albite porphyroblast growth, corresponding to the peak metamorphic growth of garnet, and a later, retrograde growth after D2.

Crow & Max (1976) considered the Kinrovar schist to be pre-Caledonian because of the

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complexity of the inclusion assemblages in the feldspar porphyroblasts, but this would not be necessary if the regional fabric here were in fact S3. Garnet and biotite in the Kinrovar schist occur both as inclusions in feldspar and in the rock matrix, and therefore provide some indication of whether the inclusion relicts formed under substantially different conditions from the rest of the rock matrix. Microprobe analyses of the different textural varieties from 3 samples are presented in table 1, and show no significant differences. On Achill Island, index minerals (garnet or biotite) may occur in plagioclase porphyroblasts while being absent from the rest of the rock. This tends to confirm the suggestion of quite extensive regional retrogression.

TABLE 1. COMPARISON OF GARNET AND BIOTITE COMPOSITIONS BETWEEN INCLUSIONS IN FELDSPAR AND MATRIX GRAINS

(Total Fe is expressed as FeO. Abbreviation: b.d., below detection. Analysed by energy-dispersive microprobe at Cambridge University.)

garnet analyses

| | JG 111 | | JG 113 | | JG 119 | |
|---|---|---|---|---|---|---|
| sample | matrix | incl. | matrix | incl. | matrix | incl. |
| $SiO_2$ | 38.20 | 37.76 | 37.65 | 37.51 | 38.25 | 38.65 |
| $Al_2O_3$ | 21.48 | 21.23 | 21.27 | 21.15 | 21.36 | 21.75 |
| FeO | 31.57 | 30.81 | 32.92 | 32.18 | 30.68 | 30.18 |
| MgO | 1.14 | 1.15 | 1.84 | 1.88 | 0.85 | 1.06 |
| MnO | 0.66 | 1.54 | 0.68 | 0.56 | 1.35 | 2.01 |
| CaO | 8.72 | 7.81 | 6.31 | 6.64 | 8.97 | 9.09 |
| total | 101.77 | 100.29 | 100.65 | 99.93 | 101.45 | 102.74 |
| no. of atoms to 12(O) | | | | | | |
| Si | 3.005 | 3.014 | 3.002 | 3.006 | 3.021 | 3.011 |
| Al | 1.992 | 1.997 | 1.999 | 1.998 | 1.989 | 1.997 |
| Fe | 2.077 | 2.058 | 2.195 | 2.157 | 2.027 | 1.967 |
| Mg | 0.134 | 0.136 | 0.218 | 0.225 | 0.100 | 0.123 |
| Mn | 0.057 | 0.104 | 0.046 | 0.038 | 0.090 | 0.133 |
| Ca | 0.735 | 0.667 | 0.539 | 0.570 | 0.759 | 0.759 |

biotite analyses

| | JG 111 | | JG 113 | | JG 119 | |
|---|---|---|---|---|---|---|
| sample | matrix | incl. | matrix | incl. | matrix | incl. |
| $SiO_2$ | 35.87 | 37.74 | 35.87 | 36.62 | 36.97 | 38.29 |
| $TiO_2$ | 1.97 | 2.40 | 1.67 | 1.65 | 1.82 | 2.03 |
| $Al_2O_3$ | 17.07 | 16.71 | 17.54 | 17.72 | 17.95 | 18.29 |
| FeO | 20.51 | 20.27 | 18.02 | 18.39 | 19.96 | 19.62 |
| MgO | 9.20 | 8.90 | 10.40 | 10.08 | 9.52 | 9.56 |
| MnO | 0.07 | 0.08 | b.d. | b.d. | 0.03 | b.d. |
| $K_2O$ | 8.17 | 8.33 | 7.85 | 8.55 | 8.71 | 8.74 |
| total | 92.86 | 94.43 | 91.35 | 93.01 | 94.78 | 96.52 |
| no. of atoms to 22(O) | | | | | | |
| Si | 5.584 | 5.756 | 5.597 | 5.634 | 5.597 | 5.679 |
| Ti | 0.241 | 0.275 | 0.196 | 0.192 | 0.209 | 0.226 |
| Al | 3.125 | 2.998 | 3.227 | 3.215 | 3.230 | 3.197 |
| Fe | 2.675 | 2.585 | 2.351 | 2.367 | 2.541 | 2.433 |
| Mg | 2.126 | 2.019 | 2.417 | 2.312 | 2.159 | 2.119 |
| Mn | 0.009 | 0.011 | 0.0 | 0.0 | 0.004 | 0.0 |
| K | 1.627 | 1.619 | 1.561 | 1.679 | 1.690 | 1.654 |

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In summary, the metamorphic history of the Barrovian metamorphic rocks of NW Mayo is comparable to that of many other medium-pressure metamorphic terrains, in that the peak temperatures were attained after the major phases of deformation were complete but progressive heating during the deformation history is attested to by the textures of garnet porphyroblasts. Plagioclase porphyroblasts in the southern half of the inlier may reflect an MP1–MS2 metamorphic peak, or be related to retrograde MP2–MS3 recrystallization. Most rocks do not retain any trace of the earliest stages of their metamorphism, but an exception is provided by the epidote pods and bands exposed south of Ashleam Bay in the lowest grade part of the area, on Achill Island (figure 2) (Gray & Yardley 1979).

*Early metamorphic phases in south Achill*

In a single small inlet at Claggan (Irish National Grid Reference L689 957), several elongate pods of epidosite occur with metavolcanic chlorite schists and grey-to-black phyllites. The epidote layers are typically 2–3 m in length and less than 1 m across; in their cores they approach pure epidote with only minor quartz, albite, haematite and amphibole, while at their margins they have a similar assemblage to that of the adjacent metavolcanic schists, i.e. epidote + albite + actinolite + chlorite + quartz + haematite. In the intervening portions of the pods, however, Gray & Yardley (1979) reported the occurrence of crossite, occurring with epidote, albite, quartz, haematite, actinolite and sphene, and also chlorite and stipnomelane in some instances. This remains the only known occurrence of blueschist (*sensu lato*) in the Dalradian Supergroup.

A typical crossite epidosite is illustrated in figure 3*a*, plate 1, and the compositions of the sodic amphiboles from six such specimens are plotted in figure 4, which demonstrates that they range from crossite to glaucophane in composition. Not uncommonly, sodic amphiboles are altered at their margins to, or may be completely replaced by, a green calcic amphibole that usually grows as bundles of acicular crystals associated with some albite. Crossite grains entirely enclosed by epidote are most resistant to alteration. This type of transition from sodic to calcic amphibole has been widely reported from circum-Pacific high-pressure metamorphic belts (see, for example, Ernst *et al.* 1970), although contradictory trends are also known. The compositional variation of the amphiboles from the Achill blueschists is illustrated in figure 5, with the trend observed in successive zones indicated by the arrows. Representative amphibole analyses are listed in table 2. In some samples the crossites show a strong linear alignment indicating that they may have grown, or been stable during D1 deformation.

Some crossite-bearing epidosites contain distinctive areas of albite crowded with aligned platelets of haematite and near end-member actinolite (figure 3*b*,*c*, plate 1). The texture suggests pseudomorphic replacement of a pre-existing phase, but no relicts of the parent mineral are ever seen. Crossite is an unlikely parent because elsewhere in the same samples it shows alteration to actinolite with only minor albite and haematite (figure 3*b*). Holland & Ray

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DESCRIPTION OF PLATE 1

FIGURE 3. Photomicrographs of epidote–crossite schists, Achill Island. (*a*) Fresh crossite (cr) with epidote (ep) and minor sphene and haematite, sample JG308. (*b*) Enigmatic pseudomorph texture (P) with oriented haematite and actinolite within albite, set in epidote; actinolite after crossite is present at (a); sample BY1537. (*c*) Back-scattered electron image of haematite (white) and actinolite (grey) in albite (dark grey) forming part of a pseudomorph texture, sample BY1537.

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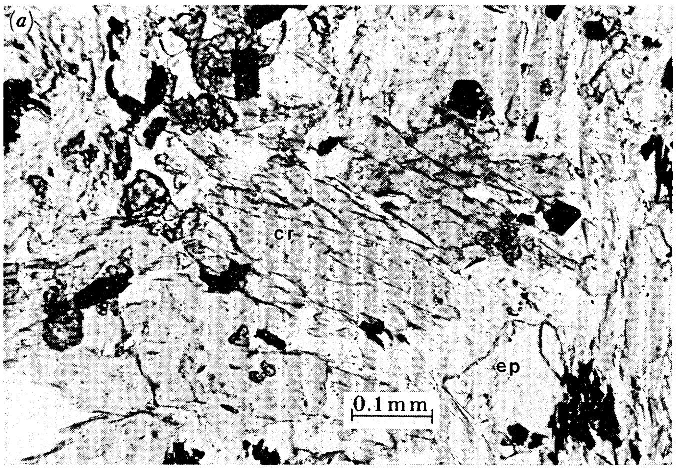


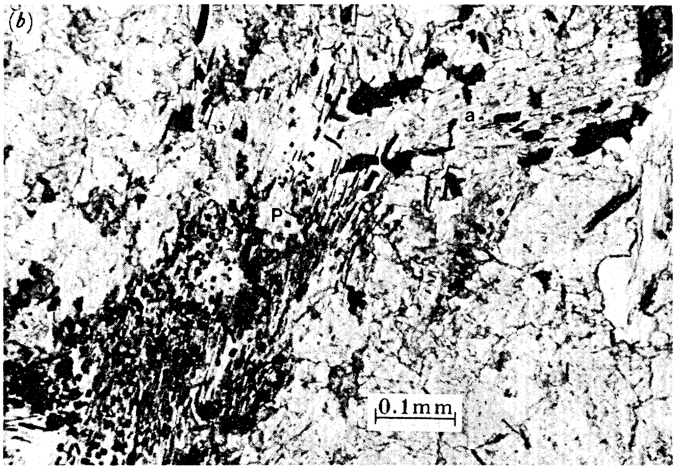


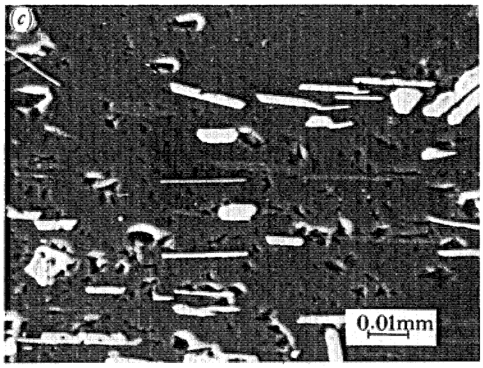


FIGURE 3. For description see opposite.

