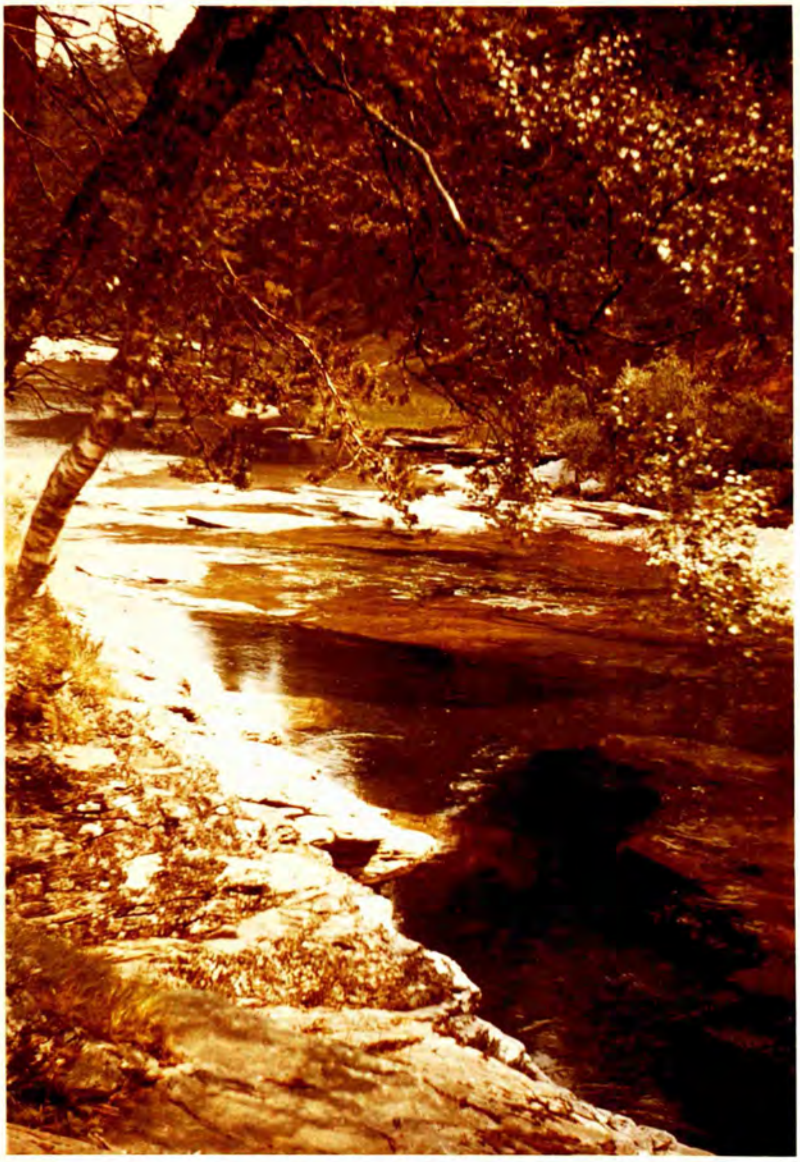

Plate 1. The Dee, below the Linn of Dee, Inverey. (Frontispiece)

THE STRUCTURAL, METAMORPHIC AND IGNEOUS HISTORY OF THE AREA AROUND BRAEMAR

THESIS PRESENTED FOR PH.D. DEGREE IN PETROLOGY,
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By

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ABSTRACT

The area is made up of Lower Dalradian metasedimentary rocks intruded by granite ("Younger Granite") and minor intermediate to basic igneous rocks, the order of intrusion being basic to acid.

Although the succession of the metasediments cannot be proved, enough evidence exists to show that these rocks were folded to form two primary recumbent nappe structures, partially bound by slides. The attitude of these structures is compatible with the general Caledonoid pattern and they are thought to belong to part of the hinge region of the Ben Lui Syncline.

Intense smaller-scale folds in the metasediments are the products mainly of two subsequent deformations, the earlier producing "cross-folds", the later producing folds on Caledonoid axes.

Both nappes, however, have a complicated history of deformation, this being borne out by petrofabric studies of the quartzite and granite intrusions of each nappe. The deformation fabric of the granite is often similar to that of the adjacent quartzite despite the fact that the granite appears in the field to post-date most of the folding. The fabric of both rock-types corroborates other evidence that the general structure of the upper nappe is simpler than that of the lower.

Metamorphic changes in the pelitic rocks can to some extent be correlated with the main phases of deformation. The rocks were firstly regionally metamorphosed, and then (in the area around and to the north-east of Braemar) thermally altered. The resulting zone of foliated hornfelses corresponds with the limit of granite intrusion around the margins of the Lochnagar Granite Complex. The "thermal" minerals are themselves affected by even later deformation - probably contemporaneous with the straining of the granite.

ACKNOWLEDGEMENTS

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The writer is greatly indebted to his supervisor, Professor B.C. King, without whose help and encouragement this thesis could not have been attempted in its present form. Other members of staff and research students in the Department of Geology, Bedford College, are thanked for their help, advice and discussion which was constantly available.

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Lastly the writer wishes to thank his typist, Mrs. Mary Humphries, whose competence and care are self-evident.

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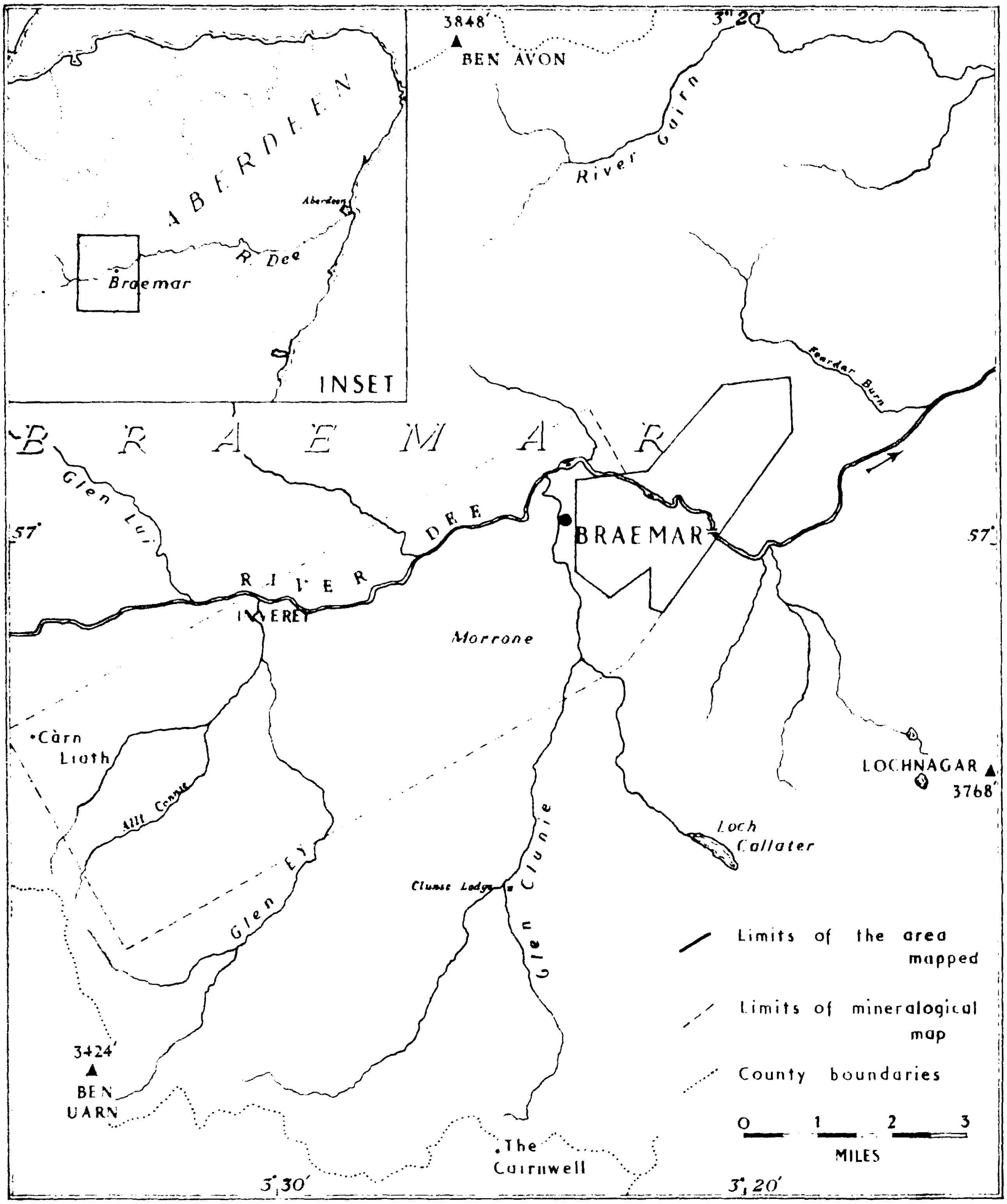


Fig. 1. The location and limits of the area, showing (inset) its position in Aberdeenshire.

PART I

GENERAL INTRODUCTION AND PHYSICAL FEATURES

General Introduction

The area, comprising some $3\frac{1}{2}$ square miles of well-exposed geology, is situated in the Scottish Highlands, immediately east and northeast of Braemar, Aberdeenshire (see fig. 1). The rocks therein belong mainly to the lower part of the Dalradian Series, but the area also includes a portion of the Lochnagar Granite (sheet 65, one-inch-to-the-mile, Geological Survey of Scotland).

The Dalradian rocks comprise metasediments of the Blair Atholl, Perthshire Quartzite and Ben Eagach series. There is also a possible occurrence of Moine rocks (upon which the others rest). Granite is the most abundant and widespread of the igneous rocks, there being also diorites, "felsites" and porphyritic micrograniies. Meta-igneous rocks are represented by rarer epidioritic types which grade to microdiorites.

The present investigation attempts to establish a relationship between the structural evolution, the metamorphism and the igneous history of the area. The main basis of research is detailed field-mapping, with laboratory studies of the mineral assemblages, microstructures and fabrics within rock samples.

Away from valleys rock exposure is excellent, allowing considerable detail to be appreciated, though many actual boundaries are all too rarely available for inspection. Mapping has been carried out with the aid of 6-inch-to-the-mile aerial photographs, the final maps being based on the 6-inch-to-the-mile Ordnance Survey sheets of the area. Two such final maps (nos. 1 and 2) cover the northern and southern halves of the area respectively at a scale of 15 inches to the mile, while part of map 2, mapped at 36 inches to the mile, is presented separately (no. 3).

Plate 2. Glen Clunie from Creag Choinnich, showing its U-shaped profile, with moraines in Corrie Feragie (foreground).

Plate 3. Creag Choinnich from Càrn nan Sgliat, showing ice-formed crags (left). Braemar lies in the valley immediately behind the hill.

Plate 4. Creag Choinnich from Craggan Rour, with the upper reaches of the Dee beyond.

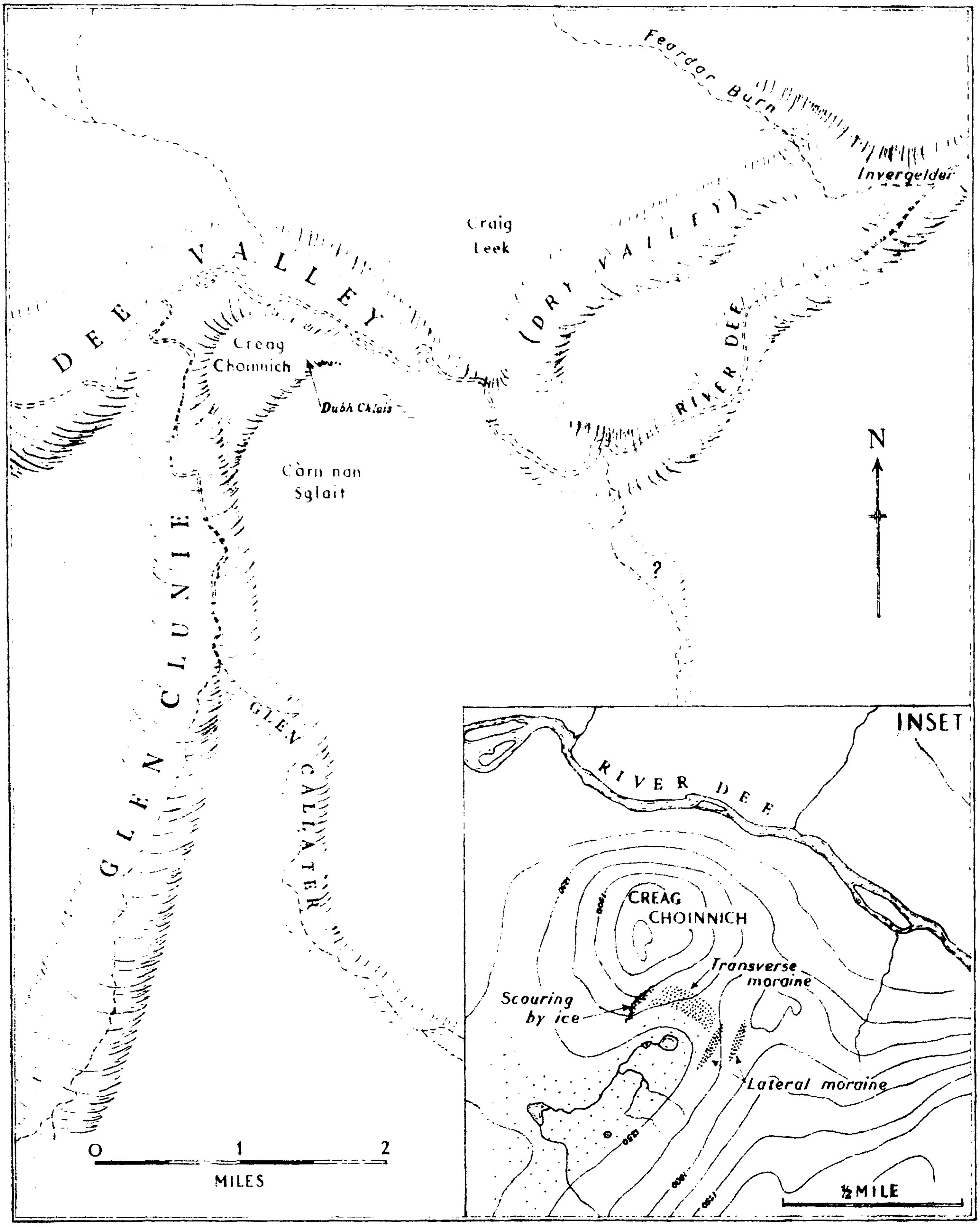


Fig. 2. The probable path of ice shown by U-shaped valleys.
Inset: glacial features around Creag Choinnich.

The history of previous research in and around the area will be dealt with separately for each part of the thesis.

Physical Features

Braemar lies on the southern edge of the highly dissected plateau of the Cairngorms. The township itself, situated at the confluence of the Dee and the Clunie, is 1,100 feet above O.D., while the principal hills of the area, Creag Choinnich, Càrn nan Sgliat and Craig Leek, reach heights of 1,764, 2,260 and 2,085 feet respectively.

The River Dee, flowing eastwards, cuts the area in two. The Clunie Water, flowing northwards to join the Dee at Braemar, forms the western limit of the southern half, while the northern half of the area is bounded on the eastern margin by an abandoned loop of the Dee.

The upper reaches of the Dee are heavily glaciated, while ice-formed deposits occur below Braemar. Glen Clunie is a distinctly U-shaped valley, and where it joins the Dee at Braemar there is further evidence of ice action to be seen in moraines of different types (pl. 2).

At one time ice must have swept down the Clunie to meet that in the Dee valley, where it was forced to pass round the northeast side of Creag Choinnich, via Dubh Chlais, thus isolating that hill from Càrn nan Sgliat (pl. 3,4). There is further evidence of this movement in the scoured southern face of Creag Choinnich, and in the lateral moraine formed on the opposing side of Càrn nan Sgliat (see fig. 2).

At Dubh Chlais the lower levels of this ice-mass met an impasse of solid rock, causing the formation in front of this of a transverse moraine (pl. 5,6), and giving rise ultimately to the development of drainage channels on either side of the rock barrier.

The main flow of ice, passing farther down the Dee Valley, turned sharply past the end of Craig Leek and into a now dry U-shaped valley,

Plate 5. Transverse moraine, with remnants of a small glacial lake below.

Plate 6. Details of the transverse moraine shown in plate 5.

rejoining the existing Dee Valley at Invergelder. The pile-up of morainic debris at the entrance to this now abandoned loop of the Dee (pl. 7) (see also fig. 2) was probably instrumental in forcing the river to find its present course, which is flanked by several levels of terraces.

Plate 7. The old valley of the Dee (now dry), with the present valley in the foreground at the Bridge of Dee.

PART II

STRATIGRAPHY

Introduction

The main stratigraphic units correspond to those first established by Bailey in the Perthshire Highlands (1922), although the order of succession is by no means obvious in this area owing to structural complications. For this reason previous workers (see PART III for details) failed to appreciate the succession. For the same reason original thicknesses cannot be reliably estimated, although it is evident that each unit represents many hundreds of feet of sediment.

The succession is:-

	Ben Eagach Schist ("Black Schist")
Dalradian Series	Perthshire Quartzite Series
	Blair Atholl Series
	Moine Series

Stratigraphic discordances would not be readily detectable, but the lateral continuity of the units and the occurrence of transitional formations suggests that no significant unconformities are represented in the succession.

Blair Atholl Series

The subdivision into "Pale Group" and "Dark Group" as recognised by Bailey farther to the southwest appears to be inapplicable, although the chief lithological types are represented, mainly limestone and schist. One minor occurrence of white limestone and a few doubtful ones of "Banded Group" are all that can be considered to represent the "Pale Group". What is more, no stratigraphical order can be established within the Blair Atholl Series around Braemar. No Boulder Bed is to be found.

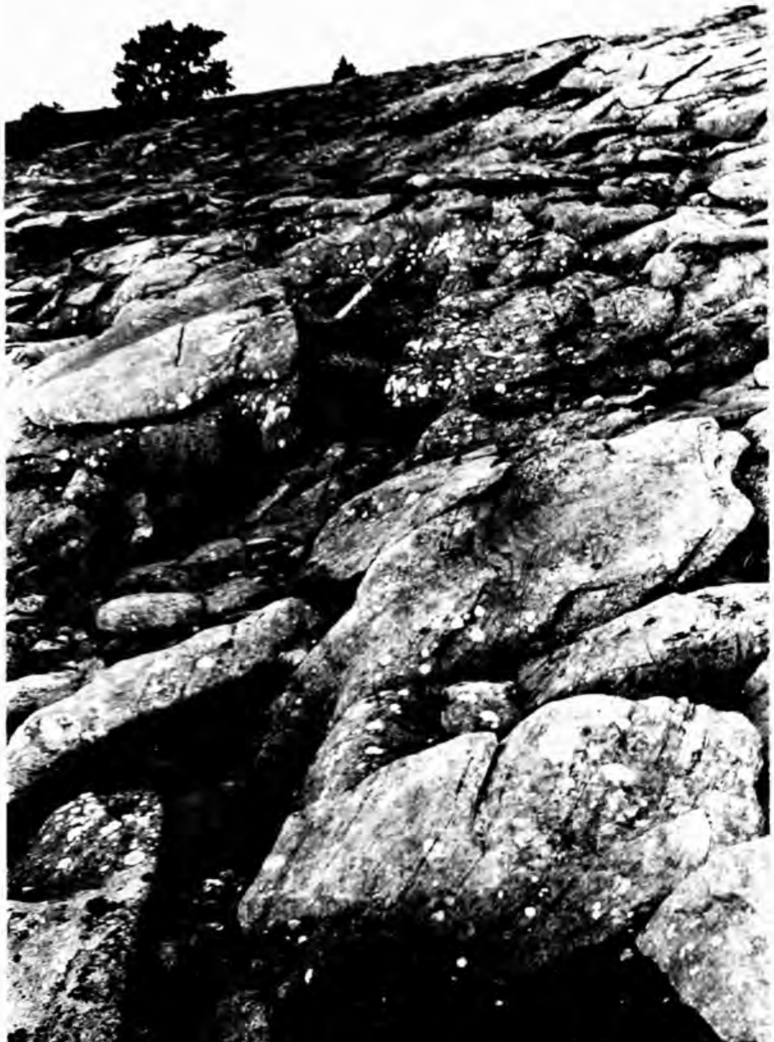


Plate 8. Main Blair Atholl Limestone exhibiting fine banding and a very crude cleavage which is approximately parallel to it.

The most abundant rock-type in the Series is schist, though as the area lies entirely within the hornfels zone the appearance of this rock-type differs from that to the southwest of Braemar. Three principal types may be recognised:- (a) dark or grey, (b) calcareous, (c) quartzose schists or hornfelses. The first of these is the most abundant, while locally there are minor amounts of a coarse dark micaceous schist, classed separately.

(a) The normal type is usually medium-grained, and except in the cordierite-rich part of the area, north of the Dee, banding is difficult to recognise. Some varieties are more graphitic and thus virtually identical to Ben Eagach schist. Others have quartz-rich or quartzite layers and are often rusty-weathering as a consequence of their pyrite content.

The true hornfelses often develop a distinct foliation and are fully described under "Metamorphism" (PART IV).

(b) The calcareous schist is much lighter in colour, softer and usually less schistose. Occasional thin bands reflect more or less calcareous-rich layers. Tremolite-schist is less common, being found in small lenses.

(c) Quartzose schists are tough, compact and brown in colour. Schistosity is poor to almost non-existent. These rocks may be very difficult to distinguish from fine-grained dioritic intrusions.

The main limestone in the area is identical to that lying outside to the southwest. When relatively pure it is massive and with little banding, though it may be cleaved or slightly schistose (pl. 8). More commonly it is slightly impure, and in this condition a very distinctive ribbing is present. This appears to have been deformed by the earliest folding, and may be a primary feature, though in its present condition it is deformed by later folding (pl. 9).

Plate 9. Ribbed limestone deformed by later-Caledonoid folds. There is evidence of earlier folding in some bands.

On Creag Choinnich certain exposures of limestone near the granite contain small garnets.

Another less common type of "limestone" restricted to Creag Choinnich is green and pink banded, and relatively compact. Technically this is a calc-silicate-hornfels, comprising alternating one-inch bands of diopsidic pyroxene, less amounts of plagioclase, sphene and sometimes biotite, with bands of porphyroblastic garnet. The garnet is ragged in outline. Associated brown patches are due to the presence of vesuvianite.

Examination of such a rock (CB 35) shows the garnet to reach almost an inch across, and even from the hand specimen alone it can be seen that the garnet is in the process of being overwhelmed by the brown vesuvianite.

Some exposures, as at the summit of Creag Choinnich, are entirely of green rock, others are richer in biotite.

Creag Choinnich also furnishes exposures of layers of pyroxene-felspar-hornfels alternating with bands composed of quartz and chlorite, occasional green amphibole and, rarely, biotite.

<u>Perthshire Quartzite Series</u>

This series was divided by Bailey into three parts, viz:-

Schiehallion* Quartzite

Killiecrankie Schist

Càrn Mairg Quartzite

However, as found by Cox (1966) in the almost adjacent area farther south in Glen Clunie, the Perthshire Quartzite Series appears to be almost entirely made up of massive quartzite. This main quartzite is almost

---

* Also spelt "Schichallion".

Plate 10. Current or false bedding in Perthshire Quartzite, Sroǹ á Bhruic.

Plate 11. Current or false bedding in less massive quartzite, Sroǹ á Bhruic.

invariably fine and never pebbly, though impure horizons may be encountered, these giving rise to a porous quartzite. No graded bedding has been found, but fine bedding is common as a pink to purple colour-banding. Other indications of bedding arise from the presence of pelitic bands, and in the variation in felspar content, while some exposures of quartzite possess a platiness or simple jointing as the only reflection of bedding (being parallel to good bedding in neighbouring exposures).

Owing to the rarity and ambiguity of observations of current bedding (pl. 10,11), the latter has not proved a reliable indicator of the order of succession in this area. Some possible examples of current-bedded layers have, for instance, been highly deformed (pl. 12).

The quartzite itself varies from pure white to cream or pink, with very thin clear quartz bands parallel to bedding. In thin section the quartz is seen to be highly strained and some specimens reveal two directions along which strain shadows are aligned. Both types of felspar are present, potash felspar (untwinned) being more abundant than the albite-oligoclase. Augen shapes are common and where bands of felspar are crowded with commonly orientated grains a sort of cleavage is imparted to the quartzite. Other common, though not abundant, constituents are chlorite and biotite, sometimes muscovite, all showing a crude common alignment. Zircon is often found as an accessory along with material having an adamantine lustre in reflected light, but with an appearance intermediate between rutile and opaque iron ore. This mineral, in small concentrations, produces a colour banding in the quartzite. Black bands are the result of the presence of pyrites. Apatite is rare in quartzite.

Micaceous bands up to half-an-inch thick may be found in the massive quartzite but their influence on the appearance of the sequence is

Plate 12. Fold in banded quartzite showing possible current or false bedding (above halfpenny).

Plate 13. Vertically-inclined transition rocks at Sroǹ á Bhruic exhibiting irregularities in the banding, which is parallel to bedding in the adjacent quartzite.

negligible owing to their scarcity.

The appearance of the quartzite is, however, altered by the presence of much felspar, particularly on Creag Choinnich, where there is very good evidence of local felspathisation caused by the presence of granite (for details see PART III B, p.45, and PART V, p.79).

On either side of the Perthshire Quartzite are gradational rocks effecting a transition downwards into the Blair Atholl Series and upwards into the Ben Eagach Schist. Typically both groups are of banded rocks, being regular alternations about an inch or two thick of quartzite and schist. While undoubtedly a sedimentary feature the banding is so highly deformed that no primary structures can be confidently identified (pl. 13). These rocks do not appear at every interformational boundary, but they are widespread, having been found at numerous localities to the southwest of Braemar* (beyond the present area).

On Carn nan Sgliat these rocks are at a slightly higher metamorphic grade on the Ben Eagach side of the Perthshire Quartzite.

Ben Eagach Schist ("Black Schist")

This is a very uniform formation, most of which is represented in the area by the typical hornfels described in detail under "Metamorphism" and "Structure". In hand specimen it is identical to the hornfelsed Blair Atholl Schist, and is identified therefore on its stratigraphic position. Consequently where the latter is doubtful it is not possible to assign dark schists (or hornfelses) to the Ben Eagach group with certainty; such is the case with the mass of hornfels around the quarry below the Lion's Face (see map 2) which cannot be ascribed to any part

* In the area mapped by Cox (1966), however, the Perthshire Quartzite is bounded by thrusts and the transitional rocks are not in evidence.

Plate 14. Moine-like rocks. Planes of movement are parallel to the very regular banding. Height of section about 10 feet.

of the main succession, as its structural position is enigmatic (see p.16 , Structure).

At the southwest end of Càrn nan Sgliat lies a zone of relatively unhornfelsed schist. Dark and graphitic, with finer cleavage, it is typical of the Ben Eagach Schist in the area to the southwest of Braemar. Also in this part of Càrn nan Sgliat are outcrops of dark green hornblende-schist, having the general form of masses of epidiorite. Tremolite limestone is less common.

Close to the summit of the hill two "horizons" of normal limestone have been found. The only other break in the lithological monotony is an occasional band of quartzite no more than a few inches thick.

Moine-like Rocks

Below the problematical horizon of hornfelses mentioned above is a sequence of banded quartzites. Apart from impurities and thin dark bands these rocks are much like Perthshire Quartzite (pl. 14). However within the Moine rocks there are horizons similar to this in Glen Ey and near the road leading uphill from Braemar to Corriemulzie. Since the low structural position of the banded quartzites must be near to that of those genuine Moine rocks they may be designated "Moine(?)" - see page 16.

Other such occurrences have been found in a zone of refolded Blair Atholl Limestone on the east side of Creag Choinnich. These rocks, about 100 feet thick, are all fine-grained, some of almost glassy aspect with lighter and darker bands. The commonest type is a micaceous granulite. The sequence is conformable with the Limestone and itself contains rare bands of limestone.

A thinner group of similar rocks is again associated with Blair Atholl Limestone at a point just above the Queen's Drive and the Lion's Face. This could well be the same stratigraphical level, and these rocks, too, are considered likely to be Moine.

PART III

STRUCTURAL GEOLOGY

A. METAMORPHIC ROCKS

Introduction

As with metamorphism, the structures to be found in the rocks of the area to the southwest of Braemar have been studied and compared with those of the area under discussion. Most of the work has been devoted to the latter, but comparisons have been necessary as no detailed work has been done previously around and to the north of Braemar. Indeed the area is too small and structurally complicated to be treated in isolation.

B.C. King (unpublished), working to the southwest of Braemar in an adjacent area, has interpreted that area on the assumption that the main stratigraphic units and sequence correspond to those even farther to the southwest, which have had a longer history of research, and where the major structures are more easily identifiable. The Dalradians in question thence continue in this belt as far as Islay and ultimately to Ireland, becoming correspondingly simpler in structure.

However, while many important structural features have been elucidated in the present area, others are far from certain. Yet detailed mapping has shown the degree of complication on a scale rarely attempted, in many cases linking major and minor structures, and demonstrating their similarities. In addition information on such aspects as petrofabric analysis has been found.

Interesting replacement structures by granite etc. have also been revealed by such detailed mapping. Most of these on Creag Choinnich are too minute to be shown on the 1-inch Survey map, and they do not appear to be mentioned in the previous literature (see PART IV).

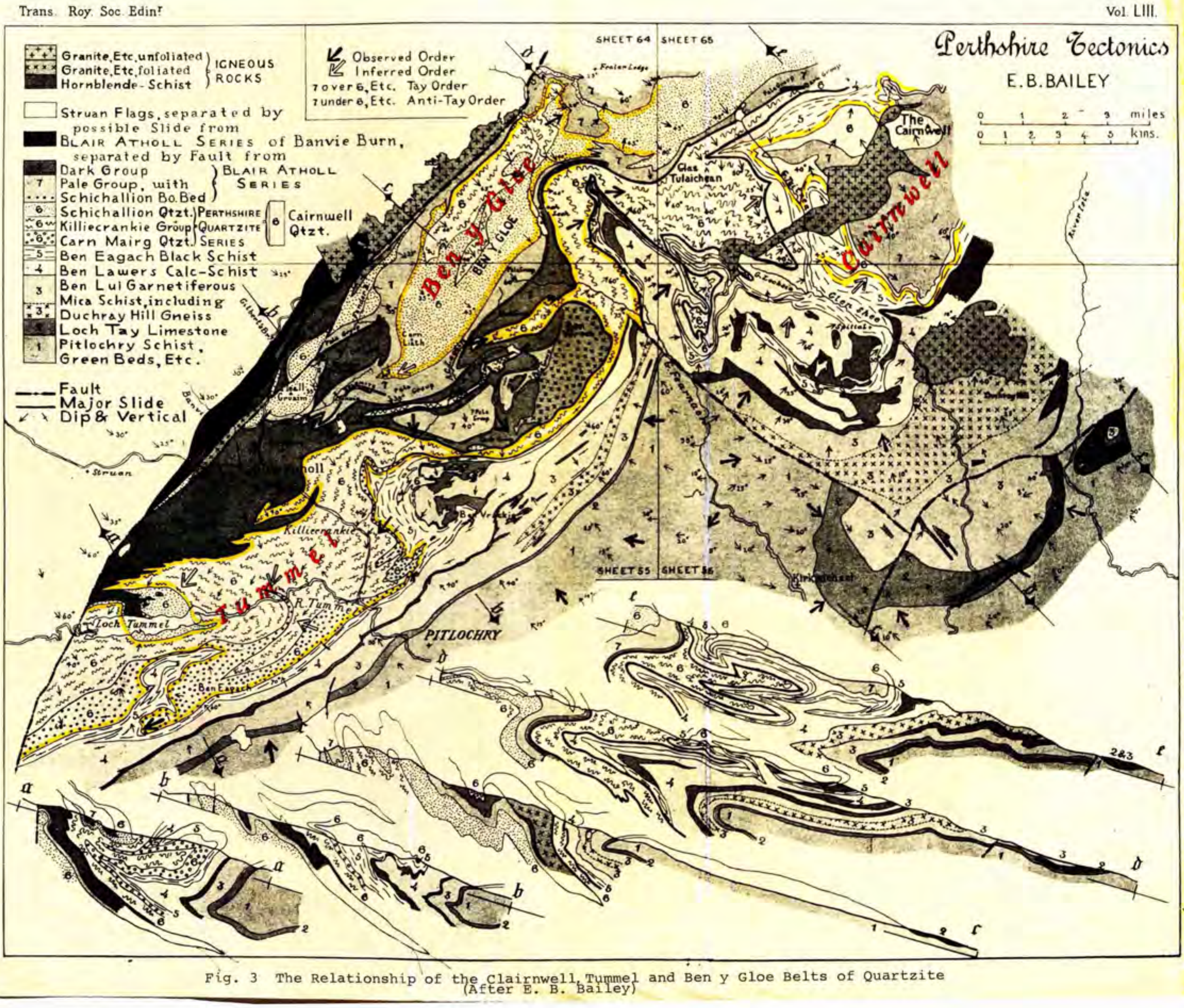


Fig. 3 The Relationship of the Clairnwell, Tummel and Ben y Gloe Belts of Quartzite
(After E. B. Bailey)

Previous Work

G. Barrow and E.H. Cunningham Craig carried out the original surveys around Braemar in 1904. While these workers recognised some of the complexities in the minor folds they did not appreciate the major ones. Consequently they did not establish the succession correctly, nor did they agree with each other. Craig in fact believed that the Perthshire Quartzite lay unconformably in synclinal folds of the Blair Atholl Series and Ben Eagach Schist. This fact is reflected in the survey map of the area in the relative position and shape of the Perthshire Quartzite outcrops.

Not until Bailey commenced his study was much achieved in elucidating the tectonic style of the Dalradians. His contribution was based on detailed local study and on broader considerations of the belt as a whole.

He, more than anyone, helped firmly to establish the concept of large-scale nappes in the southwest Highlands of Scotland. In 1922 he stated that the schists of the southwest Highlands belonged to three main divisions or nappes - the lowest being the "Ballappel Foundation", followed by the "Iltay Nappe", and then the "Loch-Awe Nappe".

Although the later work of Cummins and Shackleton (1955) and Shackleton (1958) in the Perthshire Highlands produced a more plausible picture of the major folds, linking Bailey's upper two structures into the Tay Nappe (a refolded recumbent fold), Bailey's appreciation of such large-scale phenomena enabled him to apply his ideas to the solution of the fold belt farther northeast up to Braemar.

He applied his ideas to smaller-scale structures also, and in 1925 interpreted the belt between Blair Atholl and the Cairnwell, identifying three major recumbent fold limbs of the Perthshire Quartzite Series:-

Cairnwell Limb (top)
Tummel Limb (middle)
Ben y Gloe Limb (bottom)

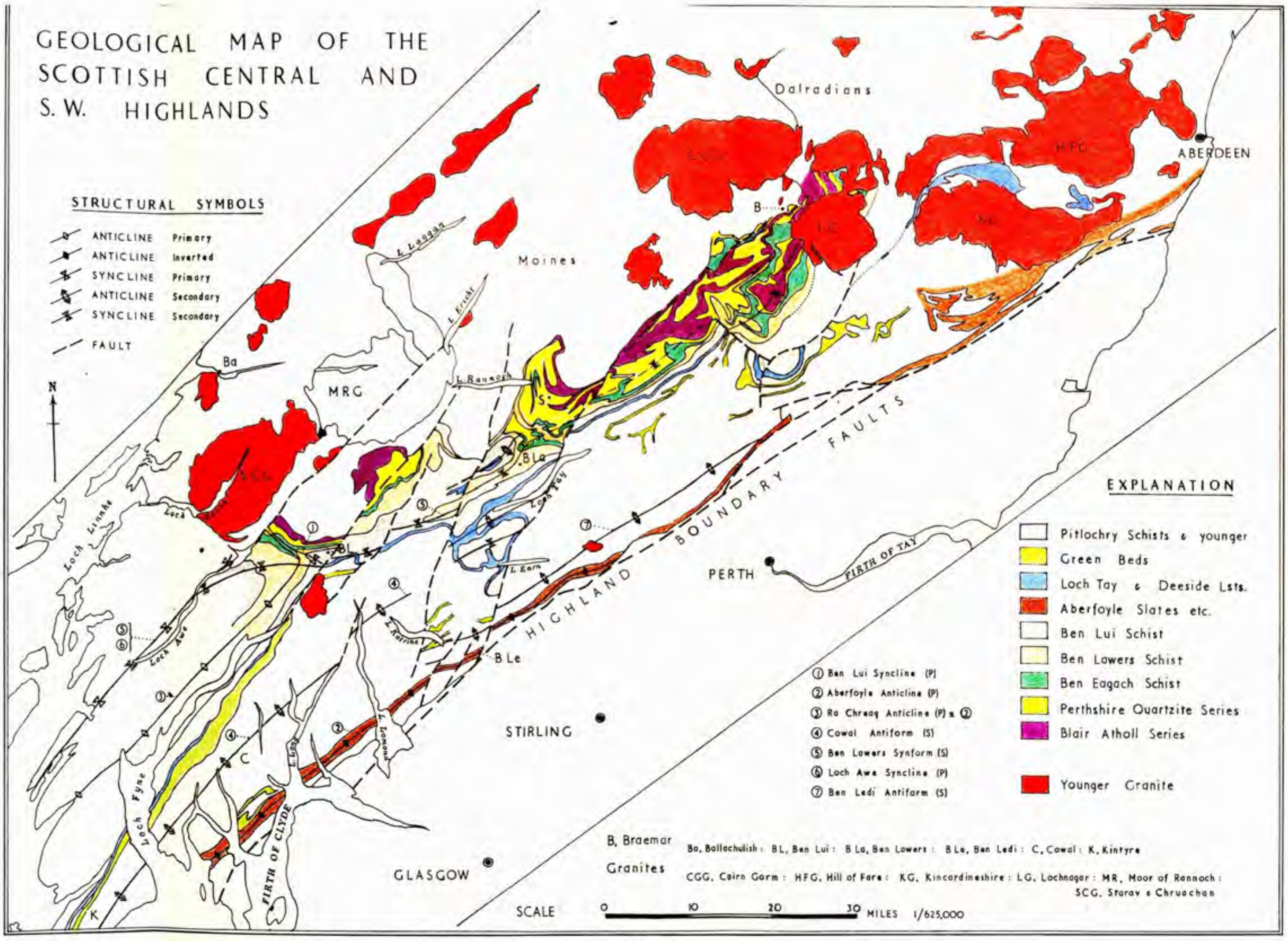

GEOLOGICAL MAP OF THE SCOTTISH CENTRAL AND S. W. HIGHLANDS
STRUCTURAL SYMBOLS
ANTICLINE Primary
ANTICLINE Inverted
SYNCLINE Primary
ANTICLINE Secondary
SYNCLINE Secondary
FAULT
N
Dalradians
Moines
MRG
ABERDEEN
PERTH
STIRLING
GLASGOW
FIRTH OF TAY
FIRTH OF CLYDE
HIGHLAND BOUNDARY FAULTS
L. Laggan
L. Ericht
L. Rannoch
Loch Tay
L. Earn
L. Katrine
L. Lomond
Loch Linnhe
Loch Awe
Loch Fyne
Ba
B
BL
B La
B Le
C
K
EXPLANATION
Pitlochry Schists & younger
Green Beds
Loch Tay & Deeside Lsts.
Aberfoyle Slates etc.
Ben Lui Schist
Ben Lawers Schist
Ben Eagach Schist
Perthshire Quartzite Series
Blair Atholl Series
Younger Granite
① Ben Lui Syncline (P)
② Aberfoyle Anticline (P)
③ Ra Chreag Anticline (P) ≡ ②
④ Cowal Antiform (S)
⑤ Ben Lawers Synform (S)
⑥ Loch Awe Syncline (P)
⑦ Ben Ledi Antiform (S)
B, Braemar
Granites
Ba, Ballachulish : BL, Ben Lui : B La, Ben Lawers : B Le, Ben Ledi : C, Cowal : K, Kintyre
CGG, Cairn Gorm : HFG, Hill of Fare : KG, Kincardineshire : LG, Lochnagar : MR, Moor of Rannoch : SCG, Starav & Chruachan
SCALE 0 10 20 30 MILES 1/625,000

Fig. 4

From considerations of pitch, and presumably also from comparison with Schiehallion to the southwest (which Bailey subsequently reinterpreted), the Ben y Glòe belt of quartzite was considered an anticlinal structure (fig. 3).

Bailey also recognised the continuation of the Ben Lawers "Syncline", passing close to Loch Tummel, Ben Vrackie, Ben Vuroch and the Cairnwell. The synformal Cairnwell Belt of Bailey is thus correlated with the Ben Lawers "Syncline", as indicated on the coloured diagram (fig. 4).

At the same time Bailey's notions enabled him to appreciate the existence, in certain areas, of smaller-scale slides, along which recumbent folds had been displaced. Consequently he was able to invoke the presence of slides to attempt to understand the Braemar area by linking up structural units which had previously been considered separate. According to his map, therefore (fig. 5), Càrn nan Sgliat and Morrone belong to the slid out portion of the Tummel Belt and the rest of the succession east of Càrn nan Sgliat and Morrone to the Cairnwell Belt.

However the elements below these two belts, i.e. lying to the west and including Craig Choinnich, did not easily fit into Bailey's simplified plan, and at this stage he ceased investigation, though it is tempting to assume that he believed Creag Choinnich to belong to the Ben y Gloe Belt.

Much later the area around Glen Shee was investigated by Matheson (Ph.D. Thesis, 1958), who agreed with Bailey's notions.

Recently Pantin (1961), on re-examining the Ben y Gloe area, recognised the Càrn Liath Slide, which he considered to cause the termination of the main Ben y Gloe quartzite mass; but more important, he considered this belt to be synclinal and not anticlinal, though the evidence is not absolutely conclusive.

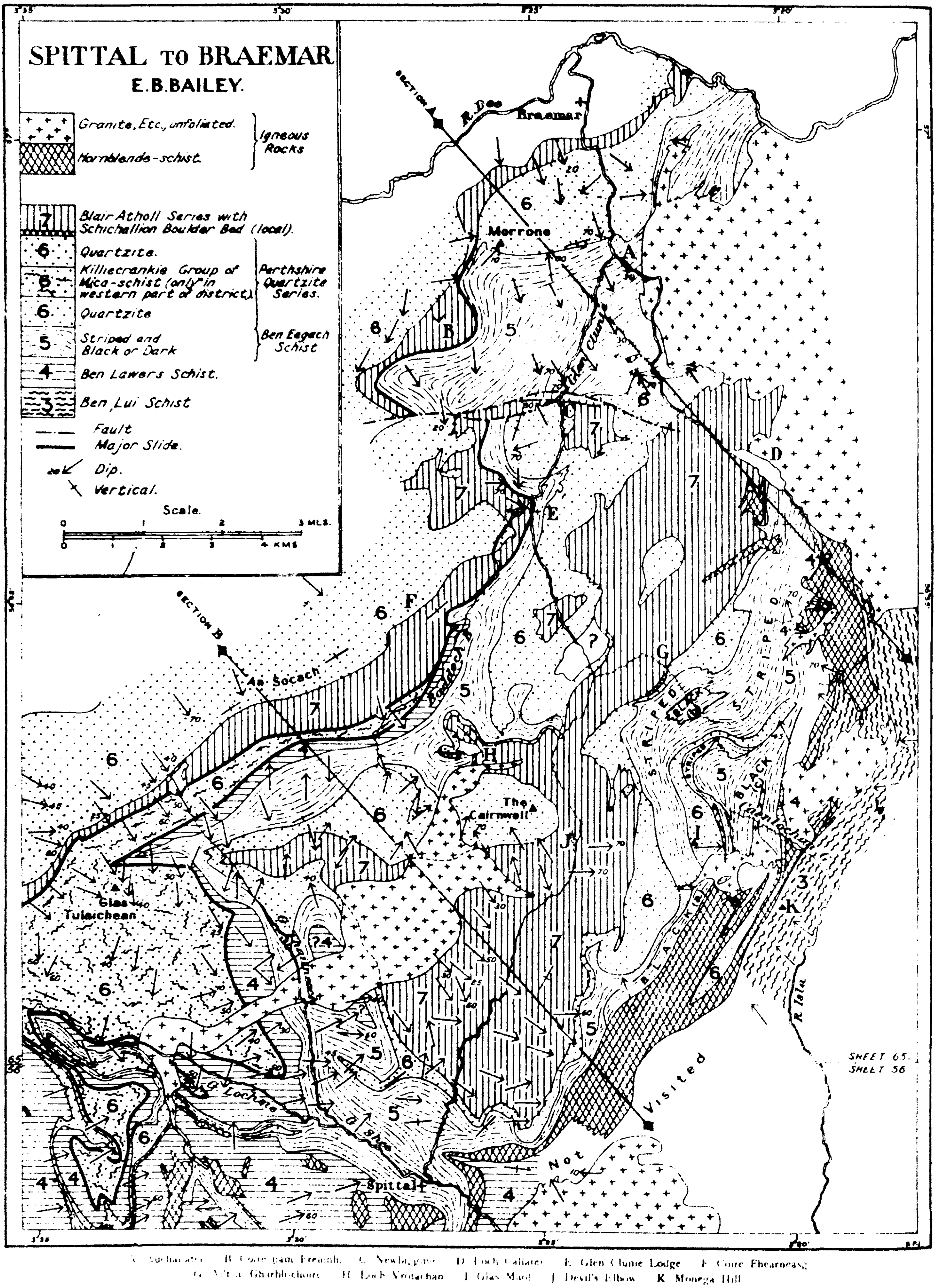


Fig. 5. Geological map of the area between Spittal of Glenshee and Braemar (after E.B. Bailey).

The recent work of King (unpublished) in the area to the southwest of Braemar has, however, revealed the presence of several major thrusts which post-date the main episodes of folding. The presence of these thrusts has been confirmed by Cox (Ph.D. Thesis, 1966) in the area south of Braemar from Clunie Lodge to the Cairnwell. Three structural levels or nappes are recognised:-

(1) The Morrone Nappe

(2) The Blair Atholl Thrust Zone

(3) The Lower Nappe

The "Morrone Nappe", comprising the remnants of a primary (recumbent) synclinal core of Ben Eagach Schist overlain by Perthshire Quartzite, can be equated by direct mapping to Càrn nan Sgliat (which comprises the lower side of the syncline with the Ben Eagach Schist overlying the Perthshire Quartzite).

The "Blair Atholl Thrust Zone", believed to be a primary recumbent structure, could correspond with the Blair Atholl belt running along the base of Càrn nan Sgliat and Craig Leek, while the principally quartzitic mass of Creag Choinnich could be the "Lower Nappe". As will be shown presently, the Blair Atholl belt in the present area is squeezed out towards the southeast in much the same way as recorded by Cox.

According to King (personal communication) and Cox the Cairnwell Synform as envisaged by Bailey does not exist, this zone being instead occupied by the "Morrone Nappe" (1) and part of the "Blair Atholl Thrust Zone" (2).

Although the findings of King and Cox have not been traced farther southwest it would appear nevertheless that the Lower Nappe (3) comprises part of Bailey's Ben y Gloe Limb.

Consequently whereas Bailey envisaged fold structures accompanied

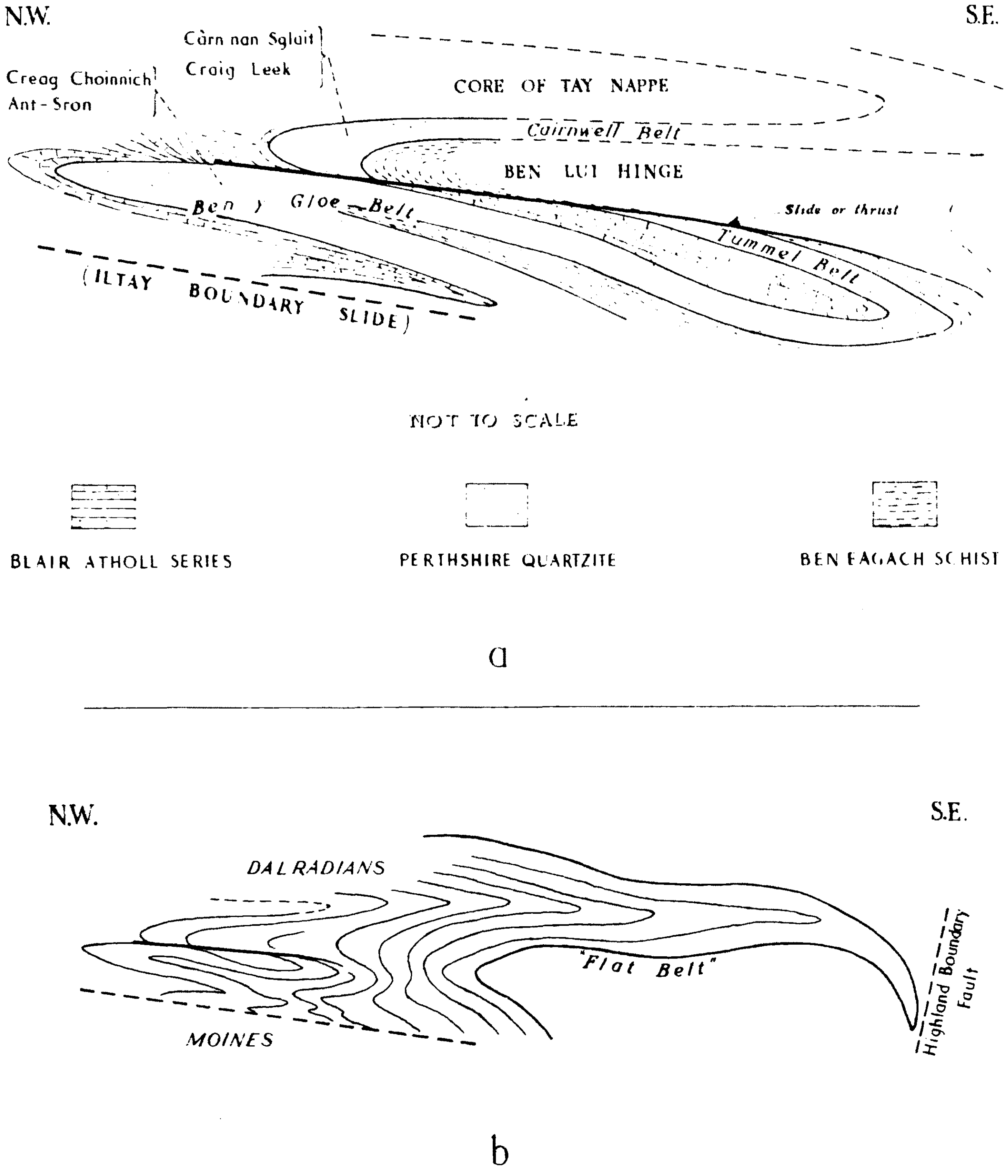


Fig. 6. (a) Simplified structure of the Braemar area in relation to the Ben Lui Syncline.
(b) Profile of the Iltay Nappe.

by sliding largely along interformational boundaries as primary features, King and Cox have found that the primary and secondary folds have been cut across by later thrusts.

Brief Summary of Structures

The area mapped may thus be considered, in the light of previous work, to contain parts of the main recumbent limbs of folds in the Perthshire Quartzite Series, established near Blair Atholl. These, in turn, are part of a major structure which stretches for the entire length of the Dalradian outcrop in Central and Southwest Scotland and known as the Iltay Nappe. This has subsequently been deformed by later major folds mainly of similar axial direction.

The complications of the Braemar area would appear to lie in the complementary Ben Lui Syncline or hinge, and to have been affected by later thrusting (fig. 6).

NOTE ON FOLD SYMBOLS:- The Dalradian rocks of the Scottish Highlands are generally believed to have suffered at least three successive phases of folding, usually designated "F_1", "F_2" and "F_3" etc. Within the area mapped, however, there is insufficient evidence to be dogmatic as to the order of fold episodes, and indeed there is some evidence that folding may have been manifest on two sets of axes simultaneously. Hence to avoid confusion early Caledonoid folds are designated F_{A1} and the resultant schistosity or cleavage S_{A1}. Early cross-folds are F_{B1}, related planar elements S_{B1}. Later Caledonoid folds are F_{A2} etc.

Primary Structures (F_{A1}, S_{A1})

It is supposed that the area mapped contains portions of at least two main nappes, because on stratigraphical grounds a structural repetition can be recognised. This idea is supported by the partial sliding or thrusting out of the Blair Atholl Limestone between the Perthshire Quartzite of Creag Choinnich and that of Càrn nan Sgliat. Though not obvious on the map the structural discordance can be observed to the east of the Lion's Face. Discordance is also found in the corresponding slide or thrust on Morrone, to the southwest of Braemar.

These nappes are parts of large recumbent folds overturned to the southeast. They were developed in the plastic stage of deformation, but also suffered a later more "brittle" deformation as witnessed by the thrust separating them.

The direction of anticlinal closure cannot conclusively be established. On Morrone part of an observed closure indicates a southeasterly direction, and certainly a more acceptable structural pattern results from considering this as the direction. This is as Bailey envisaged the major folds, projecting his ideas from Ben y Gloe.

The lower of the two nappes consists of the quartzite of Creag Choinnich (and infolds of limestone) and possibly small outcrops of quartzite on Craggan Rour, Ant-Sron and the lower outcrops of Craig Leek. The upper nappe comprises the remaining quartzite and Blair Atholl rocks of the area, and has as its core the whole of the Ben Eagach Schist.

These main units trend northeast-southwest along the Caledonoid direction, and like those southwest of Braemar, dip to the southeast. The main internal structures having approximately the same trend, it is very likely that these principle units are primary major structures. Certainly no earlier have been detected.

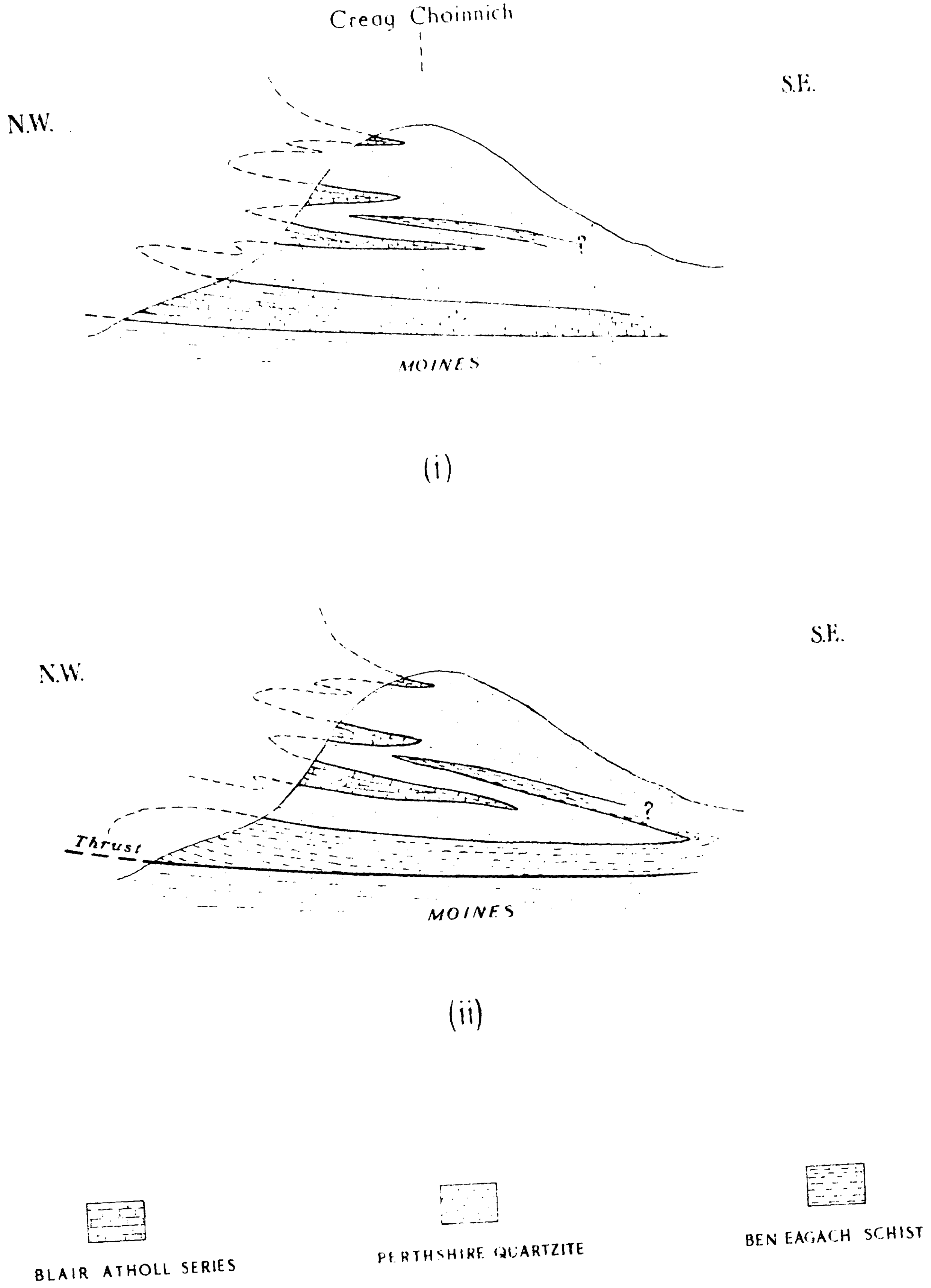


Fig. 7. Alternative structural interpretations of Creag Choinnich.

Plate 15. The supposed base of the lower nappe resting on Moine(?) rocks - - Creag Choinnich.

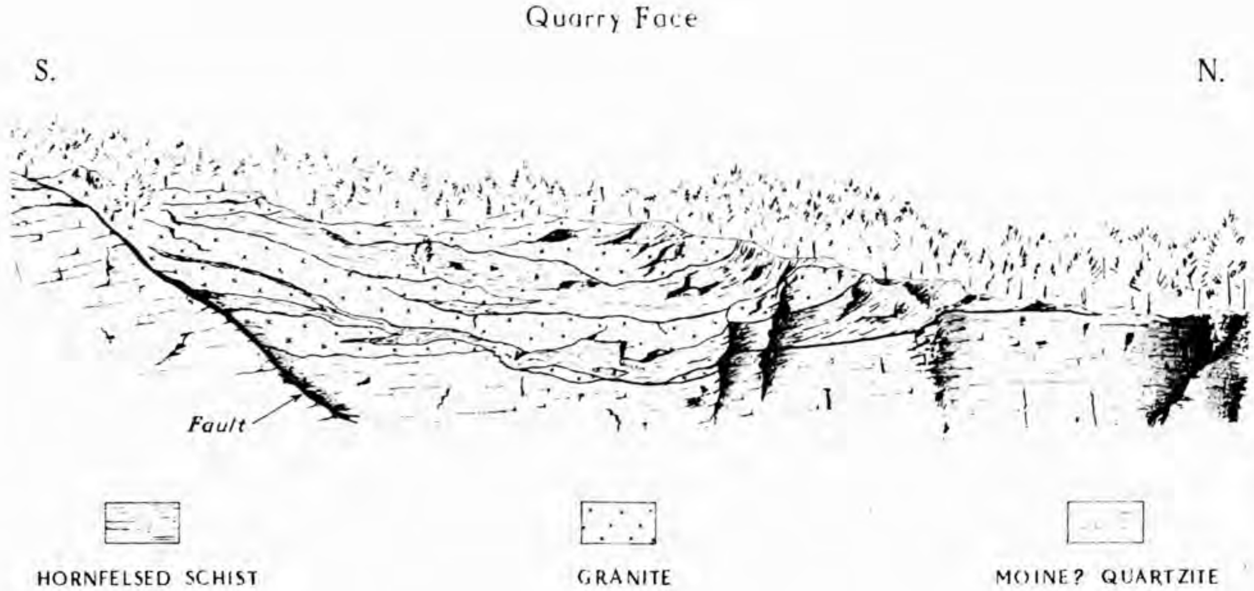


Fig. 8. Sketch-section of the Quarry Face, Creag Choinnich.

(a) Lower Nappe

Extreme complication permits only the upper part of Creag Choinnich to be interpreted confidently. This level is a recumbent anticline of quartzite with a core of limestone closing to the southeast (limestone occurs on the northwest side of the hill only).

The lowermost exposures around the northern part of the same hill may belong to the Moine Series in part. The granitised schist below the Lion's Face is part of either the Blair Atholl Series or Ben Eagach Schist, as already stated under "Stratigraphy". The two most likely structural interpretations (simplified) are shown.

The lowest rocks of all, found in the quarry 1/6th mile north of the Lion's Face, are believed to be Moines. Their simplicity of structure belies all but a trace of earlier deformation blotted out by extreme shearing and flattening. Consequently the lower nappe is believed to rest directly on the Moines at the base of Creag Choinnich, the junction being part of a slide (fig. 8 , pl.15).

This level is not exposed in the northward continuation of the lower nappe below Craig Leek and Ant-Sron.

In more detail the lower nappe of Creag Choinnich presents an outcrop pattern controlled by two trends: a "cross-fold" trend in the limestones and schists of the summit region, and a Caledonoid (late) trend in similar rocks of the northwest side of the hill. It is, however, the main masses of quartzite which retain the primary structure, that of parallelism of bedding and cleavage. The southeast half of the hill is pervaded by a planar structure dipping regularly southwestwards. The remaining half has a less regular dip to the southeast.

Primary structures, it is believed, are responsible for (a) the coincidence of bedding and cleavage; (b) the recurrence of quartzose schist horizons (Q′); and (c) the great thickness of quartzite here in

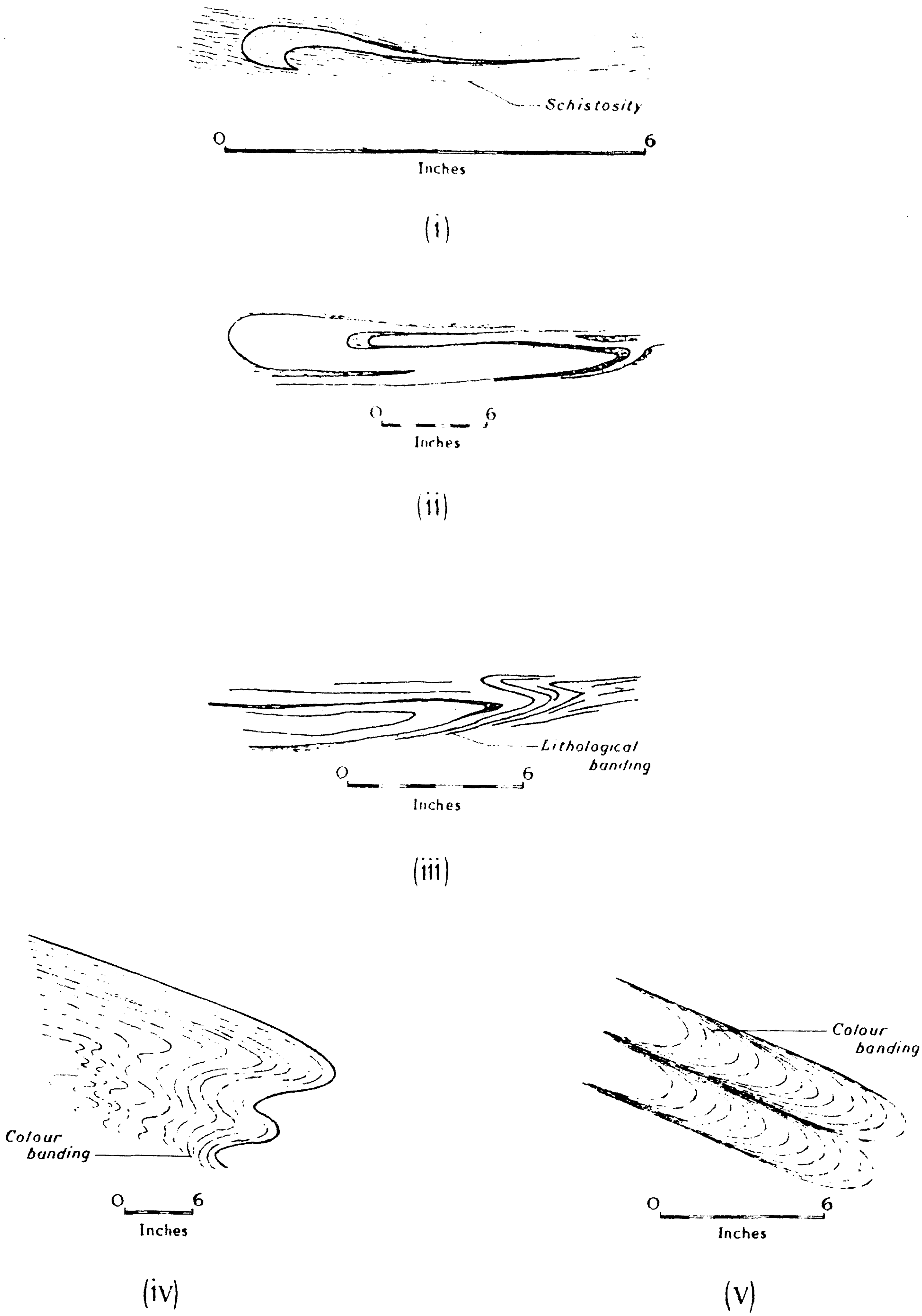
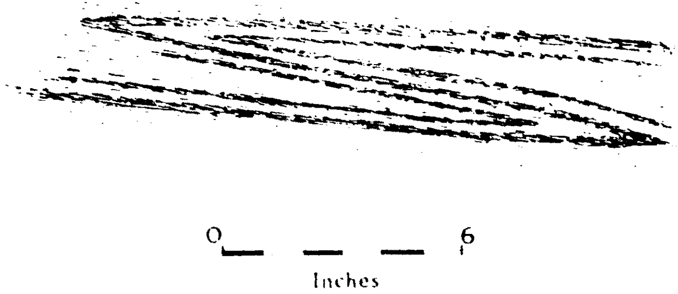
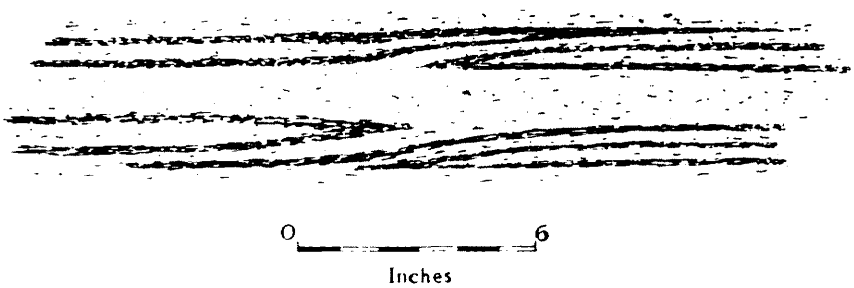


Fig. 9. Fold profiles in semipelites (i-iii) and massive quartzite (iv,v).



(i)



(ii)

Fig. 10. Highly compressed folds affecting the foliation in schists.

comparison to very little in the lower nappe on Morrone. To some extent, however, cross-folding is also responsible for this thickening.

On a minor scale early-Caledonoid folds are almost unknown in rocks other than quartzite and quartzose schist. Often structures of this generation are recognised or inferred from such evidence as relict fold noses - perhaps a compressed fold closure in quartzite less than an inch thick, "floating" rootless in a more pelitic medium.

Occasionally axial plane "cleavage" is developed in the larger minor folds in the quartzite, shown either by the elongation of felspar crystals and/or by the recrystallisation of quartz in thin strips. Shearing is manifest too.

Examples of the smallest folds in the lower nappe are only to be found in semipelitic rocks. Since such structures tend to be sheared out, as described above, the sense of overturning and direction of translation remain unknown, as does the geometry of the stacking of successive folds. The structures are typical shear or flow folds, with thickened closures (fig. 9). Individual larger folds, usually seen in massive quartzite, are better preserved, but their isolation similarly prevents a complete study from being made(pl. 16).

In schists and limestones original bedding structures are not to be found; consequently one can only infer the presence of early Caledonoid folds etc. Also, the earliest schistosity may be an axial plane cleavage to the now unrecognisable F_{A1} folds, or it may be a primary bedding-plane cleavage. Since, however, in the semi-pelites the earliest schistosity is related to relict F_{A1} fold closures, i.e. it is S_{A1}, it is likely that the same applies in schists and limestones. Fortunately rare cases have been observed where the same schistosity, deformed by later crossfolds (F_{B1}), is clearly seen to be axial planar to highly compressed folds of the type illustrated (fig.10) which can only be F_{A1}.

Plate 16. Isolated, possibly primary folding in quartzite of the lower nappe.

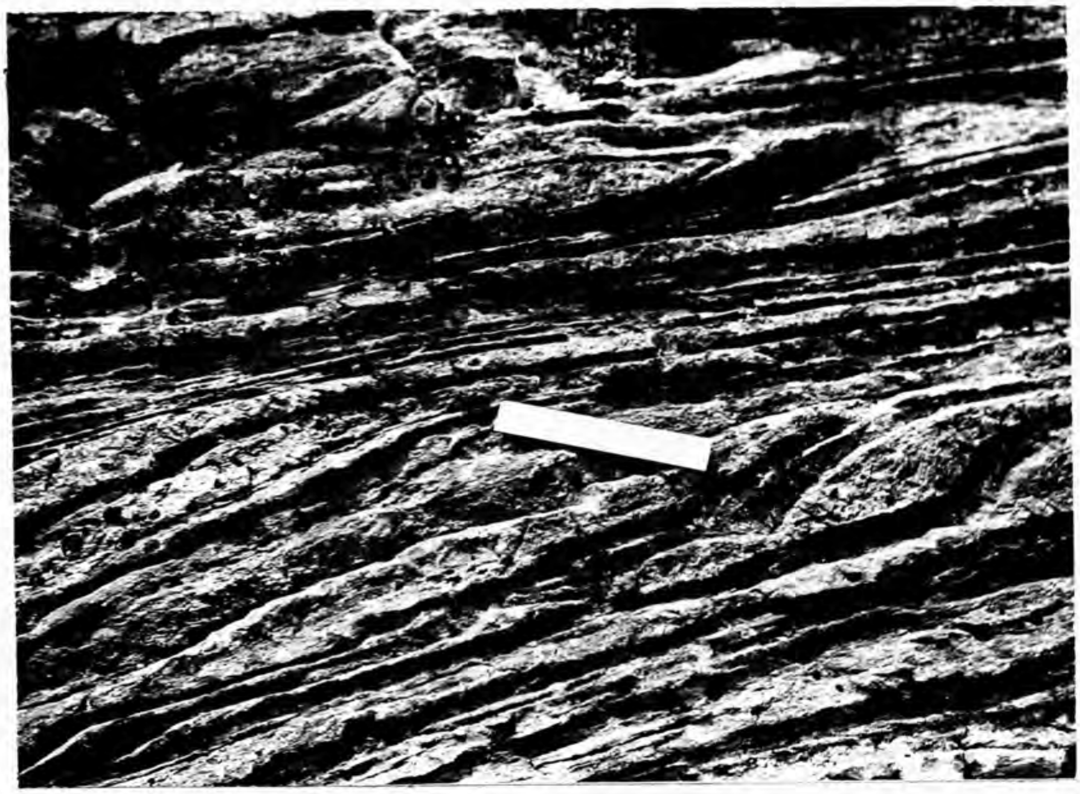
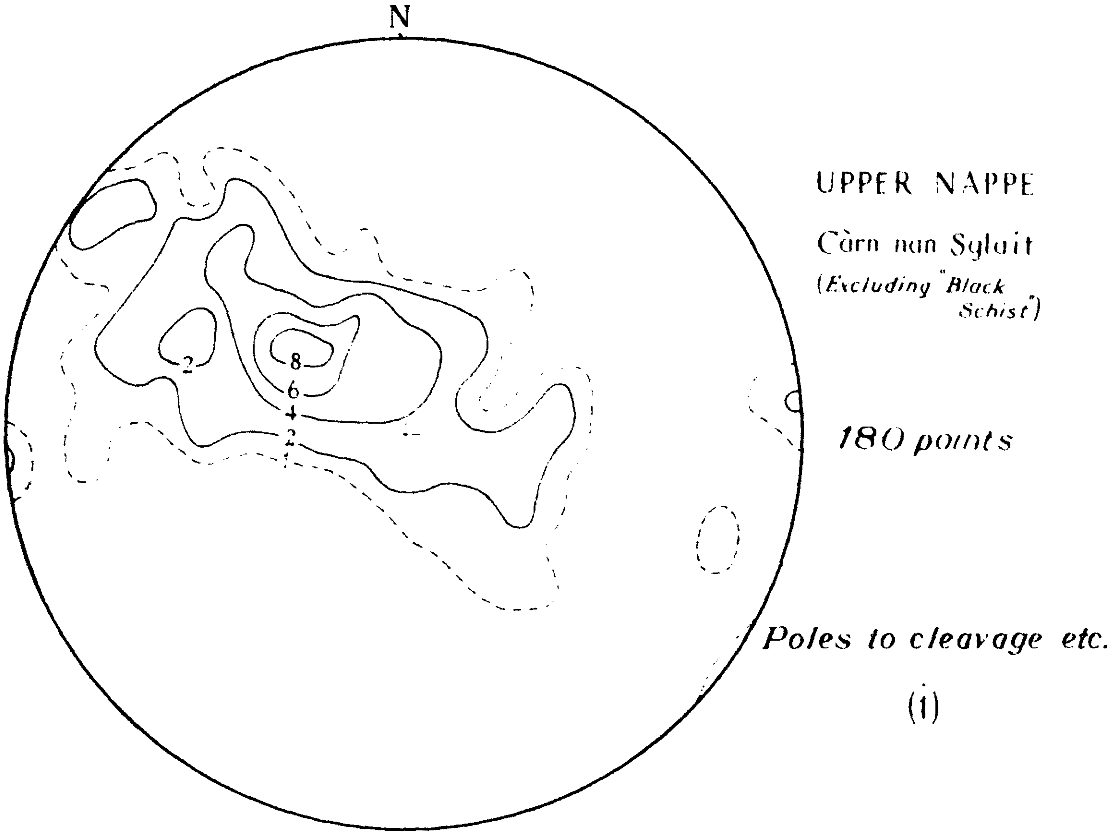


Plate 17. Early folding (top left) with later oblique schistosity in thicker pelitic bands (below and to right of ruler).



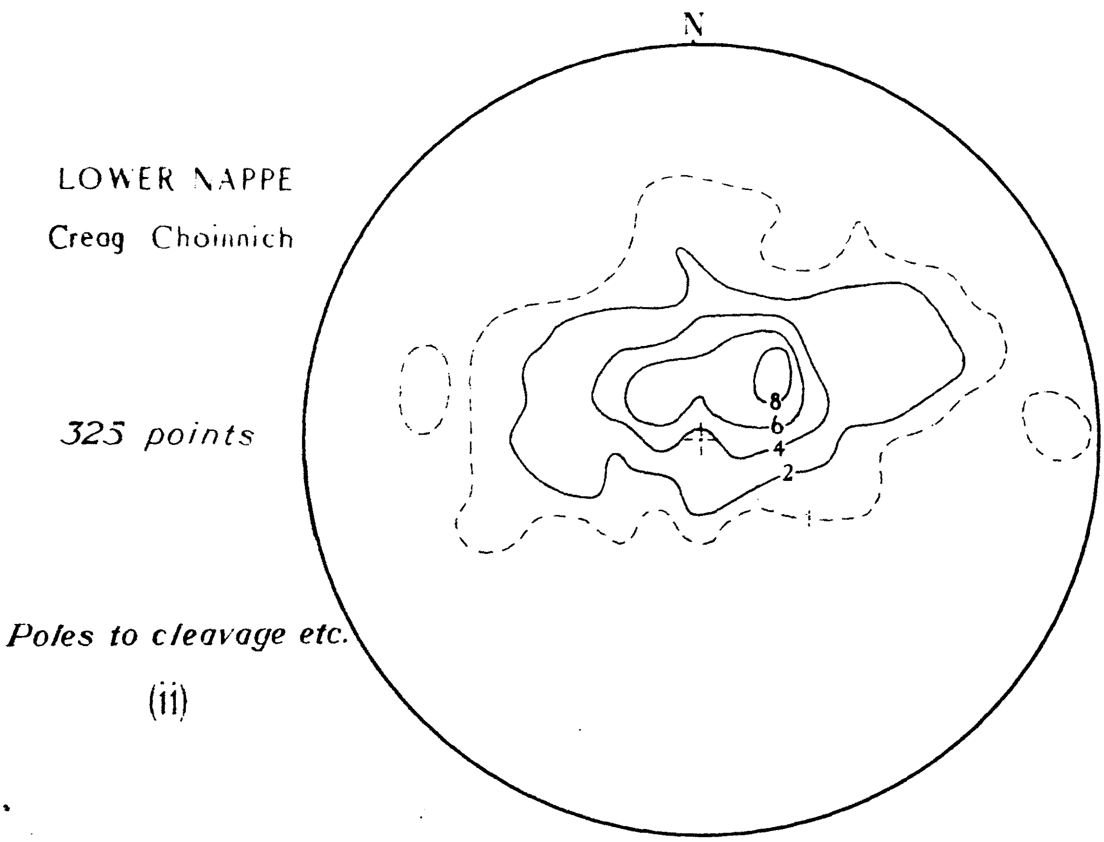


Fig. 11. Stereograms of poles to schistosity or cleavage in the upper nappe (i) and lower nappe (ii).

This early schistosity is particularly well developed in the limestone outcrops around the summit of Creag Choinnich, where it clearly pre-dates the later folds (F_{B1}) (see fig. 13). Depending on the amount of quartz present in schists, this early schistosity may be dominant or in varying degrees of replacement by a later one (S_{B1}).

As plate 17 illustrates, where the quartzitic bands are close together the schistosity is parallel and still in the axial plane of early folds, but in the thicker pelitic bands the schistosity is oblique and becoming parallel to the axial plane of a much larger secondary fold (beyond the limits of the photograph).

Statistically the poles to schistosity or cleavage planes in the lower nappe are spread out (fig. 11), but with a definite maximum in the northeast quadrant of the stereogram, corroborating earlier statements on the major structure.

The lower nappe is believed to continue northwards in the quartzite of the lower slopes of Craig Leek and Ant-Sron. In the former case the quartzite is separated from the upper mass by Blair Atholl Limestone. The relationships are thus analogous with those south of the Dee, even to the presence in the quartzite of Craig Leek of highly flattened early minor folds.

On Ant-Sron the quartzite is found in masses, more or less isolated by schist and limestone, the form of which cannot be simply related to the internal structures. Thrusting or sliding between the main lithological units has been observed, so that they are regarded as complicated "fold-wedges". The presence of minor cross-folds suggests that later deformation has contributed to the development of the outcrop pattern here.

(b) Upper Nappe

The upper nappe appears much simpler than the lower, consisting of a core of Ben Eagach Schist, an outer layer of Perthshire Quartzite, and an envelope of the Blair Atholl Series. The limestone of the latter acts as a zone of décollement. The structure appears to be that of a recumbent syncline, facing to the southeast, and partly "slid out" at the base. It must be regarded as being in the hinge region of the Tay Nappe, that is, in the complementary Ben Lui recumbent syncline.

The major trend of the individual formations is dominantly northeast-southwest, with gentle dip towards the southeast, except in the extreme north and northeast of the area. Here the trend becomes northwest-southeast, dip to the southwest - a feature that continues much farther to the northwest, beyond the present area.

South of the Dee the angle of dip in the upper nappe increases upwards towards the core of Ben Eagach Schist, in which dips may become vertical. Consequently the true closure of the nappe is almost at this point, the other (upper) side of this hinge having been completely removed by erosion. Its easterly continuation is represented in part by the down-folded Perthshire Quartzite at Sròn á Bhruic, south of the Old Bridge of Dee. Here the dip of the quartzite and adjacent "Black Schist" is vertical.

On Craig Leek, north of the Dee, the Ben Eagach "Black Schist" lies at a structurally lower level than the Perthshire Quartzite, due to the effect of a fault running parallel to the present cliffs, but the attitude of the two units is similar to that to the south.

Large early-Caledonoid folds may be recognised on the maps provided (especially map 2). An obvious one is the long tongue-like mass of Perthshire Quartzite penetrating the limestone belt at the foot of Càrn nan Sgliat, near the Queen's Drive. Another smaller mass lies immediately to the east. These have the appearance of subsidiary primary folds or digitations of the major ones.

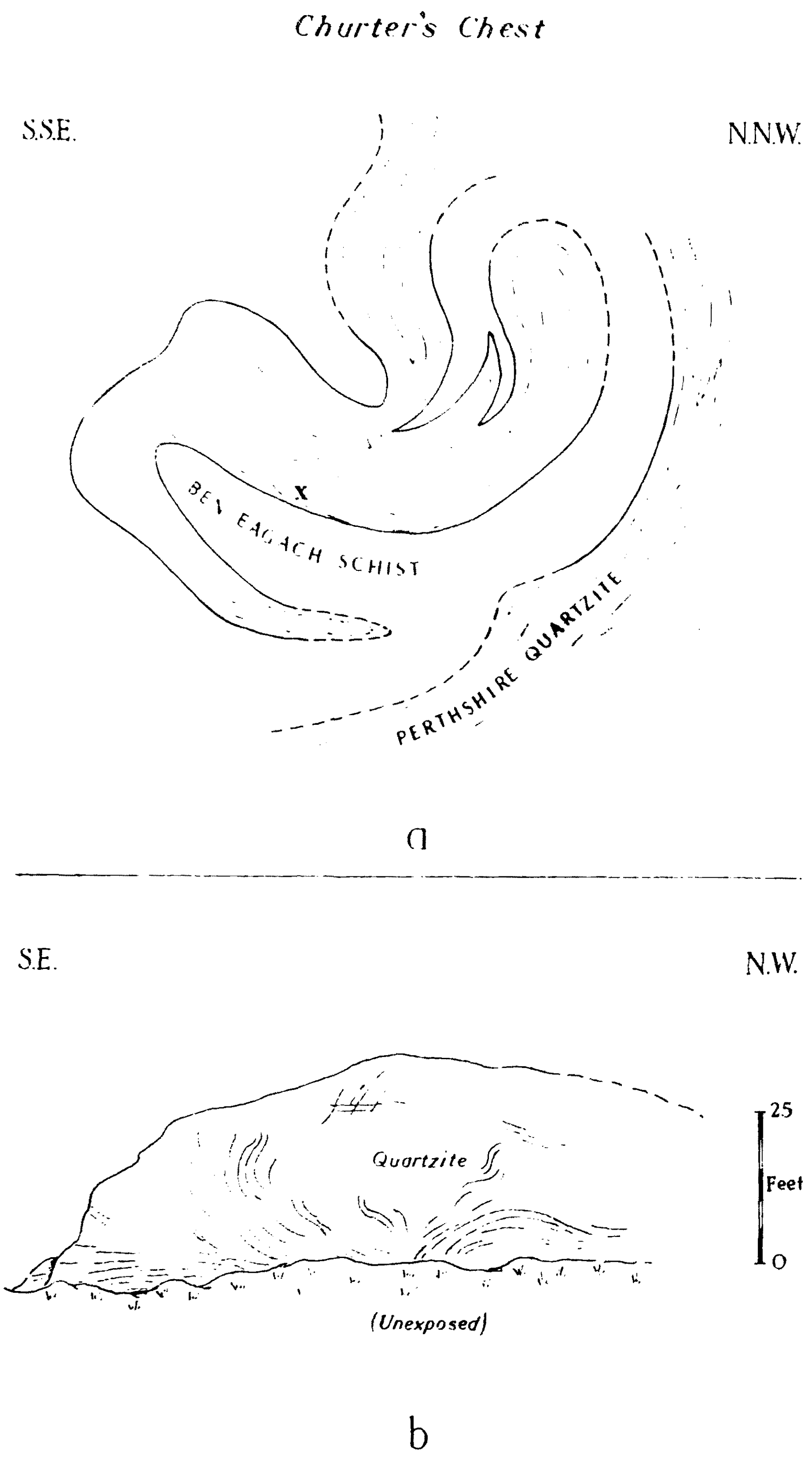


Fig. 12. (a) Possible interpretation of the structure at Churter's Chest.
(b) **Diagram** of folded joint- or cl**eavage-planes at the point** "X" in (a).

The zones adjacent to these masses have been invaded by numerous minor intrusions, clearly following horizons of structural weakness. A similar situation occurs at a higher level. Here lenses of quartzose schist in the main quartzite, and the presence of a swarm of felsites, may mark an otherwise hidden slide or fold closure, particularly as the felsite sheets are present in the area to the southwest at a level of large-scale dislocations in the upper nappe (the Morrone Nappe).

Evidence for other early folds largely depends upon the repetition of, say, limestone within schist in the Blair Atholl Series. Rapidly varying lithological sequences from place to place suggest repetition or complication by the earliest folding rather than by later. Within the uniform Ben Eagach Schist it is likely that similar folding exists, but it can only be appreciated where, as at Churter's Chest, the adjacent quartzite is also involved.

This locality, a bold feature in the cliffs at the north end of Craig Clunie, represents the junction of the Ben Eagach and Perthshire Quartzite formations, deformed by at least two sets of fold movements. Various interpretations are possible but the basis of each is that mesoscopic early-Caledonoid folds have been warped by later ones on approximately the same axes (fig. 12).

Examples of truly minor folds may be found in the centre of the quartzite of the upper nappe on the southwest portion of Càrn nan Sgliat. Tight, recumbent folds, like those of the lower nappe, are not found: in contrast, many are of several centimetres amplitude only, with near-vertical axial planes. Intense deformation has, however, been manifest (fig.9v). The above folds have a Caledonoid trend, except at the extreme southwesterly limit of the area where east-west fold-axes predominate (see stereogram, fig.17).

The latter have all the appearance of the same early phase, and are

Plate 18. Curved joint surfaces containing folded bedding planes, as indicated.

Plate 19. Evidence of highly compressed primary folds (indicated) from Sroǹ á Bhruic.

Plate 20. Complex nature of early folding in semipelites, north of the Dee.

Plate 21. Part of the exposure covered by plate 20 (top of picture).

accompanied by numerous lineations. Some are parallel to, others close, but definitely sub-parallel, to the fold-crests. The two sets of lineations are believed to belong to different generations, though only one locality provides direct evidence of this:- (see also plate 18)

The colour bands here are parallel to surfaces which are obviously folded bedding planes. Upon these are fine lineations, plunging 20° to the east at 105°, caused by the intersection of the bedding with joint-planes. The joint planes, parallel to the axial planes of the folds, are themselves warped and bear a second lineation roughly parallel to the crests of the warps, plunging 20° to the east at 80°. Clearly two phases of deformation have affected this belt of quartzite, their respective axes lying close to one another, yet possibly far enough apart in places to account for the observed variations.

Minor folds in the semipelitic rocks of the upper nappe are much as for those in the lower nappe, and show as much variation in style. Sometimes the evidence of the movements responsible is all but obliterated (pl. 19); sometimes it is complex (pl. 20,21).

Minor folds in limestones are very difficult to recognise. The extensively-developed ribbing is believed, in spite of its uniformity, to represent the first cleavage, S_{A1}: that is to say, the banding or bedding in the original rock has been folded and completely compressed to form a new layering without the production of a true cleavage. Only two neighbouring localities in the whole area provide evidence of this process in the intermediate stage.

At one (041, 077) two juxtaposed ribs are seen, when traced as parallel bands round a series of minor F_{A2} axes, eventually converge at a highly acute angle, clearly visible as a fold hinge (plate 22). The adjacent exposure, a vertical surface running parallel to the Caledonoid direction, is of massive limestone which reveals the style of the earlier

Plate 22. Closure of primary fold in ribbed limestone, indicated by blade of knife.

Plate 23. Banded limestone exhibiting early folding (immediately above and below ruler). Note boudinage, top left.

Plate 24. Variation in F_{B1} minor folds, with thickening towards hinges.

minor crossfolds. One of these clearly bends the axis of an even earlier fold, which must therefore belong to the F_{A1} phase, while close to this point a very tightly folded quartzite band carries a lineation on the same (Caledonoid) axis. This must thus be a primary lineation.

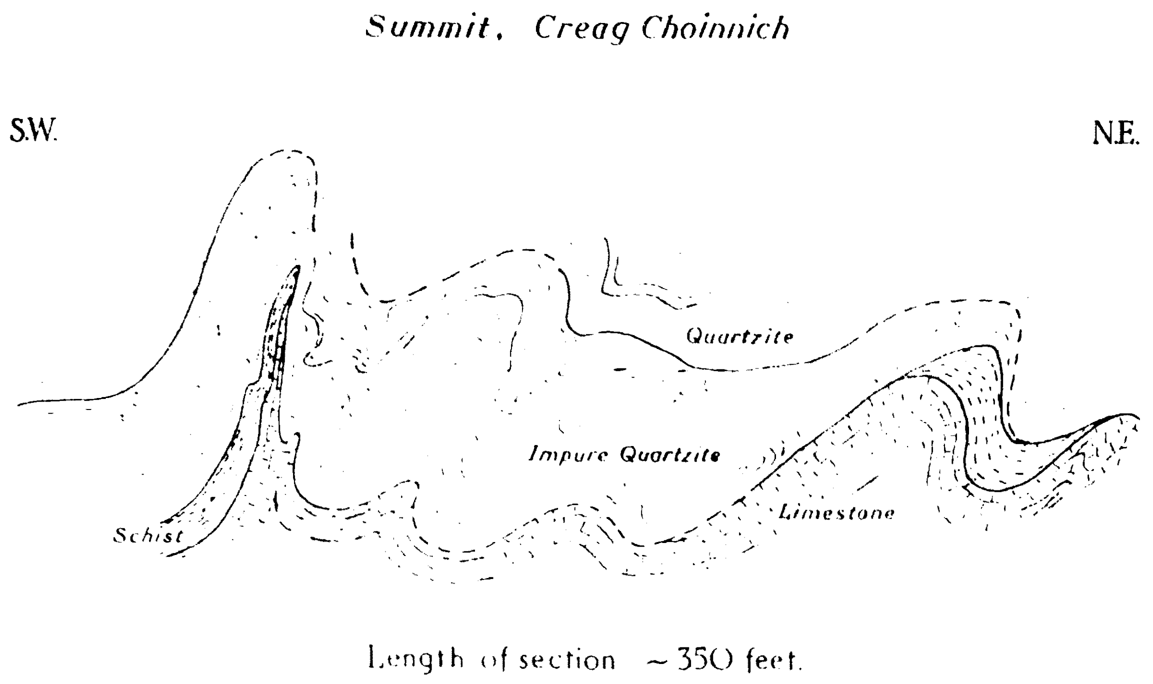
Finally, considering primary minor structures in the pelitic rocks, it has been found that in the schists adjoining belts of limestone in the upper nappe south of the Dee, a new schistosity has usually developed where F_{A2} minor folding is pronounced. It is only rarely possible to recognise an earlier schistosity rotated about these later-Caledonoid axes. It is believed to be the earliest schistosity (S_{A1}) and not S_{B1}, since no cross-folds or lineations have been found. Other early minor folds, wrapped round later-Caledonoid examples, are described under "Later-Caledonoid Structures".

North of the Dee the pelitic rocks are more highly metamorphosed and approach the condition of permeation gneisses with a distinct foliation. No primary structures are believed to have survived this transformation.

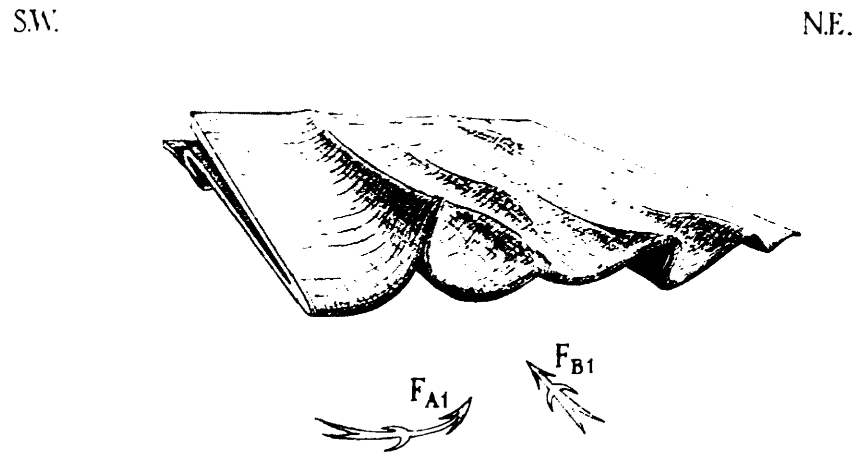
Early Cross-Fold Structures (F_{B1}, S_{B1})

No major structures of this age have been recognised. Smaller folds and lineations, being subject to the effects of later folds, do not always lie at 90° to the Caledonoid direction, so that the term "early cross-fold" is, in some cases, a relative one.

Nevertheless it is still true that many such phenomena can be recognised by this trend, usually accompanied by a slightly steeper plunge than is found in Caledonoid trends. The style of folding is recumbent, but less compressed than that of primary folds. The limbs are often stacked vertically.



a



b

Fig. 13. (a) Diagrammatic section across the summit of Creag Choinnich.
(b) Diagram of deformed limestone as in (a).

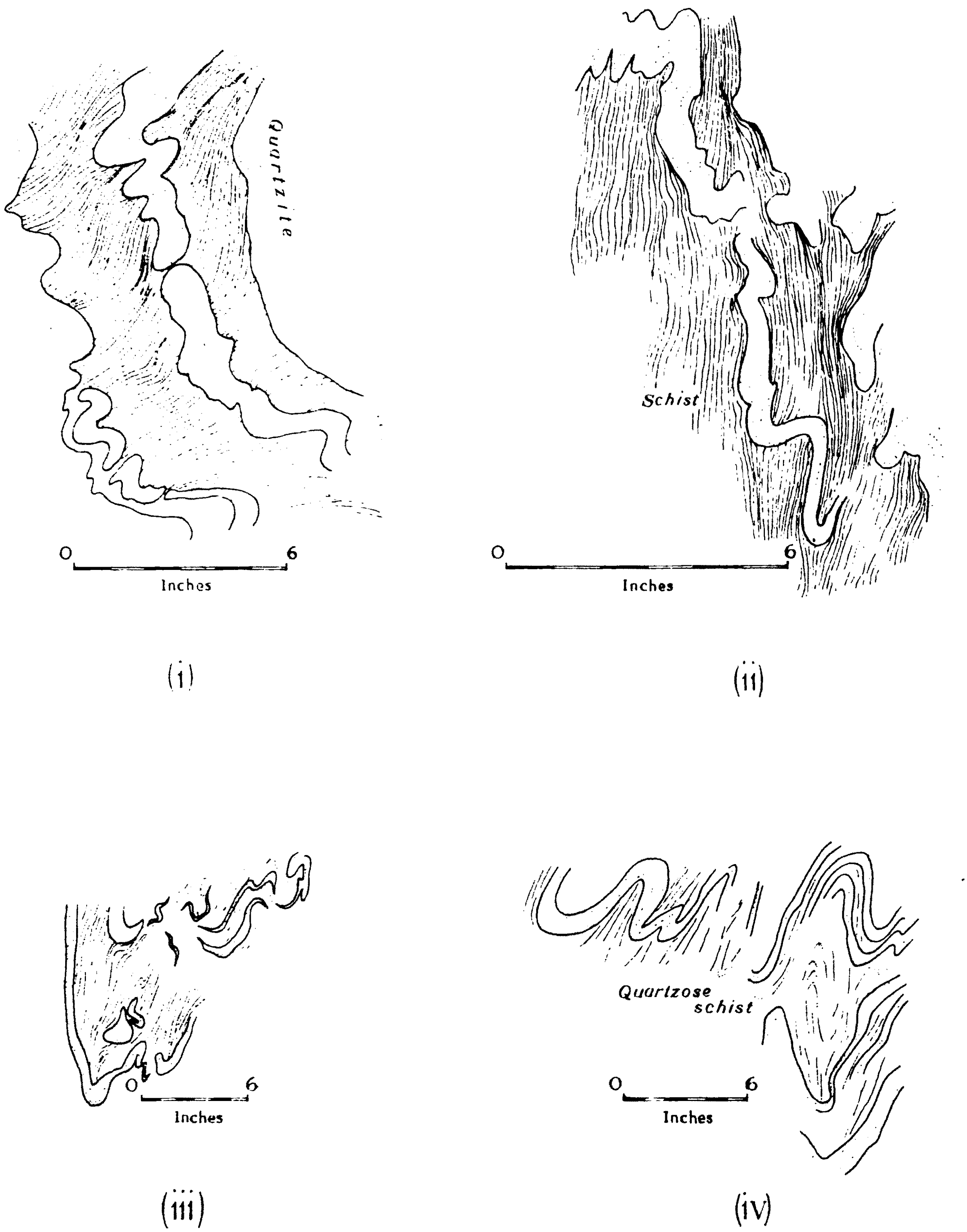


Fig. 14. The style of F_{B1} minor folds in semipelites.