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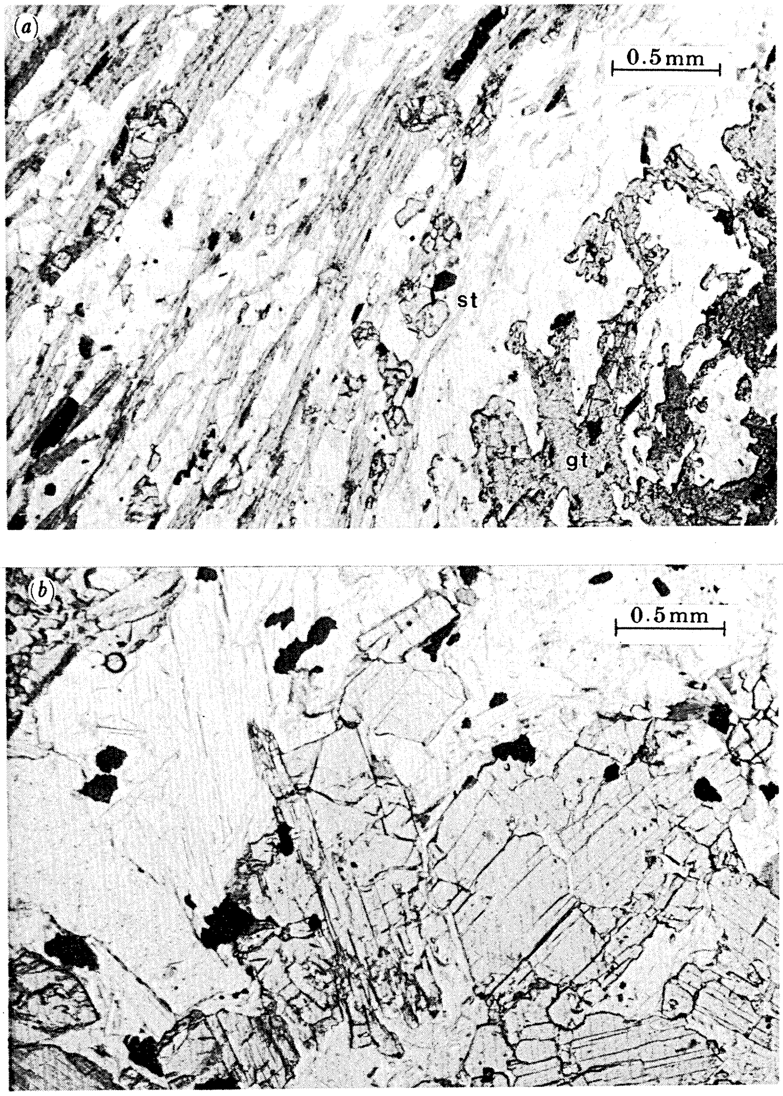


FIGURE 12. For description see opposite.

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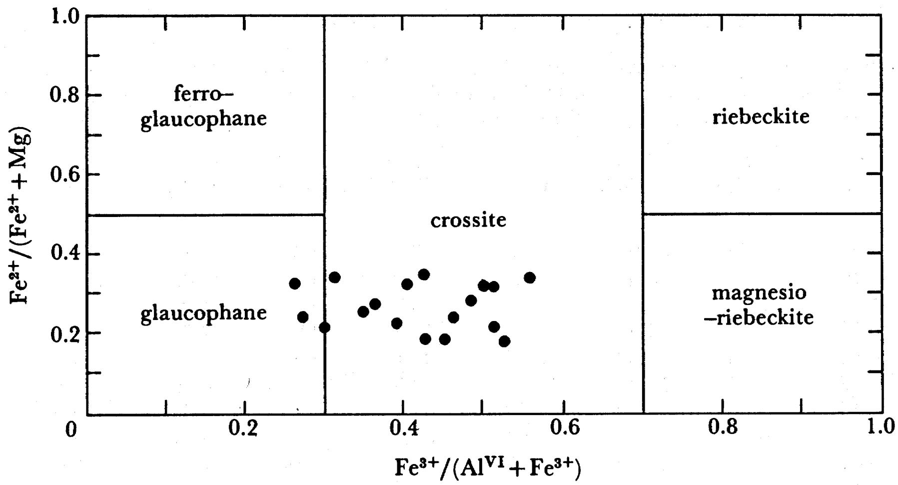


FIGURE 4. Plot of the compositional range of Na-amphiboles from 6 samples from the Achill Island locality. Analyses have been recalculated by the 'maximum $Fe^{3+}$' method of Stout (1972).

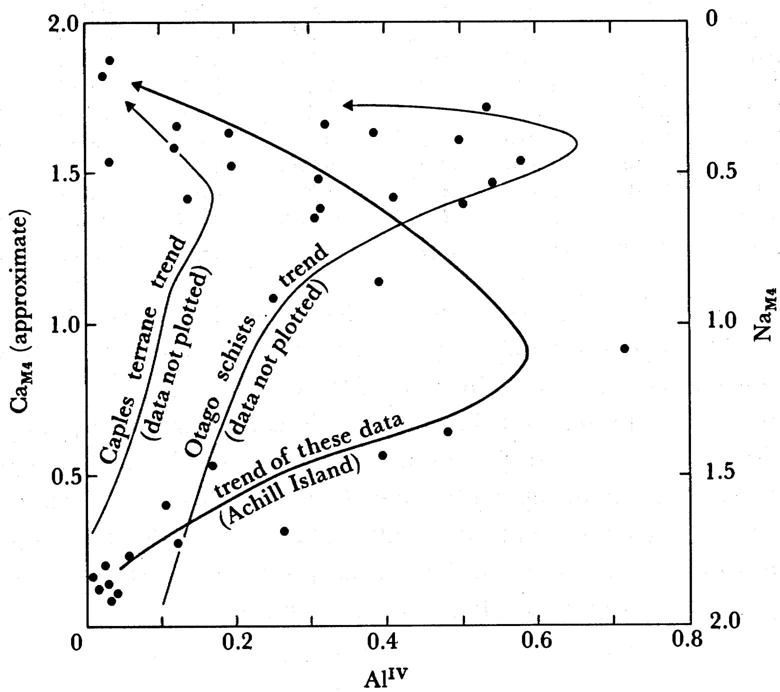


FIGURE 5. Plot of the full range of amphibole compositions from crossite–epidote schists, after Brown (1977). Many Na-amphibole points overlap. Ca-amphibole recalculated after Spear (1981). The general trend of variation in amphibole composition with time, e.g. in zoned grains, is indicated by the heavy line. For comparison, two trends obtained similarly from overprinted blueschists in New Zealand (Yardley 1982) are shown by light lines.

## DESCRIPTION OF PLATE 2

FIGURE 12. Photomicrographs of staurolite schists. (*a*) Sample BY146, Connemara (Yardley *et al.* 1980). Note fine grained staurolite (st) adjacent to a large (several millimetres) garnet. (*b*) Sample 75-1205, Ox Mountains (Yardley *et al.* 1979). Coarse composite grains of staurolite and kyanite occur in a muscovite-rich matrix. Large garnets (several millimetres) also occur in the rock.

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Table 2. Representative amphibole analyses from crossite–epidote schists

(Analysis 1, inclusion in albite pseudomorph texture (figure 2*c*); 2–6, green amphiboles after crossite; 7–9, sodic amphiboles. $Fe_2O_3$ calculated by the 'maximum $Fe^{3+}$' method of Stout (1972) for Na-amphiboles, and by setting A-site occupancy = 0.3 $Al^{IV}$ (Spear (1981) for Ca-amphiboles.)

| analysis... | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 |
|---|---|---|---|---|---|---|---|---|---|
| sample... | BY1537 | BY1539 | BY1539 | JG309 | JG18 | JG18 | JG18 | BY1539 | JG306 |
| $SiO_2$ | 54.81 | 54.69 | 52.00 | 54.78 | 50.26 | 52.91 | 57.43 | 57.05 | 56.90 |
| $TiO_2$ | 0.10 | b.d. | b.d. | b.d. | 0.19 | 0.12 | b.d. | b.d. | b.d. |
| $Al_2O_3$ | 1.60 | 2.77 | 4.58 | 5.27 | 7.59 | 6.92 | 4.91 | 6.42 | 8.45 |
| $Fe_2O_3$ | 0.0 | 2.50 | 4.48 | 0.0 | 7.84 | 8.94 | 7.69 | 7.04 | 4.78 |
| FeO | 10.63 | 10.35 | 10.57 | 9.98 | 9.91 | 8.95 | 9.20 | 7.76 | 8.98 |
| MgO | 15.62 | 15.13 | 13.25 | 14.17 | 10.83 | 10.64 | 10.39 | 10.79 | 9.98 |
| MnO | 0.16 | b.d. | 0.16 | 0.13 | b.d. | 0.19 | b.d. | b.d. | b.d. |
| CaO | 11.99 | 9.80 | 8.63 | 9.24 | 5.84 | 4.18 | 0.81 | 0.78 | 0.56 |
| $Na_2O$ | 0.62 | 1.81 | 2.50 | 3.00 | 4.94 | 5.48 | 7.05 | 7.04 | 7.37 |
| $K_2O$ | 0.10 | b.d. | b.d. | 0.07 | 0.17 | 0.11 | b.d. | b.d. | b.d. |
| total | 95.67 | 97.05 | 96.17 | 96.62 | 97.47 | 98.44 | 97.48 | 96.88 | 97.02 |
| no. of atoms to 23(O) (anhydrous basis) | | | | | | | | | |
| Si | 7.955 | 7.834 | 7.591 | 7.812 | 7.826 | 7.526 | 8.098 | 8.019 | 7.967 |
| Ti | 0.011 | 0.0 | 0.0 | 0.0 | 0.020 | 0.013 | 0.0 | 0.0 | 0.0 |
| Al | 0.273 | 0.467 | 0.788 | 0.885 | 1.297 | 1.161 | 0.816 | 1.064 | 1.395 |
| $Fe^3$ | 0.0 | 0.269 | 0.492 | 0.0 | 0.855 | 0.957 | 0.816 | 0.744 | 0.504 |
| $Fe^2$ | 1.290 | 1.240 | 1.291 | 1.191 | 1.202 | 1.065 | 1.085 | 0.912 | 1.052 |
| Mg | 3.380 | 3.230 | 2.882 | 3.011 | 2.339 | 2.256 | 2.184 | 2.260 | 2.082 |
| Mn | 0.019 | 0.0 | 0.019 | 0.016 | 0.0 | 0.023 | 0.0 | 0.0 | 0.0 |
| Ca | 1.865 | 1.505 | 1.350 | 1.411 | 0.908 | 0.637 | 0.122 | 0.117 | 0.083 |
| Na | 0.176 | 0.504 | 0.708 | 0.828 | 1.389 | 1.511 | 1.927 | 1.918 | 2.001 |
| K | 0.003 | 0.0 | 0.0 | 0.013 | 0.031 | 0.019 | 0.0 | 0.0 | 0.0 |
| total | 14.972 | 15.049 | 15.121 | 15.167 | 15.327 | 15.168 | 15.048 | 15.034 | 15.084 |

(1986) have, however, illustrated very similar pseudomorph textures that contain relicts of jadeitic pyroxene. We therefore tentatively interpret the pseudomorph textures as indicating the former presence of sodic pyroxene in some of the epidosites.

One epidosite sample preserves evidence of still earlier metamorphism. In this rock, granoblastic epidote grains generally contain submicroscopic inclusions that give rise to a clouded appearance, but are locally completely clear. The clear areas define a ghost pattern of plagioclase lathes in a typical ophitic basalt texture. From this it is concluded that the epidosites result from metasomatic alteration of pre-existing basalt before the onset of deformation and probably during sea-floor metamorphism.