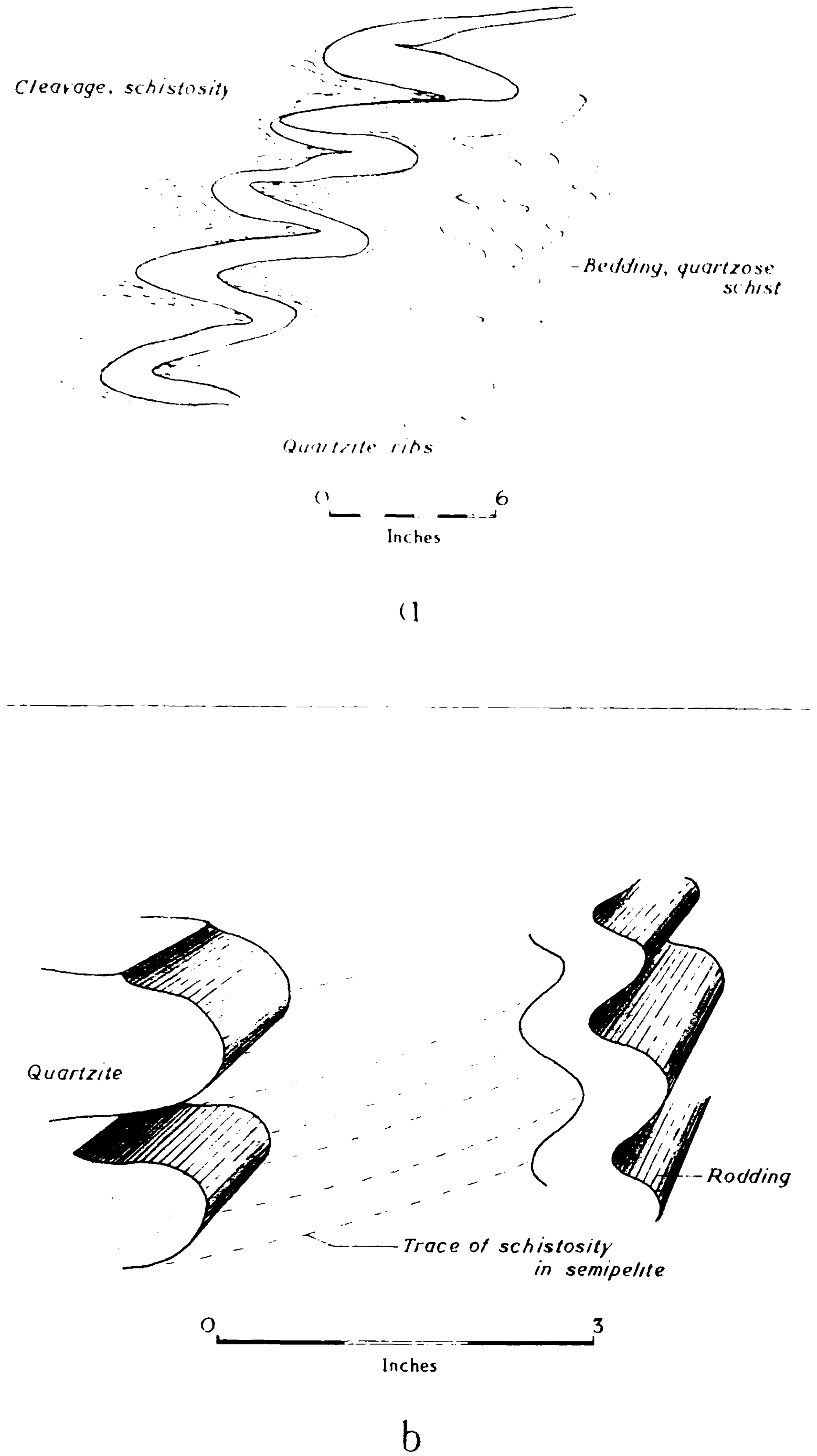


Fig. 15. (a) The relationship of schistosity to banding in semipelites.
(b) Lineations and mullion-like folds.

(a) Lower Nappe

Typical examples of F_{B1} are to be found abundantly on parts of Creag Choinnich, particularly in the rusty quartzites and quartzose schists around the summit of this hill. Here one can study the relationship of minor and mesoscopic cross-folds, for this zone represents a sharp warp in the otherwise gently dipping strata.

This structure, a broken antiform running past the southern side of the summit down towards Corrie Feragie, on a 160° trend, has tilted the axial planes of the minor folds, all of which have axes parallel to this trend (fig.13a). This implies a somewhat later age for the main structure; yet it must be F_{B1} since it is related by a complementary synform under the summit of Creag Choinnich to smaller F_{B1} folds in limestone (fig. 13b), further described below.

The style of deformation is like that of more-open primary folds (fig. 14). The amplitude of the minor ones is usually less than a foot, with tougher bands typically being thickened towards the hinge region (pl.24). The schistosity is commonly parallel to the banding except at the closures where it is an axial planar one (fig.15a). Within the banding some evidence of earlier folding has been found, particularly in the quartzite ribs. However in truly pelitic schist domains no such evidence remains, the new schistosity having taken over completely.

Cross-folding is most readily appreciated at the junction of quartzite and semipelites, where mullion-like folds may be correlated with the schistosity S_{B1}. Lineations in the massive quartzite agree precisely with the folds proving that they are lineations in b relative F_{B1} and not Caledonoid a-lineations (fig. 15b).

Deformation of the limestone of the lower nappe is not like that of the schist. The typical ribbed limestone has but little imprints of this phase, though it shows an abundance of later-Caledonoid folds.

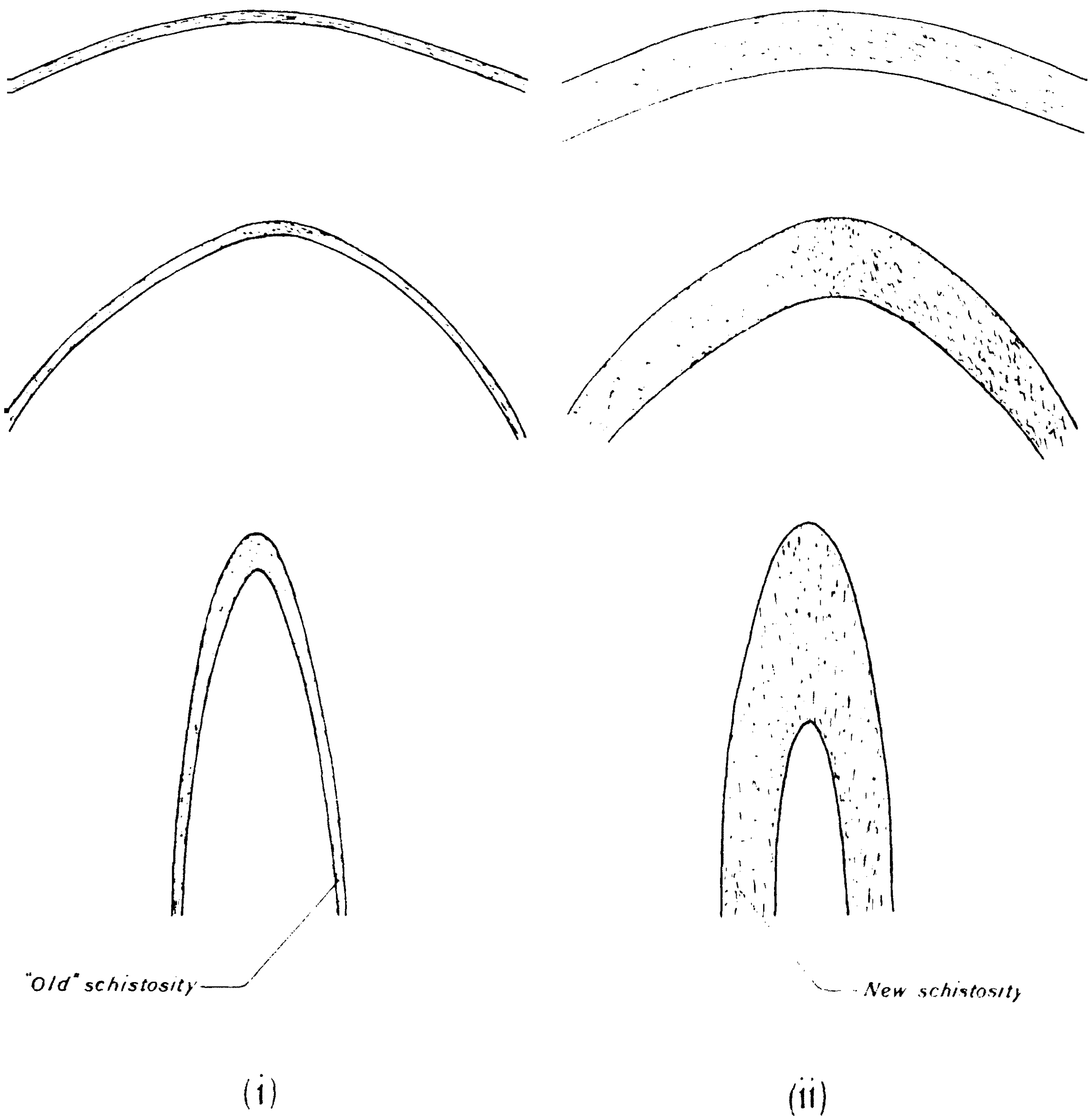


Fig. 16. Diagram illustrating the behaviour of schistosity during the folding of thin and thick bands.

The massive limestone of the summit, in contrast, shows cross-folds only. There is no ribbing; instead, a noticeable cleavage is deformed by larger, more open folds, ranging over ten feet in amplitude. These folds "bridge the gap" between truly minor and larger mesoscopic ones. The various limestone patches or inclusions in the granite at the summit of Creag Choinnich represent antiformal hinges, plunging to the southeast. The axial planes tend to be vertical, but some dip steeply to the southwest, and the limbs become isoclinal.

The granite of the summit region cuts across the fold-hinges, substantiating the notion that they are early cross-folds (see "Petrofabrics"). In addition the thermal metamorphic mineral assemblages in the adjacent pelites are little deformed, indicating that the deformation is early.

Other mesoscopic folds on cross-fold axes with steep axial planes have been recognised, with corresponding minor structures. The trend of some of these suggests a gradual swing in the general cross-fold direction.

As regards the schistosity S_{B1} on Creag Choinnich, it has been found well developed in the absence of any visible folds. This planar structure is recognised by its obliquity to lithological bands and by its steeper attitude.

It has been found, though, that when the schistosity is developed in thin pelitic bands it is nearly parallel to banding: in fact the thinner the bands the more the schistosity approaches their attitude. Often, too, the quartzitic bands are irregular, and contain evidence of boudinage instead of folding.

Sometimes, as at locality (map 3; 017,087) where two schistosities are present their mutual relations may be demonstrated. Here a sequence of semipelitic rocks is affected by tight minor folds, banding being nearly parallel to the schistosity, except in the thicker pelites, where

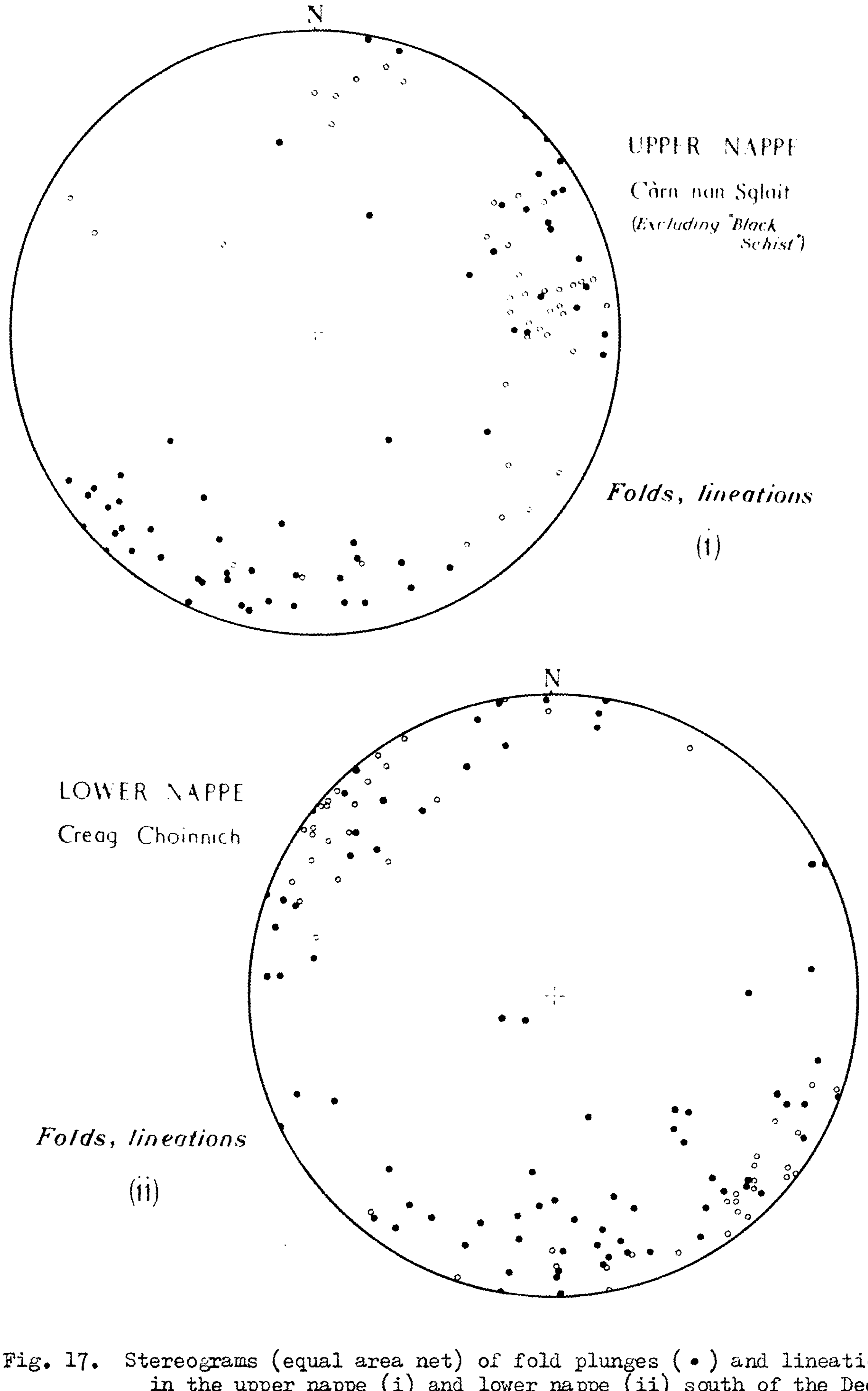
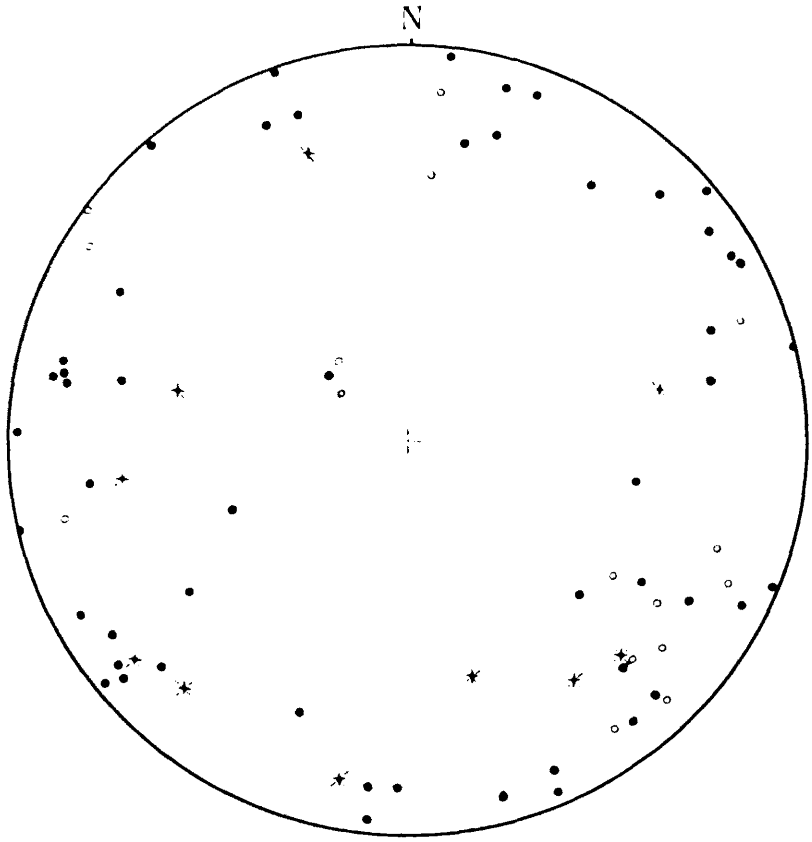


Fig. 17. Stereograms (equal area net) of fold plunges (•) and lineations (∘) in the upper nappe (i) and lower nappe (ii) south of the Dee. In (i) Caledonoid axes predominate, in (ii) cross-fold axes predominate.

NORTH OF THE DEE

(Excluding "Black Schist")



Fold plunges

Fig. 18. Stereogram (equal area net) of fold plunges (•) and lineations (∘) north of the Dee, showing possible very late folds (✱).

the two diverge by over 20°. In the tightest folds the schistosity is an axial plane one, but parallel to the quartzitic bands away from the fold closures. It is inferred that this outcrop illustrates the cleavage S_{A1} being folded and becoming at tight closures a new axial plane structure (S_{B1}). In the limbs of thin schist bands the original S_{A1} remains parallel. When the limbs of tight folds are mutually parallel the new S_{B1} is also parallel, but in the thicker bands of schist with diverging limbs, the new S_{B1} naturally becomes oblique to the banding. There are thus two schistosities parallel to lithological banding in two quite separate cases (fig. 16).

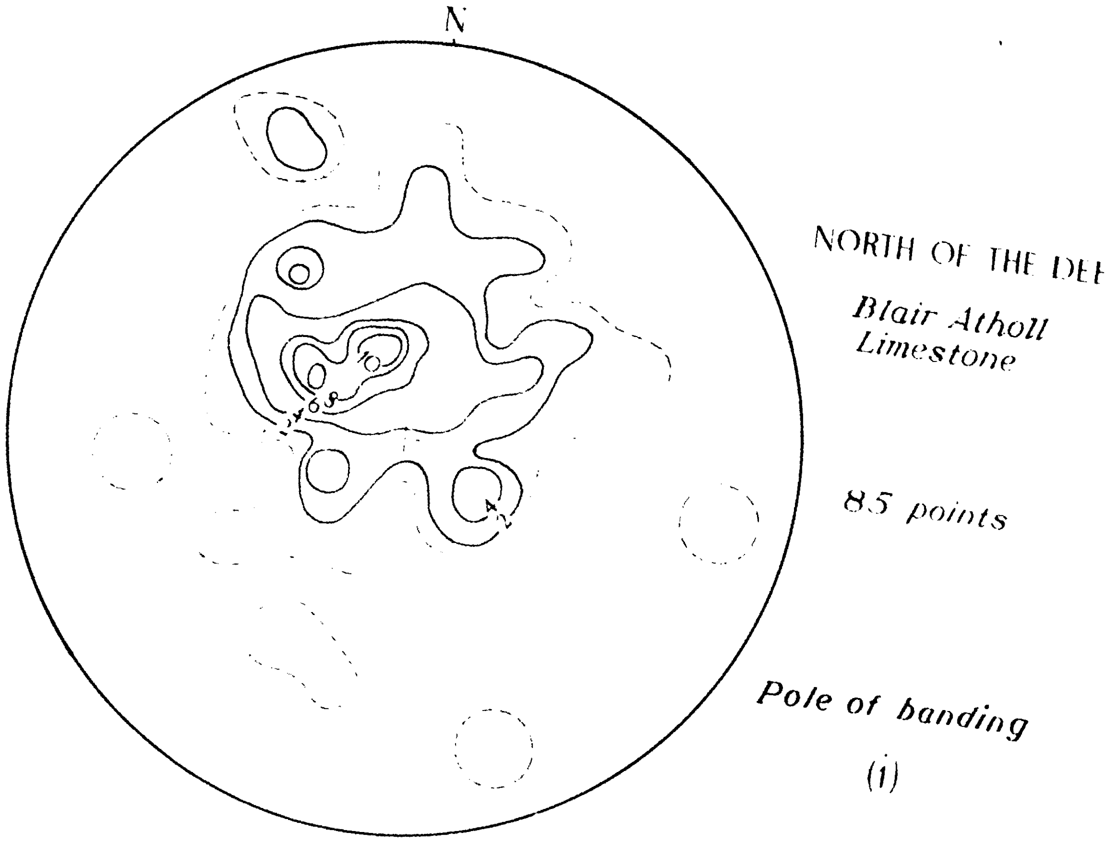
North of the Dee, on Ant-Sron, the extreme variation of both planar and linear elements of the mesoscopic structures is almost certainly due to the interplay between folds on both Caledonoid and cross-fold directions.

As may be recalled, small folds, mullions, and lineations in the Perthshire Quartzite, exposed along the bottom of Craig Leek, are clearly earlier than the later-Caledonoid warps, and trend roughly at right-angles to them. There can be no doubt that the earlier are manifestations of F_{B1} movements.

No unique axial direction for F_{B1} is obvious from a statistical treatment of measurements made north of the Dee. This is probably the result of the F_{B1} deformation being imposed on already folded layering, or from the "scattering" effect of later folds (fig. 18).

(b) Upper Nappe

The upper nappe has not revealed many cross-folds. The effects of deformation are still preserved, however, in the limestone of Càrn nan Sgliat at one or two localities. A regular fine lineation crosses the surfaces of the later minor folds (F_{A2}) at 90° to their axes. This linear feature is a miniature crumpling of the surface, and it may accompany minor folds of identical trend which produce interference



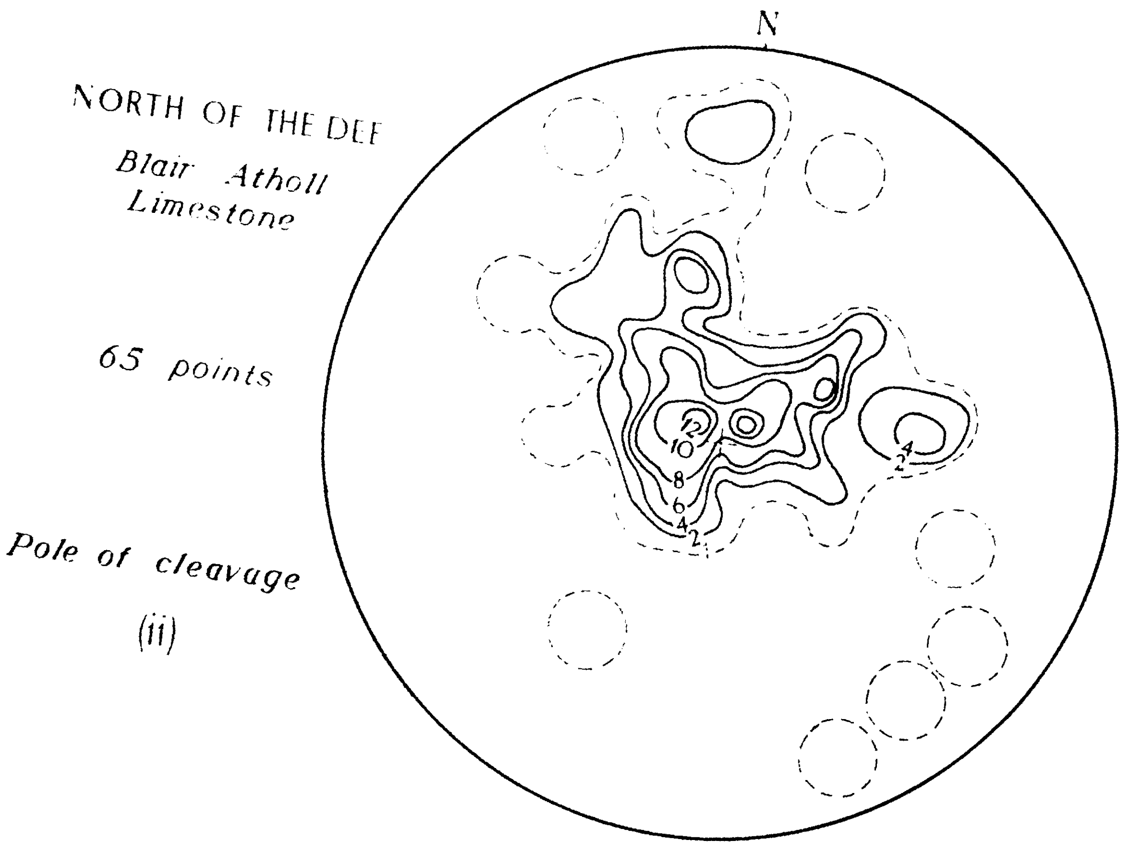
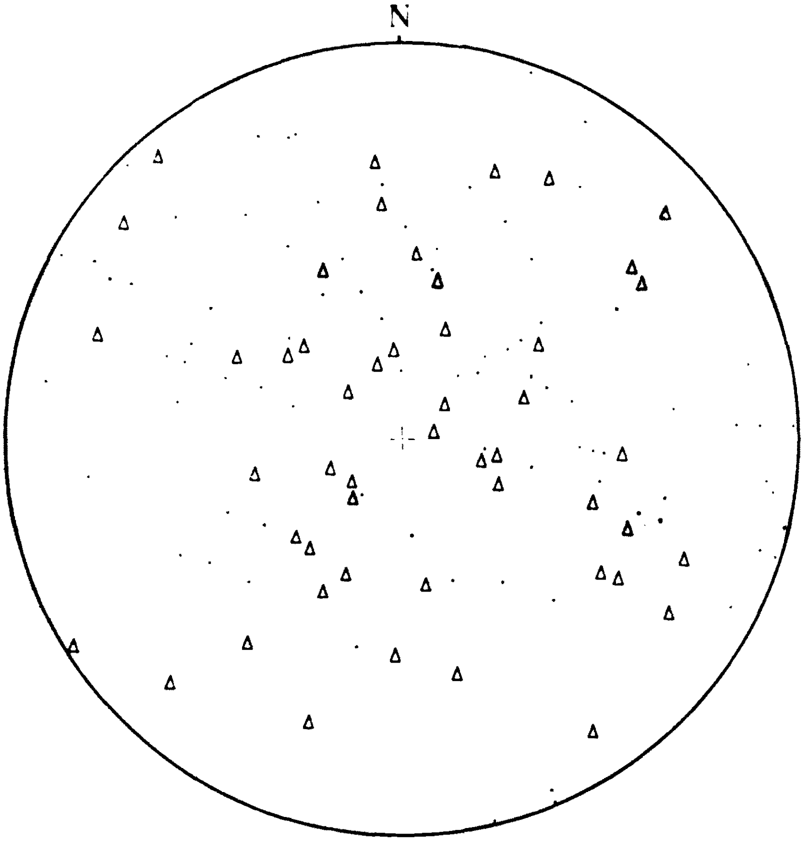


Fig. 19. Stereograms of poles to banding (i) and cleavage (ii) in the Blair Atholl Limestone, north of the Dee.

NORTH OF THE DEE

Perthshire Quartzite



Poles to cleavage and bedding

Fig. 20. Stereogram (equal area net) of poles to cleavage (Δ) and bedding (·) in the Perthshire Quartzite, north of the Dee.

Plate 25. Interference patterns in limestone caused by the presence of both F_{B1} and F_{A2} minor folds.

Plate 26. Small F_{B1} fold closures wrapped round later (F_{A2}) fold.

Plate 27. Large recumbent F_{B1} fold in the limestone of Craig Leek. Minor folds in the upper limb are apparently overturned to the left, those of the lower limb to the right.

patterns (cf. Ramsay 1962) with the later folds (pl.25).

Usually it is clear that the cross-folds are the earlier, especially when the two sets are oblique to one another (pl.26).

North of the Dee refolding relationships are not obvious in limestone. The development of cleavage is more marked, and it would appear that the various minor folds on different axial directions may be related to this same cleavage. Indeed, it is suspected that many folds are of the type shown by King and Rast (1955, p.259), whose examples have been selected from the area just southwest of Braemar. The flat-lying cleavage in the purely "schistose" limestones here is believed to be S_{A2} rather than S_{B1}.

Nevertheless one larger fold has been observed in the banded limestone at the north end of Craig Leek. Its cross-fold alignment is obvious by direct observation of the nose on two cliff-faces (pl.27). Parasitic folds on the limbs are consistent in style and illustrate the danger of interpreting the general direction of tectonic transport from minor folds alone.

A contrast in the attitude of the larger folds is to be found on Churter's Chest. An isolated outcrop of quartzite, apparently surrounded by Ben Eagach Schist and exposed in a cliff-section, can be seen to wedge into the cliff. It represents a fold closure standing almost vertically. Cleavage and fold-mullions are also vertical, the axial planes striking along the Caledonoid direction. This cross-fold has thus been superimposed on steeply-dipping strata, or has itself been refolded by later-Caledonoid deformation.

Later-Caledonoid Structures (F_{A2}, S_{A2})

Major structures of this age can be difficult to distinguish from earlier ones. Only one has been recognised. It is responsible for the

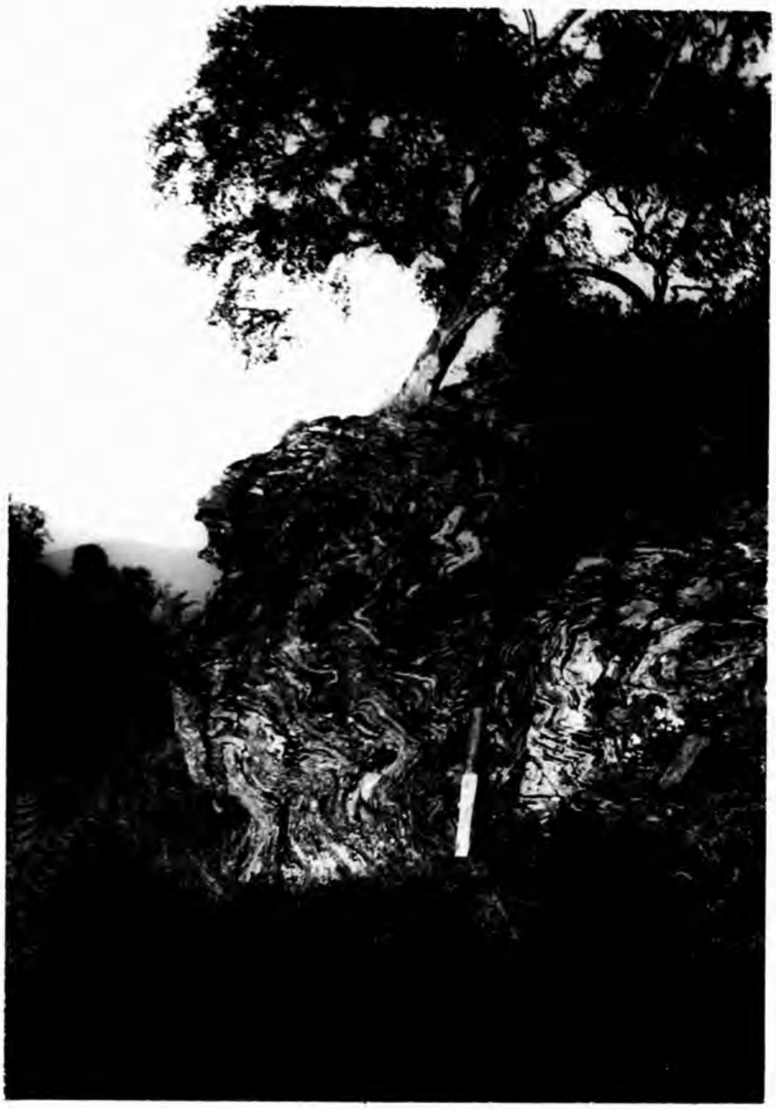


Plate 28. F_{A2} minor folds in Blair Atholl Limestone, Càrn nan Sgliat.

Plate 29. Close-up of part of plate 28.

Plate 30. Folded and unfolded domains in ribbed limestone, Càrn nan Sgliat.

vertical attitude and Caledonoid strike of the lithological units and schistosity at Sron a Bhruic where both the Ben Eagach Schist and Perthshire Quartzite are affected.

Intermediate-sized folds are similarly difficult to recognise. However, the distribution of limestone bands on the north-west side of Creag Choinnich can be seen in the field to be the result of F_{A2} folds lying axially parallel to the main contours of the hill, such that minor topographic variations have a marked effect on the outcrop pattern.

Other outcrop patterns, particularly in limestone, can sometimes be visualized only as resulting from the combination of two phases of folding, the attitude of the planar elements indicating that Later-Caledonoid folding has played an important part.

Minor folds are abundant in the ribbed limestone belts in both nappes, the style in the two being similar. The limbs are often recumbent, stacking near-vertical, but the folds are not highly compressed. They are thus similar to the minor cross-folds. The local direction of plunge may be variable (due to accommodation) but the angle of plunge is consistently low, unlike the cross-folds.

Where the ribbing is fairly even and consistent good minor folds tend to be prolific, as seen in a typical exposure (pl.28 ; close-up pl.29). In such a condition it is hard to find evidence of earlier folding in the limestone, but a few (possibly primary) minor folds have been recognised. When the F_{A2} fold-domains pass, as they may do rapidly, into unfolded ones (pl.30) it becomes less difficult to appreciate earlier structures. For instance in plate 23 there is clear evidence of earlier folding immediately above and below the ruler. The photograph also reveals boudinage structures (top left) and an F_{A2} minor fold (bottom right).

Due to variation in the competence of certain bands or zones the deformed limestone may exhibit an oak- or maple-leaf style of minor folding

Plate 31. Oak- or maple-leaf style folding.

Plate 32. Tectonic inclusions (some still defining folds) in deformed limestone.

Plate 33. Amphibolitic bodies in folded massive limestone, River Clunie.

Plate 34. Folded semipelite with fracture-cleavage in thicker quartzite band.

(pl. 31). Where, however, the tougher (less-pure) bands were originally infrequent or very thin they are now disrupted and sometimes no more than isolated tectonic inclusions (pl. 32). This is true also of epidioritic or amphibolitic bodies in massive limestone (pl. 33).

Rocks adjacent to the limestone belts often show less obvious evidence of the deformation.

In true pelites few examples of F_{A2} have been discovered, but in other banded rocks or semipelites excellently preserved minor folds can be studied. The most illustrative of these lies in a belt of rocks above the limestone on Càrn nan Sgliat at locality (038,068) (pl. 34). Full descriptions, including petrofabric data can be found on page 46 . Briefly, it may be described as a recumbent minor fold affecting a thin band of quartzite within pelite. It is apparently overturned to the southeast, with an axial plane dipping gently to the northwest. The quartzite band further exhibits fracture-cleavage, and there is reliable evidence of smaller folds within the band which may be correlated with others outside which are obviously related to the main fold. Certain irregularities are believed to reflect an earlier deformation.

This example shows precisely in miniature the deformation with the development of dislocations in tougher bands that must have occurred on the more major scale.

A more complex fold-sequence can be appreciated at (111,144) on Craig Leek in the banded schist, where quartzite bands have behaved differently from tough grey schist bands within true pelitic schist. The quartzite bands within which are early highly recumbent folds, are more tightly folded than the tough grey bands, and plunge 10° towards 55°. The schist folds plunge 25° towards 145° and in some cases have tight folds wrapped round them. Clearly the quartzite and schist bands are at a different stage of tectonic evolution, yet though in close proximity

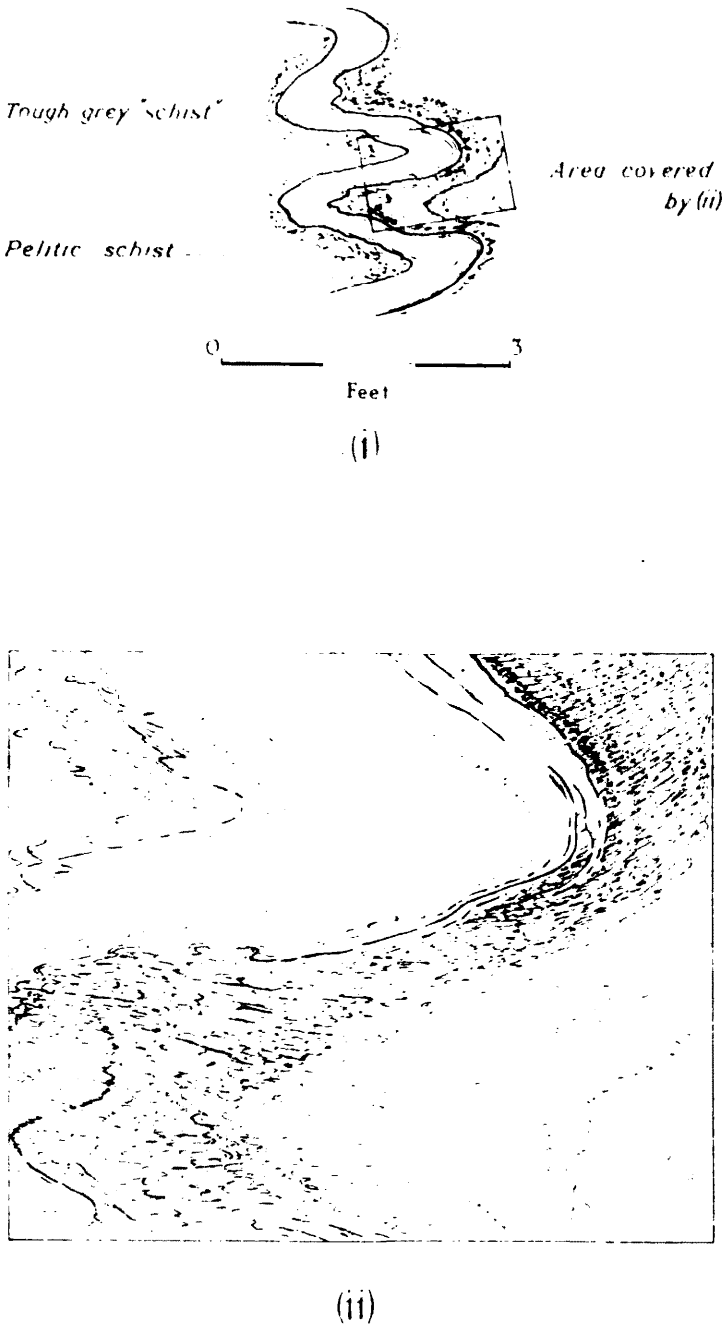


Fig. 21. Sketch-diagram of the schistosity associated with folds in Blair Atholl schist.

Plate 35. Folds in foliation of hornfelsed Blair Atholl Schist, with axial plane schistosity.

Plate 36. Later folds, deforming the schistosity shown in plate 35.

neither set of folds affects the other and so they could be inferred to be contemporaneous. Since the main folds in the schist are not primary one is forced to assume that the quartzite exhibits Later-Caledonoid deformation, and the schist a Later-Cross-Fold one (F_{B2}).

Yet another interesting feature is the schistosity associated with the tough grey schist, which, being folded on a smaller and tighter scale, is in the process of becoming a new axial plane schistosity. Its appearance is that of a strain-slip, so typical of much of the Blair Atholl and Ben Eagach schists elsewhere: see fig. 21.

If one now traverses northwest into the Blair Atholl Series proper, it is obvious that the schistosity in these "hornfelses" is simply a more advanced stage of the previous, the earlier chevron-type microfolds being now only irregularities in the new schistosity which cuts across the foliation (pl. 35). Later warps can also be found, plunging 25° towards 260° (pl. 36). These clearly deform the new schistosity.

This situation is a very important link between the deformation in the widespread hornfelses and the more easily identifiable folding in banded or bedded rocks where fold-sequences are better known.

Where, however, two equally developed schistosities co-exist the problems of resolving the deformation history become complicated, and rely more upon the study of mineral assemblages.

<u>Later-Cross-Fold Structures</u> (F_{B2}, S_{B2}). <u>Latest Structures</u>

Microscopic examination of thin sections of schists and hornfelses has established that thermal metamorphic mineral assemblages have formed after the later-Caledonoid main phase of deformation. These minerals are themselves deformed - often considerably so - and it is believed that later folds, probably cross-folds, were the principle agents.

This deduction is strengthened by the observation of several post-schistosity minor folds of variable intensity, and largely on cross-fold axes, in both hornfelses and limestones.

Some of these have already been described (p.29) in part. Others, which can be related directly to mineral assemblages, have been found north of the Dee at locality (154,178). They are recumbent Z-shaped folds with horizontal axes aligned towards 110°. They deform the pronounced foliation, preserving within it evidence of earlier deformation yet creating a new axial plane schistosity. The thermal metamorphic mineral assemblage (post-regional metamorphic assemblages) is distinctly earlier than the axial plane schistosity, so that one can say definitely that there has been a post-F_{A2} deformation in the area.

At the same locality the axial plane schistosity is affected by later warps plunging 10° towards 230°. These are thus the latest folds known. Similar post-schistosity buckles are known from other hornfels localities, where they may be accompanied by a closely-spaced jointing parallel to the axial planes.

The major equivalents of the Later-Cross-Fold Structures cannot be easily recognised. They may be represented by culminations or depressions of pitch along the Caledonoid trend, such as were recognised by Bailey (1925) in the Central Highlands. His map shows these features occurring along the line of the Ben Lawers Synform between Loch Tummel and Glen Shee. These weak structures have a cross-fold trend, but they do not match the intensity of the minor ones just described.

Around Braemar a depression occurs in the upper nappe at the Clunie, followed by a culmination just south of the Lion's Face, in line with the thick limestone belt, and then by a depression between the Dee and Ant-Sron. All of these are too ill-defined to be readily appreciated by statistical analysis of minor folds. However, from suitable vantage points in the field it becomes evident, for instance, that structural levels in the upper

Plate 37. Thick antiformal mass of limestone: view looking N.W. from quartzite mass of Craig Leek.

Plate 38. Small dome (hammer at centre) and synform (foreground) in limestone, Ant-Sron.

nappe show a gradual inclination downwards towards the River Clunie from both directions.

There may be a correlation between late mesoscopic warps and thick masses of limestone. For instance in the northwest corner of the area a distinctly thickened mass of limestone, forming a "spur", can be recognised in the field as a gentle antiform on a cross-fold axis (pl. 37). In this case (a more massive limestone), the cleavage is deformed, but the structure cannot be traced into the above-lying quartzite.

Minor warps in limestone have also been encountered, but unlike the hornfelses, limestone does not usually lend itself to the relative age-determination of the mineral assemblages, so that deformation structures cannot be correlated with those of pelitic rocks. Plate 38 illustrates a small dome (hammer at centre) in which the layering or banding but not the cleavage is folded, while immediately in front of it (foreground) is a synform with banding and cleavage equally deformed.

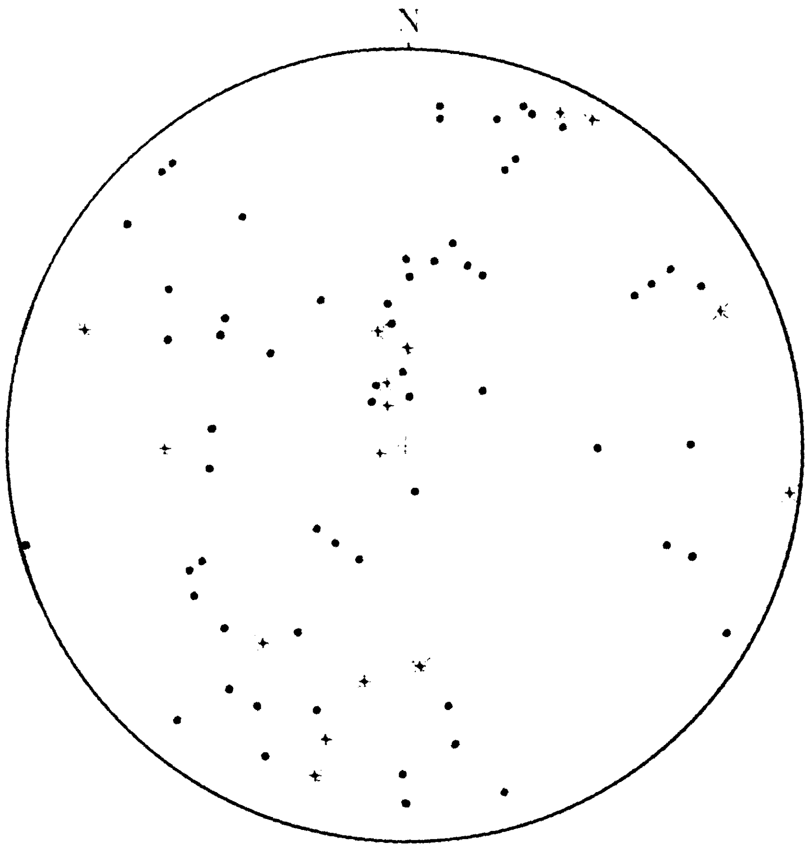
Structure of the "Black Schist" of Càrn nan Sgliat

The uniform lithology of this formation prevents the precise nature of the folding from being appreciated. Where the schist is in contact with quartzite, as at Churter's Chest, F_{A1}, F_{B1} and F_{A2} folds can be recognised, suggesting that even greater complexity exists in the core of the mass. This is borne out by mineralogical studies, especially of specimen CB 120, which indicate that the most obvious deformation is at least as young as F_{A2}.

The ubiquitous foliation, described already as typical of this zone of hornfelsed schist, is deformed by folds ranging in amplitude from a few inches upwards. The schistosity is poorly developed (mica content low), and when visible is parallel or sub-parallel to the foliation in most exposures, though a number of folds have a partially developed axial plane schistosity.

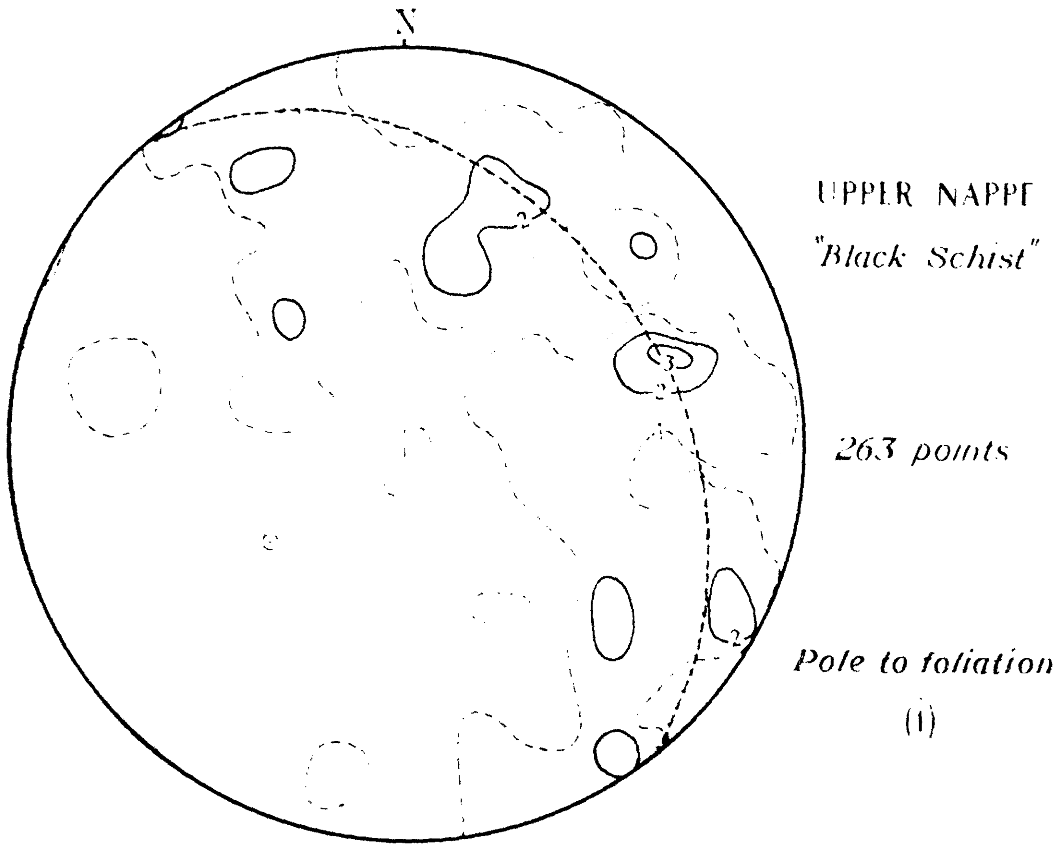
UPPER NAPPE

"Black Schist"



Fold plunges

Fig. 22. Stereogram (equal area net) of fold plunges (•) including the latest structures (✳) in the "Black Schist".