## Metamorphism in the Ox Mountains inlier

The Ox Mountains inlier is almost continuous with the NW Mayo inlier (figure 1), and provides a link to the southern part of the Donegal inlier. Here, controversy surrounds the age of the rocks and their metamorphism, and whether they are in fact a Dalradian sequence or of greater antiquity. Although Phillips *et al.* (1975) and Andrews *et al.* (1978) have claimed to have found evidence that these rocks are pre-Caledonian, we favour the arguments advanced by Long & Yardley (1979) for a Dalradian age, especially in the light of the new dates of Long *et al.* (1984), and treat the main part of the Ox Mountains inlier as a part of the Irish Dalradian, with a narrow area of basement granulites occurring to the north east.

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A metamorphic map for this inlier has been provided by Yardley *et al.* (1979) and further details are given by Andrews *et al.* (1978). The regional metamorphic grade in the western part of the inlier is comparable to that in the adjacent part of the NW Mayo inlier, i.e. garnet zone with local development of chloritoid schists. Eastward across a major fault, however, the grade is higher with staurolite and staurolite–kyanite zone assemblages in pelitic schists. At the northeastern end of the inlier, chloritoid schists reappear. In part, the lower-grade rocks here are retrograde after original amphibolites facies assemblages, but some primary garnet zone rocks may also be present. One of the most interesting features of the Ox Mountains inlier is that it parallels, and lies close to, the Highland boundary line (Max & Riddihough 1975), although the metamorphic grade is much higher than is found close to the Highland boundary in Scotland. Close to this major lineament occurrences of low-grade upper Dalradian rocks may be let down on lesser faults along this line, furthermore intense cataclasis with associated retrogression is developed near the SE margin of the inlier in many places, and may also result from pervasive deformation close to the Highland boundary line.

The metamorphic history is comparable to that of NW Mayo, in that the peak of metamorphism is MP2 and garnets in the highest-grade rocks document earlier progressive metamorphism. Retrograde metamorphism in this area was accompanied by the development of a new S3 spaced fabric in the more intensely retrogressed rocks.

## The pressure–temperature–time history of metamorphism in NW Mayo and the Ox Mountains

### *Barrovian metamorphism*

#### 1. *Temperature estimates*

The temperatures of metamorphism in NW Mayo have been calculated primarily with three geothermometers. Data for the Ox Mountains have been presented by Yardley *et al.* (1979) and are comparable to those reported here for similar grades in NW Mayo. The three types of geothermometer used in this study were the garnet–biotite cation exchange thermometer, the calcite–dolomite solvus thermometer and oxygen isotope thermometers involving fractionation between quartz and magnetite, muscovite or feldspar. Results are plotted on figure 2.

A large number of calibrations have been proposed for the garnet–biotite geothermometer over the past ten years, most of which fortuitously give similar results for amphibolite facies rocks, although there are significant discrepancies at higher and lower grades. The calibration we have favoured in general is that of Ferry & Spear (1978), since it is derived directly from experimental results, but its use is limited by deviations from ideality when the end-member compositions used in the experiments mix with other components in natural minerals. Although various attempts have been made to compensate for variation in mineral compositions when applying this thermometer, such corrections must be wholly or partly empirical and probably therefore valid only for the range of temperatures and compositions for which they are calibrated. We have used the calibration of Goldman & Albee (1977) for the lowest-grade garnet-bearing rocks, and the Ferry & Spear calibration where it yields results above 450 °C, since the two calibrations converge.

Three samples of calcite–dolomite marble have been analysed. Sample JG73 (Achill Island) yields only a small spread of calcite compositions with 8 point analyses indicating a temperature near 495 °C, with single points suggesting temperatures about 50 K above and below the rest

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respectively. Samples JG232 and JG233 from near the staurolite–kyanite isograd yield a much larger spread of calcite compositions with 24 points giving a continuous range of temperatures from 280 to 550 °C. Three points suggest temperatures from 600 to 640 °C and so the true peak metamorphic temperature is inferred to have been 595±45 °C. The calibration of Rice (1977) was used throughout.

The validity of applying oxygen isotope thermometry to regionally metamorphosed rocks is currently controversial. Various studies in the 1970s appeared to obtain consistent results for a number of mineral pairs, and the systematic variation of fractionation factors with metamorphic grade was clearly established (O'Neil & Ghent 1975; Kerrich *et al.* 1977). However, the application of recent experimental determinations of a number of critical fractionation factors for metamorphic rocks (Matsuhisa *et al.* 1979) makes it clear that the semiempirical fractionation factors used in many previous studies actually yield temperatures that are very much higher than those obtained from experimental calibrations (Clayton 1981). The discrepancy between experimental and empirical fractionation factors is particularly acute for quartz–feldspar and quartz–muscovite mineral pairs, which include most of the data obtained in this study. Except in the unlikely event that the experimental work is unaccountably wrong, the data obtained here are apparently meaningless, except as indicators of the blocking temperature for stable isotope diffusion. In this case however, unlike most other blocking temperatures, those for oxygen diffusion evidently depend very closely on the maximum temperature to which the rock was subjected, in view of the systematic variations found in this and previous studies. Hence it could be argued that the semiempirical fractionation factors do have a *de facto* validity as geothermometers, at least up to the lower amphibolite facies. A possible explanation (Gray 1981) is that the blocking of oxygen diffusion is linked to absorption of pore water due to incipient retrograde reaction as the rocks begin to cool.

For this reason, results obtained in this study have been recalculated using the fractionation factors of Javoy (1977) and are summarized in table 3. All yield results comparable to those obtained by other geothermometers applied to the same or nearby samples, except for JG149 and JG358, which lie in close proximity to the late Caledonian Termon granite.

TABLE 3. SUMMARY STABLE ISOTOPE ANALYSES FROM NW MAYO SAMPLES

(Abbreviations: q, quartz; mt, magnetite; ms, muscovite; fs, feldspar.)

| sample | $\delta^{18}O_q$ | $\delta^{18}O_{mt}$ | $\delta^{18}O_{ms}$ | $\delta^{18}O_{fs}$ | $T$/°C q–mt | $T$/°C q–ms | $T$/°C q–fs |
|---|---|---|---|---|---|---|---|
| JG 56 | 13.24 | 4.13 | 10.29 | 11.16 | 510 | 515 | 410 |
| JG 106 | 13.39 | — | 8.25 | 10.13 | — | 345 | 270 |
| JG 149 | 12.72 | — | 6.48 | 9.78 | — | 295 | 300 |
| JG 235 | 15.06 | — | 13.06 | — | — | 650 | — |
| JG 358 | 13.20 | — | 7.16 | 10.00 | — | 300 | 275 |
| JG 382 | 11.49 | — | 8.60 | 9.71 | — | 520 | 465 |
| JG 448 | 15.77 | — | 12.28 | 14.00 | — | 460 | 465 |
| JG 408 | 10.09 | — | 6.49 | — | — | 450 | — |

Combining results from all the geothermometers, (figure 2) it is clear that the garnet zone in NW Mayo spans a wide range of temperatures from 400 to 600 °C. The staurolite–kyanite zone may range up to about 650 °C. While individual determinations are subject to large uncertainty, and the calibration of the individual thermometers may be no better than ±50 °C, we nevertheless believe that the pattern of results reflect qualitatively the structure

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of temperature variation within the garnet zone. For example, there is some evidence for an increase in grade eastwards across the Corraun fault, which Crow *et al.* (1971) argued is a major regional lineament.

The wide range of temperatures within the garnet zone must, in part, reflect the fact that pelites are rare in the region. This is because the abundant semi-pelites are unlikely to develop new index minerals on heating beyond garnet zone conditions. It may well be that the higher grade parts of the garnet zone should be assigned to the staurolite zone and this has been done in the area of Bunnaveela Lough, on the basis of vein occurrences of staurolite with kyanite, but relatively low (less than 600 °C) geothermometer temperatures. The very low temperatures obtained from the low-grade parts of the garnet zone are in striking contrast to the figure of 510 °C obtained for the garnet isograd in SW Scotland by Graham *et al.* (1983). There is, however, no evidence to suggest that temperatures in NW Mayo have been reset during cooling, since independent geothermometers give similar results. Allowing for uncertainties in the geothermometer calibrations, our results suggest that the garnet isograd in Mayo lies at around 400–450 °C, which is reasonably consistent with the temperature of 400 °C obtained by Black (1974) for the garnet isograd in high-pressure rocks from New Caledonia.

2. *Pressure estimates*

Although some attempts have been made to calculate pressures for low- to medium-grade semi-pelites and pelites of the type that make up the bulk of the NW Mayo inlier, we do not believe that any of the geobarometers that would be applicable to the rocks we have found are sufficiently well understood to merit their use in this study. Only for the staurolite–kyanite zone can pressure be determined with any reliability, and the approach we have used here is similar to that of Yardley *et al.* (1979), updated in the light of new experimental work. Pressures and temperatures for this zone were determined simultaneously from the intersection of univariant curves calculated for each of the following equilibria:

anorthite = grossular + kyanite + quartz (Newton & Hasleton 1981), (1)
Fe-biotite + Mg-garnet = Mg-biotite + Fe-garnet (Ferry & Spear 1978), (2)
paragonite + quartz = kyanite + albite + $H_2O$ (Chatterjee 1972; Eugster *et al.* 1972), (3)
Fe-staurolite + quartz = Fe-garnet + kyanite + $H_2O$ (Yardley 1981). (4)

A further constraint is provided by the equilibrium

garnet + rutile = ilmenite + kyanite + quartz (Bohlen *et al.* 1983), (5)

since staurolite–kyanite schists in this area typically contain rutile without ilmenite.

Data for one sample of staurolite–kyanite schist from NW Mayo are presented in table 4 and the results of the calculations are portrayed graphically in figure 6. Note that it has been assumed that $P_{H_2O} = P_{total}$, following examination of fluid inclusions, which are of mildly saline $H_2O$ in these schists. Similar results are obtained from Ox Mountains samples following revision of the earlier-published calculations (Yardley *et al.* 1979), and it is concluded that the conditions of staurolite–kyanite zone metamorphism were $8 \pm 2$ kbar and $620 \pm 30$ °C.

### *Conditions of early blueschist metamorphism*

No direct quantitative estimates can be made for the pressure and temperature of formation of the early crossite–epidote schist assemblages. The assemblages are, however, closely

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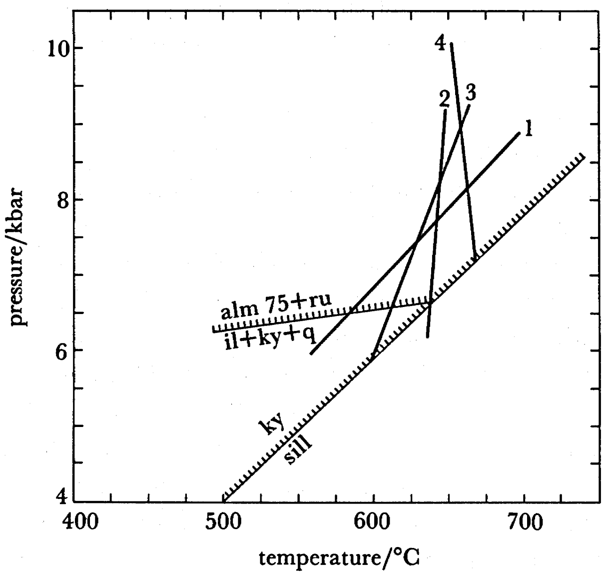


FIGURE 6. $P$–$T$ diagram illustrating the estimation of the conditions of equilibration of sample JG199 (staurolite–kyanite zone). Numbered curves are equilibrium curves calculated for each of the reactions 1–4 in the text, from the experimental equilibrium constant expressions given in the references therein, using the data summarized in table 4. Hachured lines provide lower $P$–$T$ limits to the assemblage.