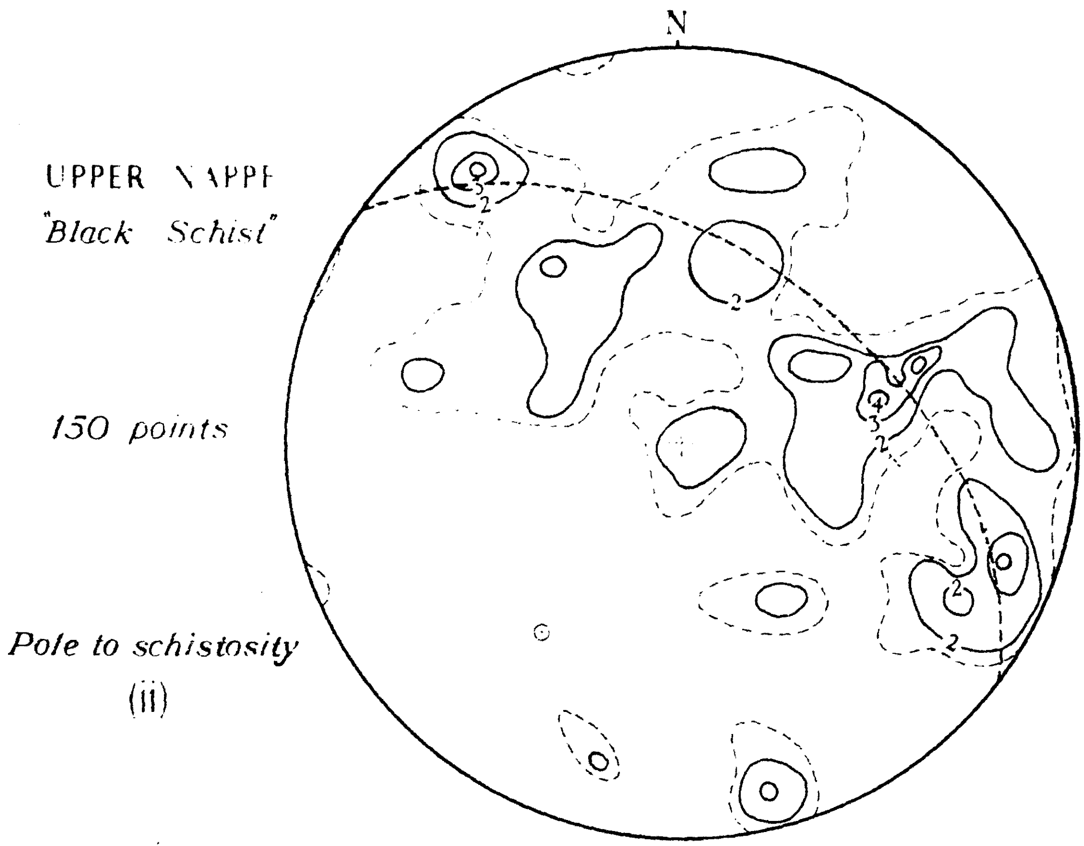


Fig. 23. Stereograms of poles to foliation (i) and schistosity (ii) in the "Black Schist". The distribution of poles in both cases corresponds with a great circle (dotted line, pole ⊙).

Plate 39. Refolded fold in foliation of the "Black Schist", Càrn nan Sgliat.

Plate 40. Two adjacent folds with different axial directions (top right).

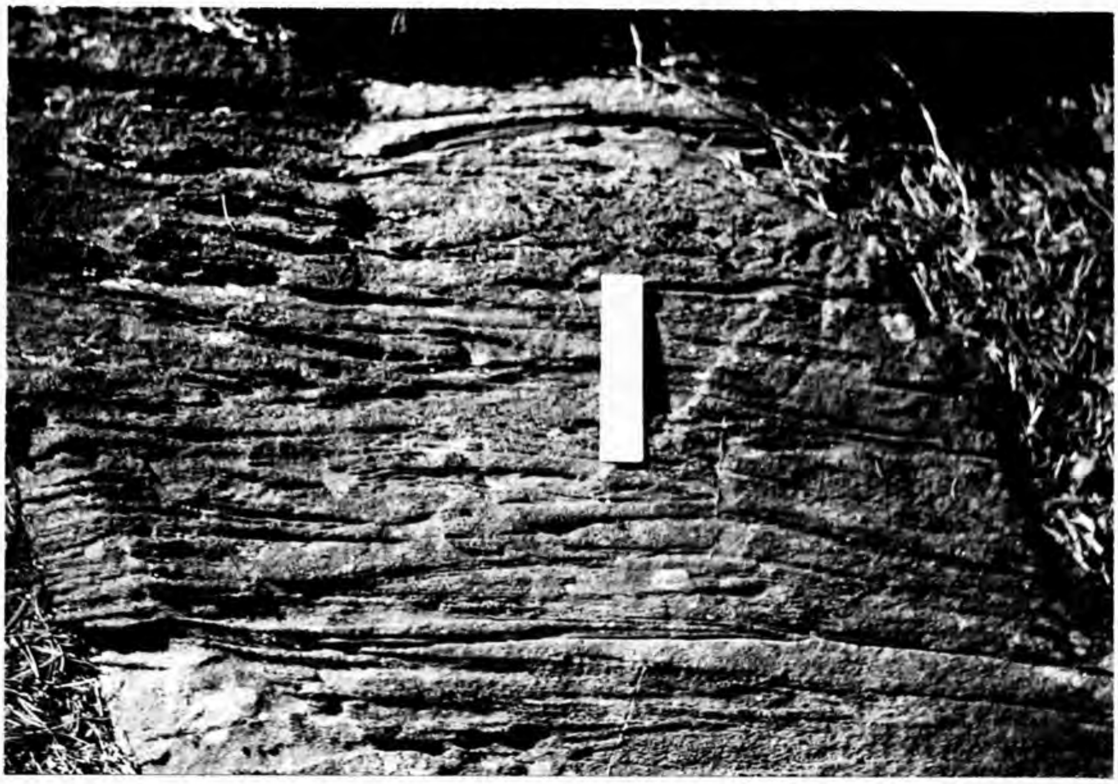


Plate 41. Evidence of early folding in the irregularities of the foliation.

Plate 42. A typical shear-fold in the foliation of the "Black Schist".

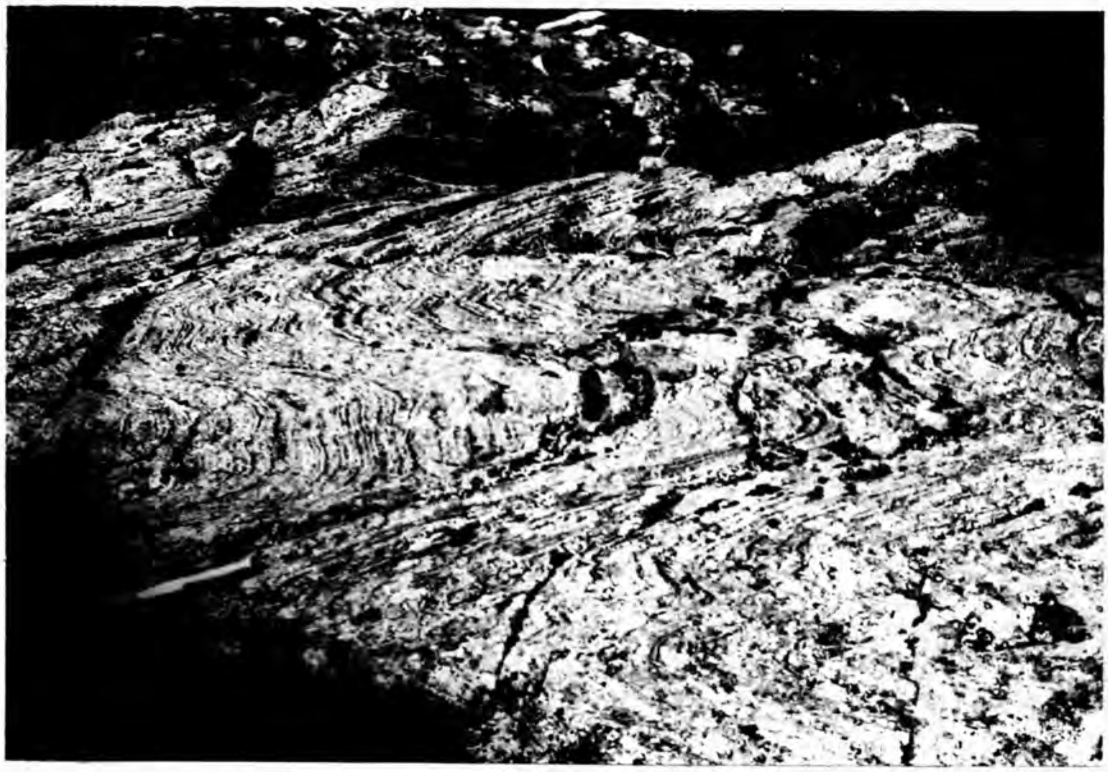


Plate 43. The similar nature of the folds continues across many layers.

Plate 44. Part of the area in plate 43, illustrating the composite nature of the folding.

Plate 45. Part of a large, and possibly later, concentric fold.

Plate 46. A concentric fold with a more brittle type of deformation.

Plate 47. Chevron-like folding associated with close-spaced jointing.

Plate 48. Shear-folds passing into chevron-like folds.

Almost all directions of fold-plunge, and almost all attitudes of foliation have been recorded (figs. 22 & 23). In addition, refolded folds of the type shown in plate 39 are extremely rare, yet folds of widely different axial directions may lie in close proximity (pl. 40). Consequently no reliable deformation history can be established.

There is limited evidence of relatively early folding (pl. 41), most of which, it is suspected, has been destroyed by the subsequent highly penetrative quasi-plastic deformation. Most of the obvious folds in the "Black Schist" belong to this phase. The commonest are shear-folds, of which that in plate 42 is typical. They are similar or parallel and the closures may be traced long distances across successive layers, as in plate 43 . In this example, shown in close-up (pl. 44), the composite nature of the folding - a common feature - is evident.

Other types, such as concentric folds (pl.45 , 46) are much sparser and may be later in origin*. This includes chevron-like folds which are associated with a close-spaced jointing often with slight displacement in the manner of a strain-slip cleavage (pl. 47). These are suggestive of a more brittle and therefore later deformation, and certainly the schistosity, when recognisable, is deformed with the foliation.

There may, however, have been continuous deformation between the earlier "plastic" and the later more brittle stage, for some shear-folds can be seen to pass into chevron-like folds in which a crude jointing is developed as just described (pl. 48). Also, the brittle stage may have been locally more intense (pl. 49).

A further reason for believing the chevron-like folds to be relatively late phenomena is that the associated micro-jointing is near-vertical and may sometimes be seen to lie parallel to the axial plane of large open flexure folds (pl. 50), found in the centre of the "Black Schist" mass

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* Note, for instance, the less-regular nature of the fold in plate 46 .

Plate 49. Illustrating the effects of intense shearing during the brittle stage of deformation.

Plate 50. A large flexure fold with jointing parallel to the axial plane.

and which plunge at about 50° towards 330°. This disposition is unlike the otherwise identical though smaller late buckles in the "Black Schist" below Craig Leek, north of the Dee; nor does it agree with the only statistical axis of deformation that can be derived from stereographic plots of all fold axes in the "Black Schist" (fig. 23).

This statistical axis, based on the general attitude of the foliation and to a lesser extent that of the schistosity, points to a relatively late deformation on a Caledonoid axis.

In summary, then, it can only be said that this belt of hornfelsed schist has suffered more than one phase of deformation.

Faults

The Dalradian rocks of the Central Highlands of Scotland are transected by large sinistral tear-faults trending N.E.-S.W. A smaller representative of this group lies just outside the area mapped, to the southwest of Braemar (see map , fig. 47), the displacement along it being of the order of one mile. Within the area no such faults have been identified. As in other Dalradian regions dislocations are relatively uncommon, either because the tectonics are too complicated to permit their recognition, or because deformation remained plastic until very late.

Evidence for other faulting has been found on aerial photographs as well as in the field. In the latter case physical features and/or the juxtaposition of mutually incompatible structures provide the main clues in the search for linear displacements. In only one instance can the true fault plane with attendant shattering be observed. Of the few faults in the area almost all lie between N-S and N.E.-S.W. in trend, and appear to be near-vertical. There is possibly a N-S set and a N.E.-S.W. set but the total number is too few to be certain of this. In most cases the precise nature of the fault is not known.

What is surely the largest fault in the area can be clearly deduced from the relationship of the Perthshire Quartzite to the Ben Eagach Schist on Càrn nan Sgliat and on Craig Leek. In the former the Ben Eagach Schist lies normally above the Perthshire Quartzite, but just north of the Dee on Craig Leek appears to lie structurally and topographically below the Perthshire Quartzite. Only a near-vertical fault with an eastern downthrow of more than 100 feet can be responsible. Such a large fault presumably is the primary cause of the cliffs here, and indeed may be responsible for the whole dry valley along which the Dee once flowed. The fault may thus curve round with the valley, or give off as a branch a smaller fault running into Ant-Sron. The smaller is recognised by the incompatible strikes of the limestone and schist on either side while the peculiar kink in its line reflects the close proximity of the two opposing masses of rocks, where the position of the fault can be fixed to within a few feet. The fault is terminated by a sharp gully - possibly the site of another fault. Also on Ant-Sron is a N.E.-S.W. high-angle fault which it is necessary to postulate to separate a huge mass of quartzite from a feature of limestone.

The other end of the main valley fault would be expected to occur on the east side of Càrn nan Sgliat, but no large fault can be found. It possibly terminates at Churter's Chest where smaller faults have been traced or it could die out along the very straight V-shaped valley (Clais Mhòr) cut into the Ben Eagach Schist and lying in line with the large dry valley. Exposures of schist at the upper end of the valley, however, prove that no displacement exists at this level. The inevitable conclusion that the main fault does not continue to the south has an important bearing on the argument for or against the existence of a ring-fault circumscribing the Lochnagar Granite.

In the quarry due north of the Lion's Face (map 2) a fault may be

Plate 51. Fault (arrowed) in the quarry-face, Creag Choinnich.

studied in detail (pl. 51). The fault-plane trends 30° east of north and dips at 60°-75° to the northwest, the downthrow on the northwest side being about 30 feet (measured vertically). On a smaller, neighbouring fault slickensides plunge 45° towards 85°. The fault zone of the former dislocation contains a crushed mass of quartzite and it is evident that schistose rocks have been locally dragged into the movement-plane. Dislocated sheets of granite may be matched up on either side of the fault.

Around the quarry jointing in the main rock-types has only local significance, but smaller faults have similar direction and behaviour to the larger, and some of these cause shearing and slickensiding in the granite.

On Craggan Rour, north of the Dee, a possible N-S fault is visible from aerial photographs as a narrow negative feature. In the field the geology alone does not reveal the fault, but as the field map (1) shows the geology is nevertheless simplified by identifying the feature as a fault even though the movement is not known. Since the fault-plane is obscured, the minor granite bodies - irregular as always - cannot be proved to pre-date the fault, but their distribution suggests that this be so in most cases.

A similar feature, lying farther east and running parallel to the cliffs, cannot be seen to affect the distribution of the metasedimentary rocks, but is considered to be a fault. Some outcrops of granite may be displaced along this line but since a sheet of microdiorite (normally pre-granite) traverses the feature cleanly it follows that this fault is earlier than some, at least, of the granites.

In spite of the many nearly circular or concentric features in the main Lochnagar Granite Complex, no direct evidence has been found around Braemar of a ring-fault.

Within the main granite margins, however, there are some interesting

Plate 52. Fault-like feature in granite (centre distance).

Plate 53. Marshy ground on the lower side of the fault-like feature.

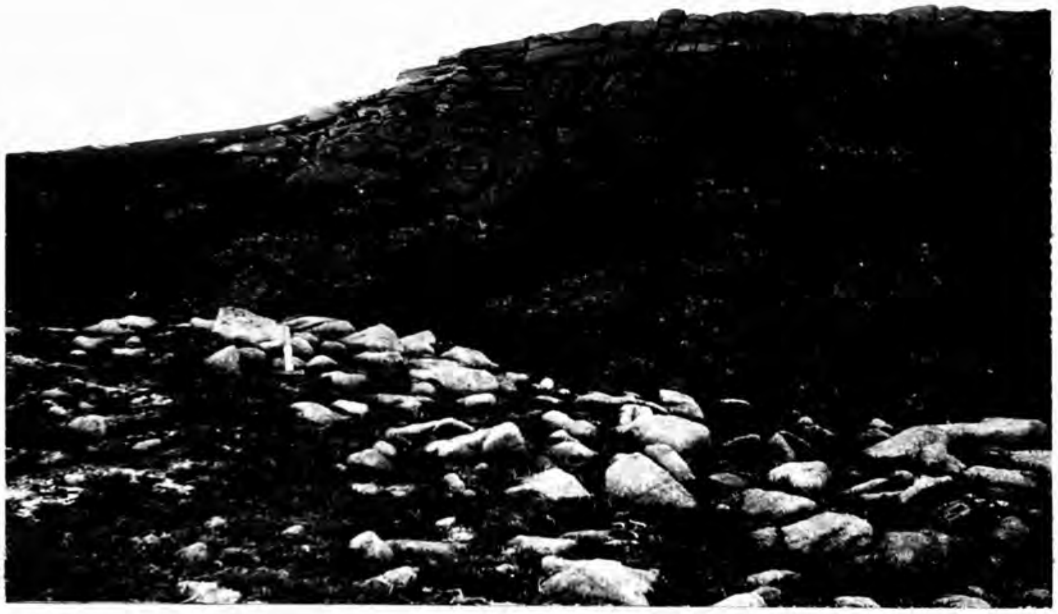


Plate 54. Light- and dark-weathering granite on opposite sides of the fault-like feature.

Plate 55. Jointing parallel to the axial plane of folds in massive quartzite.

Plate 56. Various sets of joint-planes in an outcrop of Perthshire Quartzite.

Plate 57. Sill-like mass of granite in the lower nappe, Creag Choinnich.

Plate 58. Part of detail shown in plate 57, illustrating micro-jointing in the granite.

linear structures (seen from aerial photographs). About one mile south-southwest of Sroǹ á Bhruic a set of negative features lies parallel to the edge of the mass. A smaller set lies at right angles. Of all of these only one of the latter set has several of the aspects of a fault in the granite:-

(1) It is a distinct "one-sided" feature (pl. 52), i.e. a scarp.

(2) The lower side is occupied by marshy ground (pl. 53).

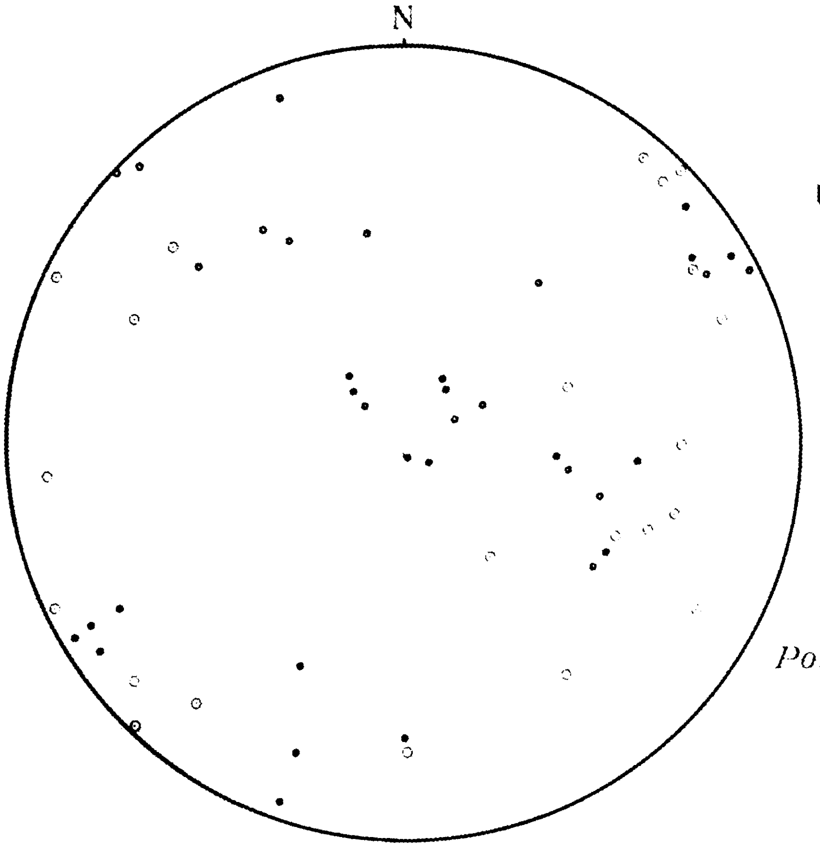
(3) It separates light-weathering granite from dark-weathering granite (pl. 54).

However, the examination of fresh material reveals no superficial petrological difference in the two granites. As no contact is preserved, and there is no evidence of crushing in the rocks near the feature, no fault can be proved. The other structures are probably part of a major joint pattern in the Lochnagar Granite.

Joints: cleavage in igneous rocks

As already described, joints may be developed parallel to bedding planes and to the axial plane of folds in quartzites (pl. 55) and quartz-rich rocks. However, joints, minor fractures, cleavage etc. that have no obvious relation to folding have been found in igneous rocks, as well as in quartzites. In plate 56 the Perthshire Quartzite of the upper nappe is seen to have several sets of close-spaced joints. The hammer and ruler rest on a joint-surface which is almost certainly a bedding-plane, crossed by a lineation. The following illustration (pl. 57) is of a sill-like mass of granite invading gently dipping quartzite and quartz-rich schist in the lower nappe on Creag Choinnich. A close-up view in the vicinity of the hammer (extreme left) reveals a pronounced micro-jointing (pl. 58).

Readings of the attitude of these planar elements have been collected

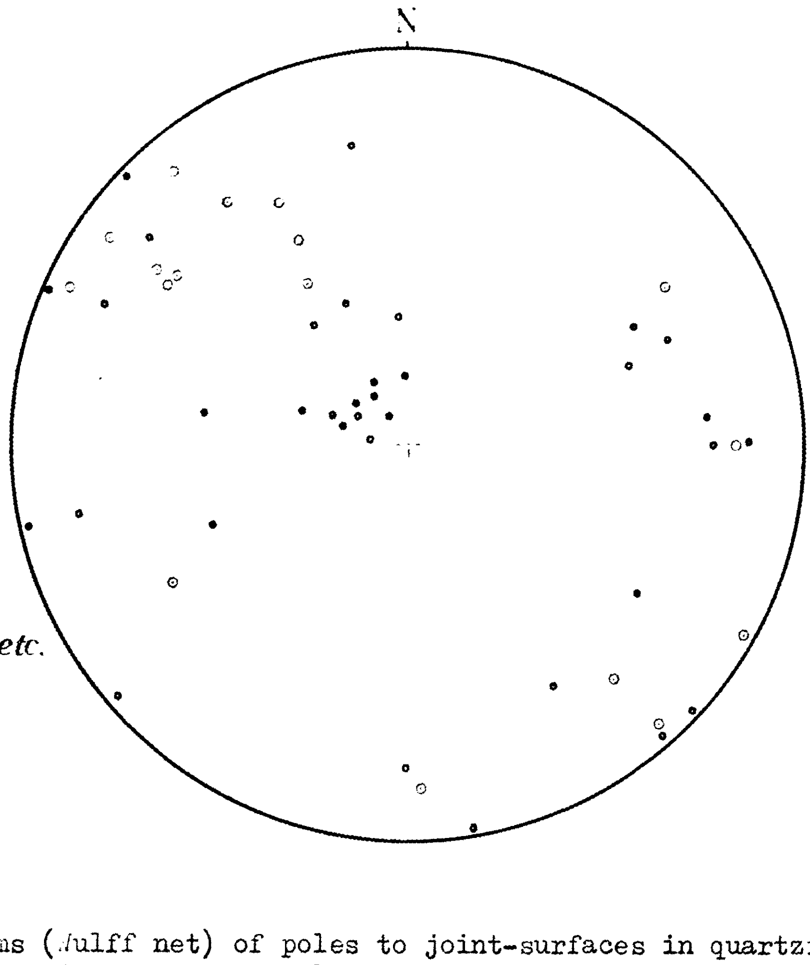


UPPER NAPPE

Poles to jointing etc.

(i)

LOWER NAPPE



Poles to jointing etc.

(ii)

Fig. 24. Stereograms (Wulff net) of poles to joint-surfaces in quartzite (⊙) and joint-surfaces, "cleavage" or foliation in igneous rocks (∘) in the upper nappe (i) and lower nappe (ii).

Plate 59. Illustrating the effect of high-angle joints on outcrops of quartzite, Càrn nan Sgliat.

Plate 60. A "felsite" affected by high-angle joints, Càrn nan Sgliat.

Plate 61. "Cleavage" or fine foliation in a body of "felsite".

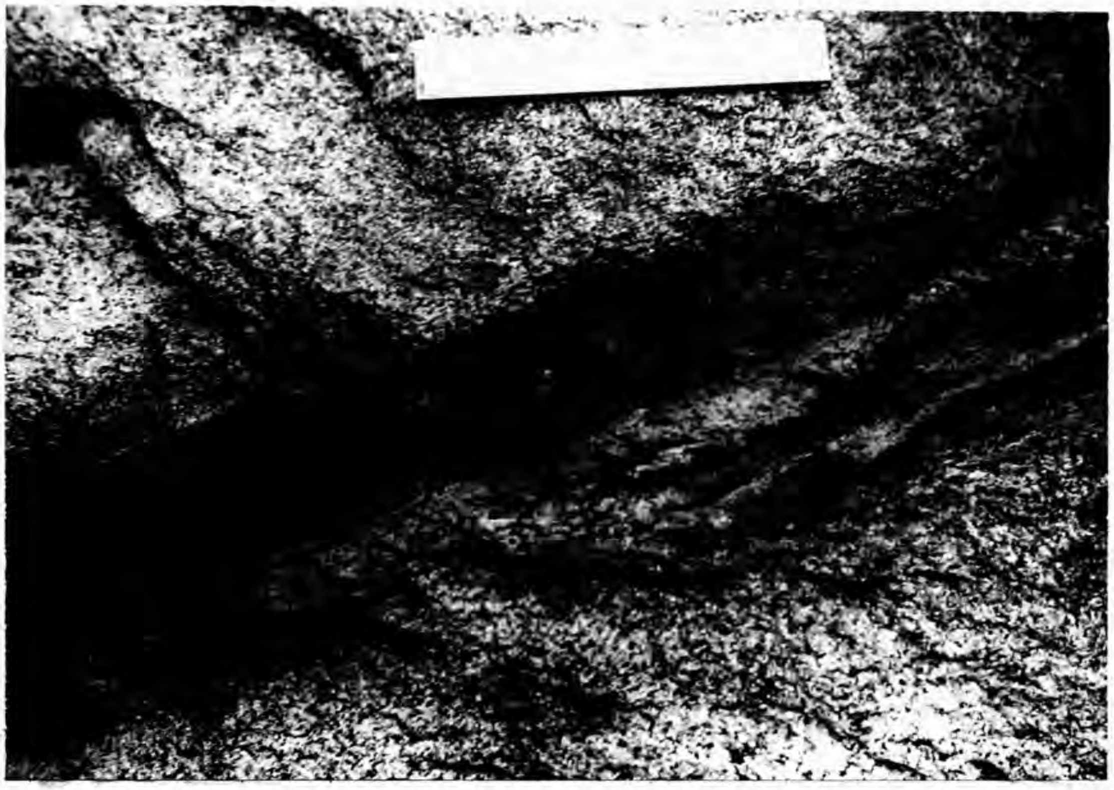


Plate 62. Crude cleavage in granite adjacent to inclusion of schist.

from both nappes. Plotted on stereograms they reveal a similar distribution pattern, and presumably therefore have had a similar origin.

Figure 24ii (lower nappe) and figure 24i (upper nappe) summarise the available readings (to the nearest 5^{o}), those from igneous rocks - but not including true epidiorites - slightly outnumbering those from quartzites. In addition readings of "cleavage" or alignment of platy minerals in igneous rocks (mostly granites) have been recorded.

While there is a considerable spread of planes, both diagrams show the joints to be near-vertical and to have a tendency to fall into two roughly equal groups, one striking between 320^{o} and 350^{o}, the other between 10^{o} and 60^{o}. The two sets thus intersect at approximately 60^{o}. There would appear to be a very slight difference in orientation between the two nappes but no significance can be attached to it. Plate 59 , looking northeast along the lower levels of the upper nappe on Càrn nan Sgliat, illustrates the effect of some of these high-angle joints on the outcrops of quartzite; while plate 60 shows their effect on the felsites from higher levels in the same nappe.

The "cleavage" in the igneous rocks, while varying in strike, dips at usually less than 30^{o}, and generally towards the south-southeast. It is thus nearly perpendicular to both sets of joints mentioned above. In granites it may be a flow alignment of dark minerals, or a deformation phenomenon. It has also been recorded, though rarely, in felsitic rocks (pl. 61).

While intensely sheared exposures of granite have been encountered no distinct shear zones or shatter belts can be recognised. The deformation is believed to be related to the main phases of folding (see "Petrofabric Analysis"), though very locally the granite may take on a cleavage parallel to that of adjacent inclusions or contacts with schist (pl. 62).

B. PETROFABRIC ANALYSIS

Introduction

The Perthshire Quartzite was first selected for study in an attempt to compare deformation in the two nappes. However the contoured stereograms produced by standard techniques revealed considerable variations in the quartz-fabrics of the various specimens.

It was subsequently decided to analyse the granite, which occurs in both nappes, in the hope that an understanding of its fabric would help to determine the nature of the older quartzite fabrics. However the variations in the granite were found to be as marked as those in the quartzite, while the stereograms appeared more complex. Further analyses to discover if the granite was of multiple intrusion led to the idea that granites of different characteristics and environments varied in their susceptibility to deformation, and that some may have inherited quartzite fabrics as a result of assimilation of the quartzite, or because the granite originated by replacement of quartzite.

Lastly it was attempted to link the fabric of deformed semi-pelitic rocks directly to the folds concerned by analysing samples from a selected minor fold. In this instance mica fabrics were also studied.

Techniques

Where granite specimens were to be examined, they were selected from contacts with quartzite to provide a direct comparison of the two. The actual contact was prepared for mounting where possible, and the orientation mark of the section transferred to the perimeter of the stereogram.

For quartz fabrics a 4-axis universal stage was employed and the c-axes were plotted directly on to a 200 cm. equal area stereographic net. Normally 200 or 300 points were plotted. By using a one-centimetre grid and a 1%

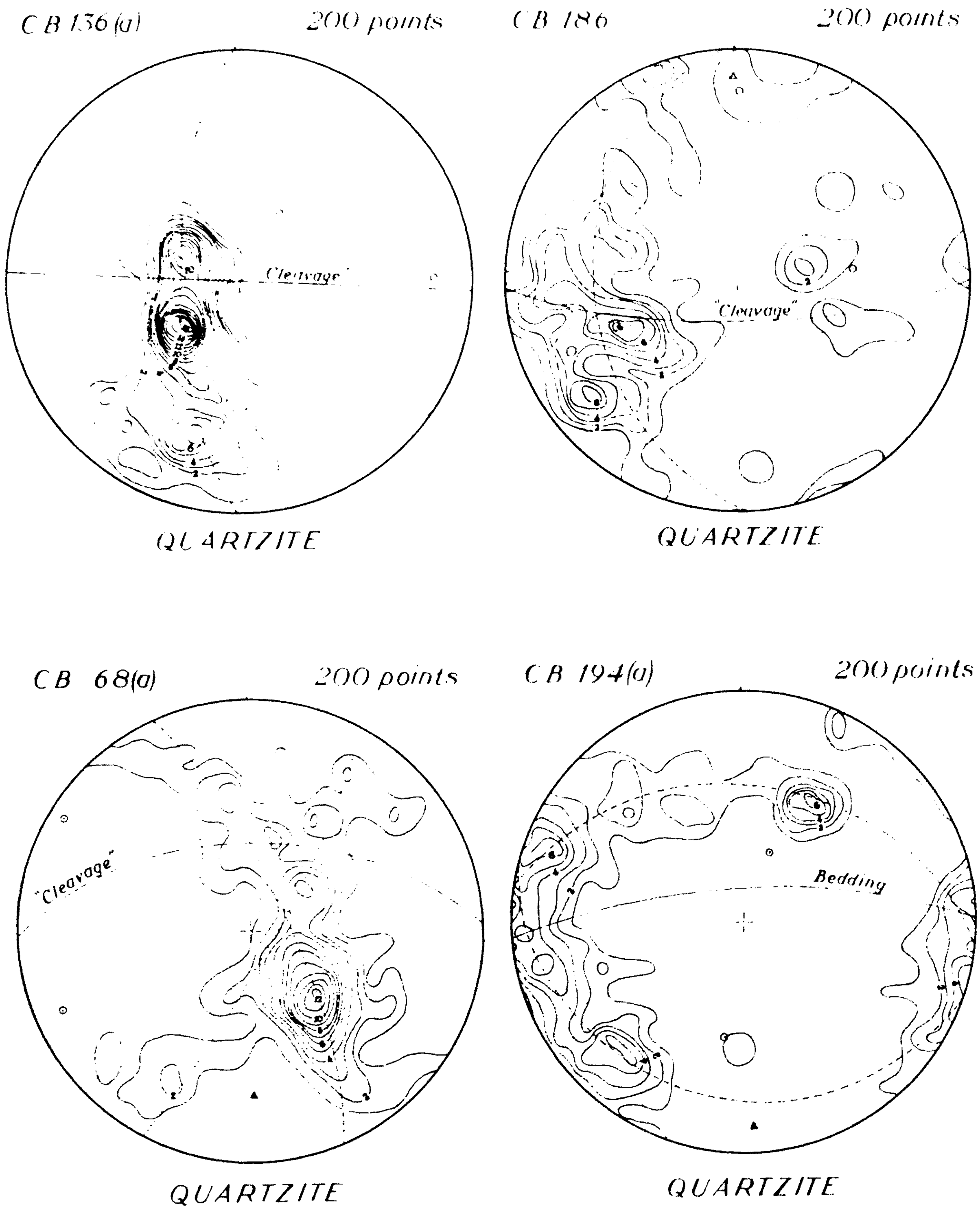


Fig. 25. Stereograms of quartzite fabrics (quartz c-axes).

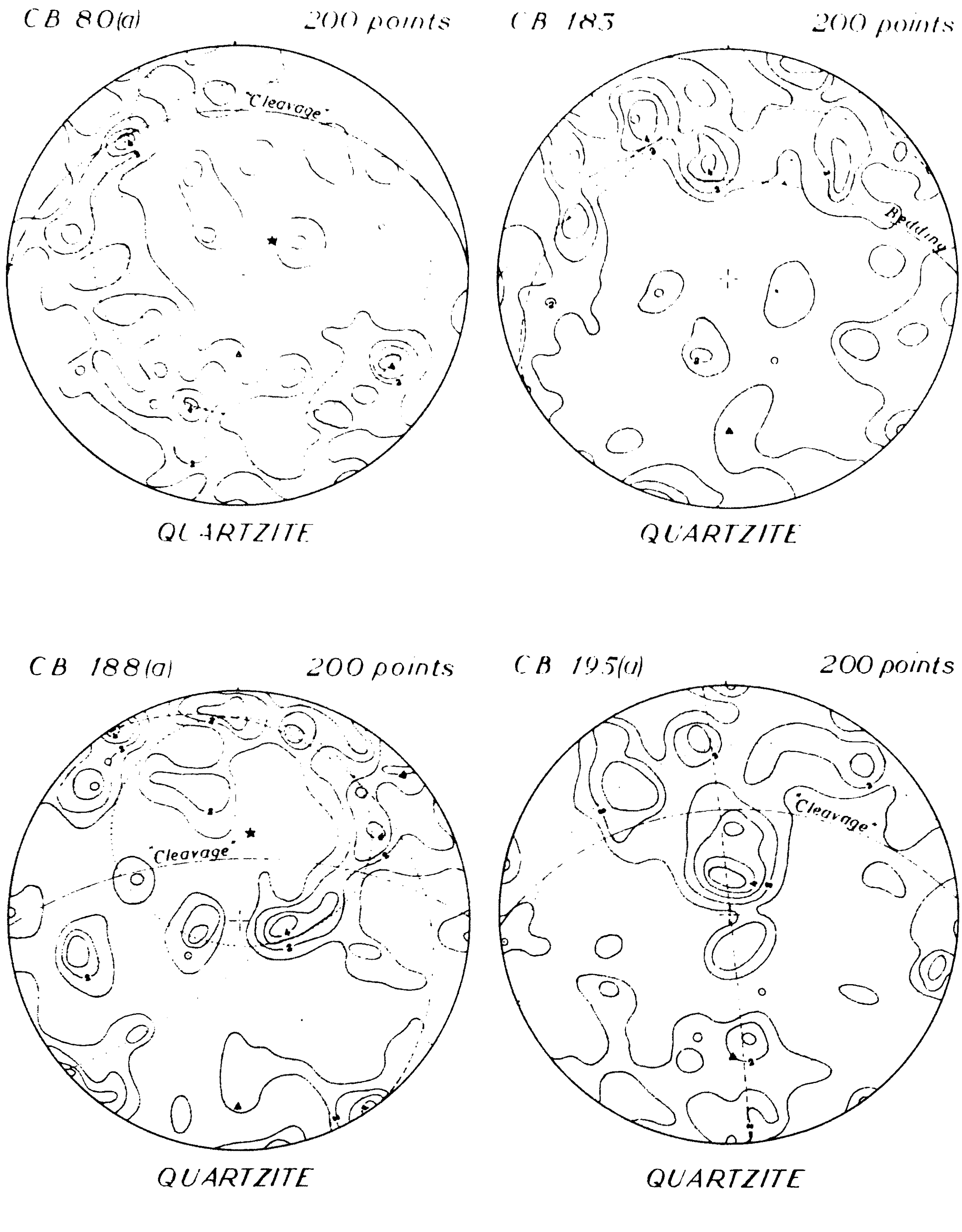


Fig. 26. Stereograms of quartzite fabrics (quartz c-axes).

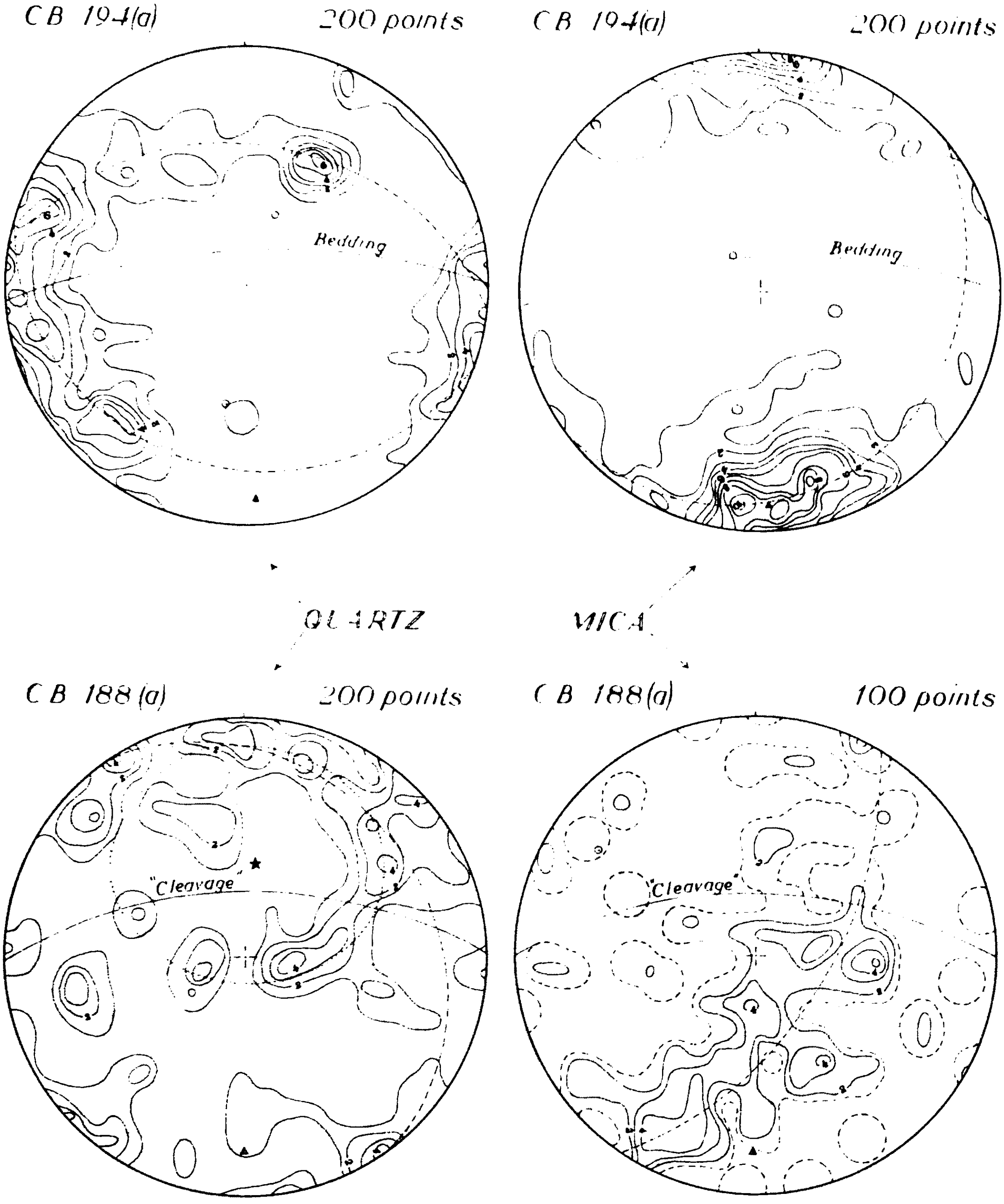


Fig. 27. Stereograms comparing quartz and mica fabrics in quartzite.

counter, as described in Emmons (1943) the final result was achieved. In mica diagrams the pole of the cleavage was used in the same way.

Practical Difficulties

Practical difficulties, which could slightly affect the results, include the measurement of strained quartz crystals with undulose extinction, the measurement of small crystals and the distinction of normal quartz from perthitic quartz in felspar. In the first case only the size of the maxima could be affected. In the two other cases the difficulties were not encountered often enough seriously to affect the pattern of the fabrics.

Examples of Quartzite Fabrics

All eight samples examined (figs. 25, 26, 27) show a preferred orientation of their quartz crystals to a greater or lesser degree. Usually one or more partial girdles are revealed by the stereograms, or less commonly, a small-circle*.

Specimen CB 136(a), from the Lower Nappe south of the Dee, has the highest recorded point-maximum of 16% within a well-defined broken girdle. The position of a possible very weak partial second girdle compares with that of the fabric of the granite in contact. Points in the stereogram of CB 186, from the Lower Nappe north of the Dee, are less concentrated, but there are again indications of two broken girdles. A third specimen from the Lower Nappe, CB 68(a), has one reasonably complete girdle with a high maximum of 12% through which passes a second, weaker girdle.

Only one specimen (CB 194(a)) from the Upper Nappe compares with those mentioned. The remaining samples are weakly orientated. There is, however, a possibility that the fabric of CB 80(a) represents a small-circle.

* On an equal area net a small-circle is not a true circle.

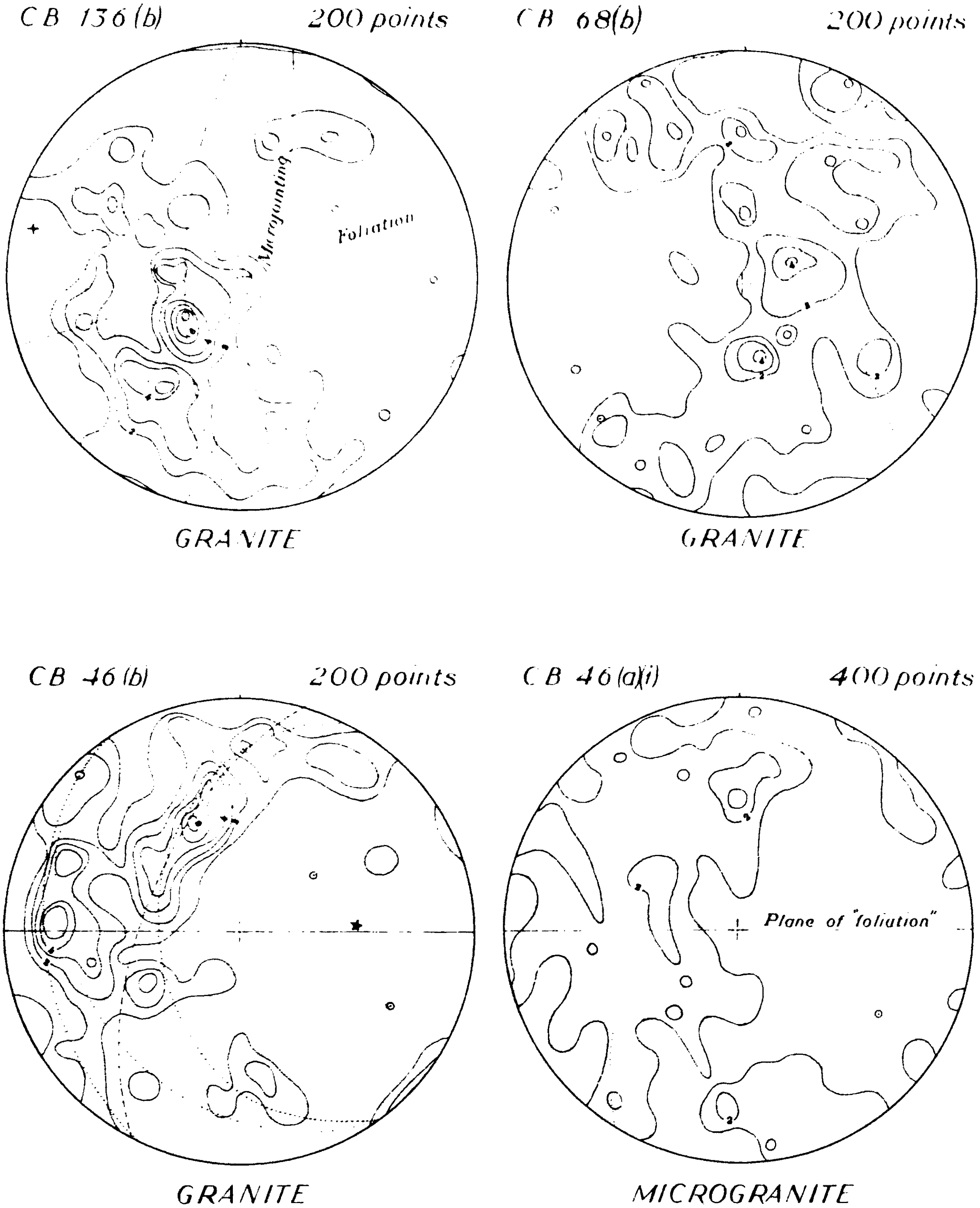


Fig. 28. Stereograms of granite fabrics (quartz c-axes).