TABLE 4. REPRESENTATIVE MINERAL ANALYSES FROM ST–KY SCHIST JG 199

| | gt | st | ky | plag | musc | bio |
|---|---|---|---|---|---|---|
| $SiO_2$ | 38.02 | 27.70 | 37.54 | 63.70 | 46.34 | 36.74 |
| $TiO_2$ | b.d. | 0.52 | b.d. | b.d. | 0.43 | 1.58 |
| $Al_2O_3$ | 21.65 | 53.85 | 63.37 | 22.47 | 36.65 | 19.08 |
| FeO | 34.42 | 12.13 | 0.14 | 0.05 | 0.93 | 18.90 |
| MgO | 4.05 | 1.52 | 0.04 | b.d. | 0.46 | 10.82 |
| MnO | 0.82 | 0.14 | b.d. | b.d. | b.d. | 0.03 |
| CaO | 1.67 | b.d. | b.d. | 3.30 | b.d. | b.d. |
| $K_2O$ | b.d. | b.d. | b.d. | 0.03 | 7.16 | 8.03 |
| $Na_2O$ | b.d. | b.d. | b.d. | 9.09 | 1.92 | b.d. |
| total | 100.63 | 95.86 | 101.09 | 98.64 | 93.87 | 95.17 |
| no. of atoms recalculated to no. of (O)... | 12 | 46 | 5 | 8 | 22 | 22 |
| Si | 3.009 | 7.579 | 1.002 | 2.838 | 6.141 | 5.509 |
| Ti | 0.0 | 0.109 | 0.0 | 0.0 | 0.043 | 0.178 |
| Al | 2.020 | 17.785 | 1.994 | 1.180 | 5.725 | 3.372 |
| Fe | 2.278 | 2.841 | 0.003 | 0.002 | 0.072 | 2.370 |
| Mg | 0.477 | 0.634 | 0.002 | 0.0 | 0.090 | 2.416 |
| Mn | 0.055 | 0.016 | 0.0 | 0.0 | 0.0 | 0.004 |
| Ca | 0.142 | 0.0 | 0.0 | 0.158 | 0.0 | 0.0 |
| K | 0.0 | 0.0 | 0.0 | 0.002 | 1.210 | 1.538 |
| Na | 0.0 | 0.0 | 0.0 | 0.785 | 0.494 | 0.0 |

comparable to those of large sectors of the Shuksan and Sanbagawa terranes (Ernst *et al.* 1970; Brown 1977) in that sodic amphiboles are restricted to specific, ferric iron-rich, lithologies, and other phases characteristic of high metamorphic pressures (e.g. lawsonite, jadeitic pyroxene) are rare or absent. Metamorphic pressures are difficult to constrain for such terranes. For relatively low temperatures (*ca.* 400 °C), jadeitic pyroxene should become relatively common at pressures in excess of 10 kbar and this probably represents an upper limit. Barovian

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metamorphism takes place close to a normal continental geothermal gradient (Richardson & Powell 1976) and this implies pressures up to 6 kbar at around 400 °C. In the absence of crossite as an equilibrium member of Barrovian metabasite assemblages this provides a minimum pressure for the blueschists. There being no more reliable criteria, the pressure of formation of the crossite–epidote assemblages is taken to be *ca.* 7 kbar, following Brown (1977). Similarly, temperatures can only be gauged by comparison with similar metamorphic terranes elsewhere, but a value of $400 \pm 50$ °C is probably reasonable.

Despite the limitations to the precision of the absolute *P–T* estimates, some interesting observations can be made about the changes in *P–T* conditions in going from blueschist metamorphism to the Barrovian assemblages. Figure 5 is a plot of amphibole compositions as $Na_{M4}$ against $Al^{IV}$ (see Brown 1977). The vertical axis is essentially a pressure indicator, with lower-pressure assemblages above higher-pressure ones. The horizontal axis represents temperature increasing with $Al^{IV}$, for the calcic amphiboles. All sodic amphiboles plot together on this figure, but there is some evidence that their compositions are also temperature dependent in natural metabasite assemblages. Yardley (1982) reported that Na-amphibole relicts in the lowest-grade, sub-greenschist facies part of the Otago Schist belt (Caples terrane) are of riebeckite, whereas crossite relicts occur in similar rocks now in the greenschist facies portion of the same belt. High-temperature blueschists, transitional to the eclogite facies, have glaucophane rather than crossite (see, Holland & Ray 1986). When the trend of changing amphibole composition with time is plotted on a diagram such as figure 5, different terranes with apparently identical transitions from Na-amphibole to Ca-amphibole assemblages reveal significantly different trends. The Achill Island samples (figure 5) progress from crossite to barroisite along a trend towards hornblende, that is comparable to that described by Holland & Richardson (1979) from overprinted blueschists of the Tauern window, Austria. In contrast, metabasite amphiboles from the Otago greenschists, New Zealand, progress directly from crossite through winchite to actinolite with moderate $Al^{IV}$ content and finally to near end-member actinolite. In the adjacent Caples terrane, riebeckite passes directly to actinolite with low $Al^{IV}$ (Yardley 1982). The two trends from New Zealand and the trend determined by Holland & Richardson probably reflect 'greenschist' overprinting of 'blueschist' at different temperatures. The Achill Island trend suggests that these blueschists were initially depressurized and/or heated at relatively high temperatures, but were then cooled during uplift so that their medium-pressure metamorphism took place at lower temperatures than those of the Otago greenschists, i.e. they were more strongly cooled during uplift.

Independent evidence to support this conclusion comes from oxygen isotope analyses of two samples from the blueschist locality (Gray 1981). A crossite-bearing sample gave $\delta^{18}O_{qutz} = +14.43$, $\delta^{18}O_{amp} = +11.27$, whereas an adjacent sample in which the amphibole is actinolite yielded $\delta^{18}O_{qutz} = +16.69$, $\delta^{18}O_{amp} = +12.09$ (all relative to SMOW). According to Garlick (1966), the large difference in quartz–amphibole fractionation factors between these samples cannot be explained by the difference in amphibole chemistry alone; it appears likely that the crossite–quartz pair equilibrated at higher temperatures than the actinolite–quartz pair, or that at least one pair is not equilibrated.

## *Summary and discussion*

The Barrovian metamorphism of the NW Mayo and Ox Mountains inliers was superimposed on an earlier, higher pressure, regional metamorphism. The first event may have produced

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blueschist or transitional blueschist–greenschist assemblages over a large area, although they are now preserved only where the subsequent overprint was minimal and in lithologies that were resistant to later deformation. High-pressure metamorphism was probably contemporaneous with early major deformation, while the peak Barrovian metamorphism postdates most fabrics. Metamorphic conditions during the high-pressure event were in the region of 400 °C and 6–10 kbar, while the Barrovian metamorphism ranged from 350 °C in south Achill island to over 600 °C, with pressures of around 8 kbar at the highest grades. There is some evidence that, in the lowest-grade Barrovian zones, the overprinting of blueschist assemblages was accompanied by cooling rather than heating, and this might reflect proximity to a zone of uplift against cooler rocks near the Highland boundary line. *P–T*–time paths for the different zones in these inliers are shown in figure 7.

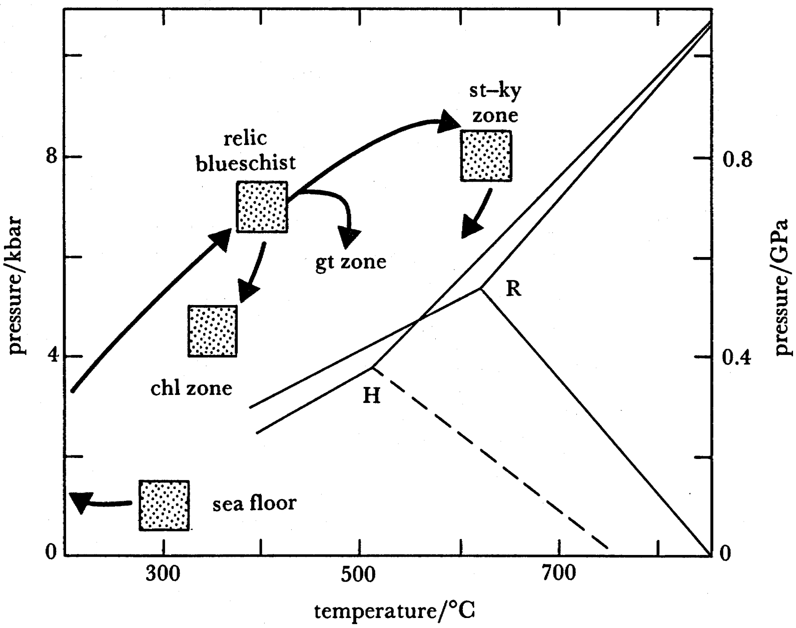


FIGURE 7. *P–T* paths illustrating the inferred metamorphic history of the Barrovian metamorphic zones of NW Mayo. The Al-silicate phase boundaries of Holdaway (1971) (H) and Richardson *et al.* (1969) (R) are shown for reference.

It should be noted that the pressures we have quoted here for low-grade Barrovian metamorphism are substantially lower than the values of 9–10 kbar quoted by Graham *et al.* (1983) for SW Scotland. This almost certainly reflects the reliance on different barometric indicators in the two areas. We consider that the complete absence of sodic amphiboles from the Scottish Dalradian is not consistent with the high pressures that have been reported, by comparison with pressure estimates from blueschist belts elsewhere, but in view of the undoubted uncertainties in geobarometry of these rocks the most important point to make is that the low-grade rocks of NW Mayo appear to be closely comparable in most respects to their Scottish equivalents.

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## Metamorphism of the Connemara massif

The Connemara inlier has been one of the most intensively studied parts of the Caledonian orogen, and detailed mapping has permitted the erection of a reliable stratigraphy and the unravelling of complex structures (Leake 1981; Tanner & Shackleton 1979). Details of the metamorphic history, metamorphic zoning and conditions of metamorphism have been published by Barber & Yardley (1985), Ferguson & Al Ameen (1986), Treloar (1982, 1985), Yardley (1976, 1986) and Yardley *et al.* (1980). This paper is concerned primarily with the interpretations and implications of the previously published results.