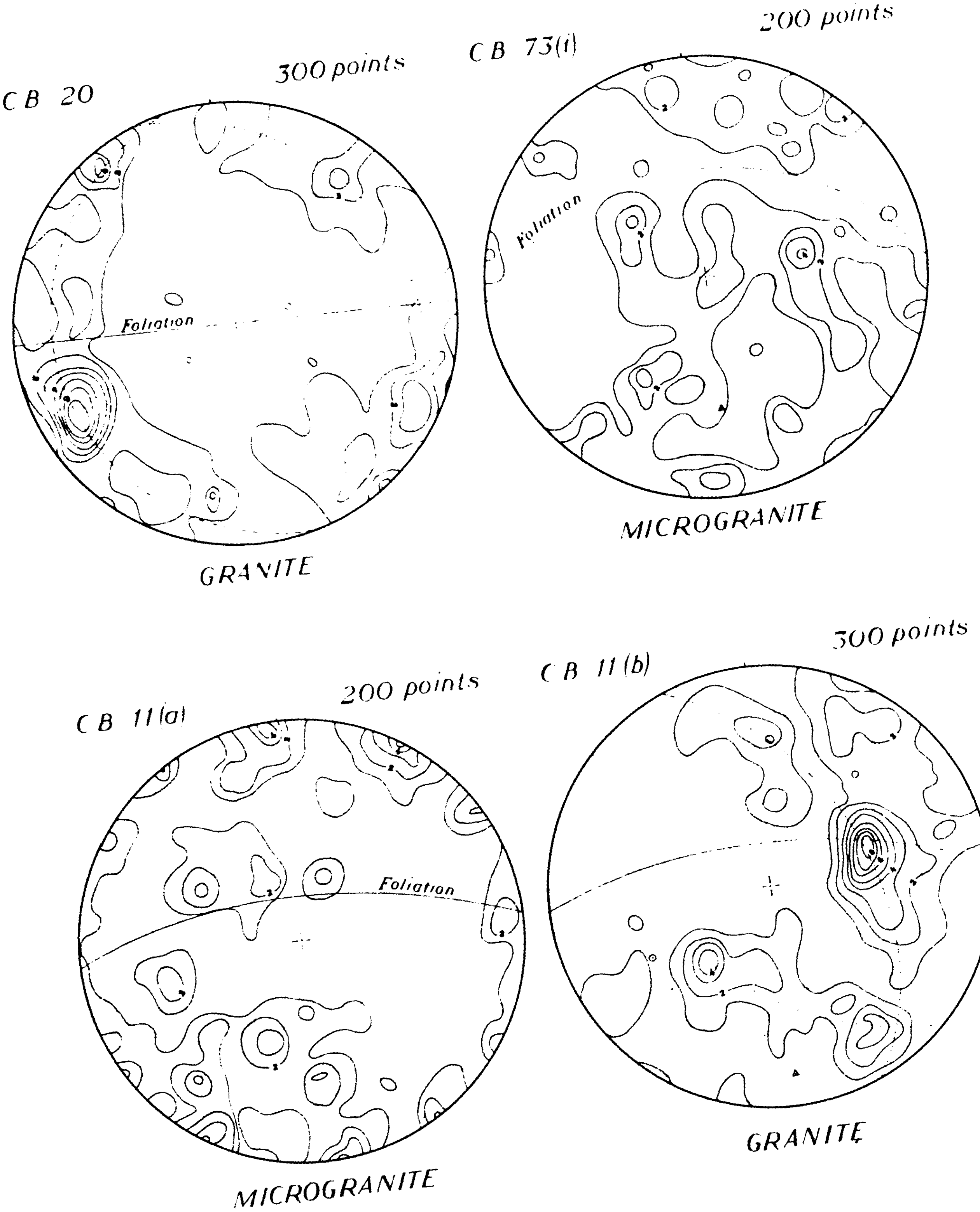


Fig. 29. Stereograms of granite fabrics (quartz c-axes).

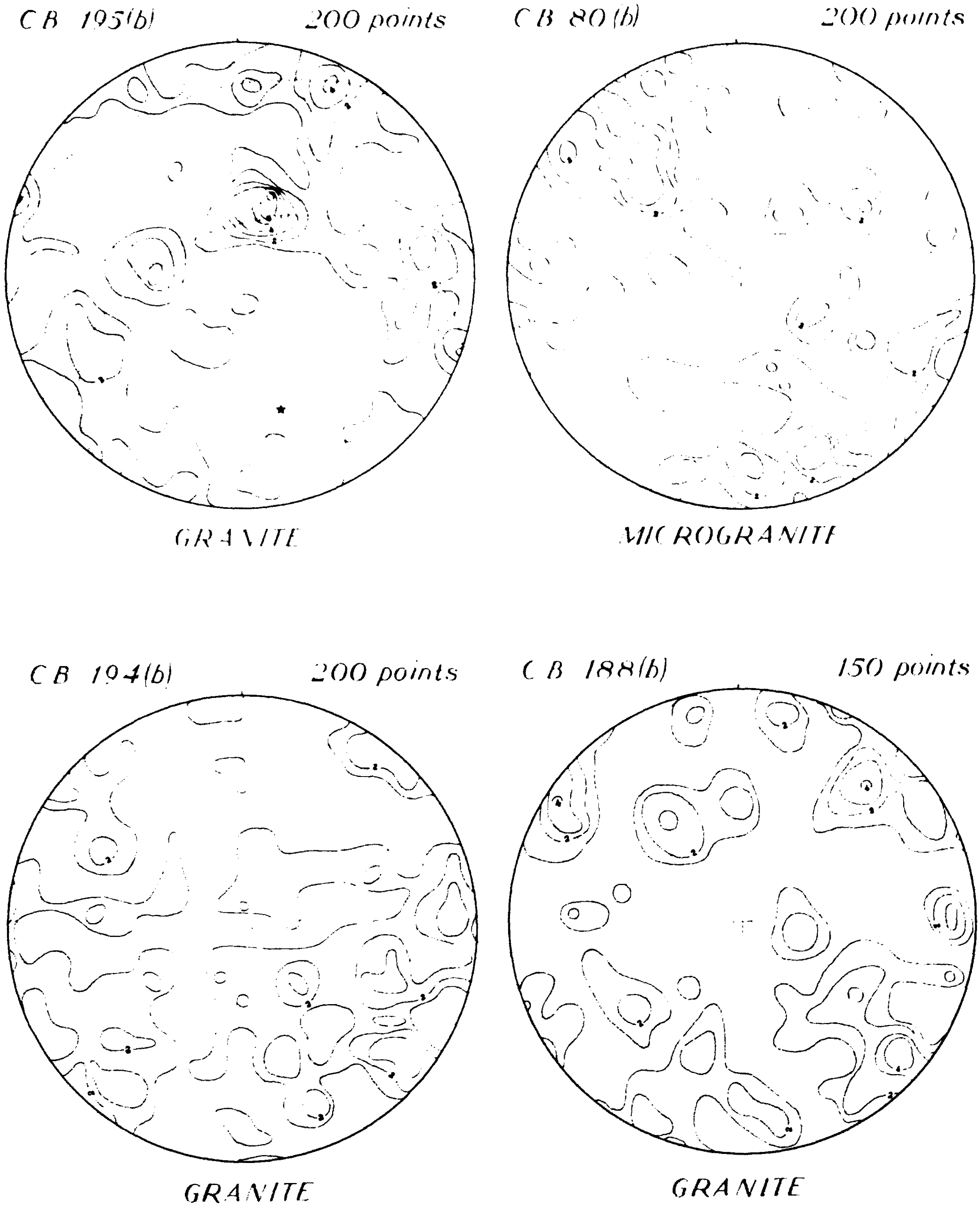


Fig. 30. Stereograms of granite fabrics (quartz c-axes).

This is supported by evidence of a well-defined small-circle in the fabric of CB 128 (see later).

When bedding or cleavage in these quartzites is visible it is always oblique or nearly perpendicular to the main girdles, but at a low angle to the greatest maxima. Generally no simple relationships between fabric and external features are obvious.

Mica diagrams may exhibit girdles in agreement with quartz girdles. The mica maximum in CB 194(a) agrees excellently with the pole of "bedding".

Examples of Granite Fabrics

None of the samples analysed shows the strong preferred orientation of some of the quartzites, while microgranites are less affected still by deformation. Of 12 specimens 4 are apparently isotropic, the most strongly deformed being from the lower nappe.

The stereograms of granite/quartzite contacts are perhaps the most interesting. Specimen CB 136(a) - the quartzite with the greatest degree of quartz preferred orientation - is in contact with a granite having an intervening chilled phase of 5 mm. width. The stereograms of both igneous phases reflect the main girdle of the quartzite (fig. 31). CB 68 is a similar case but without the chilled phase. The fabrics of CB 194(b) and CB 188(b) are apparently isotropic, and thus contrast with their associated quartzites.

In the granite CB 11(b) a distinctive fabric can be seen, but that of the possibly younger microgranite in contact is isotropic: yet CB 73, microgranite, is virtually isotropic while a later quartz vein cutting it exhibits a strong preferred orientation, the majority of the quartz c-axes lying near, but not in, the plane of the sheet itself. This suggests a different mechanism of orientation. See fig. 32.

Other interesting features are to be seen in a contact specimen of granite and porphyritic microgranite (CB 46). The quartz from the granite

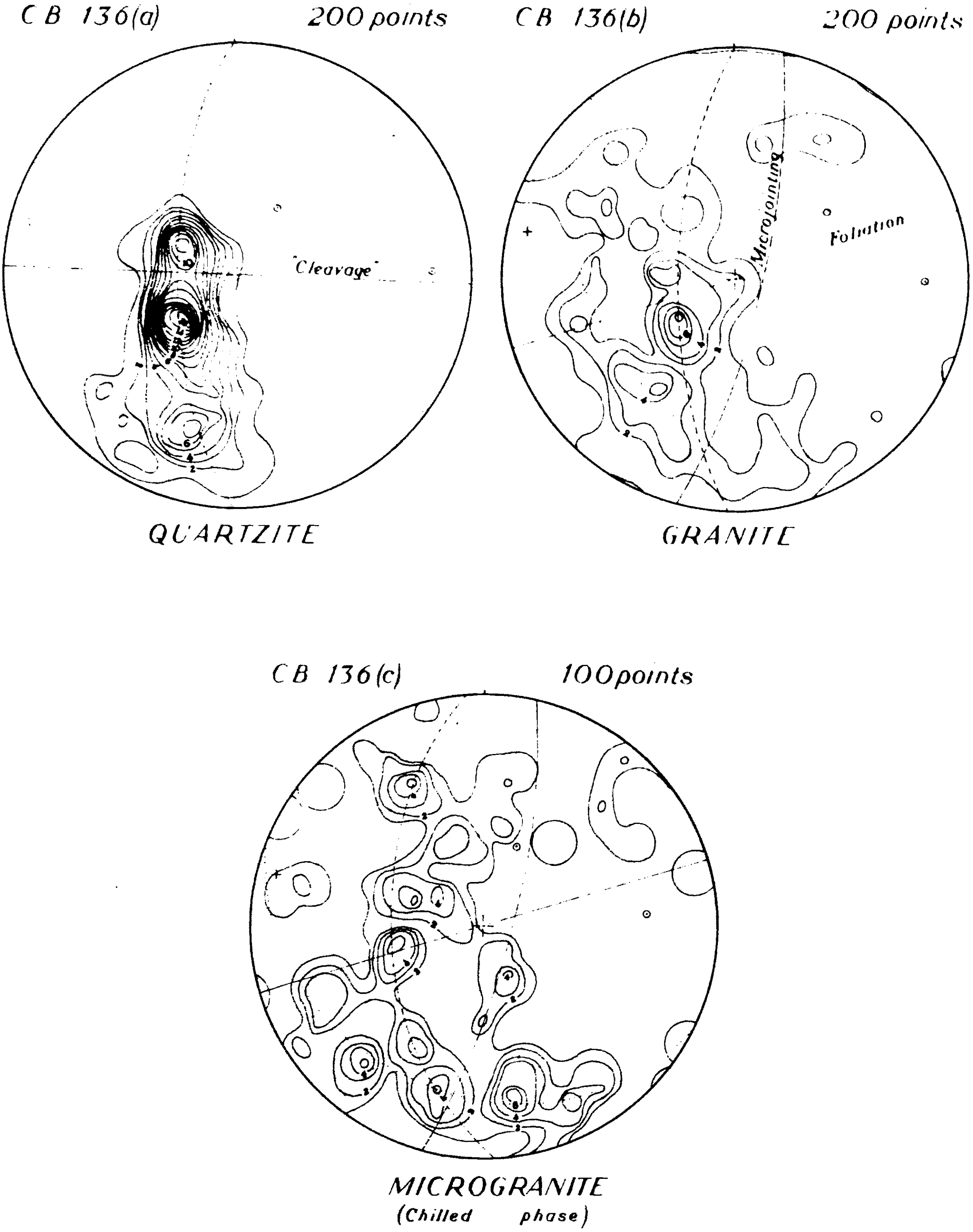


Fig. 31. Stereograms of granite/quartzite contacts (quartz c-axes).

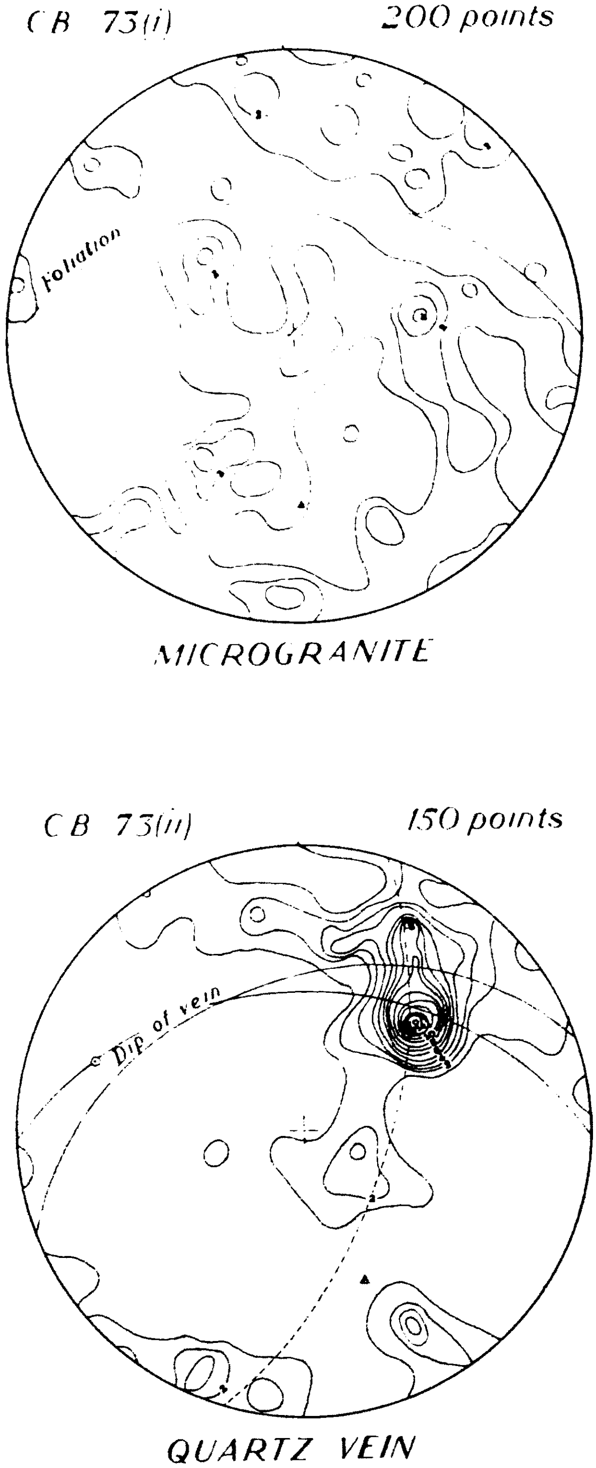


Fig. 32. Stereograms of a micrograrite and a later quartz vein (quartz c-axes).

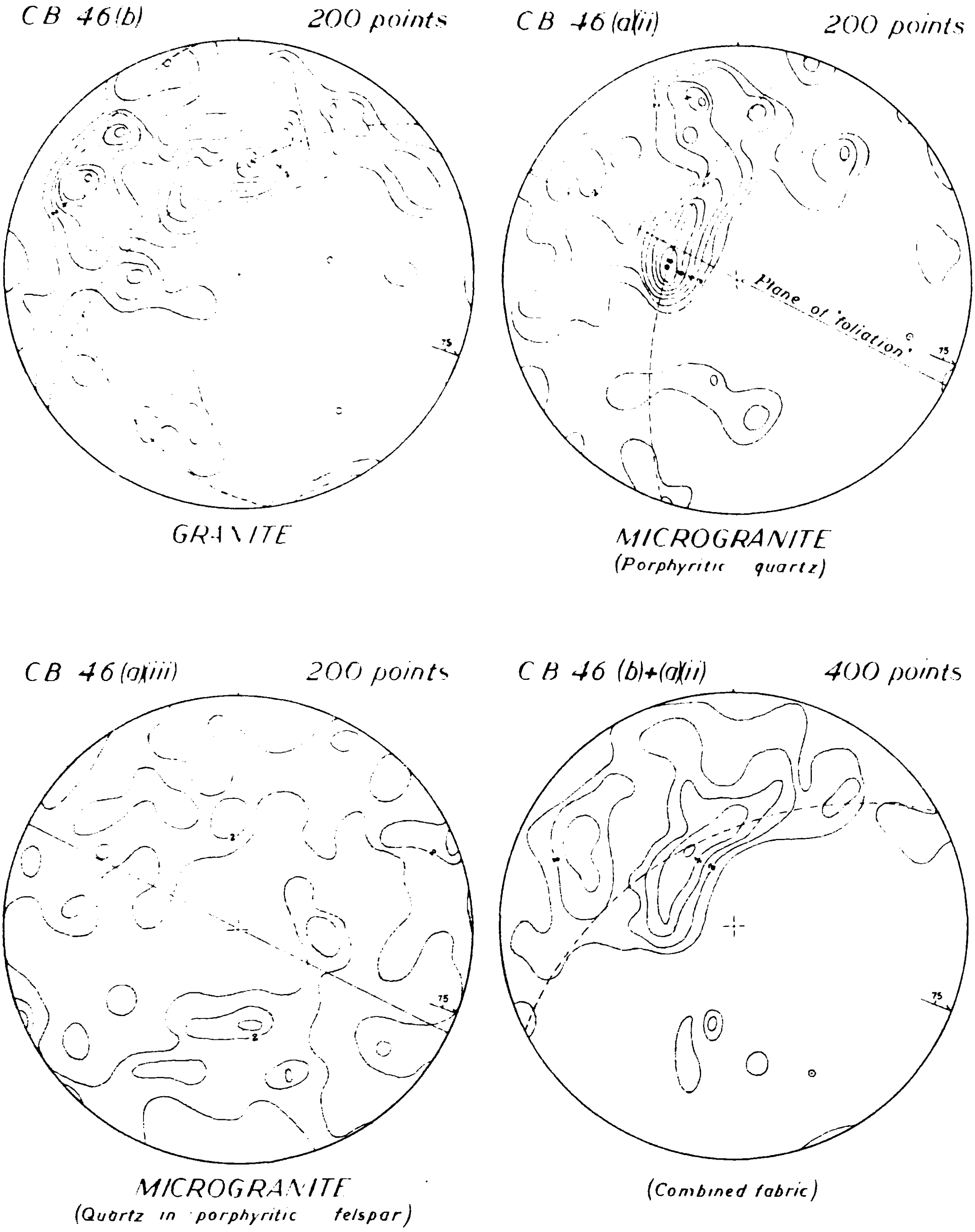


Fig. 33. Stereograms comparing the quartz fabric of a granite with those of different phases of a porphyritic microgranite (the groundmass is shown in figs. 28 and 34).

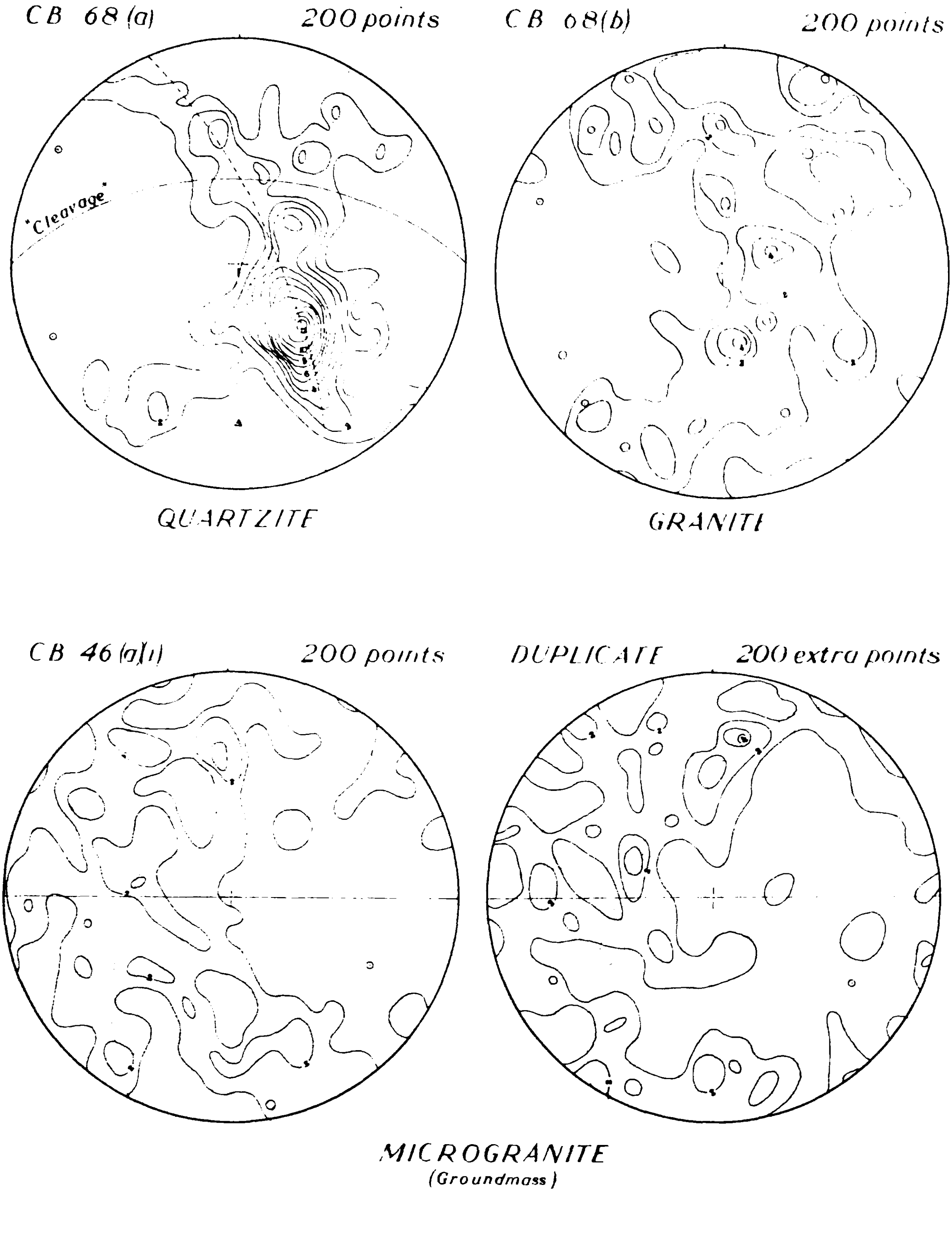


Fig. 34. Stereograms (quartz c-axes) showing two methods used to test the validity of interpretations.

has a 6% maximum: the fabric in the porphyritic quartz of the microgranite, though not identical, has a maximum of 8%. The groundmass of the microgranite has a much weaker fabric (fig. 28,34) but its pattern agrees more favourably with that of the granite than with the porphyritic quartz. Finally the fabric of the quartz inclusions in the porphyritic felspar can be seen to be virtually isotropic.

Examples from Other Rocks

For comparative purposes it is worth mentioning here the stereograms of a mica-bearing quartzite band within pelitic schist. The folded quartzite (see "Later-Caledonoid Structures", p.28) was sampled in three places (CB 128, 129, 130). Stereograms of the quartz and mica fabrics are presented in figures 42-44, with a full description on page 46 et seq. In summary, the fabrics show a strong degree of preferred orientation. The quartz diagrams represent small circles centred about the main fold axis. The poles to partial mica girdles almost coincide with this macroscopic axis.

Interpretation of Results

It has been assumed that stereograms with maxima of not less than 5% and/or with well-defined girdles or small-circles, are valid expressions of the fabric. This includes half the samples analysed. The remainder have had all possible girdles etc. marked on them before comparison with other examples and before reorientation into their field positions. If such possible girdles coincide with those of other rocks in contact, they are deemed to have some significance. For instance, CB 68(a) has one well-defined girdle and one weak. When compared with the two assumed girdles of CB 68(b), the agreement between the fabrics indicates that the assumptions are correct. Another method was employed with CB 46(a)(i), which was analysed once and then repeated with a further 200 quartz c-axes.

QUARTZ FABRICS

Reorientated on Wulff Net

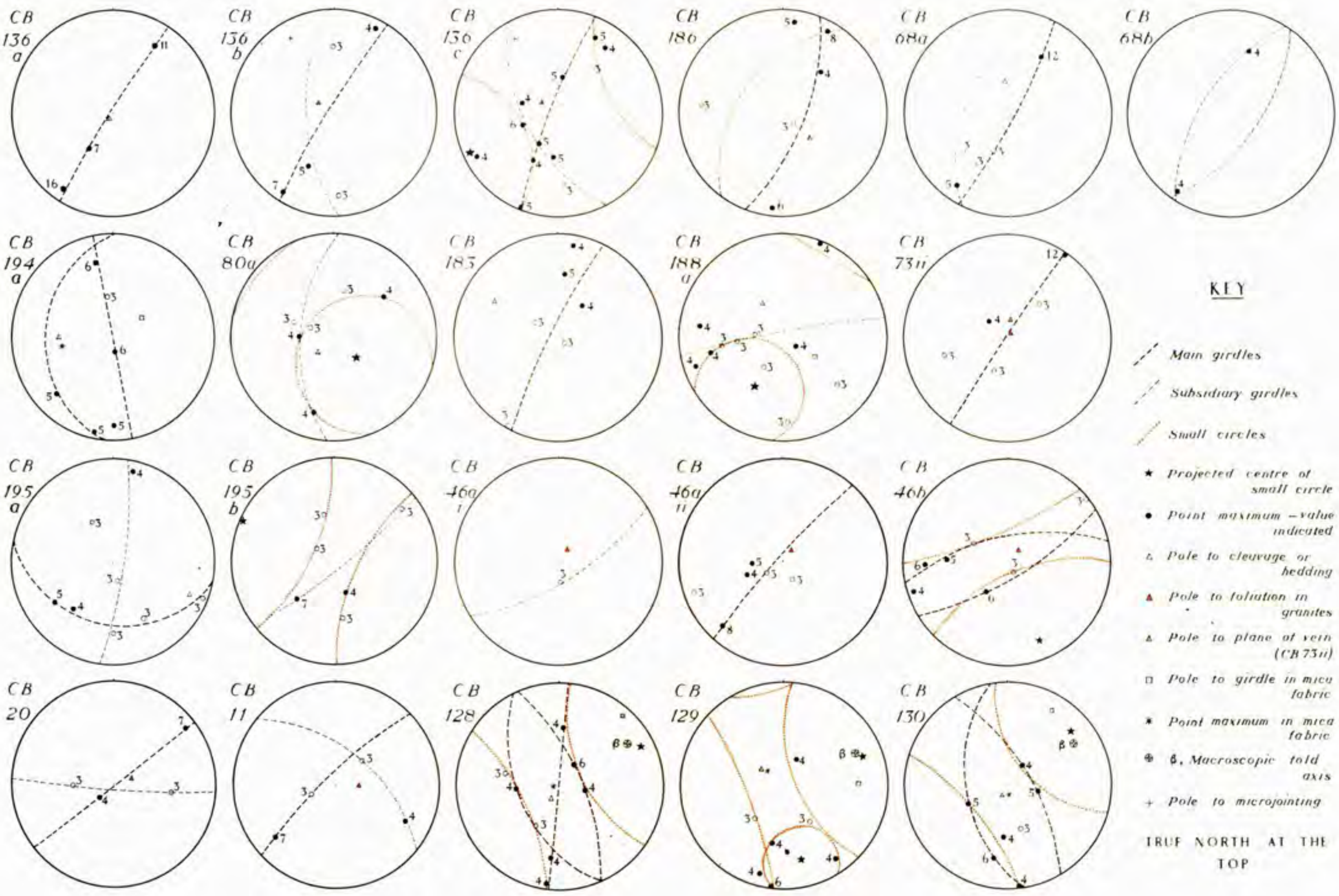


Fig.35 Syn-optic diagram of all non-isotropic stereograms showing principal elements only, reorientated on Wulff net.

The two stereograms were then "interpreted" separately, whence it was found that the results agreed reasonably well.

In order to interpret the results collectively all non-isotropic stereograms have had their principal elements reorientated into their actual field positions, with true north at the top of the diagrams, as shown. Immediately it can be appreciated that in the quartzites CB 136(a) and CB 68(a), both from the lower nappe, an excellent correspondence exists between the main girdle of each and between the distribution of points themselves. CB 183(a) and CB 186 are broadly similar to this pattern also, while CB 195(a) is partly similar. With a slight clockwise rotation the fabric of CB 194 fits the same picture.

Of the granitic rocks not paired with quartzites, CB 20/1 and CB 73(ii) compare well, CB 46(a)(i), (ii) and CB 46(b) all having certain similarities with the picture mentioned.

From these results it is interpreted that a quartz girdle, lying within the limits shown in figure 36 , and with the maximum values tending to occur as indicated, is a fairly common feature in the quartz fabric of rocks throughout the area. In addition, its position is apparently independent of the attitude of bedding or schistosity in most cases.

This generalised girdle may be an ac-girdle to a cross-fold, the a-direction being indicated by the position of the point-maximum. Girdles with transverse trends are less easy to interpret. They could be later features, yet according to D.M. Ramsay (1962), cross-girdles may under certain conditions be produced in the plane of active shearing simultaneously with ac girdles.

Other interesting features deserve mention. For instance, although the fabrics of CB 136(a) and the igneous phases associated, (b) and (c), are basically similar there are additional features in the fabrics of the igneous rocks which do not appear to be due to random scatter, and may in

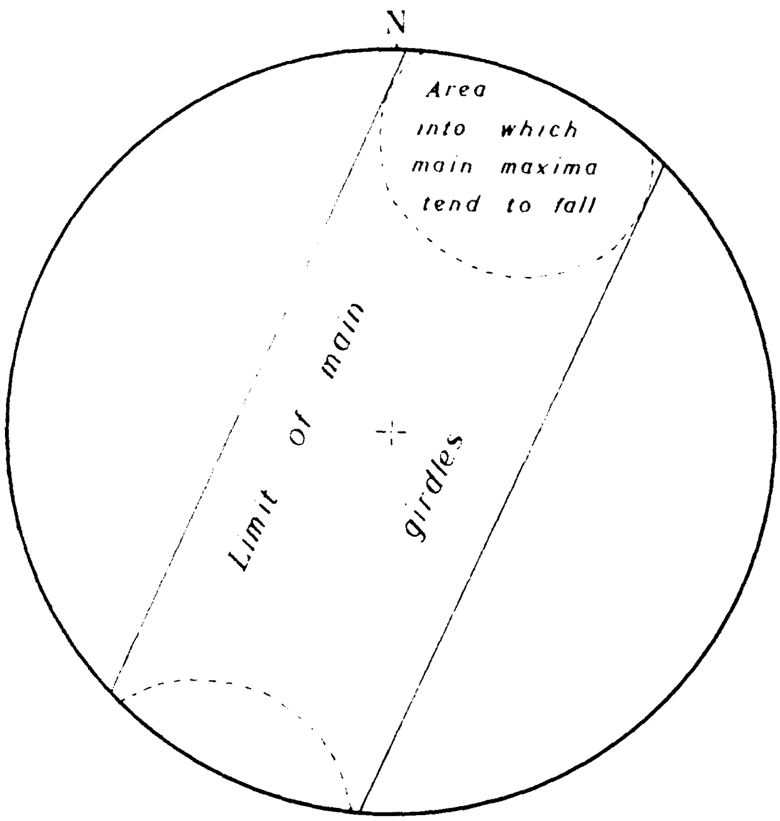


Fig. 36. The general distribution of maxima and girdles in quartz fabrics.

fact be the result of another deformation, either earlier or later. If the main girdle common to the quartzite and granite is indeed a cross-fold (F_{B1}) ac-girdle, the other structures in the granite are likely to be reflections of F_{A1} or F_{A2}. As the granite is known to be post-F_{A1} if not post-F_{B1}, the granite has either acquired an even earlier quartzite fabric by inheritance, or is more susceptible to later deformation than the quartzite. The same may be true of CB 46(a) and (b) where similar relationships exist.

If, as Voll (1960) believes, there is no primary deformation fabric in the Dalradians, it would appear that the granitic rocks are more susceptible to deformation than the quartzites. This in turn points to the cooling of the granite under stress.

If Voll is correct it also means that the transverse elements in the fabrics of some quartzites, which do not lie in shear-planes, must be of later origin, like those of the granites. Thus while no folds later than the main cross-folds have been found in certain areas where quartzite has been studied, nevertheless the evidence suggests that a later stress system was established. The general weakness and inconsistency of these transverse fabric structures implies that the quartzite masses have preserved older F_{B1} deformation features in contrast to the more responsive belts of pelitic rocks.

Regarding the granites in general it is difficult to draw conclusions as to how they acquire their fabrics. Since some granites undoubtedly have preferred orientations, some have none, yet all differ in varying degrees from the quartzites in contact (which are never isotropic), various mechanisms can be postulated:-

FABRIC INHERITED FROM QUARTZITE

(i) with subsequent deformation of either rock.

(ii) with initial quartzite fabric destroyed or altered by intrusion of granite.

FABRIC NON-INHERITED

granite starts with isotropic fabric "out of phase" with quartzite.

NO FABRIC

(i) because granite intruded post-deformation.

(ii) because fabric later destroyed.

(iii) because near centre of intrusion and thus protected from deformation.

Plate 63. General view across junction of quartzite (left half) and microgranite (right half), showing also the area covered by plates 64 and 65. CB 68; x-nicols; x 18.

Of the granites with isotropic fabrics it seems reasonable to assume that their protecting environment has played a part: the main mass of the Lochnagar granite appears to be undeformed (Oldershaw, 1958), yet thermal metamorphic mineral assemblages in the surrounding schists have suffered considerably. It is also noticeable that the finer-grained "granites" show weaker preferred orientation than the coarser. Environmental control on the crystallisation of quartz is strongly suggested. At the same time the evidence of shattered felsites along the line of major slides and of the local development of shear zones in granite proves that in some cases the fabrics must have been affected by post-crystallisation deformation.

One is thus left with two possible mechanisms, that of deformation and that of inheritance.

Inheritance of Quartz Fabrics by Granites

The idea of inheritance has been considered to account for the fact that granites which appear to be post-F_{B1} possess structures which appear to be F_{B1} in age.

The actual mechanisms considered are twofold: (1) that granite magma may marginally assimilate, without complete melting, orientated quartz crystals from the quartzite host, thereby obtaining an "impression" of the quartzite fabric; (2) that marginally a granite magma may cause felspathisation of a quartzite to the extent of producing "granite" retaining the old fabric.

At first sight there is evidence to support case (1). In a number of localities small quartzite rafts are concentrated within the granite at contacts. Less commonly small masses of quartz have been found. However the smaller xenoliths tend to be disturbed and none less than an inch or so in length have been seen. All sizes should be found if the complete assimilation theory is to be valid. Furthermore, pegmatitic felspar, with graphic texture, is often associated with the quartz masses, suggesting a

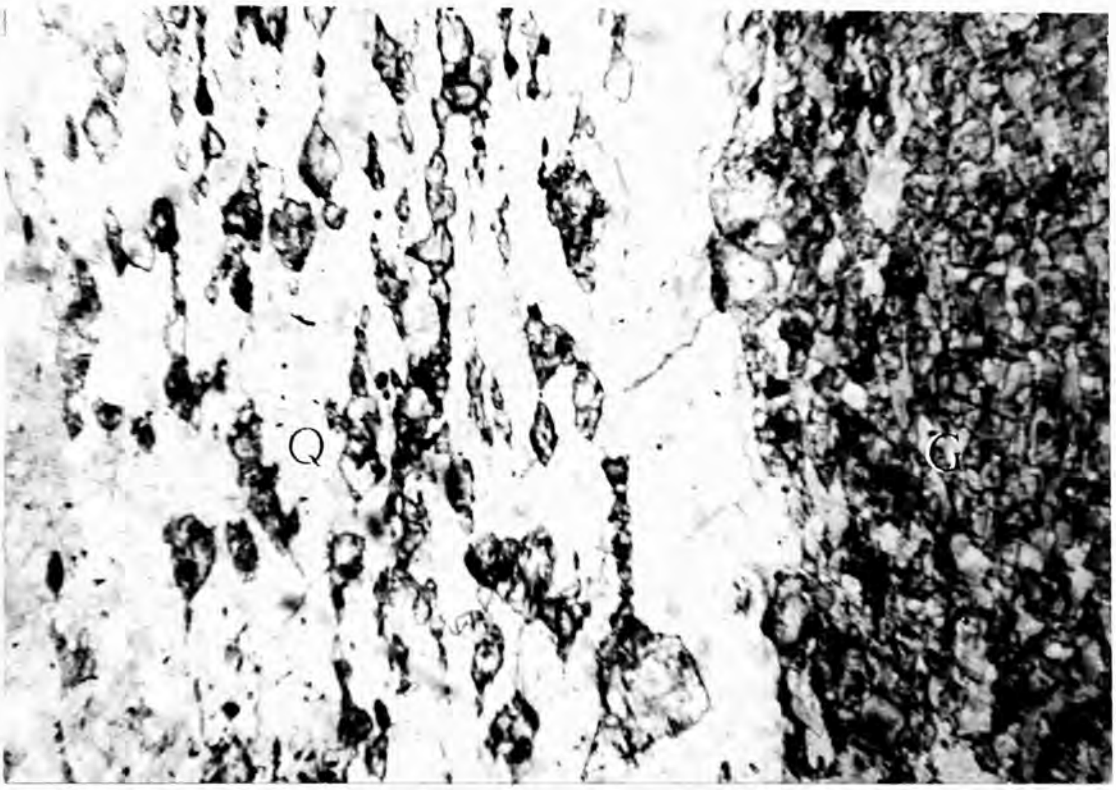


Plate 64. Closer view of junction between quartzite (Q) and microgranite (G). CB 68; x 40.

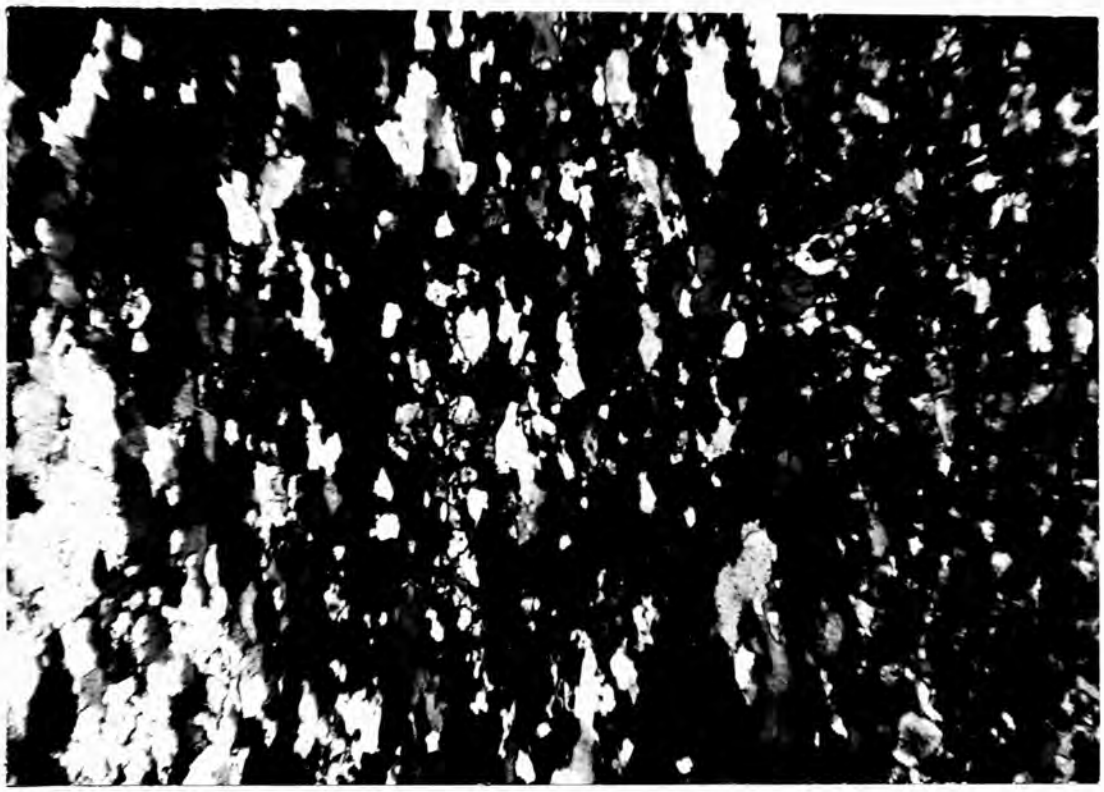


Plate 65. As for plate 64, x-nicols.

simple pegmatitic origin for the quartz.

Case (2), that of felspathisation, has been recognised only very locally. Although certain fine-grained acid intrusives may bear close resemblance to felspathic quartzite, structural discordance between the two has left the writer in no doubt that the field evidence is against the granites being produced in situ on a large scale from quartzite.

It remains only to discuss the microscopic evidence for or against fabric inheritance.

There is a superficial resemblance between some quartzites and the granites in contact in that the former are rich in felspar while the latter are rich in quartz. Sometimes, too, the felspar in each shares a common alignment, usually parallel to the contact. There might, therefore, be reason to believe that in some cases quartz assimilation has occurred, while in others "granite" has been produced by felspathisation of quartzite. In CB 136, for example, the only obvious difference between the chilled zone and the quartzite is percentage of quartz.

However, detailed comparison shows that the contacts are sharp, with, in some instances, an intervening finer igneous phase, as in CB 136. Igneous zoning in the plagioclase crystals may help to distinguish the granites, while the proportions of the alkalis in the felspars are different. Examination of areas of quartz in granites shows them to be texturally and mineralogically distinct from quartzite even though the style of deformation may be identical.

As there is only one possible exception (CB 68), it is concluded that neither assimilation of quartz(ite), nor felspathisation of quartzite are tenable explanations in general. It appears therefore that the state of the fabrics in granite is the result of later deformation which is possibly also responsible for the foliation in some granites. Unfortunately, however, there is no indication of just how much later this deformation might be.

Plate 66. Deformed quartzite band in semipelite, Càrn nan Sgliat.

Dynamics and Petrofabrics of Deformed Quartzite Band in Semipelite

The folded quartzite band from locality (038,068), already referred to on pages 28 & 41, is illustrated in plates 66 – 68 and subsequent diagrams. The most obvious feature of this F_{A2} fold is the abundance of closely-spaced shear-planes, which are sub-parallel to the main axial plane and clearly associated with the same deformation.

Apart from petrofabric aspects, there are very interesting dynamic problems involved:

(1) Has the effective thickness of the quartzite band been increased by folding as shown in extreme condition by figure 38a ; or has the band simply been deformed marginally (fig. 38b); or has shearing acted to cause thinning?

(2) Does each fold-like nose represent a single fold; or is each the result of shearing and boudinage?

(3) What is the relative direction and age of movement along the shear-planes?

(4) Is the deformation of the top surface directly related to that of the bottom of the band?

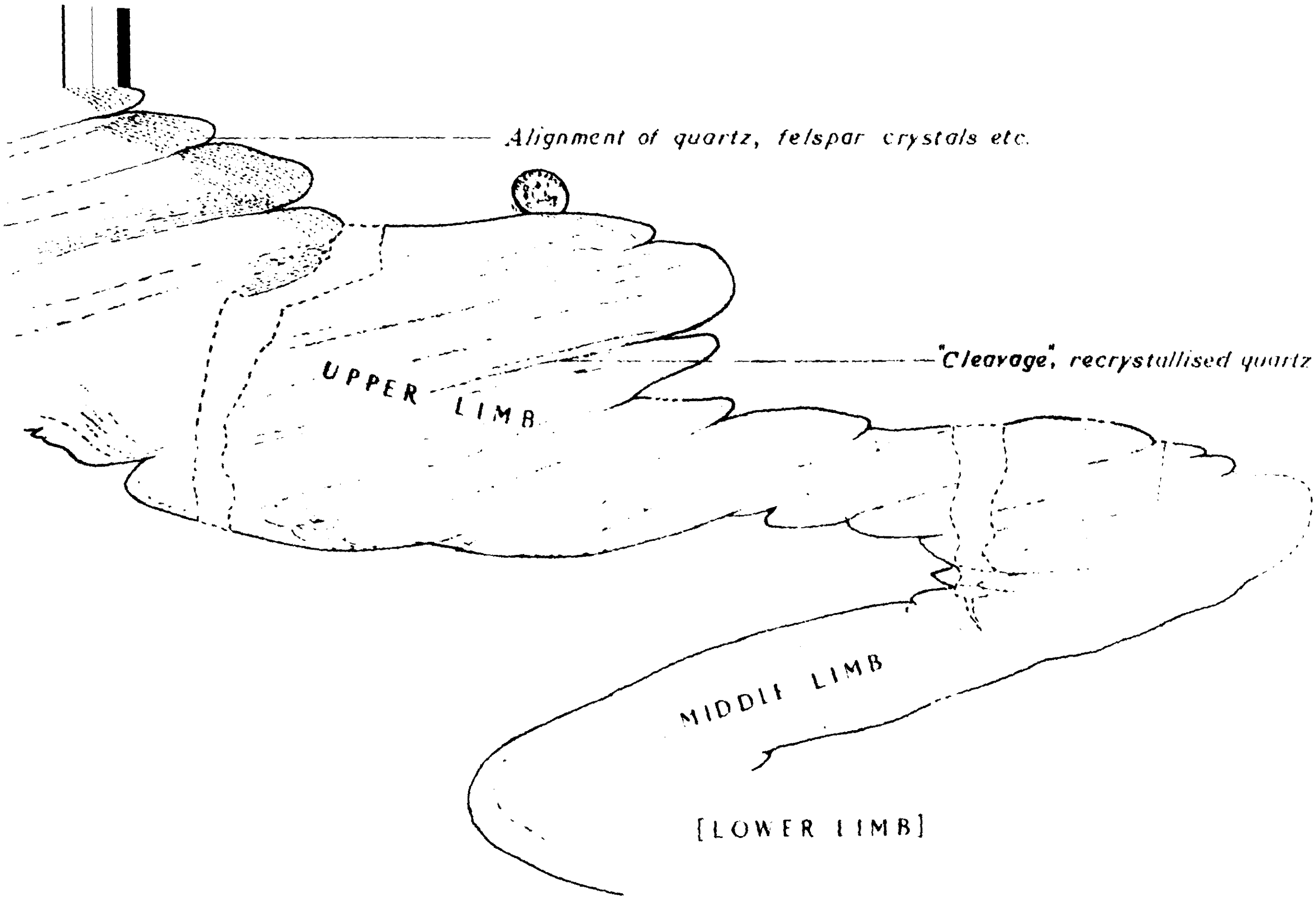
(5) Are two ages of folding in fact represented?

In detail the quartzite reveals more features which help to solve these problems (fig. 37), though bedding-like structures give ambiguous indications of folding. This, however, is due to the presence of real bedding as lines of felspar crystals or layers of quartz, and to an axial plane "cleavage", seen as lines of elongated felspars. The refraction of this cleavage from the surrounding semi-pelite through the quartzite is also evident.

In the crucial central zone most of the detail is not visible, but it is evident that strong folding has occurred at the margins, as bedding in the quartzite and adjacent schist demonstrates (pl. 67). Lack of

PLATE 67

PLATE 68



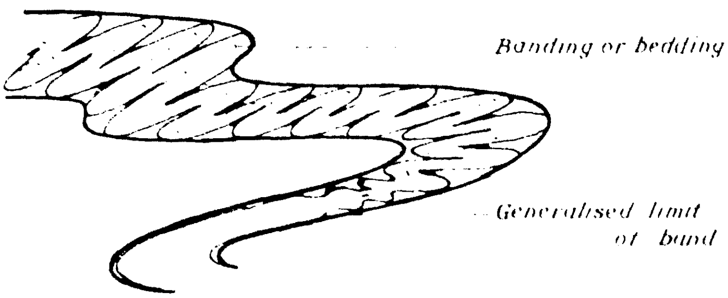
(Based on plate 68)

(i)

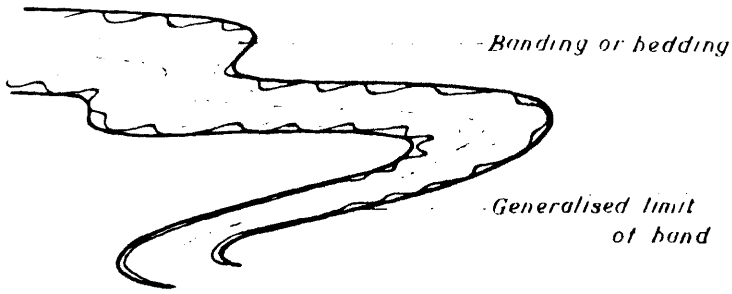
(Based on part of plate 67)

(ii)

Fig. 37. Structural details in quartzite, based on plates 67 and 68.

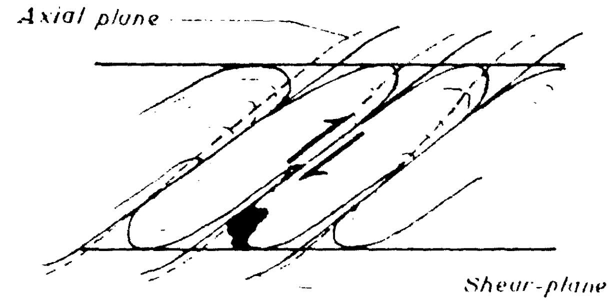


a

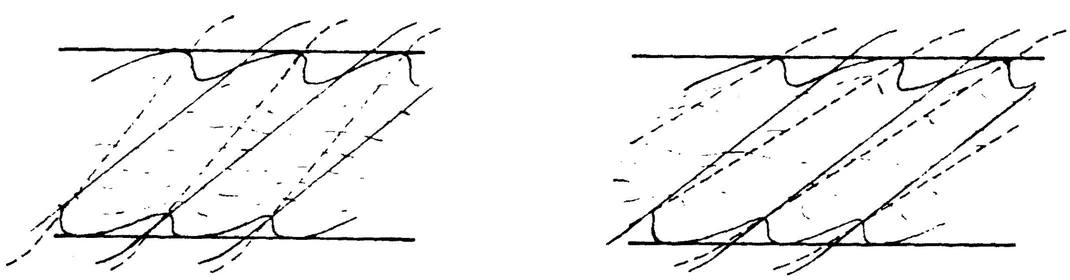


b

Fig. 38. Possible interpretations of the degree of deformation in the quartzite band: (a) strong "internal" folding; (b) marginal folding only.



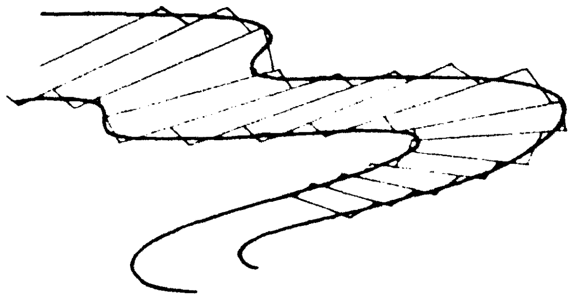
a



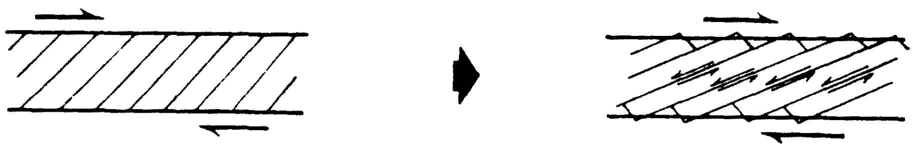
(i) (ii)

b

Fig. 39. (a) Idealized picture of folding with shearing.
(b) Two likely interpretations of the relationship between the axial planes and shear-planes.
NOTE:- Both (a) and (b) refer to the upper limb.



a



b

Fig. 40. (a) The general attitude of shear-planes, "fanning-out" round the main hinge.
(b) The development of wedges by sinistral movement along shear-planes (upper limb).

deformation, obvious in some parts of the quartzite, proves that some "noses" are not fold closures at all.

Furthermore, the absence of strong central folding is indicated by the fact that small folds on one margin do not always have counterparts on the other. This would suggest that deformation commenced as shear-folding. However, virtually every "nose", whether it represents a fold-closure or not, has a corresponding real fold in the adjacent schist. Thus a close relationship between folding and shearing can be established.

The relationship between the top and bottom surfaces of the quartzite cannot be properly ascertained, however. This depends upon whether the axial planes of the minor folds dip more or less steeply than the shear-planes. If the whole of the quartzite were evenly deformed one would expect that movement along the shear-planes would tend to slide out the "middle-limbs" of the minor folds to bring the axial planes towards coincidence with the shear-planes (fig. 39a). If one accepts the indications that the centre of the quartzite band is less deformed, the above pattern cannot be the case. Moreover, it would imply a dextral movement along the shear-planes, whereas a sinistral movement is indicated (see below). Thus the relationship must be as in either figure 39bi or figure 39bii though it cannot be established which is the case.

The contrast between the quasi-plastic minor folding and the brittle-type shearing points to a time-gap between the two in spite of their close relationship. Shearing must be the later; the shape of the slices cannot be explained otherwise, while if shearing preceded folding the shear-planes would be rotated round the main fold-closures. As it is they are merely "fanned out" (fig. 40a).

The displacement of relatively undeformed parts of the quartzite band in the upper limb of the main fold (pl. 68 and fig. 37) proves moreover that in the upper and lower limbs movement along the shear-planes was

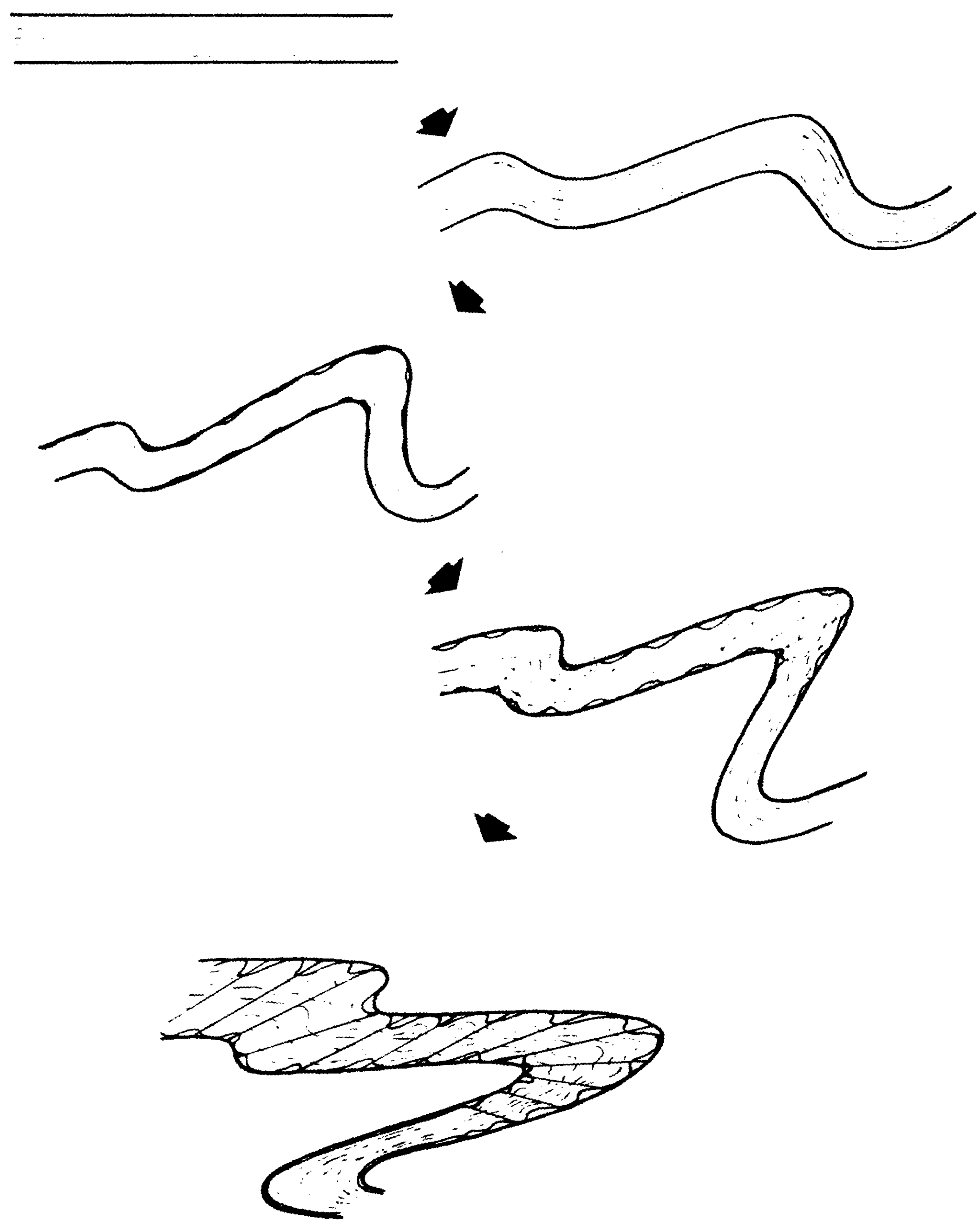


Fig. 41. Reconstruction of fold-history (idealized).

sinistral, in the middle limb dextral (fig. 40).

Measurements of thickness, where small-scale folding is absent, indicate a minimum value of two inches. The original thickness is likely to be more than this, allowing for the thinning effects of shearing and plastic flow. However it is believed that the present thickness at the main closures exceeds the original.

The smaller and the larger folds are likely to be of the same age, for three reasons: (1) the style and attitude are closely similar; (2) the absence of minor folding in the middle limb of the larger fold cannot reasonably be attributed to coincidence and points to simultaneous evolution; (3) the shear-planes are dependent upon both types of fold for their present attitude.

Reconstruction of Fold-History

The quartzite band, under oblique shear pressure, began to deform as one unit. A single large and many smaller shear folds developed, the latter partly as drag-folds relative to the larger. At a later and more brittle stage the shearing stress overcame the plastic limit of the quartzite causing the development of shear-planes and permitting rotational movement of the segments with transport en bloc towards the main fold closures. Concomitantly boudinage operating along the margins of the shear-planes gave rise to further minor "noses". See fig. 41.

Interpretation of Petrofabrics of Deformed Band

In each of the three samples of quartzite analysed (from the upper, middle and lower limbs), the quartz c-axes form a small-circle of radius about 60° (fig. 42). This is centred about the b-axis of the main fold. The fabric is thus a rotational one.

The small circles, it is suggested, are made up of at least two partial girdles. This is best seen in the stereogram of CB 128, where three parts

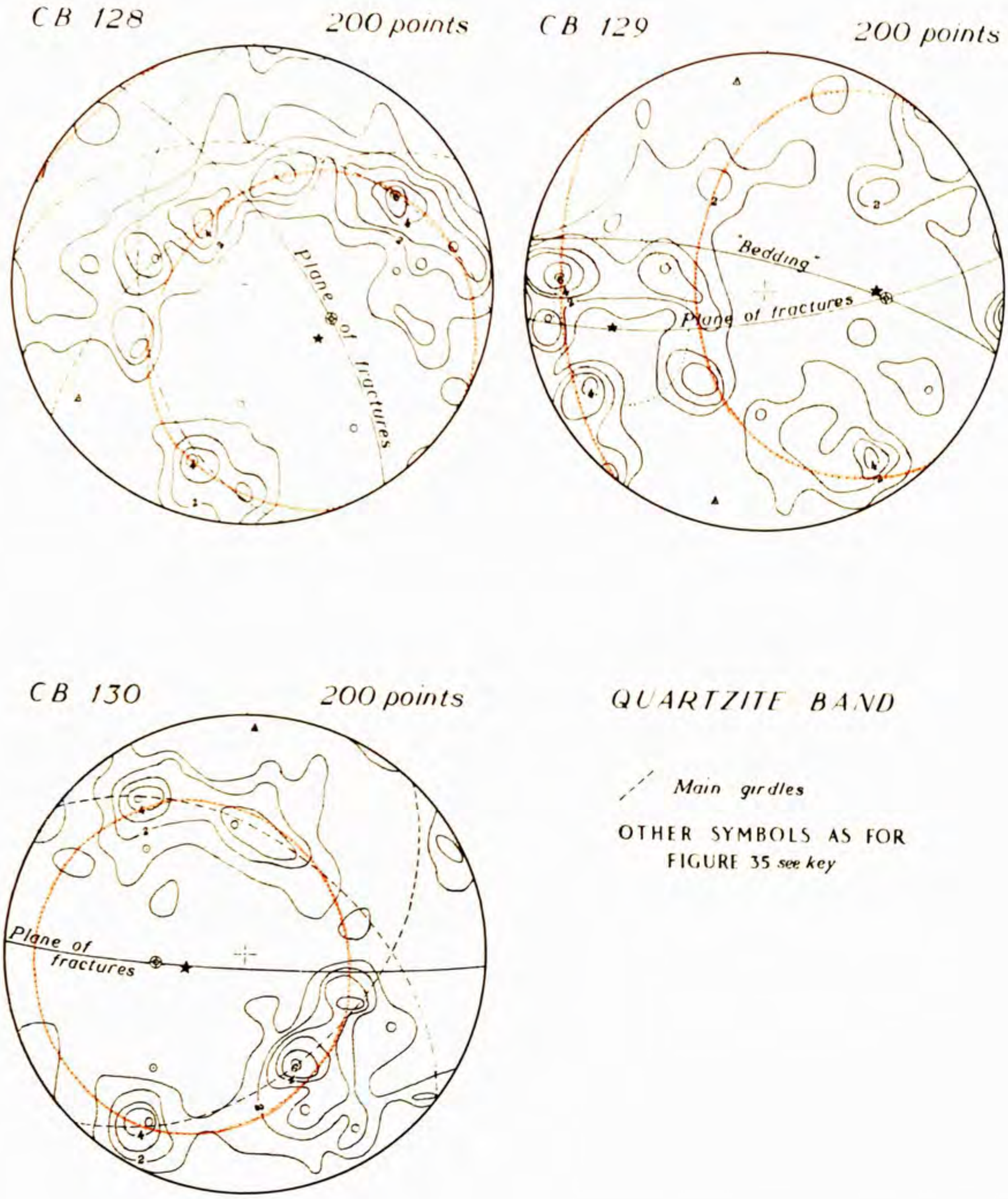


Fig. 42. Stereograms (quartz c-axes) of the deformed quartzite band. CB 128 upper limb, CB 129 middle limb, CB 130 lower limb.

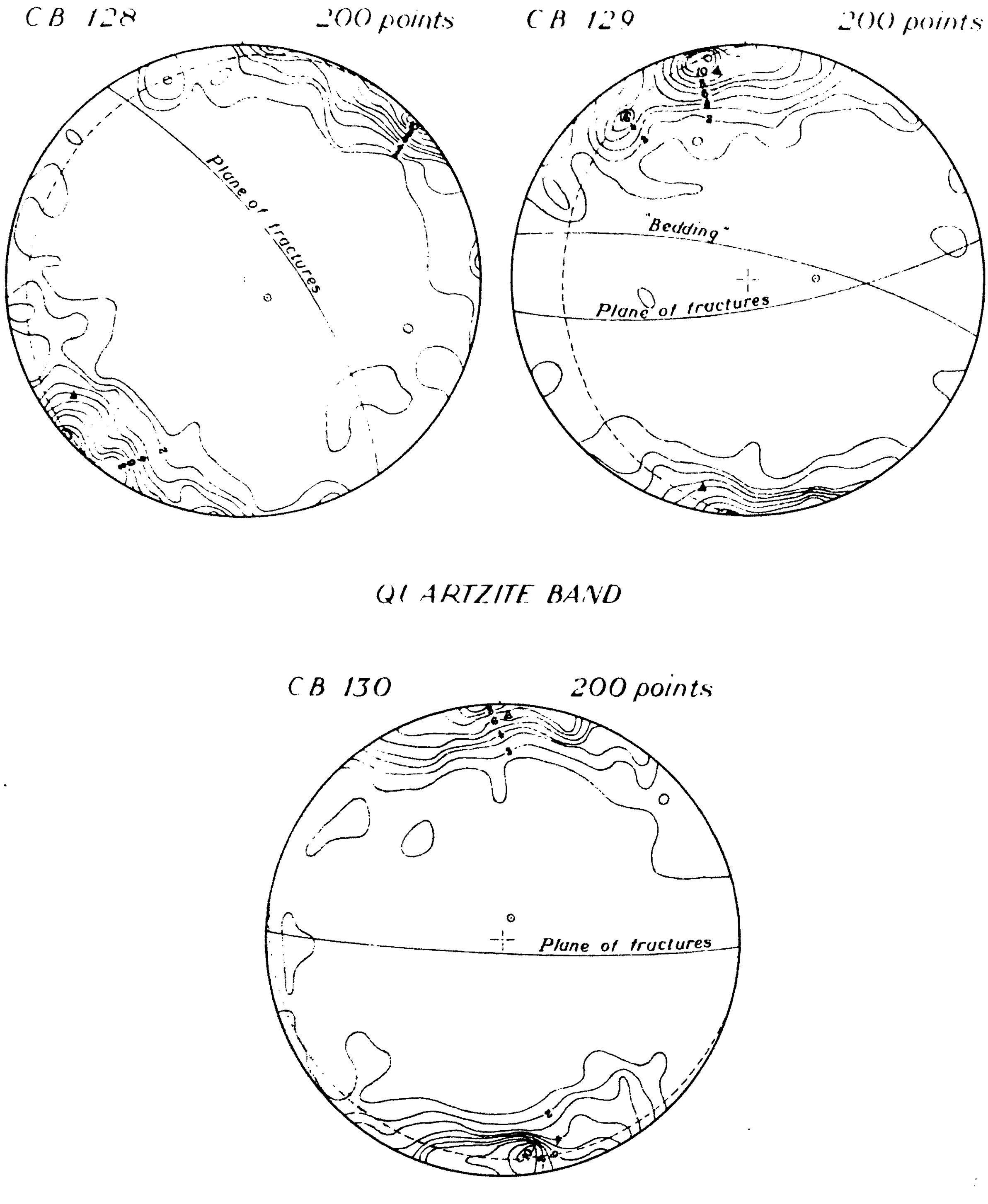


Fig. 43. Mica diagrams of CB 128, 129 and 130.

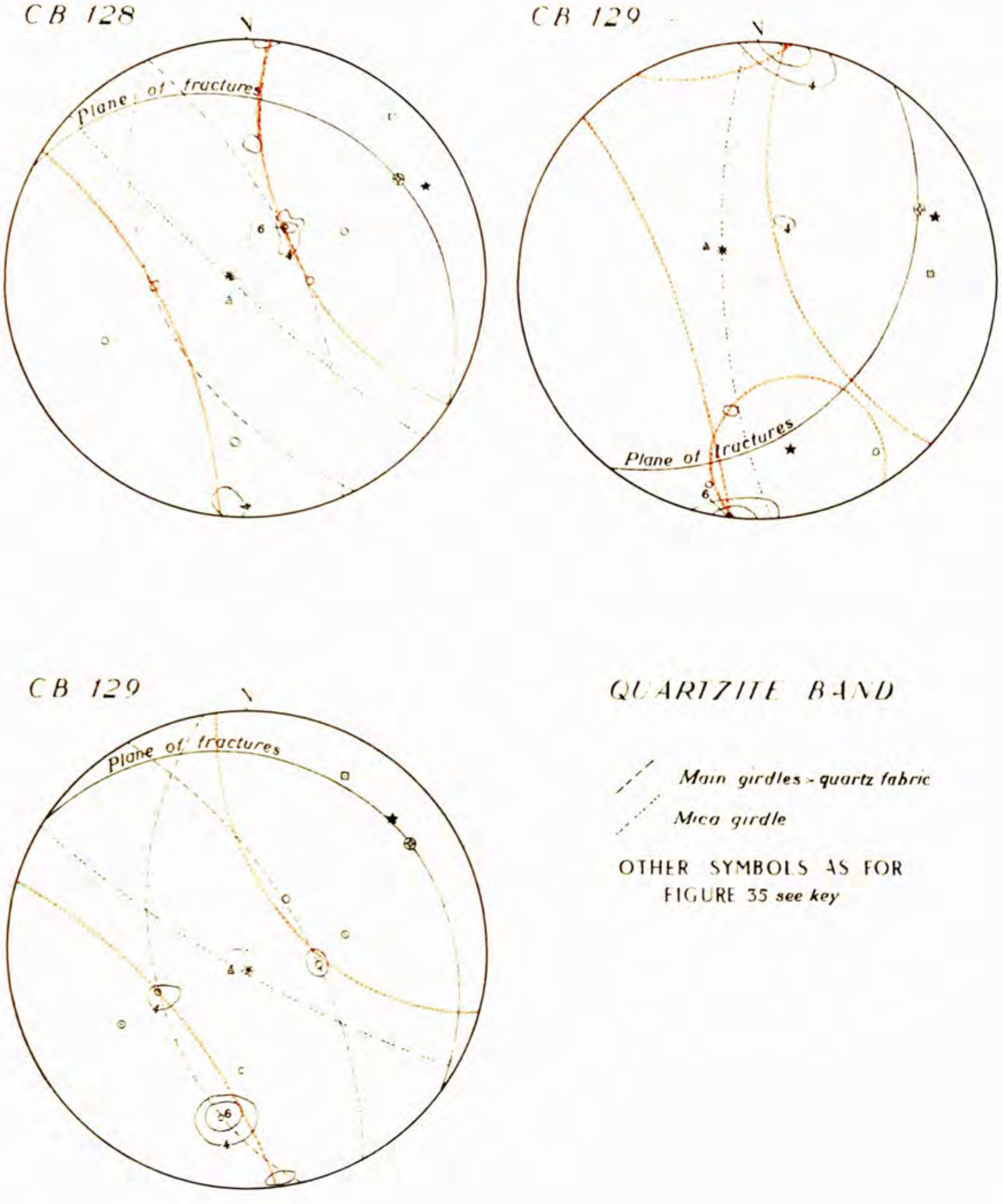


Fig. 44. The principal elements of stereograms in figs. 42 and 43 rotated to their true position on Wulff net.

of the small circle fall more readily into separate partial girdles. When rotated to their original geographic position these suggested elements show fair agreement among the three samples (fig. 44).

In CB 129 there is a concentration of points around the position indicated by a star. This position lies $90^{\circ}\pm 5^{\circ}$ from the centre of the small-circle (= the b-axis) and within the plane of shearing. In the sense that this structure is of small radius and lies at 90° from the megascopic fold axis there is agreement with the findings of Dhonau (1961) in a study of a refolded Moine rock. He, however, identifies an earlier incomplete girdle rather than a small-circle, but the general pattern of his contoured diagrams is very similar to those now presented. Dhonau concluded that as the folding progressed so the early girdle was replaced by a small-circle about a. From the position of the shear- or fracture-planes, and of the mica maximum in the mica fabric, the a-axis in CB 129 can be fixed. It coincides precisely with the centre of the structure just described.

The mica diagrams of the three specimens reveal additional interesting features. Each shows a good point-maximum on a broken girdle. Two examples are identical in every respect, but the third, from the middle limb of the fold (CB 129), differs in the position of both the point maximum and the girdle. The specimens all agree, though, in having their mica-maxima centred on the corresponding pole to the shear-planes. Figure 45 demonstrates that after a hypothetical rotation of these shear-planes to a coincident position, the point maxima would all agree, but the appropriate girdles would still be oblique to one another, rather than being centred on the b-axis. At this stage the conclusion must surely be that the orientation of mica is controlled mostly by linear shear-movements, as against the rotational ones responsible for the quartz fabric.

However, the rotation mentioned above would affect the common attitude of the great-circles in the quartz-fabric. It is concluded that the great-circles are not directly related to the formation of the shear-fractures,

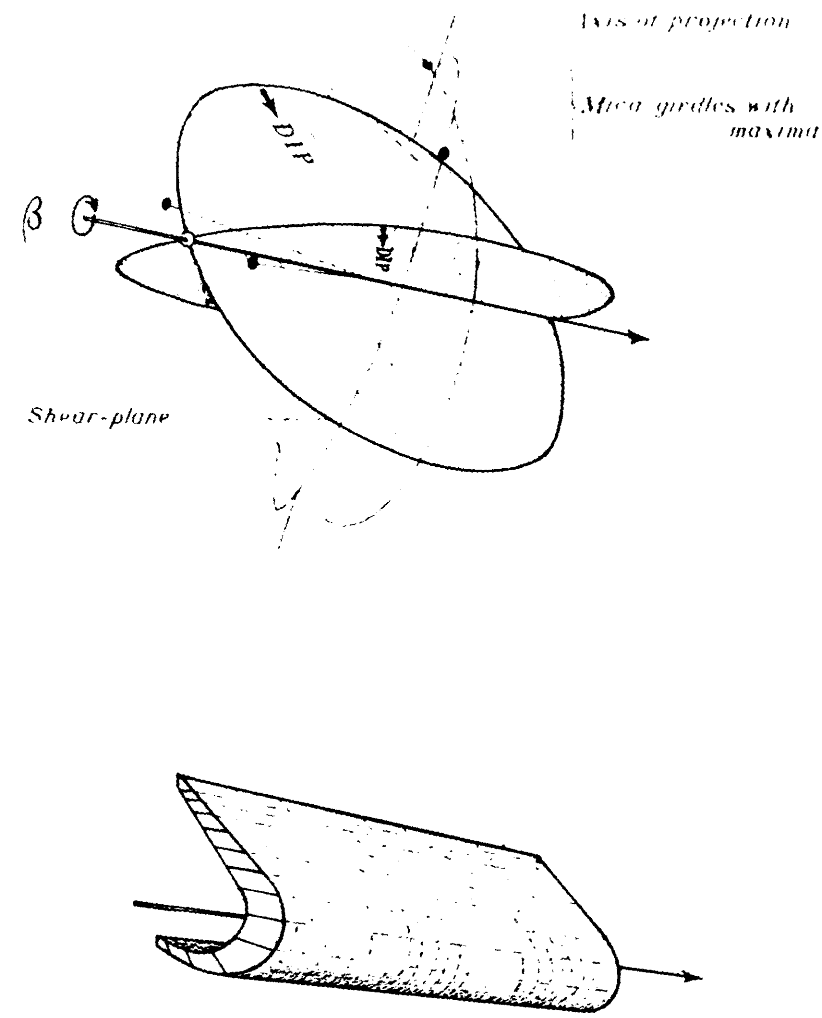


Fig. 45. Three-dimensional view of the shear-planes and mica girdles with respect to the upper and middle limbs of the fold.

and are thus of early origin, possibly remnants of F_{B1} deformation.

Finally these results indicate that at least for the F_{A2} deformation the petrofabric data can be directly related to the field evidence of folding and shearing, and that more than one deformation is required to account for the variation in rock fabrics in the area.

PART IV

METAMORPHISM

Introduction

Metamorphic effects in the rocks around Braemar have been studied mainly in the pelites. These belong to the Ben Eagach "Black Schist" and the Blair Atholl Series. Some are associated with the Perthshire Quartzite Series.

In order to illustrate the metamorphic history of the area comparisons will be made between the schists of the area mapped and those of the wider region extending to the southwest. This will facilitate a study of the metamorphic changes to be found going from seven miles to the southwest of the Lochnagar Granite, right to the margins of the granite mass itself.

Previous Work

One of the few significant works published on this aspect of the geology is the "Explanation of Sheet 65", in the Memoirs of the Geological Survey, Scotland, by G. Barrow and E.H. Cunningham Craig, in 1912. This account does not present a clear picture as regards the main cause(s) of the metamorphism, and the distribution of the varying types of metamorphism. It is recognised that the metamorphism is generally easily separable into an earlier regional phase, and a later thermal phase (caused by the Newer Granite of Lochnagar, etc.). The regional metamorphism is considered to be "Dynamic and Constructive"; the former depends largely on the type of folding, producing such features as foliations, strain-slip schistosities, and mullions; the latter is almost entirely thermal in nature (not to be confused with the later contact thermal episodes).

The memoir leaves the impression that while some of the Newer Granites

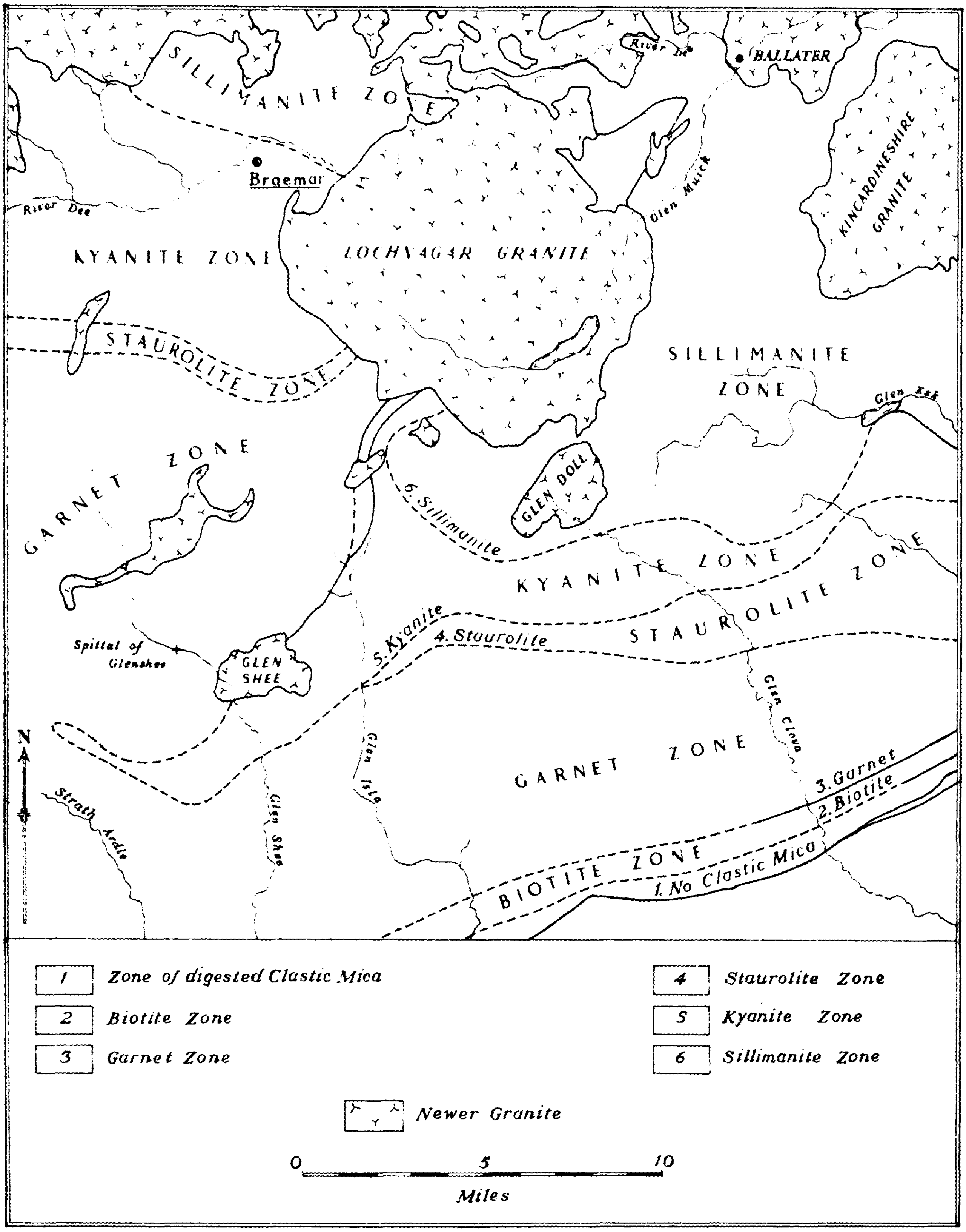


Fig. 46. Barrow's Zones: after G. Barrow (1912).

in the Scottish Highlands have good contact aureoles, with the development of hornfelses, the Aberdeenshire masses do not. Sillimanite-bearing rocks (those raised to the highest temperatures) have a regional distribution independent of the granites, while the sillimanite-hornfels, although more restricted, "... does not occur as a continuous aureole round any of the Newer Granite Masses of this area" (p. 110).

However, as regards the Braemar area, it is apparent that Barrow in fact discovered otherwise, for he found that "... the effects of a subsequent contact metamorphism ... supersedes (sic) and obliterates (sic) any effects of the regional type in the neighbourhood of the plutonic intrusions of the newer magma" (op.cit., p.103). Furthermore, in the "Summary of Progress", 1897, p.49, and in the main publication of 1912, p.106, it is observed that around, and to the north, northeast, west and southwest of Braemar, "... the area over which the alteration into hornfels occurs, has been found by Mr. Barrow to be substantially identical with that in which the intrusions of granite have taken place". It must be noted, though, that sillimanite is not mentioned here, but on p.105 of the 1912 publication it is stated that sillimanite, cordierite and andalusite are typical of the hornfelsed "Black Schist" to the north and northeast of Braemar.

According to the Memoir, andalusite occurs up to four miles south of Braemar, kyanite up to six miles, and staurolite up to seven miles. Garnet is abundant in most areas of sheet 65, south of Braemar, while cordierite is rare. Sillimanite is not mentioned here. The Memoir further states that the mineral isograds meet the Lochnagar Granite almost at right angles. These are presumably the boundaries of Barrow's Zones, as shown on his map of 1912 (fig. 46). The Survey would thus seem to have recognised the zones of regional metamorphism, with the superimposed zone of contact thermal metamorphism, as witnessed by the occurrence of andalusite and cordierite, and also by the development of hornfelsing.

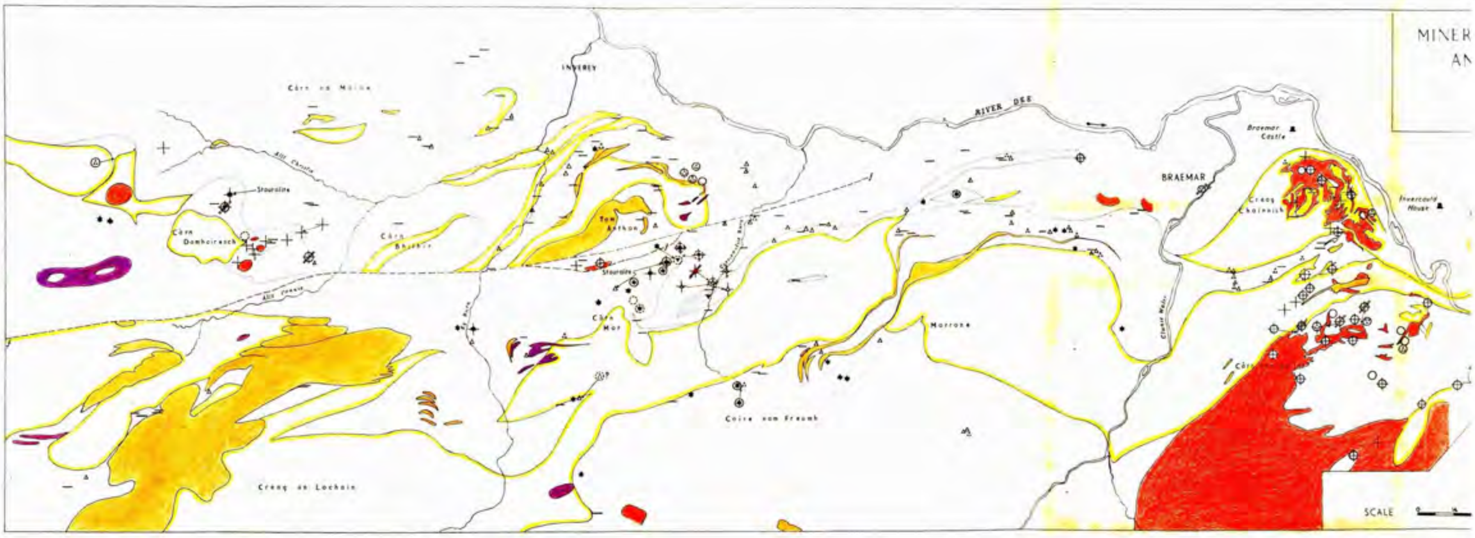
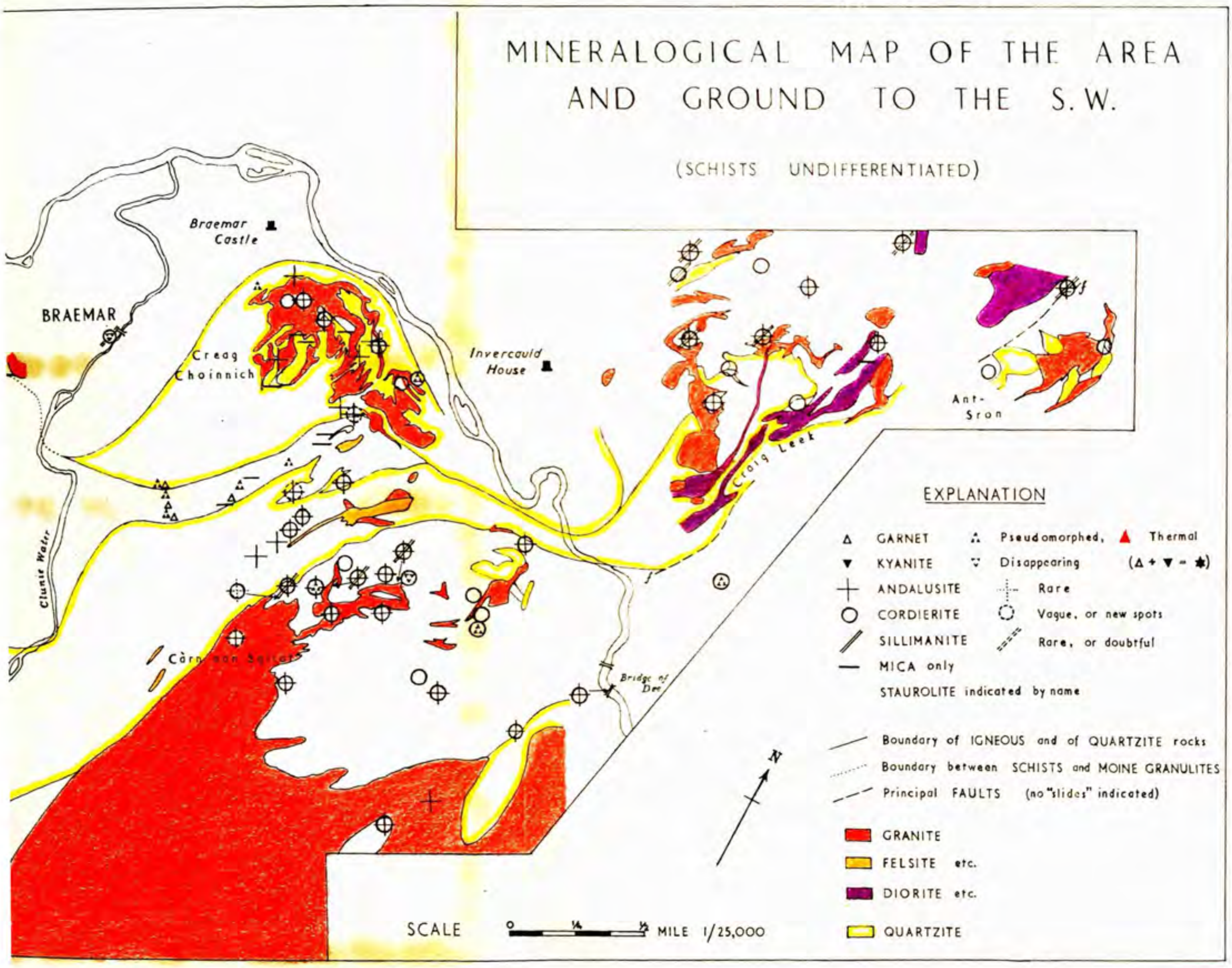

MINER
AN
INVEREY
Càrn na Moine
Staurolite
Càrn Dambhaireach
Càrn Bhithir
Tom Anthon
Staurolite
Càrn Mor
Creag an Lochain
Coire nam Freumh
RIVER DEE
Morrone
BRAEMAR
Braemar Castle
Creag Choinnich
Invercauld House
Clunie Water
SCALE

Fig. 47


MINERALOGICAL MAP OF THE AREA AND GROUND TO THE S.W.
(SCHISTS UNDIFFERENTIATED)
Braemar Castle
BRAEMAR
Creag Choinnich
Invercauld House
Ant-Sron
Craig Leek
Clunie Water
Càrn nan Sgitar
Bridge of Dee
EXPLANATION
GARNET
Pseudomorphed,
Thermal
KYANITE
Disappearing
ANDALUSITE
Rare
CORDIERITE
Vague, or new spots
SILLIMANITE
Rare, or doubtful
MICA only
STAUROLITE indicated by name
Boundary of IGNEOUS and of QUARTZITE rocks
Boundary between SCHISTS and MOINE GRANULITES
Principal FAULTS (no "slides" indicated)
GRANITE
FELSITE etc.
DIORITE etc.
QUARTZITE
N
SCALE
0
¼
½ MILE 1/25,000

recognises purely contact-thermal effects in the hornfelsed limestone, caused by Newer Granite.

In his paper on the composite gneiss of Glen Shee, W.O. Williamson (1935) mentions that while his area has many similarities with Cromar, no andalusite, cordierite, and sillimanite have developed during the regional metamorphism associated with the emplacement of the Older Granite there. Instead, these three minerals are produced in the thermal aureole of the Younger Granite (granodiorite), where it cuts the composite gneiss and associated schists, and hornfelses them.

Thus, previous work presents conflicting ideas regarding both the nature of, and sequence of events in, the metamorphism. The present study brings a great deal of additional data to bear on these problems, and, although many uncertainties remain, a more precise picture has emerged.

General Distribution of Metamorphic Minerals

Strictly speaking a mineral facies classification within the area is not possible, due to the polyphasal nature of the metamorphism. For instance, assemblages characteristic of regional metamorphism exist in unstable equilibrium with those characteristic of thermal metamorphism, while often retrogressive metamorphism has added to the complications.

The map (fig. 47), showing the distribution of the principal metamorphic minerals, illustrates that as a first approximation the "regional" minerals found to the southwest of Braemar give way somewhere around Braemar itself to "contact" minerals east and northeast of Braemar. In other words, as a generalisation, almandine-amphibolite facies schists are transformed to pyroxene-hornfels facies rocks towards the granite masses of Upper Deeside.

In more detail it is evident that pyroxene-hornfels conditions exist very locally within the "regional" area (part of the classical Barrovian area), as the distribution of cordierite, andalusite, and rare sillimanite

Plate 69. Garnet porphyroblasts enclosing quartz crystals aligned at right angles to the external fabric. KB 18; x 80.

Plate 70. S-shaped fabric in garnet porphyroblast. KB 167; x 80.

indicates. The map further shows that this fact has often a connection with visible outcrops of granite (Newer Granite) and/or hornfelsing.

The Barrovian area also exhibits a poorly-defined mineralogical zoning, which will become more evident as changes are traced in the schists.

Of the minerals indicated on the map, only sillimanite is not readily reliable as a reflection of the conditions of its growth, since (1) it can occur as the highest grade of both regional and thermal metamorphism, and often simply recrystallises when the latter is superimposed upon the former; and (2) its general occurrence here may, in fact, be considered to coincide with the highest zones of both types of metamorphism.

Petrography and Petrology of Pelitic Rocks

There are three convenient sub-areas: (a) the main Barrovian area, between Càrn Damhaireach and Braemar; (b) the intermediate zone of Creag Choinnich and Càrn nan Sgliat; and (c) the "contact" zone, mainly north of the Dee.

(a) Càrn Damhaireach to Braemar

(i) "Garnet" zone. A typical specimen, KB 496 is a medium/coarse grey schist with alternations of granulitic and micaceous foliae. Suitable surfaces reveal numerous small garnets.

In thin section banding and schistosity are seen to be parallel, though in the coarser, micaceous part a new strain-slip is developing. Here the mica (muscovite) mostly lies parallel to the fold limbs but is cut by new growth parallel to the embryonic strain-slip cleavage. Some of the older mica has been influenced by earlier fold directions which have not quite been destroyed. The hyp-idioblastic garnets contain small quartz inclusions. These, and unidentifiable dark ones, impart a distinct fabric to their hosts, lying oblique to the external fabric (pl. 69) and occasionally forming an s-shape (pl. 70). Rarely the internal garnet fabric is seen to have been

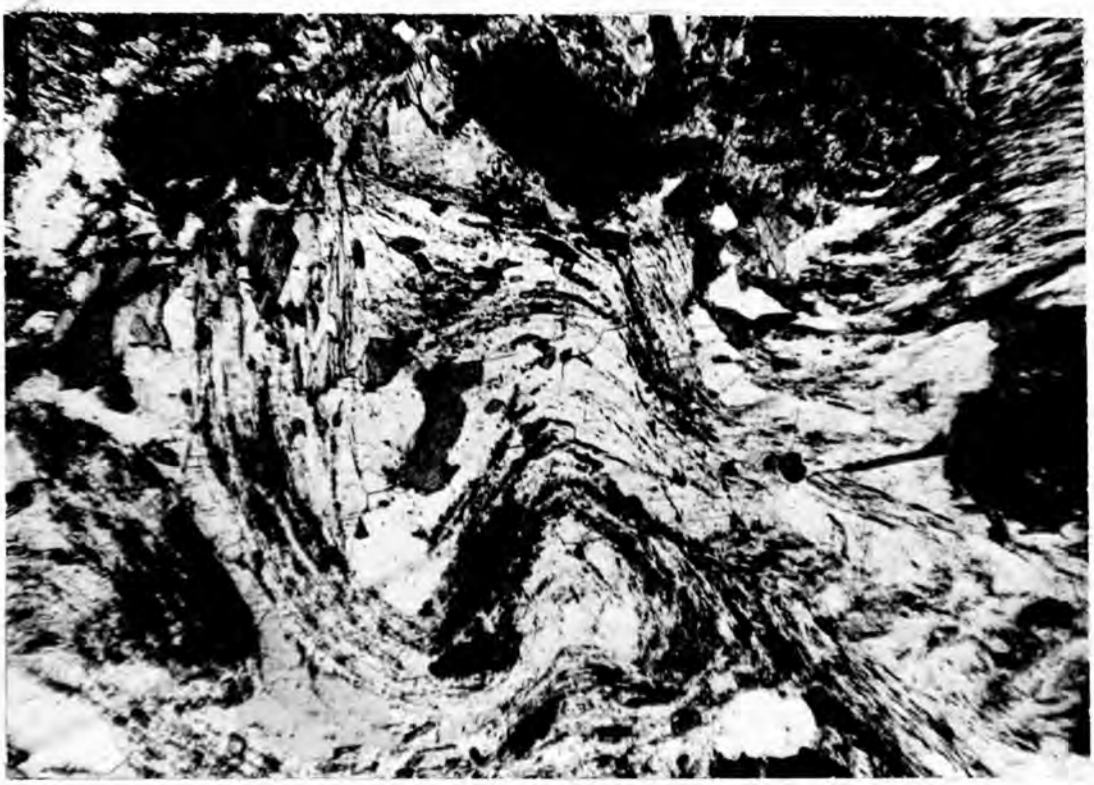


Plate 71. Typical "Black Schist" in which fine inclusions, overgrown by plagioclase, reveal evidence of early folding. KB 127; x 35.