The predominant trend of bedding, fold axes and isograds in Connemara is east–west, so that the region can be considered as a series of zones seen in north–south traverse (figure 8). The central part of the inlier is occupied by complexly folded Dalradian metasediments which display an increase in grade southward into a migmatite belt. South of this is a belt of synorogenic intrusive gneisses ranging from basic to acid compositions. The migmatites are primarily developed in a structurally complex 'steep belt' in which there are repetitions of the stratigraphy along syn-metamorphic slides. Occasional later faults in this belt have let in blocks of lower grade rocks. Whereas the bulk of the Dalradian sediments form a continuous refolded series with the same formations reappearing at different grades, at least two successive units in the northern part of the inlier are in tectonic (slide) contact with the rocks to the south and do not have unequivocal correlatives in the main part of the succession, although it is unlikely that any great thickness is omitted (Cruse & Leake 1968; Leake *et al.* 1984). Major divisions and structure of the Connemara inlier are illustrated on figure 8, and in section on figure 13*b*.

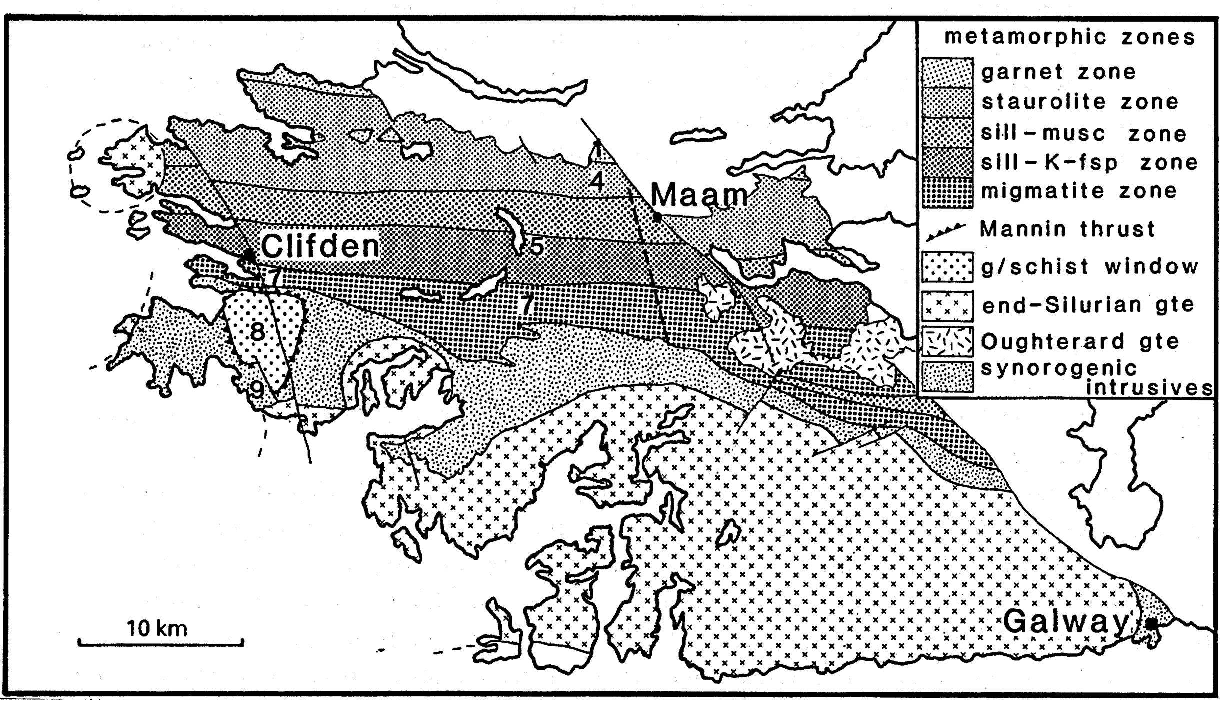


Figure 8. Metamorphic map of Connemara showing zones in the Dalradian rocks, intrusives and the Mannin Thrust after Leake (1981) and Barber & Yardley (1985). Numbered regions correlate with structural features illustrated in section in figure 13*b*; 1, 4, 5, 7, 9 are regions of varying orientation of F3 folds due to later folding; 7 is a 'steep belt' with near-vertical bedding structures; 8, low-grade rocks exposed in a window through the Mannin Thrust. Heavy line is that of the section in figure 9.

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The early metamorphic history of Connemara has close parallels to that of NW Mayo and the Ox Mountains. An early major phase of folding (D2) produced the dominant regional schistosity, but pre- to syn-D2 garnets are widespread and preserve an earlier fabric as inclusions. No relicts of any earlier (pre-garnet grade) metamorphism remain. Staurolite, and very rarely kyanite, grew after the D2 deformation and were inferred by Yardley *et al.* (1980) to have grown together. In contrast sillimanite, and more rarely andalusite, grew subsequently at the expense of staurolite. Near the present northern limit of sillimanite, breakdown of staurolite and growth of sillimanite was apparently initiated around the beginning of the major D3 folding event, but further south, in rocks of the migmatite zone, there is good evidence that sillimanite growth largely predated D3. In other words, relative to deformation, higher-grade rocks to the south were heated through any particular isotherm before lower-grade rocks further north. Whether deformation was synchronous throughout the area is, of course, unknown. Peak metamorphic temperatures were maintained during D3 folding throughout Connemara, except for the lowest grade rocks in the north, and gave rise to a series of isograds marked by the progressive appearance of sillimanite, sillimanite + K-feldspar and migmatites. The isograd pattern appears to be unaffected by D3 folding (figure 8) despite the best endeavours of workers in the 1970s to find evidence for deformation of the isograd surfaces (Badley 1976; Yardley 1976).

A particularly significant feature of the zonal sequence is the regional development, in east Connemara in particular, of andalusite. Andalusite occurs most commonly in veins in pelitic schists on both sides of the sillimanite isograd, but also forms very coarse porphyroblasts in the pelitic schists. Within the sillimanite zone, coarse vein andalusites show partial to complete replacement by sillimanite and the distinctiveness of such pseudomorphs where present led Yardley (1976) to recognize a distinct southern limit to the region in which andalusite has grown, the significance of which is developed below. Although Yardley (1976) suggested that there might have been two phases of andalusite growth in total, early MS3 and MP3, further work suggests that all occurrences are of broadly MS3 growth.

### *Temperatures and thermal causes of the metamorphism*

Temperatures of formation of the various metamorphic zones in the eastern half of Connemara have been determined by Yardley *et al.* (1980), Barber & Yardley (1985) and Treloar (1985). Most results are based on garnet–biotite cation exchange thermometry (Ferry & Spear 1978), but with supporting evidence from garnet–cordierite cation exchange thermometry, and equilibria (3) and (4) (above) modified according to the Al-silicate polymorph present (Yardley *et al.* 1980).

Staurolite zone temperatures are inferred to range from 520 to 560 °C, with the sillimanite isograd near 580 °C. These values may be underestimates because the garnet grew before the metamorphic peak and may not have re-equilibrated (Yardley 1977). At higher grades, new growth of garnet lends greater credibility to the calculations, which suggest temperatures near 650 °C for the second sillimanite isograd, and migmatite formation from 700 to at least 750 °C. Ferguson & Al Ameen (1986) claim higher temperatures for apparently comparable grades in western Connemara, but this is because they have used a different calibration of the garnet–biotite thermometer; distribution coefficients are the same as those found at the same grade further east.

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The temperatures determined in a north–south traverse across eastern Connemara (see Yardley (1986) for a detailed zonal map of the area) are plotted in the form of a temperature profile in figure 9. Such a profile will of course be influenced by post-metamorphic D4 folding, which produced an east–west trending arch (figure 13*b*) and has been shown on palaeomagnetic grounds to significantly postdate the onset of cooling (Morris & Tanner 1977). The effect of this event will be to steepen up the thermal gradient in the northern part of the traverse. When allowance is made for this effect it is clear that the temperature profile corresponds closely to that which would be expected to result from the emplacement of the voluminous calc-alkaline synorogenic intrusives now seen in southern Connemara into rocks that were already at amphibolite facies temperatures, and is sublimely independent of the structural complexity of the region. Specifically, an east–west trending vertical sheet of intermediate magma, 2 km thick, at 1200 °C emplaced at the site of the synorogemic intrusives into country rocks initially at 500 °C would produce a zonal sequence closely comparable to that found today in Connemara, although of course the actual history of magmatism is much more complex. It should be noted that the 'flattening-out' of the temperature profile at the highest grades occurs because the contact between migmatites and intrusive gneisses is not a simple one; rather they inter-finger over a zone. Much higher temperatures (greater than 850 °C) can be obtained from xenoliths and from the immediate hornfelses of the basic intrusions (Treloar 1981). This model of essentially regional scale contact metamorphism for the zones in Connemara accords well with the textural evidence suggesting that over most of the area the metamorphic grade and crystallization history was actually very similar until after the D2 deformation.

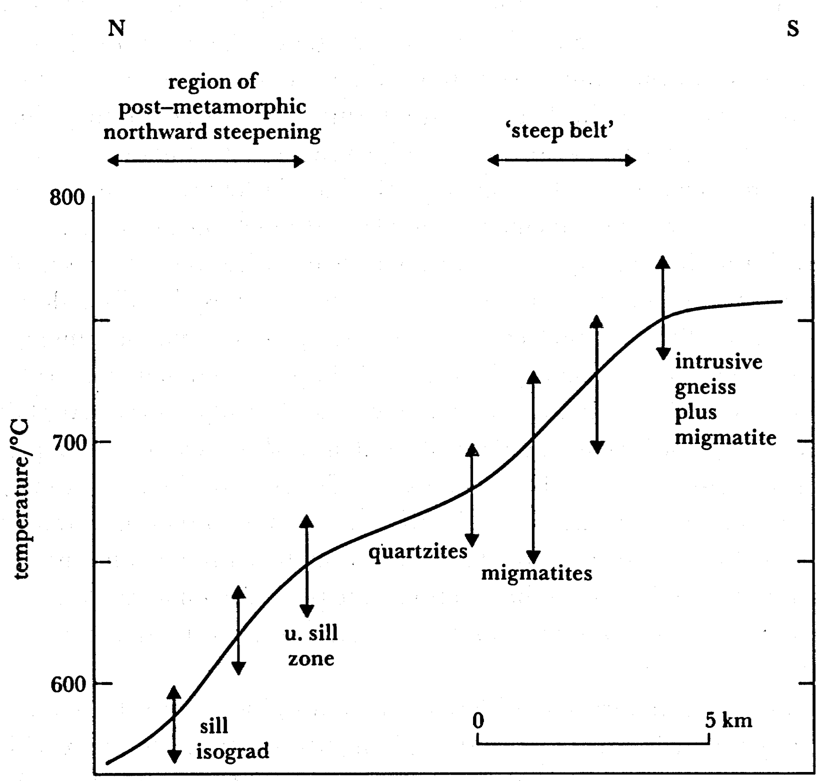


FIGURE 9. Temperature profile along a N–S section in E Connemara (figure 8). Vertical bars indicate uncertainty in the temperature estimates.

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*Pressures of metamorphism*

There are two complementary approaches to understanding the pressures of the metamorphism in Connemara. Quantitative pressure estimates can be made using pressure-sensitive equilibria, notably (1) (above), corrected for the $Al_2SiO_5$ polymorph present. These provide absolute estimates subject to large uncertainties. On the other hand, textural evidence is present that suggests the occurrence of reactions involving changes in pressure, even though these may not be of sufficient magnitude to be unequivocally demonstrated by numerical calculations.

We have recalculated metamorphic pressures from equilibrium (1), applying the more recent calibration of Newton & Hasleton (1981), with the correction reported by Ganguly & Saxena (1984), to the data of Yardley *et al.* (1980). Upper sillimanite-zone pelites yield pressures close to 5.5 kbar for $T = 650$ °C from those samples with extensive peak-metamorphic growth of garnet. A similar pressure (5.8 kbar) is obtained at lower grades in the staurolite–sillimanite zone from sample BY111, which was figured by Yardley (1977) as providing an example of re-equilibration of early formed garnet during subsequent heating. Some other lower-grade (staurolite-bearing) samples yield unrealistically high pressures outside the sillimanite stability field. Very similar pressure estimates of 5–5.5 kbar were made by Barber & Yardley (1985) and Treloar (1985) from higher-grade rocks corresponding to the southern part of the profile (figure 9). These results tend to confirm the thermal model proposed above, i.e. one of essentially isobaric heating due to magmatism.

Previous studies have suggested that there was regional uplift and erosion between MP2 growth of staurolite and kyanite, and the subsequent thermal maximum (Yardley 1976). Relatively high pressures in the range 7–10 kbar were proposed by Yardley *et al.* (1980) for the staurolite–kyanite event, on the basis of equilibrium (1). Recalculating the data after Newton & Hasleton (1981) yields pressures around 8 kbar, in agreement with the findings of Ferguson & Al Ameen (1986) using the same equilibrium. At first sight, therefore, it appears that staurolite–kyanite growth in Connemara took place at much the same pressures as in NW Mayo and the Ox Mountains. There is, however, one clear indication that this is not the case; staurolite–kyanite schists in Connemara contain ilmenite, whereas the Ti-oxide phase in the northerly inliers is rutile. Since garnet compositions are similar, and the ilmenite does not contain appreciable haematite, it follows from equilibrium (5) (Bohlen *et al.* 1983) that the metamorphic pressures in Connemara were in fact significantly lower. Applying typical Connemara mineral compositions to the data of Bohlen *et al.* yields an upper pressure limit for the kyanite–garnet–ilmenite rocks of close to 6 kbar. The reason why the garnet–plagioclase geobarometer yields incorrect results is probably that at the rather low temperatures of formation of these rocks the garnet and plagioclase that had grown initially at still lower temperatures, failed to re-equilibrate at the metamorphic peak. A similar explanation probably accounts for the high pressures obtained for many samples close to the sillimanite isograd further east.

Pressure–temperature estimates for different metamorphic zones in Connemara are illustrated in figure 10, and it can be seen that if the Al-silicate triple point of Richardson *et al.* (1969) is accepted it could be argued that the sequence of Al-silicate phases developed in Connemara, i.e. kyanite followed by andalusite, then sillimanite, could be achieved by isobaric heating with none of the uplift proposed previously. There are, however, a series of arguments

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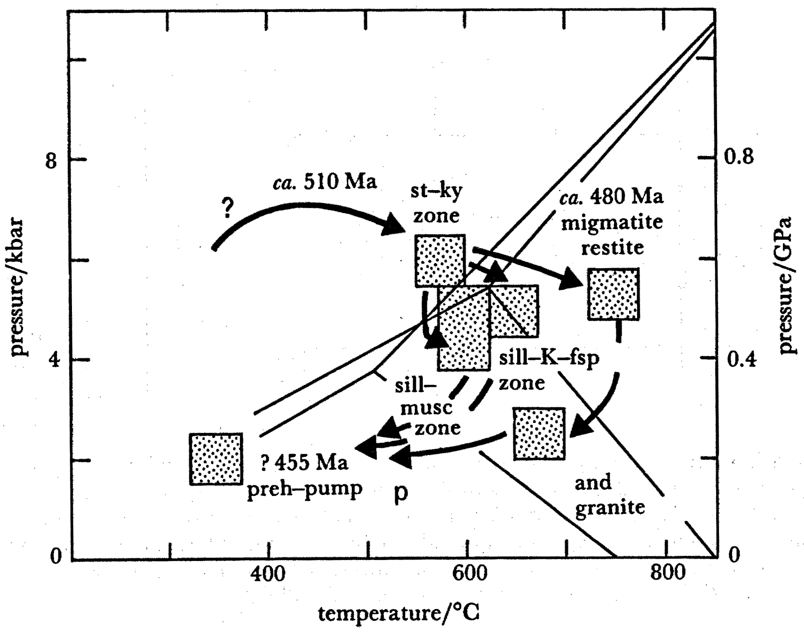


FIGURE 10. $P$–$T$ paths illustrating the inferred metamorphic history of the metamorphic zones of Connemara. Note the different heating paths for diferent zones. Ages are 'best-estimates' for the dating of different stages, discussed in the text.

to suggest that uplift did occur during the high-grade metamorphism, although perhaps less than has been supposed hitherto.

Firstly, Barber & Yardley (1985) pointed out that the occurrence of apparently magmatic andalusite in some migmatite leucosomes, and of uniformly low-density $CO_2$ fluid inclusions in others, indicates that crystallization of the anatectic melts occurred at pressures of around 2.5 kbar even though restite assemblages suggested generation at about 5.5 kbar.

Secondly, there is clear textural evidence that two pressure-sensitive equilibria have been crossed in a down-pressure sense. These are as follows,

$$\text{garnet} + \text{muscovite} = \text{biotite} + \text{sillimanite} + \text{quartz}, \tag{6}$$

$$\text{garnet} + \text{sillimanite} + \text{quartz} + H_2O = \text{cordierite}. \tag{7}$$

Reaction (6), proceeding at the same time as staurolite breakdown, produced widespread replacement of garnet by sillimanite. Reaction (7) resulted in the growth of cordierite enclosing garnet and sillimanite in the migmatite zone (Barber & Yardley 1985, figure 2).

Finally, the restricted distribution of regional andalusite is not compatible with simple isobaric heating. In a section across eastern Connemara, sillimanite is the only $Al_2SiO_5$ phase to have formed in the southern part of the section. Further north, andalusite grew first but was replaced by sillimanite, while northwards again andalusite occurs with no sillimanite. This section is modelled in figure 11*a*, which treats the regional cross section (modelled as a rigid rectangular block) as a fixed reference through which isotherms migrate northward with time while isobars move downward due to progressive uplift and erosion. Figure 11*b* shows the implications of this model for the sequence of dehydration reactions and $Al_2SiO_5$ phases formed at three points. It can be seen that the predicted reaction sequences at points A and B correspond to what is deduced for the evolution of the upper sillimanite zone and sillimanite–

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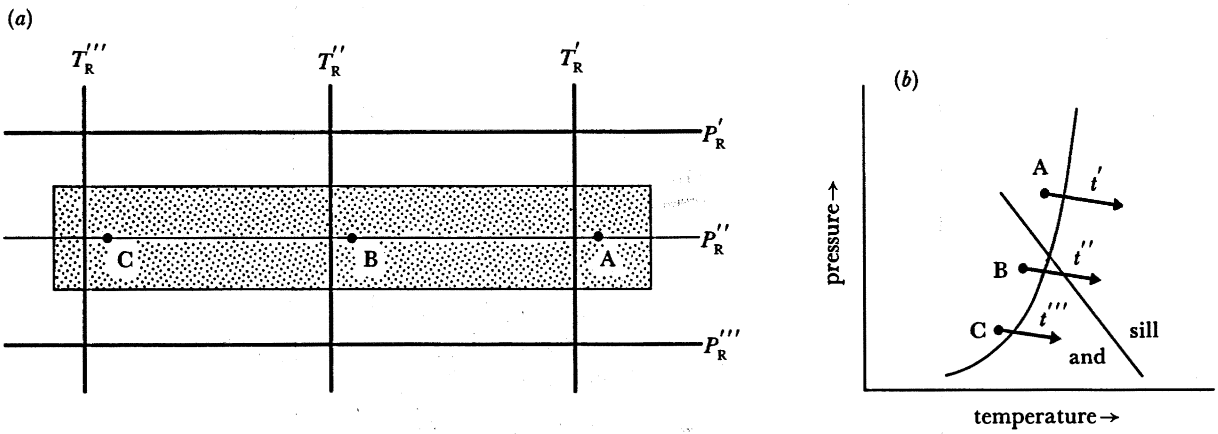


FIGURE 11. Schematic representation of the variation in $P$ and $T$ with time inferred for Connemara. (*a*) Block representing a N–S section with reference points A, B, C. An 'isotherm' approximately vertical, corresponding to the temperature for Al-silicate production is shown at three successive times $t'$ to $t'''$. The location of a reference isobar, apparently moving down through the block as it is uplifted, is also shown for the same times. (*b*) $P$–$T$ diagram showing the intersection of the andalusite–sillimanite boundary with the Al-silicate-forming reaction and the $P$–$T$ paths inferred for each of points A, B and C.

muscovite zones respectively (figure 10). Point C (figure 11*b*) corresponds to staurolite zone rocks with andalusite, found rarely in east Connemara. The additional complexity of differential movements within the reference frame has not been considered. One further observation that this model also does not attempt to explain is that regional low-pressure assemblages, i.e. cordierite and andalusite, are abundant only in eastern Connemara. Perhaps the simplest explanation of this is that the later phases of the synorogenic magma suite, i.e. the granitic K-feldspar gneiss, and perhaps also the Oughterard Granite (Senior 1973), were emplaced primarily in the east, so that this part of the region was heated later, after further uplift, than areas to the west.

*Summary: the P–T–t history of the Connemara Dalradian*

The thermal evolution of different metamorphic zones within the Connemara inlier is summarized on figure 10, together with tentative absolute ages for the different phases of the metamorphism.

The initial metamorphism of Barrovian garnet-zone type is approximately the same age as the intrusion of the Cashel basic intrusion (*ca.* syn-D2), which has been dated at 510 Ma (Pidgeon 1969). This age is similar to Grampian metamorphic ages from Scotland (Dempster 1985). After this event different parts of the area followed different uplift and heating paths to produce the present zonal sequence. The precise age of this stage is not well known but the age of 480 Ma suggested here is based on a series of 2 point Rb–Sr isochrons for adjacent migmatite restite and leucosome pairs (J. P. Barber & A. N. Halliday, in preparation).

A date for the end of the high-grade metamorphism is provided by the results of Leake *et al.* (1983, 1984) on the age of the Mannin thrust (figure 8). High-grade rocks above this major thrust exposed in SW Connemara, underwent retrogression and did not substantially heat the underlying low-grade acid volcanics. Lineated hornblendes from the thrust zone yield K–Ar ages of 454 Ma (Leake *et al.* 1984), which is similar to the Rb–Sr isochron age of 460 ± 25 Ma obtained by Leake *et al.* (1983) from the rocks below the thrust. Retrograde effects are

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widespread, although minor, throughout Connemara and while some may date from the time of the Mannin Thrust, others may be much later.

One interesting consequence of the $P$–$T$–$t$ paths shown in figure 10 is that they appear to indicate that the higher-grade rocks of the migmatite belt were derived from greater depths than lower-grade rocks further north. The relation of this conclusion to tectonic models is explored below.