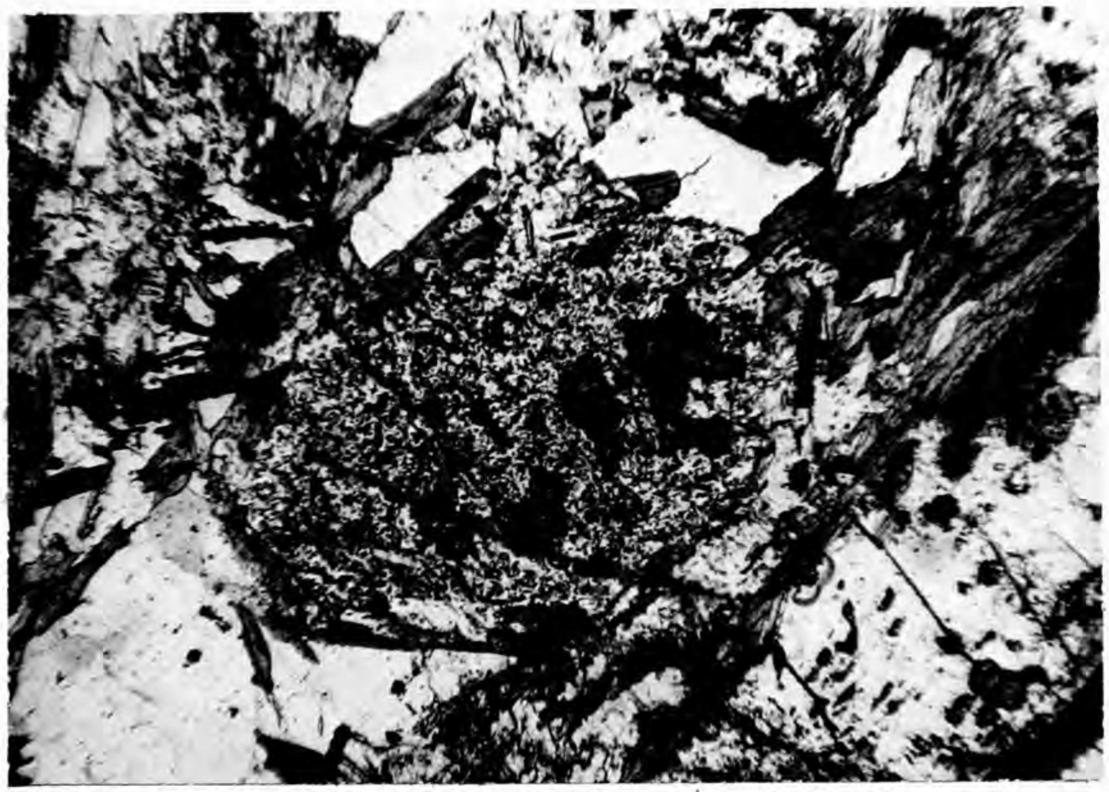


Plate 72. Oval-shaped porphyroblast of cordierite after garnet, fragments of which are still preserved. KB 127; x 80. (See also plate 98).

folded. Relative to the growth of garnet, therefore, there has been an earlier and a later deformation. The rest of the coarse part of the rock is quartzo-felspathic, but of coarser grain than the internal fabric of the garnets, as a result of post-garnet recrystallisation. Small apatites and tourmalines complete the assemblage.

In the finer, more siliceous part of the rock, the garnets are much more skeletal due to the abundance of quartz inclusions, but the evidence of deformation is similar to that just described.

In other specimens garnet is often more replaced by chlorite, and inclusions tend to be less revealing of the pre-garnet fabric, but the micas still lie in curved paths round the garnet crystals confirming the existence of a later deformation.

(ii) "Kyanite" zone. KB 127 is representative of a black schist from this zone. In essence it has a complex microstructure overgrown by large plagioclase felspars which are beginning to be replaced by cordierite. The hand specimen has a leafy appearance due to the presence of much medium/coarse biotite lying in the limbs of ill-defined micro-folds or parallel to a new schistosity.

It is seen in thin section to be mesocratic rather than melanocratic due to the abundance of quartz forming a sort of groundmass to the reddish brown biotite. Throughout this are areas of highly folded trails of graphitic(?) inclusions with parallel prismatic tourmalines and specks of magnetite. Xenoblastic kyanite crystals up to 5 mm. long cut across and enclose these folded trails, some kyanites having already cleared themselves of foreign material. This index mineral is thus post-first folding, but the occurrence of ragged, fragmentary, and even warped kyanites proves that deformation has been relatively late also. Some syn-kinematic crystals have been found lying parallel to the new schistosity.

Garnet presents a similar situation though it is less abundant and

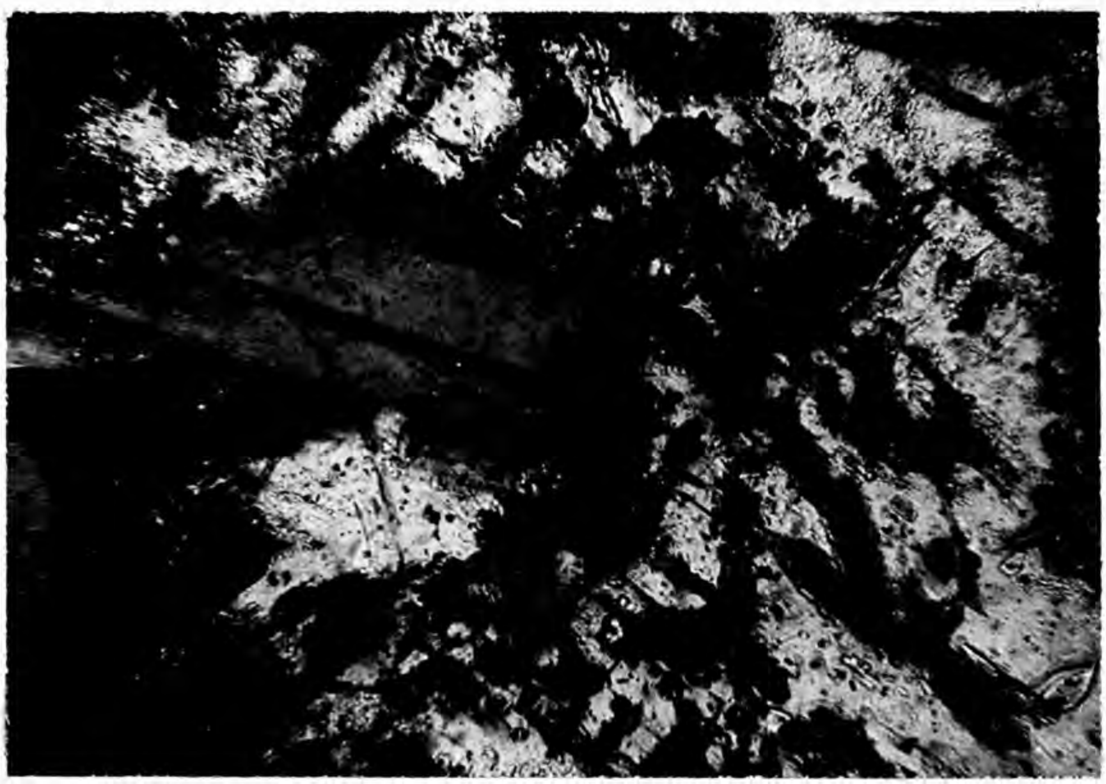


Plate 73. Cordierite (light grey) surrounding plagioclase (twinned). Note the blunt fingers of cordierite penetrating the felspar, top left. KB 127; x 230.

Plate 74. Small diamond-shaped sections of andalusite overgrowing the opaque material. KB 140; x 35.

highly skeletal. Its original extent is shown by replacive biotite and sericite, while dark inclusions are found concentrated only in the centre of garnet crystals, many of which are themselves set in larger crystals of plagioclase and/or cordierite (pl. 72).

The late predatory nature of the large plagioclases (oligoclase) is proved by their clear, straight, multiple pericline twins. In the absence of the latter the felspar and the cordierite may be impossible to distinguish, but other characteristics help to separate the two. Higher birefringence, susceptibility to replacement, rare sectorial or irregular twinning and the presence of weakly pleochroic yellowish inclusions are all indicative of cordierite. In contact with felspar its boundaries are irregular, the impression gained being always that it is "replacing" the felspar, in the sense that it is taking over the structural position of the latter (pl. 73).

Difficult though cordierite is to identify positively, the writer is satisfied that cordierite is present in many other "Black Schist" specimens, as indicated on the mineralogical map.

Another "Black Schist", KB 140, helps to complete the picture. Microscopic examination shows it to be broadly similar to the above, but with garnet dominant over kyanite. The most interesting differences, though, are the presence of staurolite, the greater development of cordierite, and the presence also of very small idioblastic andalusites which cross-cut the graphitic particles (pl. 74). Some of the latter are enclosed in the shape of the chiastolite cross. Ragged earlier crystals of andalusite have slightly different optical properties from the later ones.

Folding in this slide is more regular, possibly later than that in KB 127, since the few remaining kyanites are smaller and clearly "follow" the folds. The staurolite rods also have this tendency. Some of the cordierite is strained, while gently bent twin lamellae prove that some of the plagioclase has been slightly deformed.

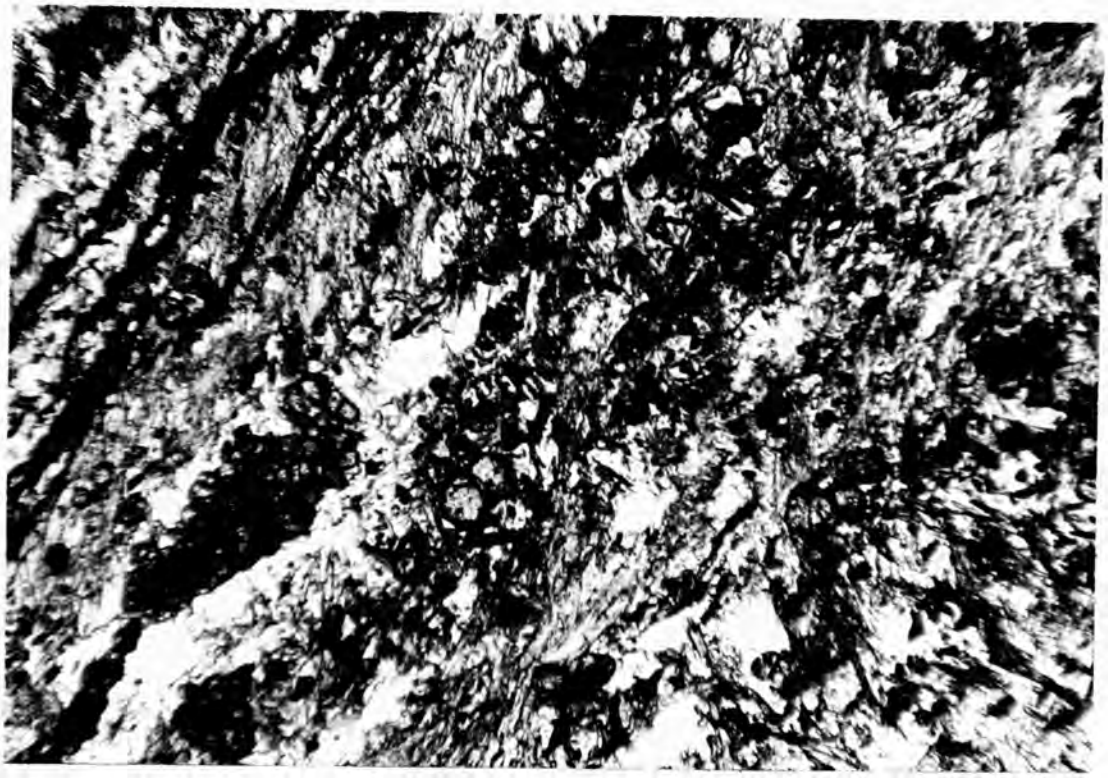


Plate 75. Clusters of small thermal garnets. KB 115; x 80.

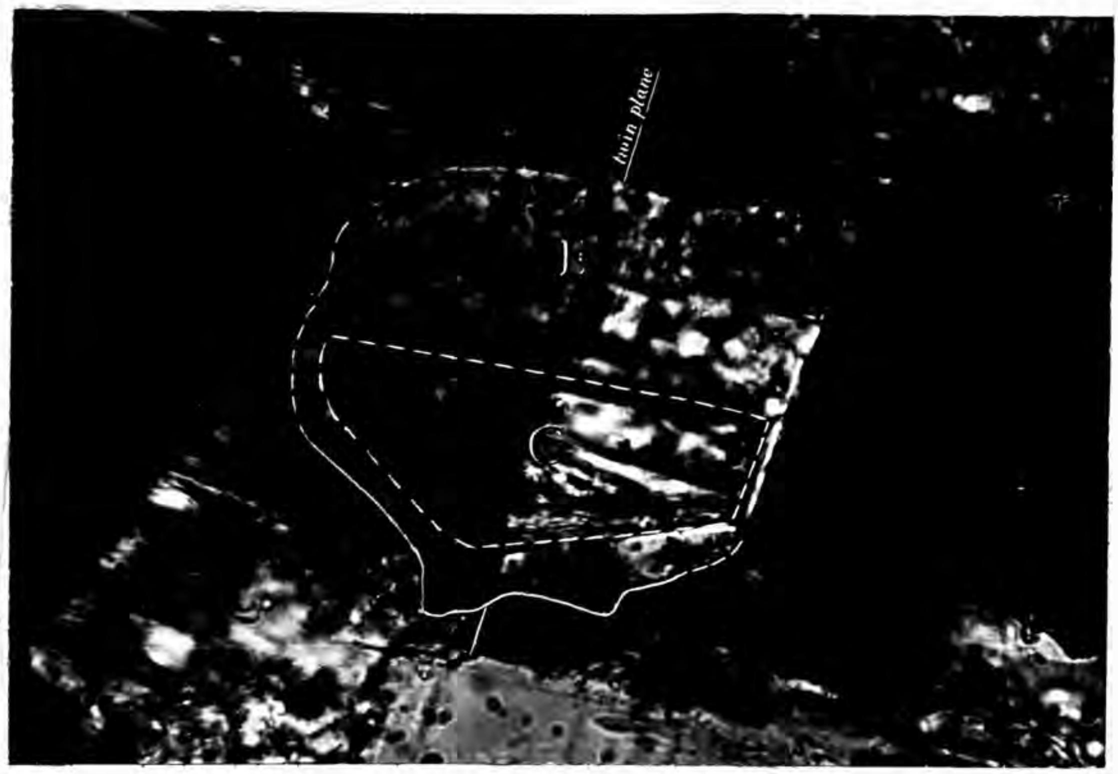


Plate 76. Twinned clinozoisite (C) overgrown by epidote (E). This is probably a fragment of a much larger crystal. KB 78; x 1200.

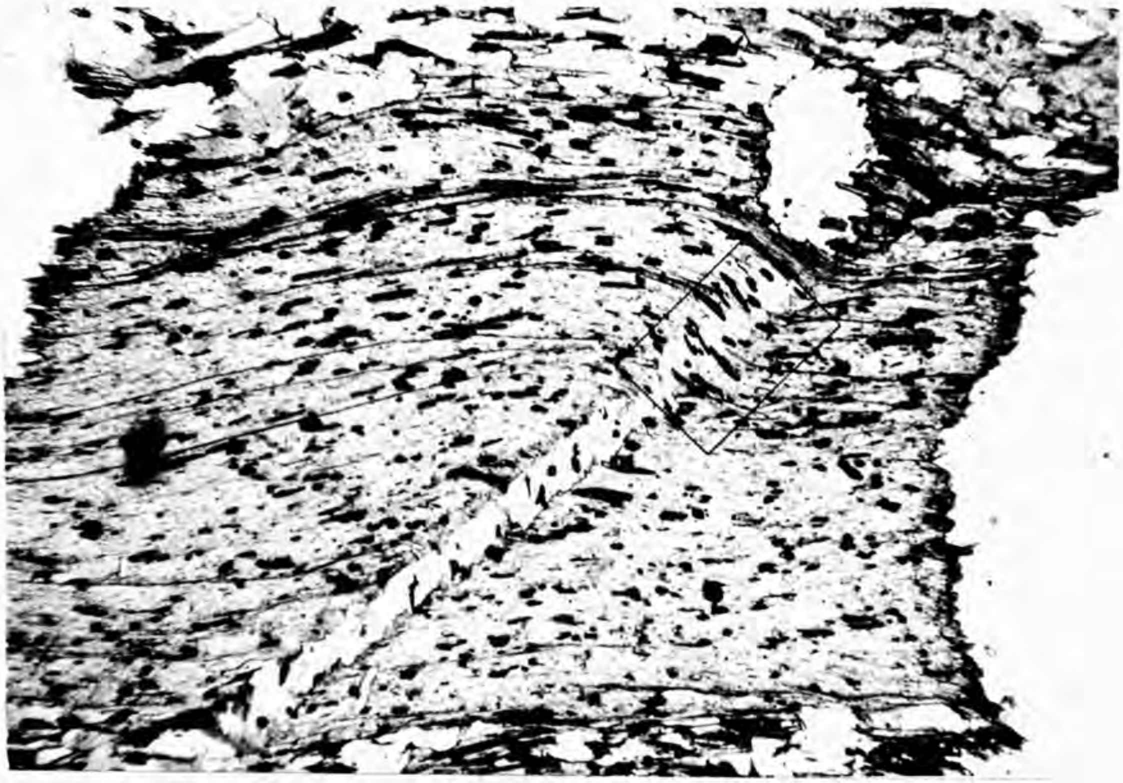


Plate 77. Deformed megacryst of muscovite enclosing small prisms of epidote, also deformed. KB 80; x 35. (For close-up of area shown, see plate 78.)

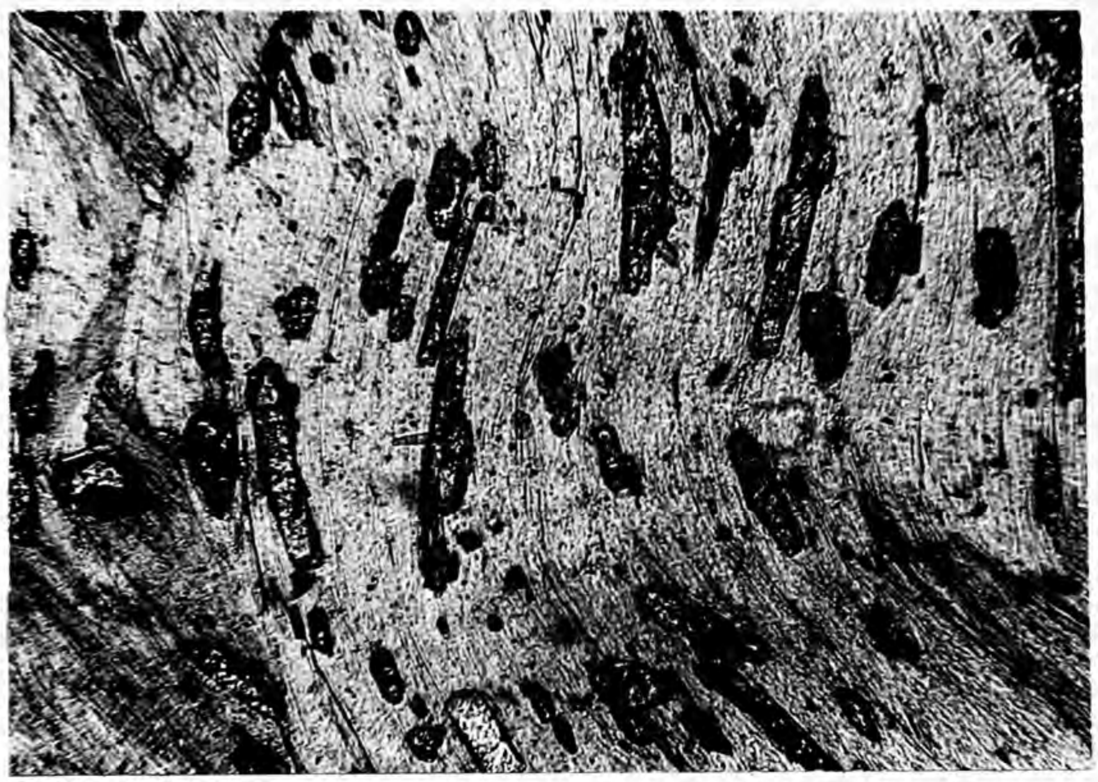


Plate 78. Close-up of area shown in part of plate 77, to illustrate the bending of epidote crystals. x 230.

Lastly, particular mention must be made of KB 115, as it shows a unique feature among the thin sections examined - the presence of numerous small thermal garnets, which, unlike the "regional" garnets, are almost idioblastic and clearly stable. Also, the crystals are grouped in clusters distributed along the banding in the rock, and individuals show no signs of rotation during growth (pl. 75). Not only is the garnet of thermal origin in this case, it is also of later development than the "usual" garnet. Other minerals in the assemblage include reddish biotite flakes of haphazard orientation. These lie within large ragged andalusites and within material much like strained cordierite but which is in fact orthoclase*.

Turning now to the more calcareous and often granulitic rocks, it has been found in the kyanite zone that minerals of the epidote family have developed in place of staurolite, kyanite and probably some garnet. Two generations have been observed, for instance in KB 78, in which twinned crystals of clinozoisite are overgrown by epidote (pl. 76). It can sometimes be proved that folding took place after the formation of parallel lines of epidote. In KB 80 large plates of sharply flexured muscovite enclose such lines of epidote, and when examined closely the epidotes, too, can be seen to have suffered bending (pl. 77, 78).

(iii) Contact-altered schist near granite. The typical rocks just described under (i) and (ii) give way at certain localities to tough grey hornfelsic schists, which contrast with the dark leafy schists. The salient points of difference are typified by KB 474, in which 20% of the section comprises fine alteration material of sericitic and pinitic affinities, which encompasses most of the other minerals. Kyanites, small and sporadic, occur only in this material. Their size and imperfect shape point to their waning nature. Staurolite and garnet, when found, are also fragmentary and unstable. On the other hand the micas show development under conditions of hornfelsing.

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* It is interesting to note here that orthoclase is often intimately intergrown with cordierite in the hornfelses north of the Dee.

Plate 79. Partially skeletal porphyroblast of andalusite enclosing biotites (grey to very dark) and terminating some but not all crystals of biotite at its margins. KB 463; x 80.

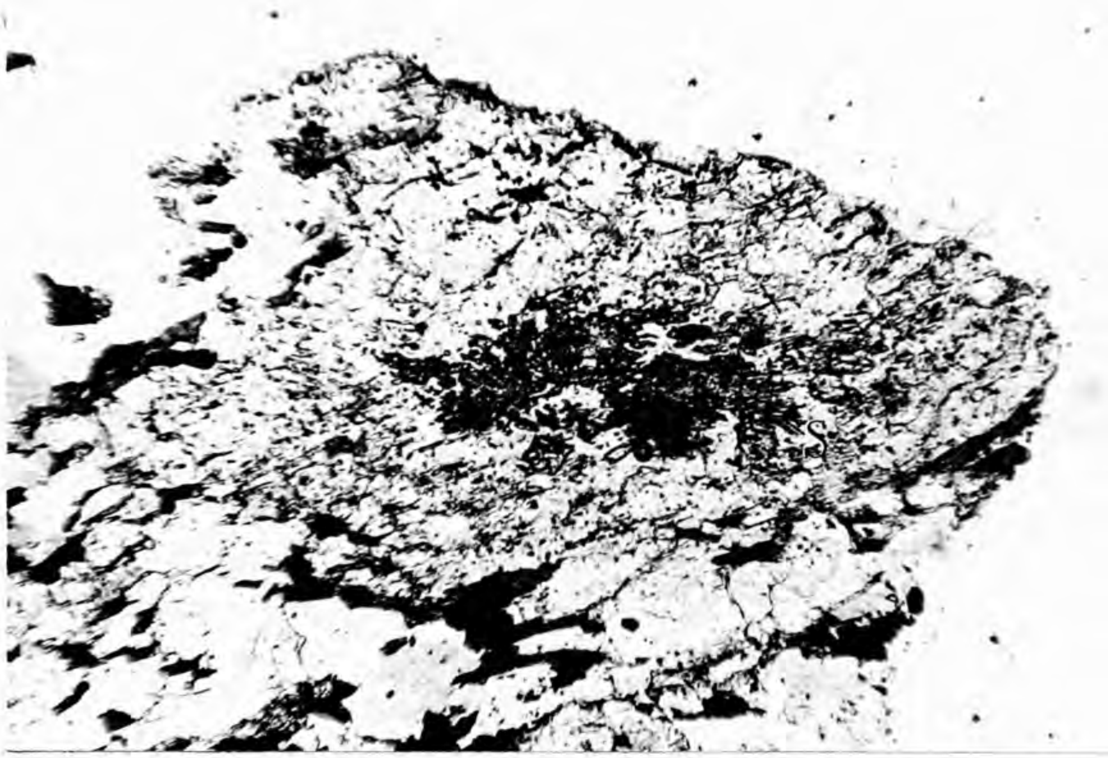


Plate 80. Remnant fragments of kyanite (centre, high relief) enclosed in a large porphyroblast of cordierite. The needle-like crystals (S) are of sillimanite. KB 152; x 80.

Most important, however, is the presence of large post-schistosity andalusite porphyroblasts. This mineral has mostly absorbed inclusions of biotite, and must have developed after the main regional metamorphism. Taken in conjunction with the fact that a number of neighbouring specimens show similar late crystallisation features and all lie near to small outcrops of granite, the evidence supports the view that a late contact thermal metamorphism has been superimposed upon a regional one.

There is also evidence to show that schists hornfelsed by close proximity to granite have not been entirely free from stress during the thermal process. In KB 463 large andalusite porphyroblasts, overgrowing folded trails of biotites, are themselves gently influenced in their growth-direction by later warps. Many biotites are abruptly terminated at the edges of the andalusite crystals, but some continue as selvedges into these porphyroblasts, and many are completely enclosed (pl. 79). This is interpreted as an illustration of the very fine balance between the disappearance of biotite to produce andalusite and the continuing growth of the existing biotite.

Other possible thermal effects can be seen in KB 152, from under Braemar Bridge, which cannot be far from in situ granite. In this schist, fine granular and irregularly-shaped kyanites are enclosed in slightly strained cordierite crystals. From out of the kyanites there emerge needles with straight extinction and other properties much like sillimanite. In view of its situation the sillimanite is regarded as of thermal and not regional origin. See pl. 80 .

(iv) Retrograde schist. As well as being chloritic, a considerable percentage of schist specimens in this part of the area are partly composed of pinitic and sericitic alteration material. On account of the general distribution of such schists these features are believed to result from retrogressive metamorphism, rather than advanced weathering. Although nothing

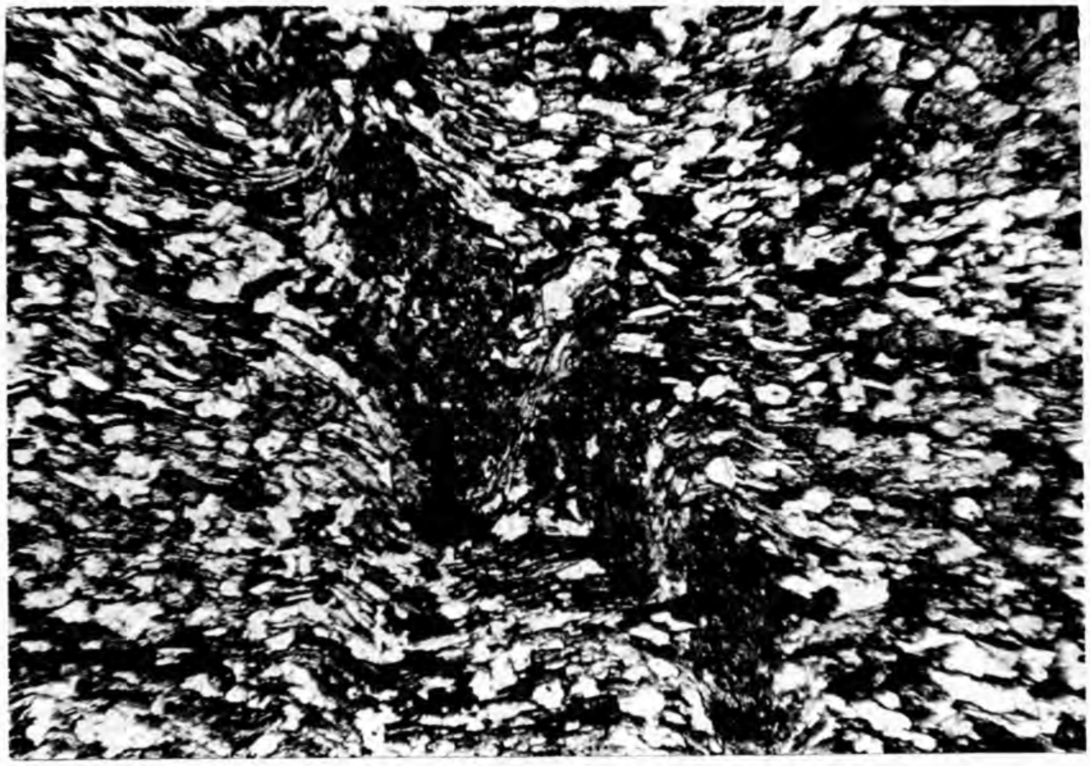


Plate 81. Pseudomorphs after garnet deformed by extension in the axial planes of microfolds. KB 224; x 35.

remains of the original minerals, they are believed, from the shapes preserved, to have been garnet and cordierite. Retrogression can reach the stage where everything in the rock but quartz has been replaced by chlorite (eg. KB 119), yet the fabric of the rock is still beautifully preserved as seen in thin section.

(b) Creag Choinnich and Càrn nan Sgliat

(i) Schist from "intermediate" zone. The uninteresting aspect of most schists here has its cause in the disappearance of most of the diagnostic regional metamorphic minerals before the appearance of thermal metamorphic ones. The history of events can be traced in a study of thin sections from specimens selected at varying distances from the zone of granite intrusion on Càrn nan Sgliat.

Farthest away, KB 224 still contains pseudomorphs after garnet porphyroblasts. These are evenly distributed and made up of sericite and biotite. Their shape shows them to have been gently deformed by extension in the axial planes of regular flexure-like microfolds (pl. 81). The original garnets caused deflection of the schistosity and are thus pre- to syn-kinematic.

In KB 363, nearer the granite, "ghost" porphyroblasts are much fewer and of two types. The first are concentrations of mica after garnet. The second are also circular in outline but variable in size, being composed of flesh-coloured alteration typical of cordierite. These structures cross-cut and overgrow the schistosity, and hence must be relatively late in origin. The presence of andalusite in this rock further convinces the author that these pseudomorphs are after cordierite. Around and within them, and the larger andalusites, biotite takes the form of rod-like crystals, which, from their common orientation and fine elongated profile are suggestive of new biotite. This problem will be dealt with later.

Lying even closer to the intrusive zone, CB 109 presents evidence of further metamorphic developments. These can be well appreciated in hand specimen in the presence of large ovoid spots (new cordierite) up to 1.5 cm. long, and "flattened" in the plane of the schistosity. In thin section the spots can be seen to be defined by an outer zone in which biotite has been partially or completely absorbed. The next zone is, in contrast, richer in biotite and muscovite than even the main part of the section. The biotite is also of redder colour and coarser. In the innermost zone andalusite porphyroblasts cross-cut the now sparser biotites, but are themselves highly sutured by the fine-grained cordierite and associated quartz. These two minerals occur throughout the "spots". The alignment of the micas within is as for the rest of the section, but that of the andalusite crystals is seemingly haphazard.

The problem here (as will be found in schists from north of the Dee) is the apparent absorption of the micas - particularly biotite - by the growing cordierite spots in such a way that the outermost zones of the spots are always devoid of mica yet the inner ones are not. If the spots grow constantly (outwards) then the mica must be peripherally dissolved and then later regenerated. The fact that the central micas are coarsest cannot reasonably be explained otherwise. Where, however, the coarser mica is not present in certain other cordierite-bearing rocks, it could be argued that zones with biotite result from fast growth of the cordierite porphyroblasts, while zones bereft of biotite indicate that cordierite has had time to digest the mica.

Another difference between this slide (CB 109) and the previously-described one (KB 363) is the great abundance of small idioblastic tourmalines which have developed rigidly parallel to the schistosity and which, too, are sparser in certain zones of cordierite spots.

(ii) Contact hornfels. CB 120, in contact with marginal parts of the Lochnagar granite, is a distinct, partially granitised hornfels, consisting

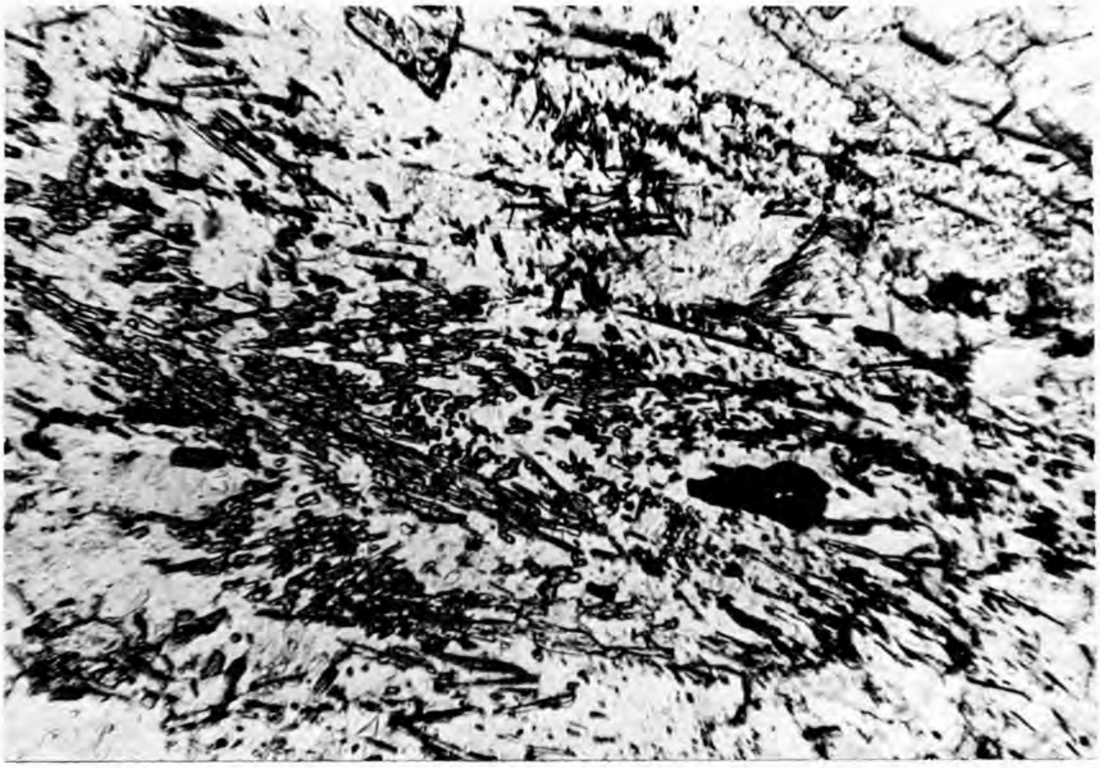


Plate 82. Later sillimanite needles intimately associated with granules of andalusite, all set in cordierite. KB 179; x 80.

of irregular alternations of quartz-rich areas, granitic areas and areas of alumino-silicate minerals. In the last-mentioned parts, the highest-grade thermal metamorphic effects of this locality can be found.

The distribution of the different areas mentioned indicates the presence of tight isoclinal folds. Within the pelitic parts most of the material is of highly granulated andalusite rods, closely "following" the folded schistosity, and often broken down into myriads of tiny granules. Gradations in size and shape indicate that andalusite porphyroblasts have been folded and crushed to effect both a schistose and a sugary appearance at the same time. A small proportion of sillimanite needles is intimately associated with this mass, the two minerals being at first sight identical. In contrast to the andalusite, however, the sillimanite needles prove to be of perfect prismatic shape and to diverge in their orientation from their isomorphs. At certain points in the section sillimanite has grown from andalusite. Cross-cutting relations can also be seen, with sillimanite invariably the later mineral. The above evidence is believed to result from the sillimanite developing from the already deformed andalusite, but being controlled to some extent in its orientation by that of the andalusite rods and granules.

Hornfelsic biotite in this environment is now in the process of being digested, and is significantly sparser than in the quartz-rich parts of the section. This is almost certainly due to the presence of cordierite which encloses all other minerals in the alumino-silicate zones. The cordierite lies in poorly-defined strips of different optical orientation which have beyond doubt been folded. A few such folds are wrapped round circular areas devoid of andalusite and sillimanite. Instead, randomly-orientated micas are abundant, and enclosed by cordierite. Rarely xenoblastic garnet may be present in the centre, having been reduced by replacement to one fifth the diameter of its original crystal size.

CB 120 thus summarises the metamorphic history of the schists around

Braemar, garnet representing the initial regional phase, cordierite and andalusite the thermal phase followed by further deformation, and sillimanite the largely post-kinematic thermal phase of greater intensity. It has been observed that in such rocks as this, tourmaline has now completely disappeared. Further evidence of this mineral and of the development of andalusite must be sought on Creag Choinnich.

(iii) Creag Choinnich. Schists from this zone are structurally and metamorphically behind those on Càrn nan Sgliat. A noticeable lack of hornfelsing is evident even in hand specimen. Sections of various specimens reveal some of the andalusite present to be highly sutured and strained, yet other crystals of it may have excellent crystal outlines. Some of the latter class have a central zone which, viewed between crossed nicols, has an anomalous blue colour. This implies two periods of growth, whereas the other type of crystal implies that all of it formed before the development of the schistosity. A review of many schist specimens has led the author to conclude that andalusite is mostly later than the main schistosity on Creag Choinnich but that slight late deformation has been felt. Possible variation in the age of the dominant deformation prohibits definite conclusions from being drawn. Some of the andalusite may be as late as post-F_{A2} since in CB 69 the crystals overgrow new biotites lying parallel to the axial planes of regular folds with a Caledonoid trend, which can hardly be primary.

(c) North of the Dee

Typical hornfelses. In this part of the area the pelitic members of the Blair Atholl Schist and the Ben Eagach Schist are virtually identical. CB 150 is representative of both and can be seen in this section to be mesocratic and to consist essentially of "hornfelsic" biotite, quartz and cordierite. Andalusite in CB 150 is eclipsed by the cordierite which encloses it. There are also occasional porphyroblasts of plagioclase. Although it

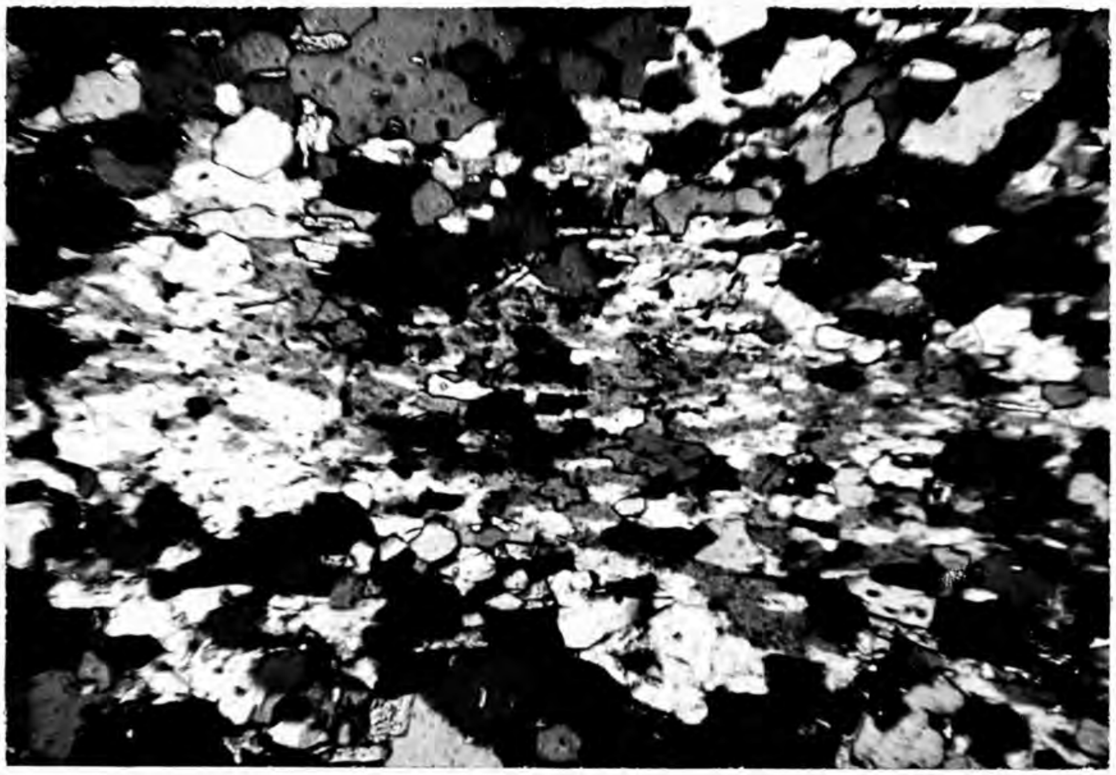


Plate 83. Strained cordierite exhibiting typical mottled effect.
CB 174; x-nicols; x 80.

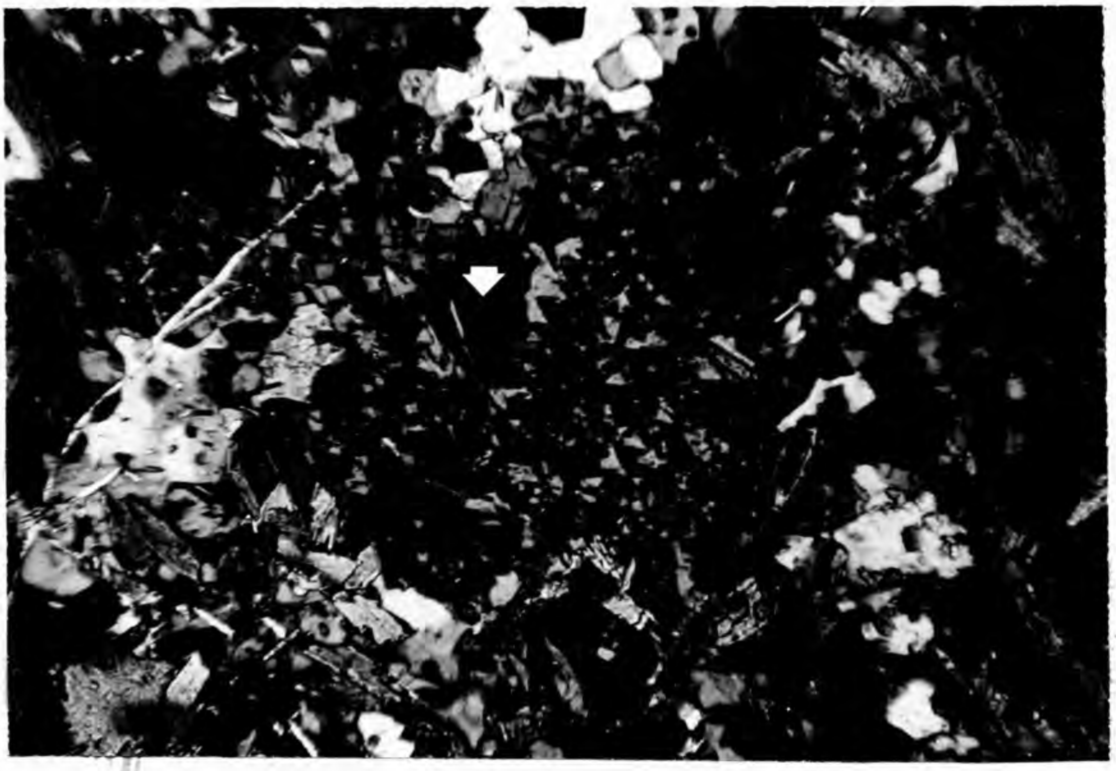


Plate 84. Smaller individuals of cordierite with good sector twinning (arrowed), associated with larger crystal(s) showing more complex twinning. CB 150; x-nicols; x 80.

has undoubtedly been hornfelsed, the rock is weakly foliated and has a moderate schistosity.

The foliation results from an imperfect segregation of biotite from the other two dominant minerals, and is parallel to the main direction of biotite alignment. To some extent quartz crystals are elongated with the schistosity but are not unduly strained. They are accompanied by small amounts of oligoclase, which may poikiloblastically enclose the smaller quartzes.

Locally up to 20% of the section examined is of cordierite. It is frequently elongated in strip fashion parallel to the planar elements, and is typically mottled or strained-looking (pl. 83). Sometimes crystals exhibit a well-defined but complicated twinning - a "multiple" form of sector-twinning (pl. 84) - which is suggestive of recrystallisation. Smaller, more circular individuals are well-defined in shape, with normal sector twinning, and hence resemble the cordierites from the hornfelses in the Buchan area proper. It has also been observed that these smaller crystals have a slightly lower birefringence than the strained ones, which again points to recrystallisation unless they are of an entirely new generation.

Biotite in CB 150 is of reddish brown aspect, grades in size down to minute rods and particles, and has various modes of occurrence. The small size of the biotites enclosed by cordierite points to partial digestion, but their good crystal form and variable alignment poses problems. For instance, a typical oval cordierite porphyroblast contains two zones of biotites. The marginal ones are rod-like and follow curved paths round the central area, as if rotated by the growing cordierite. In the centre, slightly larger and seemingly unorientated biotites are anhedral, while the tiny rods here lie parallel to the main schistosity or at 90° to it. In some porphyroblasts the biotites continue parallel and undeflected to the edges. The biotite outside the cordierite, forming the main schistosity,

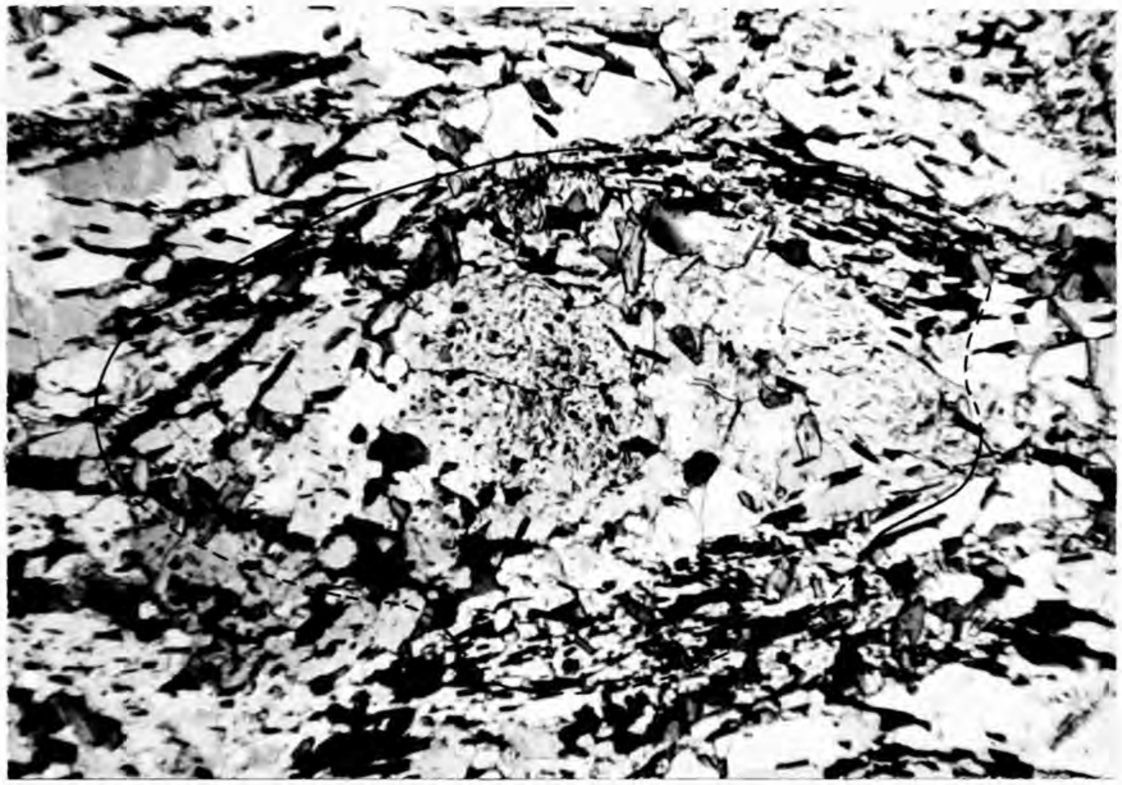
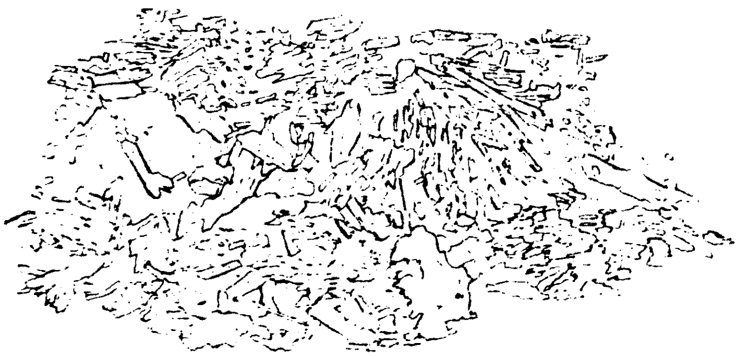


Plate 85. An ovoid porphyroblast of cordierite, lying roughly within the limits indicated, and containing zones of biotites of different shape, size and degree of orientation. CB 150; x 80.