## Comparisons between Connemara and NW Mayo

Despite the distinct differences in the final metamorphic assemblages, it is clear that there are a number of similarities in the earlier metamorphic histories of Connemara and the more northerly inliers. For example, peak temperature conditions (except possibly in south Achill) were obtained after the main D2 fabric-forming deformation, and in the higher-grade zones garnet growth was, in part, pre- to syn-D2. What is not clear is whether the early high-pressure event that is recorded from south Achill also affected the Connemara inlier (or indeed the rest of NW Mayo). Unfortunately there appears little prospect of answering this question because of the intensity of the subsequent overprint, but we consider it possible that the whole of the Irish Dalradian was initially metamorphosed under relatively high-pressure–low-temperature conditions before the Dalradian thermal maximum.

It has been suggested previously (Yardley 1980) that the distinctive thermal evolution of Connemara dates only from the time of the onset of the D3 deformation; however, the new results presented here make it clear that the conditions of the MP2 staurolite–kyanite metamorphism in Connemara were distinct from those for the similar event further north. This implies that the thermal evolution of Connemara began to diverge at an earlier stage, which is in accord with the syn-D2 onset of basic magmatism in Connemara (Leake 1969).

A less tangible distinction between the Dalradian of Connemara and Mayo, although one that is readily apparent in the field, is that the grain size of pelitic schists of comparable grade is finer in samples from Connemara. In eastern Connemara in particular, staurolite grains seldom exceed 0.5 mm and are often much smaller than this, even in rocks with a high modal abundance of staurolite. Only garnet can be considered to truly form porphyroblasts (figure 12*a*, plate 2). Westward along strike there is a distinct tendency for the grain size to coarsen, and millimetre-scale staurolites become quite common. The only occurrences of relatively coarse-grained schists in the east are in the tectonically distinct uppermost and northernmost unit, the Benlevy formation (Tanner & Shackleton 1979). In contrast, even low-grade rocks from NW Mayo are often relatively coarse grained and in the staurolite–kyanite zone matrix grains may exceed 1 mm. Staurolite–kyanite schists from the Ox Mountains include still coarser bands with centimetre-sized garnet, staurolite and kyanite (figure 12*b*, plate 2).

The finer grain size of the Connemara rocks is tentatively attributed to a more rapid rate of heating, because the textures imply that the amount of diffusive mass transfer was less than for rocks metamorphosed at the same grade elsewhere.

Hence the balance of the evidence suggests that the Connemara inlier followed a similar structural and metamorphic history to the northerly inliers up to about the end of the D2 deformation, when they began to diverge. The D3 deformation in Connemara has no parallel in the other inliers. Despite this, by the end of the Silurian, Connemara was once again behaving in the same manner as the other Dalradian inliers.

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## Tectonic settings of Irish Dalradian metamorphism

The metamorphic histories developed here involve three distinct phases, which probably correspond to different tectonic settings. In this instance there appears to have been a rapid transition between each phase of the metamorphism.

In the first phase, Dalradian sediments were buried in a low heat flow setting below at least 20 km of cover. This can most readily be envisaged as the result of underthrusting below an over-riding slab, because if such great pressures were to be attained below a purely sedimentary cover, the setting is unlikely to have been one of low heat flow. The early, major structures developed in this phase.

The second phase involved thermal relaxation to produce Barrovian assemblages, which have been shown by Richardson & Powell (1976) to correspond to a normal crust thermal gradient. This metamorphic phase persisted beyond the last major episode of penetrative deformation in most of the inliers. The only locality where phase-one assemblages have survived the phase-two overprint lies close to the Highland boundary line. Their preservation requires cooling during uplift rather than thermal relaxation, and this could be accounted for if the Highland boundary line was already active. Although there is no independent evidence to suggest early movement on the Highland boundary line as such, large-scale differential vertical movements during the Ordovician have been documented from the Scottish Dalradian by Dempster (1985).

In the third phase, high-temperature–low-pressure metamorphism is superimposed on the earlier phases. This phase was restricted to Connemara where it was accompanied by further folding. Yardley & Senior (1982) suggest that this phase developed in the roots of a volcanic arc.

Recently Leake *et al.* (1983, 1984) have suggested that, for the latter part of its metamorphic history, Connemara lay above a deep thrust system on which it was directed southward while rising due to ramp climb. The Mannin thrust represents the surface expression of the upper part of the thrust system where temperatures were relatively low. Deeper down the initial uplift took place at amphibolite facies temperatures. This model is illustrated schematically on a north–south cross section in figure 13. The bottom part of the figure indicates the proposed deep thrust and ramp system, and above this the dominant D3 fold structures and dominant tectonic slides are indicated schematically. This interprets the steep belt as a region of compression above the ramp, and indeed the northward-directed D3 fold nappes could similarly be a response to the ramp climb.

We believe that our data provide further support for the model that Leake & co-workers have proposed. In particular, the ramp-climb model predicts greater uplift for the highest grade rocks of the migmatite belt (figure 13*c*) and this accords with our petrological deductions.

One problem is why the deep derived synorogenic magmatism should not be more markedly displaced through time if the Connemara rocks were moving over a deep magma source on a thrust system. The most likely explanations are that either the amount of movement on the thrusts was only small, or it largely took place after the end of the magmatic activity. There is, in fact, some evidence for displacement of the magmatic centres of south Connemara. The most extensive developments of basic bodies are in the south and west (Leake 1970) while more granitic gneisses are chiefly developed to the east. If the Oughterard Granite is related to the other synorogenic intrusives as suggested by Senior (1973), then its present position to the north

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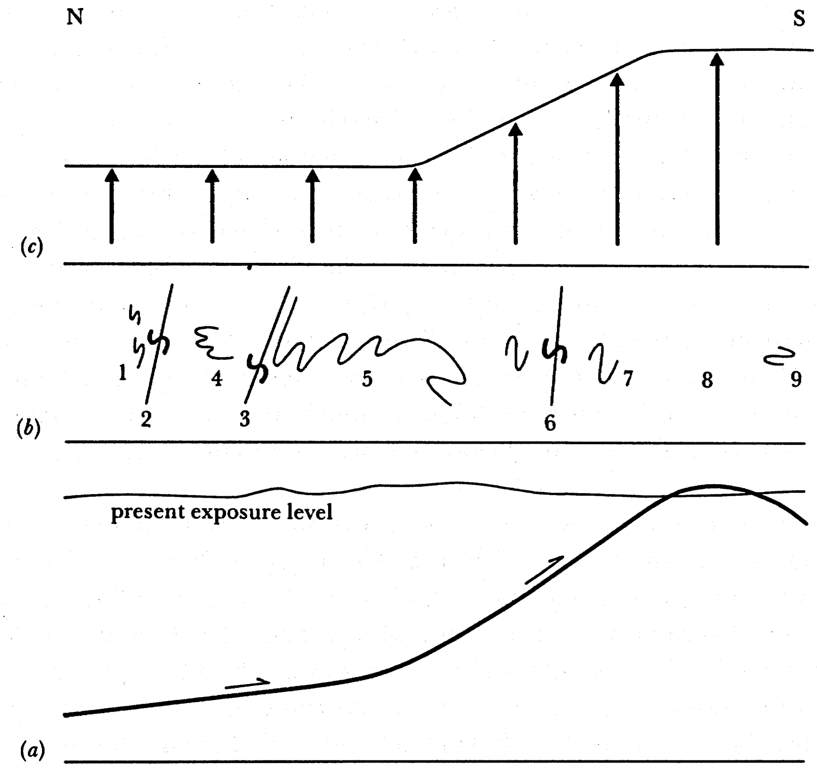


FIGURE 13. Schematic N–S section of Connemara to illustrate (*a*) underlying thrust system proposed by Leake *et al.* (1983) with (*b*) the corresponding late-stage fold structures. Folds sketched are D3 folds whose attitudes change due in part to D4. Numbering in (*b*) corresponds to figure 8 with additionally 2, 3 and 6 being tectonic slides of approximately D3 age; (*c*) shows the differential uplift of the southern part of the section inferred petrologically and implied by the thrust model.

of the intrusive gneisses is consistent with southward thrusting during the igneous activity, although only on a scale of a few kilometres.

A final point made by Leake *et al.* (1984) which merits comment is that in their thrust model the uppermost Dalradian rocks of north Connemara, occurring in tectonically distinct slices, need not be rooted in the steep belt to the south, as figured by Yardley (1976), but could have been emplaced on parallel, southward-directed thrust-ramp systems. In the case of the Benlevy formation rocks, such an interpretation would accord better with their rather coarser textures.

## General conclusions

In addition to the regional results presented here, two rather general conclusions can also be drawn. The first point is that, despite the long and complex metamorphic history of some of the rocks described, all the metamorphism predates 460 Ma and therefore is appreciably older than the closure of the Iapetus Ocean at the end of the Silurian. Hence we are dealing with metamorphic events taking place at a continental margin before final closure of the ocean. The only metamorphic effect that can be related to the final collision in this part of the orogen is the formation of small aureoles around end-tectonic granite plutons.

Secondly, we propose that the most suitable way of attempting to understand the thermal evolution of terranes with complex metamorphic histories, such as Connemara, is to treat their

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development in terms of a series of 'modules' corresponding to the metamorphic phases outlined above. The precise sequence in which the modules are superimposed on one another, and the time intervals between events, may vary very considerably between different metamorphic belts, but each module has a distinctive signature related to a particular tectonic setting of metamorphism. For example, the thermal evolution of the NW Mayo rocks (figure 7) can be interpreted in terms of a first module corresponding to burial in a low heat-flow setting followed by a second module corresponding to thermal relaxation and uplift. For Connemara, a distinctive module corresponding to heating and uplift in association with arc magmatic activity has been further superimposed and has obliterated any evidence for module 1. It so happens that in Connemara all the metamorphic events follow one another continuously but there is no *prima facie* reason why this should be the case. Many active arc volcanoes today are built on basements of high- or medium-pressure rocks, which were last metamorphosed some considerable time before the onset of recent arc volcanism.

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## *Discussion*

W. E. G. Taylor (*Luton College of Higher Education, Park Square, Luton, U.K.*). I congratulate Dr Yardley on a very fair summary of the problems and our state of knowledge regarding the regional metamorphism within the Caledonides of western Ireland. Recent work by Dr Neil Crane and myself in the Ox Mountains has indicated that subsequent to the emplacement of the Slieve Gamph igneous complex (dated at about 500–450 Ma) there was widespread retrogression of the biotites within the complex followed by the growth of relatively small spessartine garnets and/or epidotes. This latter crystallization event was initially observed in

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the xenoliths and enclaves within the complex and was thought to be a local phenomenon (Taylor 1966). Completion of the investigation of the complex (Crane 1985) indicates that the event is more widespread. This metamorphic event appears to pre-date the late-stage cataclastic and retrogressive event within the country rocks.

In view of this, how confident is Dr Yardley that he is correlating metamorphic events of the same tectono–thermal cycle within the Caledonides of western Ireland and particularly between NW Mayo and the Ox Mountains?

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B. W. D. Yardley. We thank Dr Taylor for his kind comments on our work. Of course he is correct to point out that the correlation of the rocks of the Ox Mountains with the Dalradian is not universally accepted. The Slieve Gamph igneous complex almost certainly post-dates the peak Barrovian metamorphism of the metasediments around it, but it is now clear that the complex itself is made up of two phases of plutonism: a main phase at *ca.* 480 Ma and a later phase at *ca.* 400 Ma (Long *et al.* 1984). It is the main phase that is affected by retrogression and by shearing. If the age of peak Barrovian metamorphism in the Dalradian is *ca.* 500–510 Ma, then emplacement of a late-metamorphic granite at 480 Ma is not inconsistent with its envelope being composed of Dalradian rocks. Thus the subsequent recrystallization that Dr Taylor describes must in any case post-date the main metamorphism of the Dalradian, and therefore cannot be taken to be equivalent to that event.

This controversy will only finally be resolved by careful isotopic investigation of the metasediments of the Ox Mountains and of undoubted Dalradian rocks. At present, however, we find the similarities in metamorphism and lithology between the two groups of rocks to be the most compelling evidence available.

R. Mason (*Department of Geological Sciences, University College London, U.K.*). Could the differences in textures of staurolite and other minerals, in rocks from the northern and southern parts of the Connemara metamorphic terrain, which the author describes as an 'unsolved problem', be due to differences in the composition of the fluids present during metamorphism? Burton (1986) has shown that textures and growth rates of garnet porphyroblasts, and of their matrices, are profoundly influenced by different degrees of dilution of $H_2O$ by $CO_2$ in the metamorphic fluid. Dilution by $CH_4$ is also possible, and the concentration of hydrogen ions may vary. The $H_2O:CO_2$ ratio changes over distances of the order of 0.1 mm in the rock, and there are examples where $H_2O:CO_2$ ratio can be shown to have changed during the period of garnet growth. How much is known of the variations in metamorphic fluid composition, in both space and time, in Connemara?

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Burton, K. W. 1986 Garnet–quartz intergrowths in graphitic pelites: the role of the fluid phase. *Mineralog. Mag.* (In the press.)

B. W. D. Yardley. The textural differences between staurolite schists in Connemara and elsewhere in the Dalradian are primarily differences in grain size rather than in the nature of the inclusion and intergrowth patterns, such as distinguish porphyroblasts in graphitic and

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non-graphitic rocks. Fluid inclusion studies in both NW Mayo (Gray 1981) and Connemara (Yardley *et al.* 1983) indicate that, although other fluids were present, the metamorphic fluid in the pelites during prograde metamorphism was mildly saline $H_2O$. It should be noted that graphitic schists are rare at high grades and the pelites whose textures are compared here are not graphitic.

In so far as the textures in the Connemara rocks formed with less extensive mass transfer than those at comparable grades further north, we believe that they probably reflect more rapid growth due to more rapid heating.

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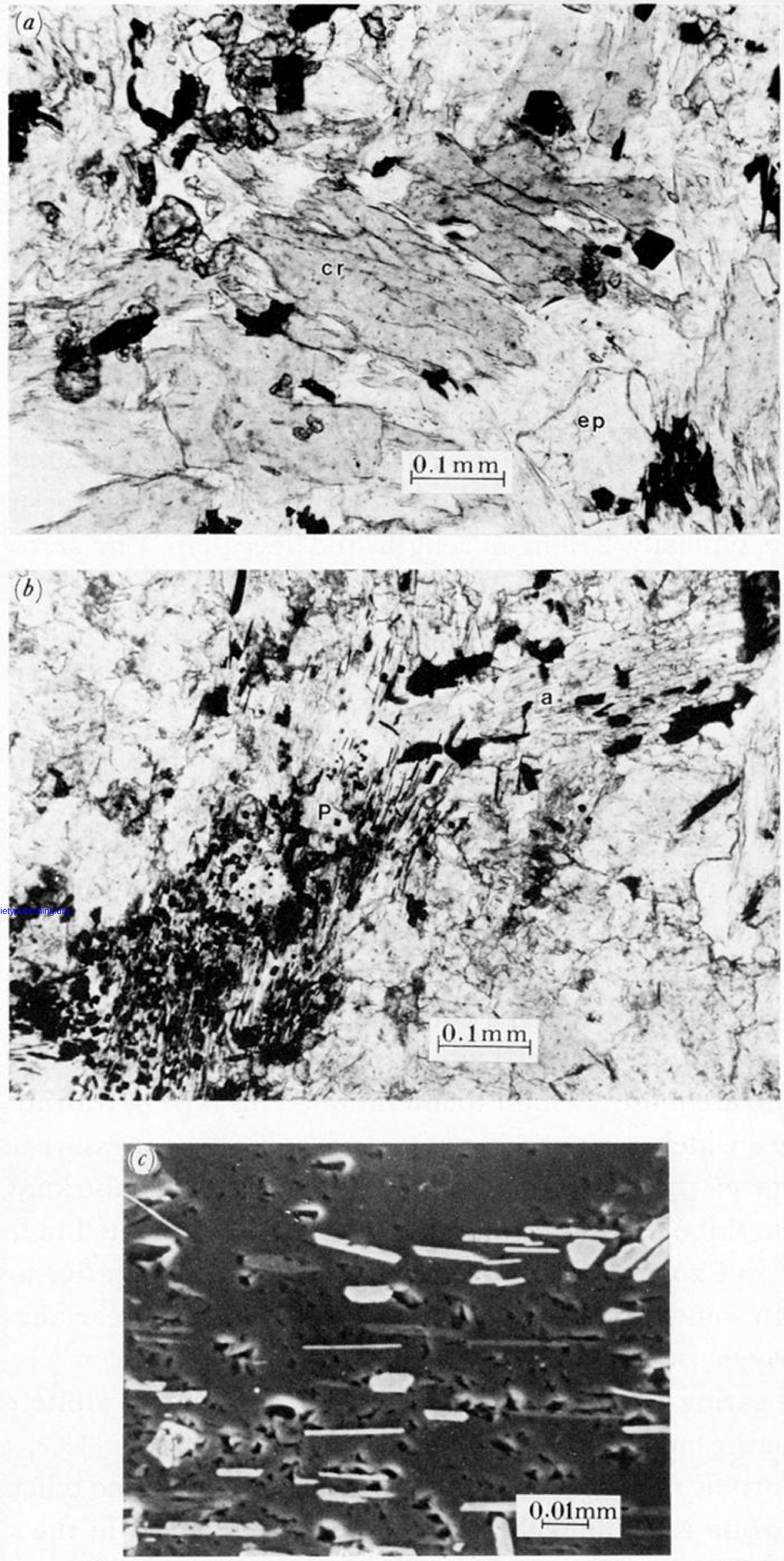


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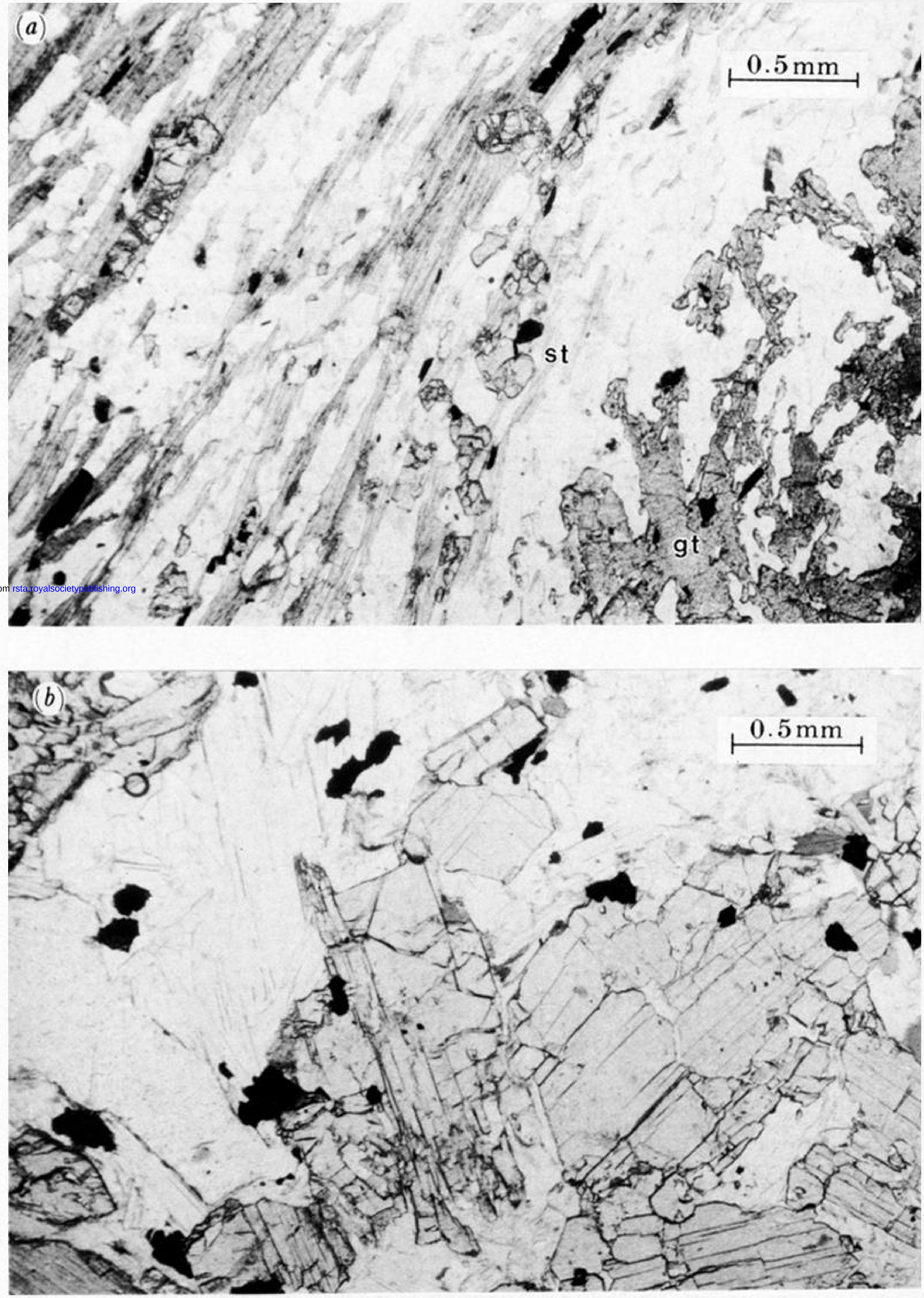
Figure 3. Photomicrographs of epidote–crossite schists, Achill Island. (*a*) Fresh crossite (cr) with epidote (ep) and minor sphene and haematite, sample JG308. (*b*) Enigmatic pseudomorph texture (P) with oriented haematite and actinolite within albite, set in epidote; actinolite after crossite is present at (a); sample BY1537. (*c*) Back-scattered electron image of haematite (white) and actinolite (grey) in albite (dark grey) forming part of a pseudomorph texture, sample BY1537.

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IGURE 12. Photomicrographs of staurolite schists. (*a*) Sample BY146, Connemara (Yardley *et al.* 1980). Note fine grained staurolite (st) adjacent to a large (several millimetres) garnet. (*b*) Sample 75-1205, Ox Mountains (Yardley *et al.* 1979). Coarse composite grains of staurolite and kyanite occur in a muscovite-rich matrix. Large garnets (several millimetres) also occur in the rock.