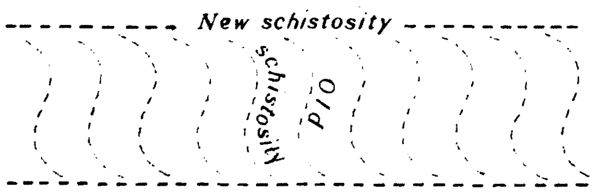
(x 25)

(i)



(x 25)

(ii)



(iii)

Fig. 48. Small biotites in cordierite (i) defining a structure similar to that elsewhere in the fabric (ii) where two schistosities are present (iii). CB 150; x 25.

has evolved by strain-slip of a previous schistosity at 90°. Thus the tiny biotites that cross the new direction must surely be relics of this previous alignment; but why do they form apparently euhedral rods?

The other problem is whether the central biotite flakes represent a pre-schistosity fabric, or whether they are the latest to form, under conditions of no-stress. Some of the larger tabular flakes have all the attributes of late thermal growth, but it is possible to discern in some cordierites flakes of biotite with a hint of a structure which equates them with the earlier schistosity (fig. 48).

Whatever conclusions are drawn the structural and metamorphic complexities of the rock are evident. Cordierite, for instance, not only appeared at a relatively early stage of hornfelsing, but continued its growth throughout at least one phase of deformation, perhaps suffering recrystallisation also. Initially cordierite must have developed partly at the expense of biotite, but regeneration of biotite may have played its part also.

Regarding age-relations it is interesting to note that the larger plagioclases preserve the same structures in trails of graphitic particles as have been found in the biotites. By analogy with the plagioclase-bearing schists to the southwest of Braemar, where folded trails of graphitic particles are common, the deformation responsible for the new schistosity may be F_{A2} or even later: i.e. the main schistosity in CB 150 is S_{A2} or later.

It is worth noting here the relationship of the microstructure of CB 150 to that visible in the rock in the field. The latter has already been referred to on page 29. The schistosity - more of a micro-foliation - has been described as having passed through the strain-slip stage still dominant in nearby exposures (see fig. 21). Its irregular nature, however, is not a legacy from earlier folds - these have been destroyed and have left only the limbs described in the thin section - but reflects a lenticular structure. This is seen in section to be caused by the (possible rotation and)

Plate 86. Two schistosities, each parallel to a micro-foliation. In this view one is apparently dominant. Cordierite occurs in the mica-rich foliae. CB 178; x 35.

Plate 87. Two schistosities in hand specimen of CB 178. Note suggestion of a tight recumbent fold in the centre.

flattening of cordierite crystals during growth and envelopment of inclusions, but without deflection of the surrounding micas.

Irregularities in the schistosity can thus arise by genuine microfolding (as in CB 120, p. 62) or by rotation of cordierite to deflect the schistosity or by the development of ovoid cordierites which assimilate the micas and thereby interrupt the schistosity.

Of less-typical hornfelses, some have a foliation and no schistosity or vice versa, while more rarely two schistosities may be present. It must be stressed that in most cases the rocks in question would be called schists but for the presence of "thermal" minerals. Structureless hornfelses in this area are virtually unknown.

Where two schistosities are present they tend to intersect at 20°; and when they cannot be genetically related even after microscopic study they are considered to have arisen contemporaneously. In CB 178 there is a microfoliation parallel to both schistosities. It is defined by very thin bands richer than normal in biotite. Here cordierite is controlled in its growth by both directions of alignment (pl. 86), thus strengthening the view that the two are contemporaneous. The hand specimen of CB 178 (see pl. 87) does, in fact, give the impression of containing very tightly compressed folds which have affected the schistosity and possibly produced parasitic accordion folds to the limbs of which the micas and cordierite are now parallel.

Even in the non-typical hornfelses there is evidence to show that cordierite is not a purely "thermal" mineral. In CB 174 some of the cordierite porphyroblasts consist of a central area of smaller unaligned biotites contrasting with those well-orientated elsewhere. This implies that biotite recrystallised during the growth of cordierite, which was partly under conditions of stress and partly not. A further development is suggested in the local presence of small rounded cordierite crystals, some in clusters,

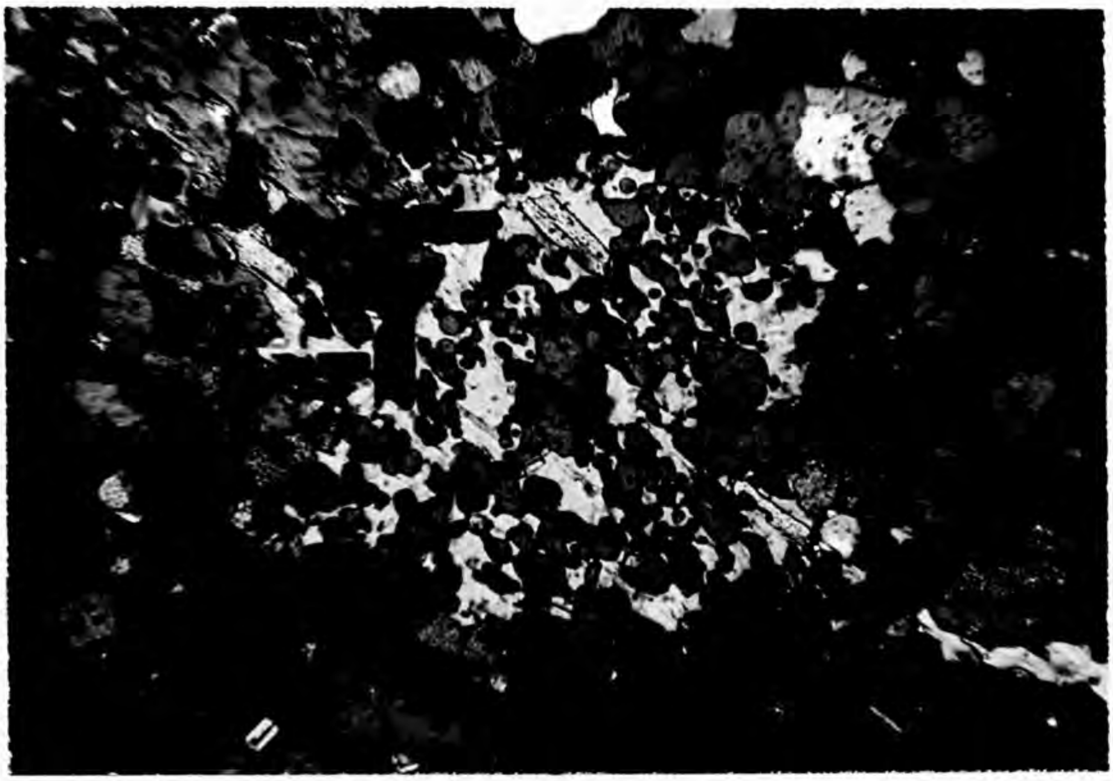


Plate 88. Clusters of small rounded cordierite crystals. Twin-planes when present are distinct. Also the birefringence is slightly less than that of the larger strained cordierites (see p.64). CB 174; x-nicols; x 80.

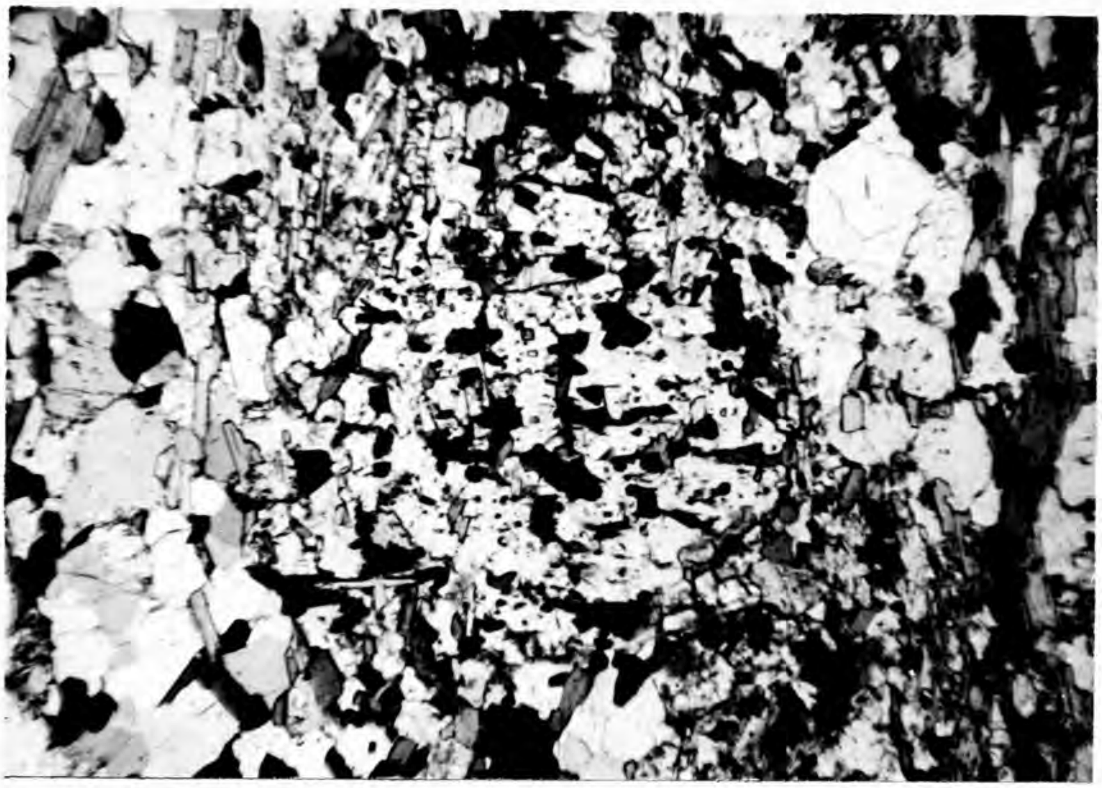


Plate 89. Porphyroblast of cordierite enclosing micas aligned at 90° to the external schistosity. CB 146; x 35.

Plate 90. Cordierite enclosing folded trails of magnetite particles etc., as indicated by the broken line. CB 168; x 80.

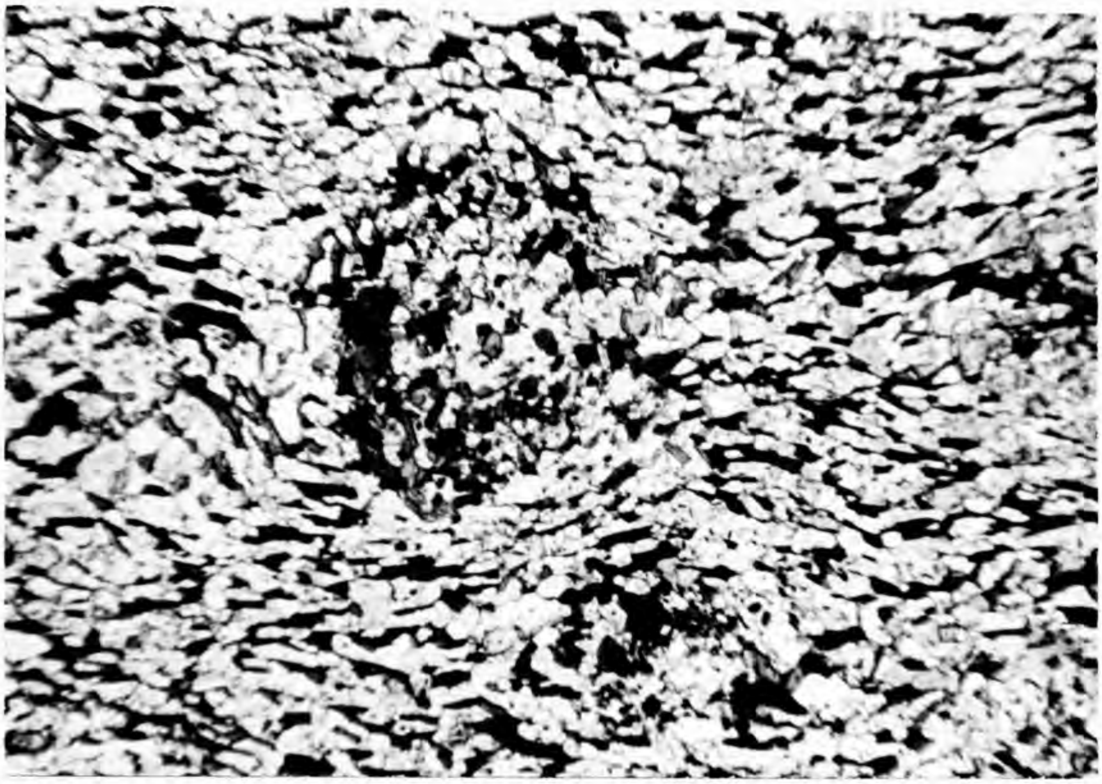


Plate 91. Porphyroblast of cordierite which has caused "bowing-out" of the schistosity, presumably due to rotation. CB 139; x 35.

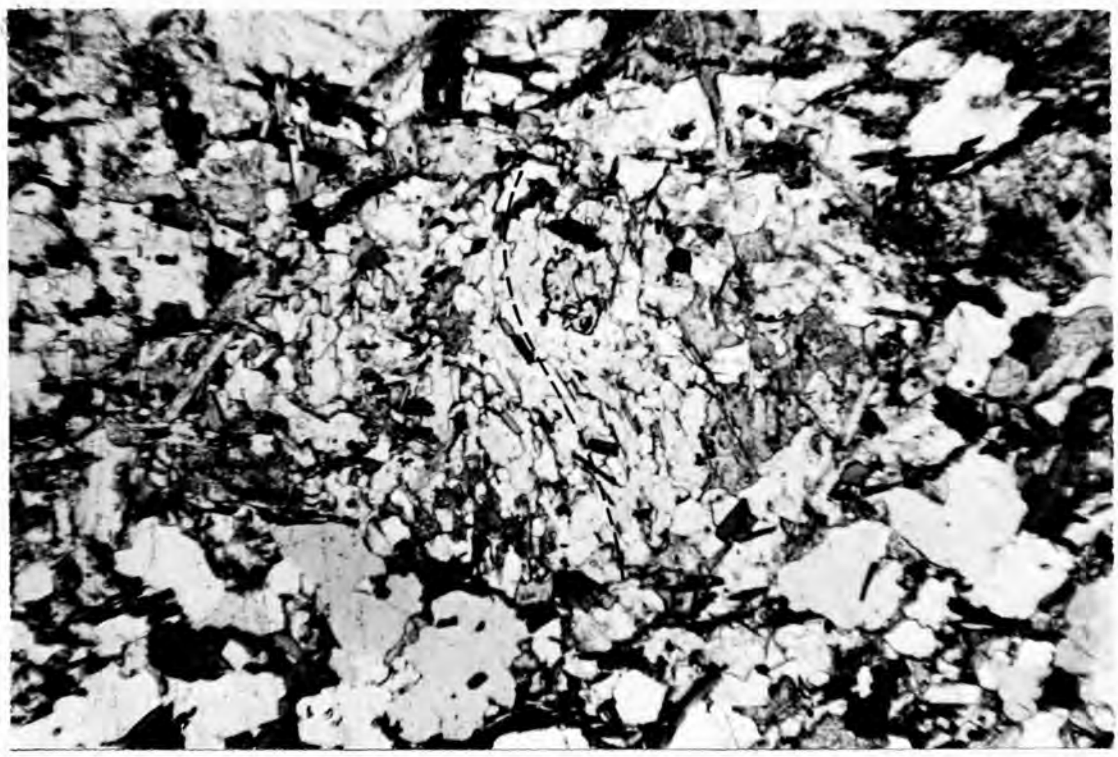


Plate 92. Porphyroblast of cordierite exhibiting a slightly S-shaped fabric (see broken line) suggestive of rotation during growth. CB 185; x 35.

with distinct twin-planes, as if recrystallisation had occurred(pl. 88).

In the more leucocratic, less micaceous hornfelses one can still perceive evidence of dynamic effects operative during rise in temperature. Sometimes there were two dynamic phases, as described before. For instance, in sections of CB 184 the micas are seen to be obviously influenced by two directions of alignment. This is true also of andalusite which is distributed in two sets of thin lines almost at right angles and which consists of highly sutured and embayed crystals enclosed by cordierite and quartz. The andalusites themselves enclose biotites which are similarly orientated. The cordierite, intergrown with orthoclase, extends mainly in one direction of schistosity and may also enclose folded trails of magnetite particles. This, and the presence of relict epidote grains, are features left over from the "black schist" stage. They contrast strongly with needles of new sillimanite which are growing out of the andalusite and which must represent a late high level of thermal metamorphism. This accords well with features of granitisation in the schist including enrichment in microcline and acid plagioclase.

In a neighbouring locality to that of CB 184 schist specimens are distinctly spotted, due to porphyroblasts of cordierite which have been rotated during growth and which contain remnant andalusite porphyroblasts, partially converted to sillimanite. Here, as in the rest of this zone, sillimanite is one of the latest minerals to form.

Nature of Changes in Metamorphic Effects

In the schists just described the weakest metamorphic effects are present in the extreme southwest of the area, the strongest in the northeast. The former are also associated with the earliest structural changes that are preserved, the latter with the latest.

Neglecting local variations, the schists to the southwest of Braemar clearly represent the end-products of the widespread regional metamorphism in the Scottish Highlands. Very rough limits to metamorphic isograds may be recognised, virtually coinciding with those defined by Barrow himself in the lower zones. A separate staurolite zone has not been established, but the garnet/kyanite isograd, occurring near the southwest border of the mineralogical map (fig. 47), runs eastwards towards the southern margin of the Lochnagar Granite.

The first appearance of sillimanite, east of the Clunie, although in the position one might expect, is not related to the sillimanite isograd around Braemar. Instead it has been found to be connected primarily with the presence of the Lochnagar Granite, but also with other granite masses both exposed and unexposed. The area mapped thus must represent part of a zone of cordierite and andalusite centred round the Lochnagar Granite, with the development of sillimanite at local contacts. This concept is at variance with that envisaged by Barrow and others.

The nature of the changes in the various zones will now be discussed in more detail and integrated with the structural history of the area.

(a) Càrn Damhaireach to Braemar

In this part of the area the earliest structures have been studied in exposures at Clunie Lodge (eg. KB 324). Small compressed micro-folds affecting the micas and dark inclusions can be correlated with a strain-slip schistosity found in the same schist-group at the Spittal of Glenshee. This and its associated larger-scale folding are secondary and pre-date the main phase of regional metamorphism.

Furthermore the folds in KB 324 are themselves deformed by larger, more open micro-folds which are typical of many "black schists" from around here. From considerations of style, axial direction and relation to the compressed folds, it would appear that the later are Later Caledonoid in age (F_{A2}). This is in keeping with the accepted contention that garnet and kyanite are essentially "post-F_2" (F_{B1}) in the Central Highlands.

Plate 93. Sheared out fragments of garnet gently flexured by folds believed to be of the F_{A2} period. KB 75; x 80.

Plate 94. Idioblastic staurolite containing the remains of a garnet crystal (arrowed). KB 151; x 80.

(i) Garnet zone. Garnet has not been observed in its development stages. All the evidence, including the small size of individual crystals, points to this mineral being unstable. The fabric of the quartz inclusions proves that the garnet grew after one phase of deformation had occurred, yet rotational fabrics indicate the garnet to be syn-tectonic. It can best be fitted into the tectonic history if considered to have grown during and after F_{B1} deformation, which in the main pre-dates that of F_{A2}.

Most of the quartz present has undergone recrystallisation during the same period while biotite and muscovite have been folded and regrown to form a new schistosity S_{B1}. Thereafter these minerals were rotated by the garnet porphyroblasts or enclosed by them during the subsequent F_{A2} deformation.

(ii) Kyanite zone. This lies nearer Braemar and represents a higher regional metamorphic grade in which garnet becomes increasingly unstable. Although idioblastic staurolite has been found it is mostly ragged in outline and looks to have been rotated into line with the evolving schistosity. The folding responsible is thus post- main metamorphism. Most of the kyanite, too, is seen after this folding, has had its effects, and it is thus difficult to find evidence of the rise of kyanite. Five specimens do, however, reveal small kyanites so intimately "bound up" with garnet as to appear to be replacing it. Some of the kyanite may develop from the opaque inclusions and small micas within the garnet.

The mineralogical and structural changes involved in this zone have not all been continuous, for while kyanite continued to form after garnet growth ceased, there must have been a static phase of growth between the two phases of deformation. This can be proved in KB 239 where neighbouring porphyroblasts together preserve the original folded fabric. The open style and axial direction of the later folds (pl. 95) support an F_{A2} interpretation, and can be correlated with F_{A2} in the field. The preserved fabric

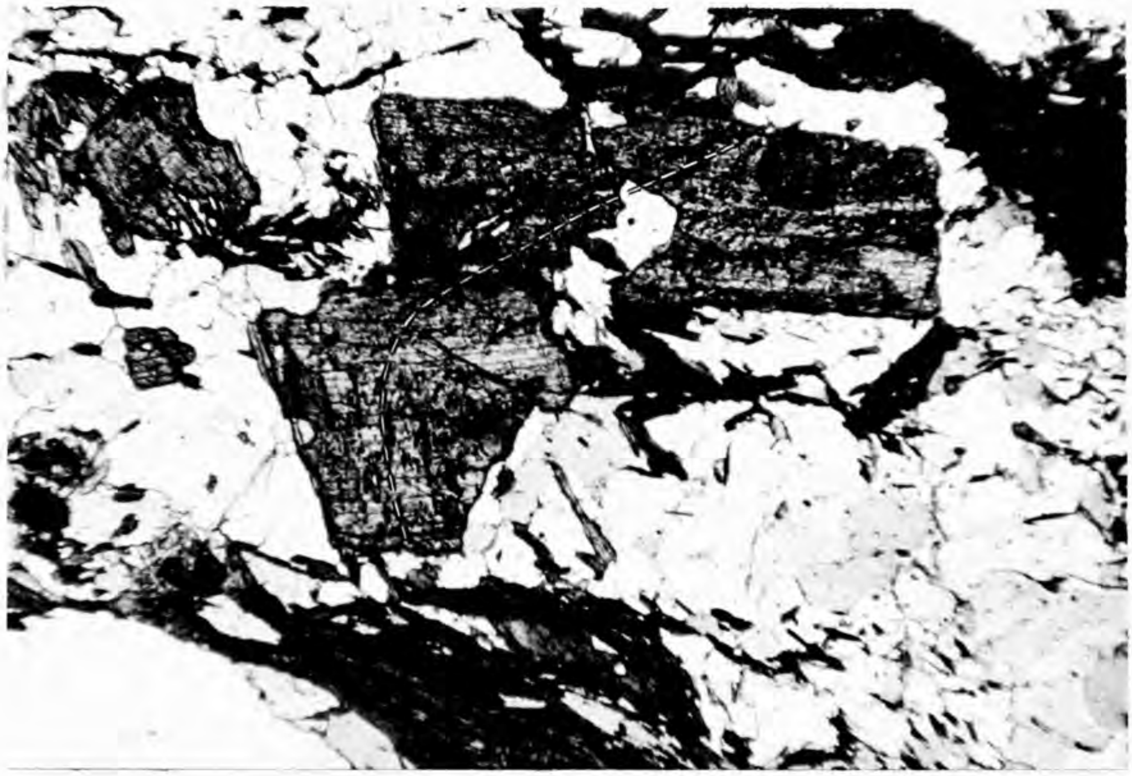


Plate 95. Two adjacent porphyroblasts of kyanite together preserving a folded fabric (broken line). KB 239; x 35.

Plate 96. Deformation of a large porphyroblast of kyanite, shown by zones with different extinction position. KB 482; x-nicols; x 35.

can thus be regarded as F_{B1}. By and large kyanite and garnet must have arisen between F_{B1} and F_{A2} and must have cleared themselves of opaque inclusions before the close of the F_{A2} period of movement.

During this period it is possible that there was further growth of kyanite. This consideration arises from the recognition in some schists of elongated porphyroblasts, usually small and lying in the schistosity. They appear unaffected by deformation, yet must have developed during the operation of a stress-field.

Some of the changes in the kyanite zone can be attributed to the F_{A2} phase of folding that occurred after the main growth of kyanite. As well as the recognisable F_{A2} folds affecting kyanite one can find in some schist-sections already granulated fragments of that mineral rotated round pseudomorphs after garnet. Quite apart from considerations of crystal form etc. it is unlikely that this is new kyanite growth since the garnet has already completely gone. The fragmentation of kyanite clearly commenced with its microfolding (pl. 96) and splitting along cracks and cleavages.

The partly deformed plagioclase porphyroblasts, which may engulf kyanite, are also witness to post-kyanite deformation. This may be late F_{A2} or the beginnings of a completely later phase, though correlation with hand specimen and field evidence indicates late F_{A2} deformation (cf. late Caledonoid folds and lineations, S.W. Càrn nan Sgliat, p. 20). See plate 97.

The general examination of schists in the kyanite zone indicates that first garnet and then kyanite gradually diminish in size towards Braemar, while plagioclase tends to increase. The growth of felspar, however, is local and probably related to proximity of granite intrusions, particularly around Corriemulzie. Towards Braemar itself the ultimate decline of plagioclase corresponds with the general onset of thermal metamorphism which is responsible for the replacement of felspar by cordierite <u>before</u> the appearance of sillimanite.

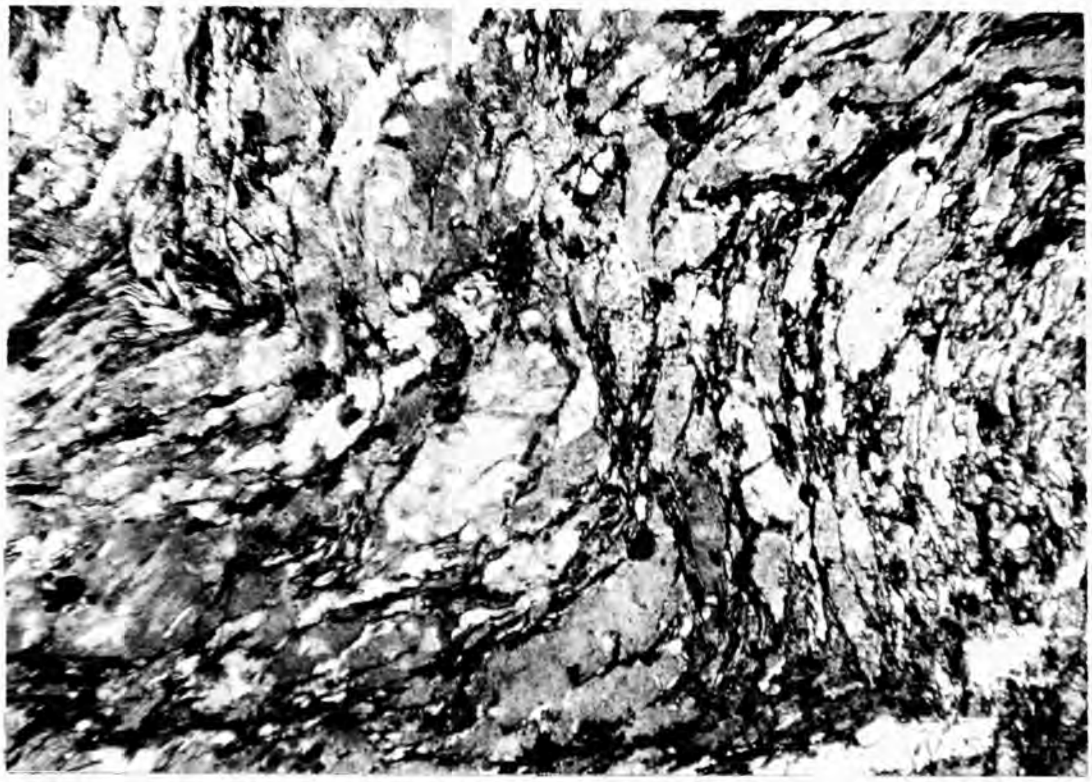


Plate 97. Illustrating the deformation of felspar (cloudy) by folds believed to be of the F_{A2} period. KB 71B; x 35.

Plate 98. Porphyroblast of cordierite (white) with tiny inclusions and a remnant of garnet (dark patch, centre). The "bowing-out" of the schistosity round the cordierite was accomplished prior to the growth of cordierite. KB 127; x 80.

(iii) The development of cordierite and andalusite. After the F_{A2} deformation the area from Càrn Damhaireach must have been subjected to local increase in temperature. This alone can explain the development of spots of cordierite often within pseudomorphs after garnet. This event may have been rapid, for even in its apparently initial stages the cordierite may be well formed though seldom abundant, and it appears to have quickly begun to replace the earlier plagioclase. It is at this stage taking the structural place of the felspar, enclosing the many trails of opaque inclusions originally incorporated in the felspar: cf. pl. 90.

Andalusite seems to be even more abrupt in its appearance, having rarely been observed as incipient crystals (p. 57). The evidence of different generations of andalusites is taken to indicate that local intrusion of granite occurred over a period of time. Early deformed, or commonly aligned, crystals prove that the tectonic episode continued after the emplacement of granite (as has already been demonstrated from petrofabric considerations).

(b) Creag Choinnich, Càrn nan Sgliat, and North of the Dee

(i) Cordierite/Andalusite Zone. This zone represents the stage at which increased thermal metamorphic effects give rise to the widespread development of new minerals, and to the appearance of a metamorphic segregation foliation in the schists.

Cordierite, in its initial stages of development, can be observed in such rocks even where the pseudomorphs after garnet have all but disappeared. The evidence also shows that garnet had suffered a prior rotation by folds almost certainly of F_{A2} age before the appearance of cordierite (pl. 98).

This part of the area also records the transformation of biotite from that typical of schists to the foxy red biotite of hornfels conditions, confirming that the rise in temperature responsible was not just a local one. A study of mica development has shown it to be a complex one and to have been influenced by tectonic forces during or after hornfelsing. It is evident that much of the new mica was formed within areas or potential areas of cordierite growth before or during the deformation of these porphyroblasts.

The corresponding transformations in the quartzo-felspathic domains of the hornfelses do not appear to have been significant at this stage, but during the subsequent deformation(s) potash felspar in particular influenced the structure of the recrystallising cordierite.

On Càrn nan Sgliat andalusite has been found in close association with cordierite, though its initial growth stages have not been preserved at all. Again, however, the conflicting evidence of the age of andalusite relative to local schistosity (which itself may be variable) and the evidence of its sutured margins against cordierite point to its development prior to cordierite, but with later or continuing growth of crystals broadly simultaneous with the evolution of a new schistosity, believed to be S_{A2}.

That the schists on Creag Choinnich are metamorphically behind their counterparts on Càrn nan Sgliat can be equated with the increased distance of this hill from the Lochnagar Granite.

Those parts of Càrn nan Sgliat lying nearest the granite margins, and the area north of the Dee, represent by contrast the culmination in the growth of cordierite if not of andalusite, during the evolution of a widespread foliation. The appearance of the foliation has locally been intensified by the introduction of granitic material, which further strengthens the view that the mineralogical changes are on the whole contact thermal ones.

This part of the cordierite/andalusite zone displays slightly later changes than that around Braemar. Some of these have involved further deformation, shown by the highly strained nature of the cordierite and by evidence of recrystallisation. The strip-like appearance of much of it, and its response to two schistosities in some localities combine to prove that cordierite was active during these metamorphic changes.

Actual micro-folding of cordierite and andalusite (CB 120, p. 62) speaks for itself as regards post-thermal dynamic effects.

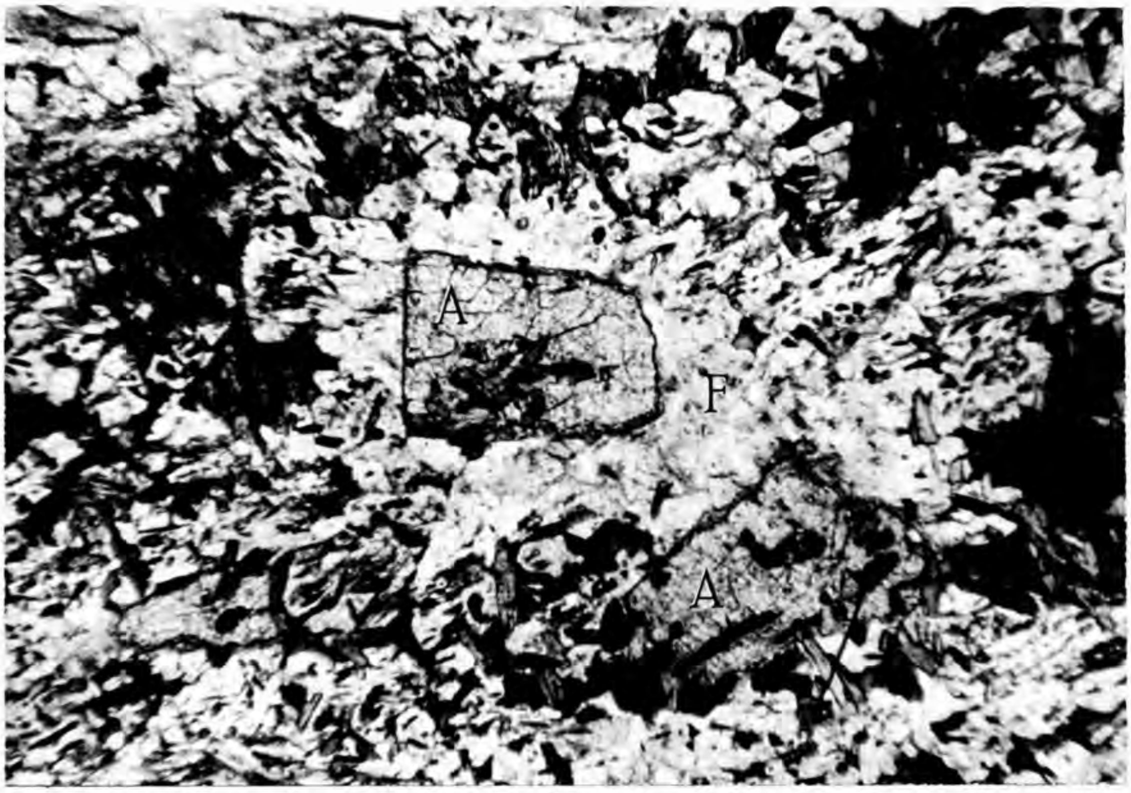


Plate 99. Andalusite (A) in a zone of felspar (F), around which are concentrations of mica + chlorite. Biotite and muscovite are also found within the andalusite, which has very thin rims of sericitic mica indicating that some at least of the mica is post-andalusite. KB 276; x 35.

(ii) Sillimanite-bearing rocks. Apart from one known locality sillimanite reached only the initial phase of growth in even the most advanced hornfelses. Its overall distribution correlates well with areas in which granite intrusions are present. It is believed to be a post-tectonic result of high-grade contact thermal metamorphism, and not simply a recrystallisation of sillimanite formed under regional metamorphic conditions. As the evidence presented implies, there is both a lateral and a temporal gap between the kyanite zone and the general appearance of sillimanite, which is later than that of cordierite and andalusite.

The Problem of Chlorite

The occurrence of chlorite has not yet been dealt with. It is usually a relatively late, replacive mineral, as if a result of retrogressive metamorphism. In spite of a fairly widespread incidence, however, it is restricted in importance to definite areas. It is not found in areas of high-temperature hornfelses and in fact gradually dies out towards such zones. It thus would seem likely that chlorite developed regionally before the main thermal phase of metamorphism which ultimately destroyed the chlorite of the higher zones. Some direct evidence supports this contention: in one or two schist specimens andalusite cross-cuts micas, including chlorite.

In the Barrovian zones chlorite is almost entirely younger than biotite, muscovite and garnet, and can be seen to post-date the F_{A2} folding in some "black schists"; but relative to andalusite its age is uncertain (pl.99).

Exceptionally chlorite is not the latest of mineral developments, for in KB 181 hornblende crystals transect those of chlorite. This chlorite is optically distinct from that normally encountered here, and is, in the writer's opinion, "left over" from the chlorite grade.

Infrequently finer chlorite of a different sort again has replaced cordierite in "retro-schists". This is a completely later effect which is present in certain types of cordierite hornfels only. It would, in consequence, seem safe to assume that most chlorite is post-regional metamorphism, but pre-contact thermal metamorphism.

These assumptions, however, are not quite compatible with the evidence from the igneous rocks believed to be responsible for the thermal metamorphism. In granites and diorites chlorite commonly replaces biotite and hornblende. Consequently one is forced to consider the formation of this chlorite as indicative of late retrogressive metamorphism possibly dependent upon the late dynamic metamorphism, which affected the igneous rocks themselves (some chlorite is itself deformed). There are thus possibly two periods of chlorite-formation.

Metamorphic Synthesis and Correlation with Banffshire

It is generally accepted that in the Central Highlands of Scotland the Dalradian rocks have suffered three main stages of metamorphism, the middle one being the greatest. The last-mentioned is considered to be largely "post-F_2", and can be recognised also in Banffshire where it appears to be later than the so-called Buchan type, though possibly broadly contemporaneous as a static phase.

According to Johnson (1962), andalusite grew in a static phase of thermal activity after the F_2 but before the F_3 period, though it probably suffered late recrystallisation. Cordierite, formed at the same time, was rotated by F_3 folds. Kyanite, rare, is considered to have formed after andalusite, sometimes by replacing it. Sillimanite is of uncertain age though post-F_2.

In Braemar it would appear that andalusite and cordierite are one complete phase of folding later in their growth than in Banffshire. Even if this must remain an hypothesis it is indisputable that these minerals around Braemar are later than the garnet-staurolite-kyanite assemblage. It is further indisputable that sillimanite here appeared one fold phase later

TABLE SHOWING THE HISTORY OF METAMORPHIC EVENTS FROM THE SOUTHWEST OF THE AREA STUDIED TO THE NORTHEAST OF BRAEMAR

(not to scale)

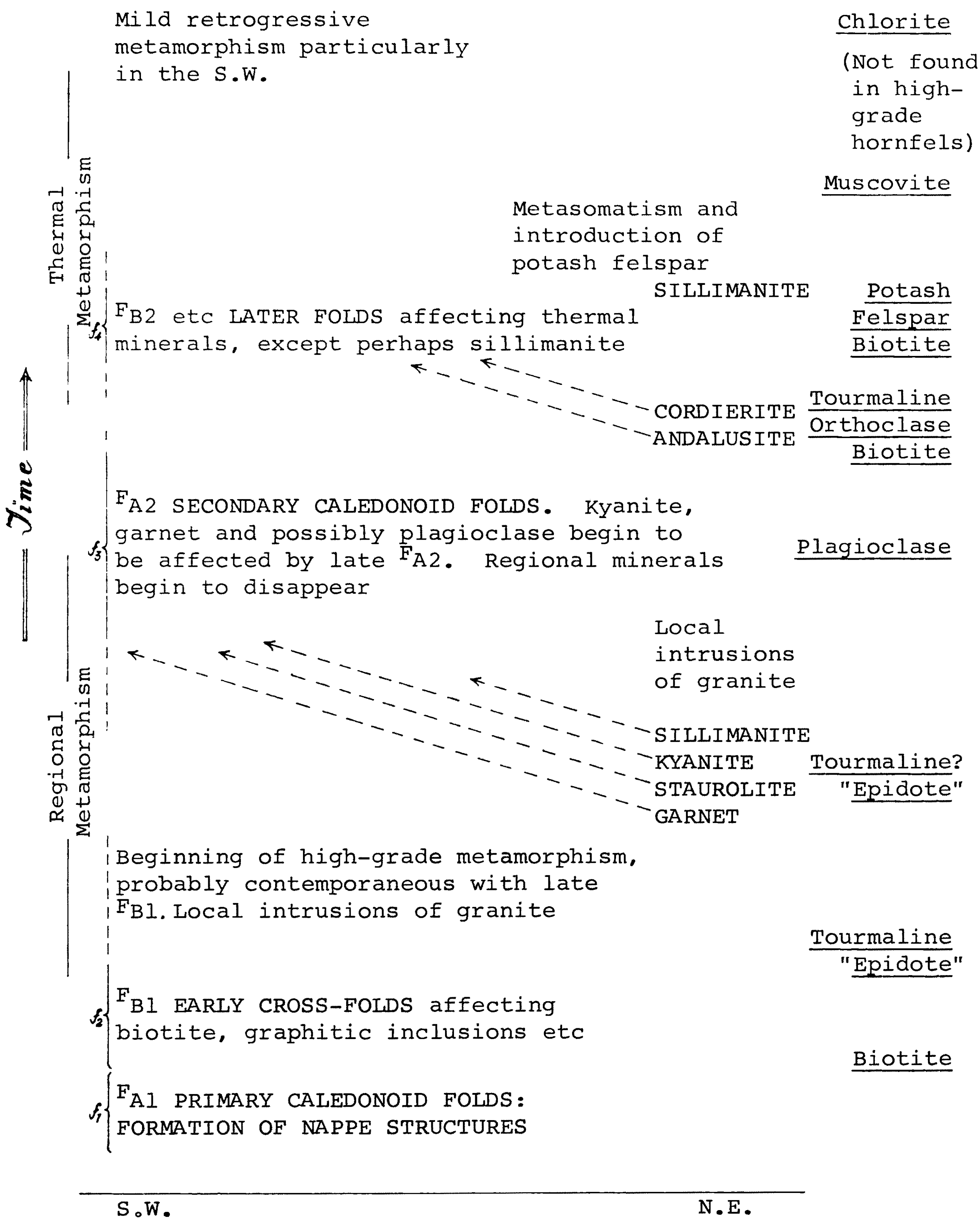


Fig. 49. Table of Metamorphic Events.

than some of the andalusite and cordierite (eg. CB 120), nor has it arisen by chemical breakdown of the micas as found by Chinner (1960) in Glen Clova.

According to Johnson the Buchan-type of mineral assemblage formed at a higher structural level in Banffshire than the Barrovian-type. In the Braemar area this cannot possibly be so as the two assemblages occur within a mile of one another along the Caledonoid trend. Consequently the Braemar area must be considered to be metamorphically distinct from Banffshire, and to have undergone a phase of Barrovian-type regional metamorphism upon which was superimposed a later phase of contact thermal metamorphism before deformation finally ceased in the area.

A table, summarising the main metamorphic events, is presented in figure 49.

PART V

IGNEOUS ROCKS

Introduction

The igneous rocks of the area range in decreasing order of areas occupied from granites to diorites to epidiorites. The area borders on the Lochnagar Granite Complex, and it is pertinent briefly to describe that mass since the igneous rocks of the area are related to it. These will then be dealt with in order of abundance.

History of Previous Research

The most important recent work is that of Oldershaw (1958, Ph.D. thesis) on the "Lochnagar Granitic Ring Complex", in which he recognises a concentric series of diorites (3 phases) and later granites (3 phases) which pushed aside the country rocks.

The diorites form the outer set, occurring now as six or so isolated masses of different grain size and composition, with a trend from basic to acid. The rarely-zoned oligoclase varies from An_{20} to An_{30}. Quartz constitutes 13% of the rock, mafics 20%. A faint foliation has been identified as a flow structure since the minerals have not suffered crushing.

The granites become finer in sequence of intrusion, and none have any microcline present. The coarse granite is outermost, being undeformed, homogeneous and porphyritic. The phenocrysts (two types of orthoclase perthite up to $\frac{1}{4}$" long) are set in a matrix of clear quartz, orthoclase, plagioclase and mafics. Only this granite contains xenoliths, and these, of schist and diorite, die out towards the central portion.

The medium granite is also homogeneous but non-porphyritic in the main, and is made up of quartz, orthoclase perthite, oligoclase (An_{26}) and biotite.

This granite has sharp contacts against the coarse, but is rarely chilled. It is believed to have come in by magmatic stoping.

The finer granite is equigranular and has itself two phases, the inner chilled against the outer. The perthitic structure is well developed in the orthoclase, and again oligoclase has the composition An_{25}. Shatter belts are noted in the fine granite.

A. ACID ROCKS

Field Relations

(a) Granites

Granite is by far the most abundant of the igneous rocks, occurring throughout most of the area. The three field maps adequately portray granite distribution, but even they do not indicate all the outcrops, many of which are too small to be mapped.

The map of Creag Choinnich is particularly illuminating in its evidence of granite occurrence which may be visualised as both large and small, regular and irregular masses often directly linked up. Occasional variations in the outward appearance of granite bodies (see p. 87) suggest phases of intrusion, but other field evidence and petrological considerations suggest that all granite masses can be related to one parent magma.

A relict stratigraphy, created by the presence of granite rather in the manner of the Donegal Granodiorite, is well seen on Creag Choinnich. The isolated fold closures in limestone at the summit, and the remnant patches of schist etc., clearly indicate that some at least of the granite has been emplaced at the expense of the country rock without disturbance of the remainder.

The same map reveals a distinct tendency for the granite to form concordant sheets etc., especially in areas of quartzite. Here there is evidence of forceful intrusion, especially where the host rock has a platy habit; but it is evident that the Blair Atholl Series offered less resistance to granite emplacement than the Perthshire Quartzite.

On Càrn nan Sgliat (upper nappe) the margins of the Lochnagar granite often "feather out" into a replacive network, being ultimately reduced to seemingly isolated masses. Although lack of exposure prevents direct observation of it the author is in no doubt that the main Lochnagar Granite (relatively undeformed) links with that of Creag Choinnich (deformed).

The boundary of the granite at the southern end of Carn nan Sgliat follows the original Perthshire Quartzite/Ben Eagach Schist boundary, the granite lying on the schist side, showing that this schist, too, has offered less resistance than the quartzite to the incoming of the granite.

North of the Dee, granite forms the same sort of irregular masses and fine networks. It is presumably the same granite as before. Some bodies, however, pass into dyke- and sill-like mass (see p.84) but as before the granite may preserve a relict stratigraphy or may be charged with small inclusions.

Inspection of the field maps alone does not entirely solve the problem of the age-relations of granite to other igneous rocks. Even individual contacts in the field may be unhelpful, and small-scale and laboratory evidence must be considered.

Like those of certain other Caledonian granite masses, the small-scale granite relations to its host provide evidence both for forceful intrusion and for emplacement by replacement of the country rocks in situ. This evidence will now be presented.

The southeast side of Creag Choinnich provides good exposures of granite in contact with quartzite, pelitic and semi-pelitic rocks. It can never be conclusively proved that any of the granite sheets are completely dilationary, but while much evidence to the contrary exists, some localities furnish direct evidence of disturbance of the country rock. Many occurrences

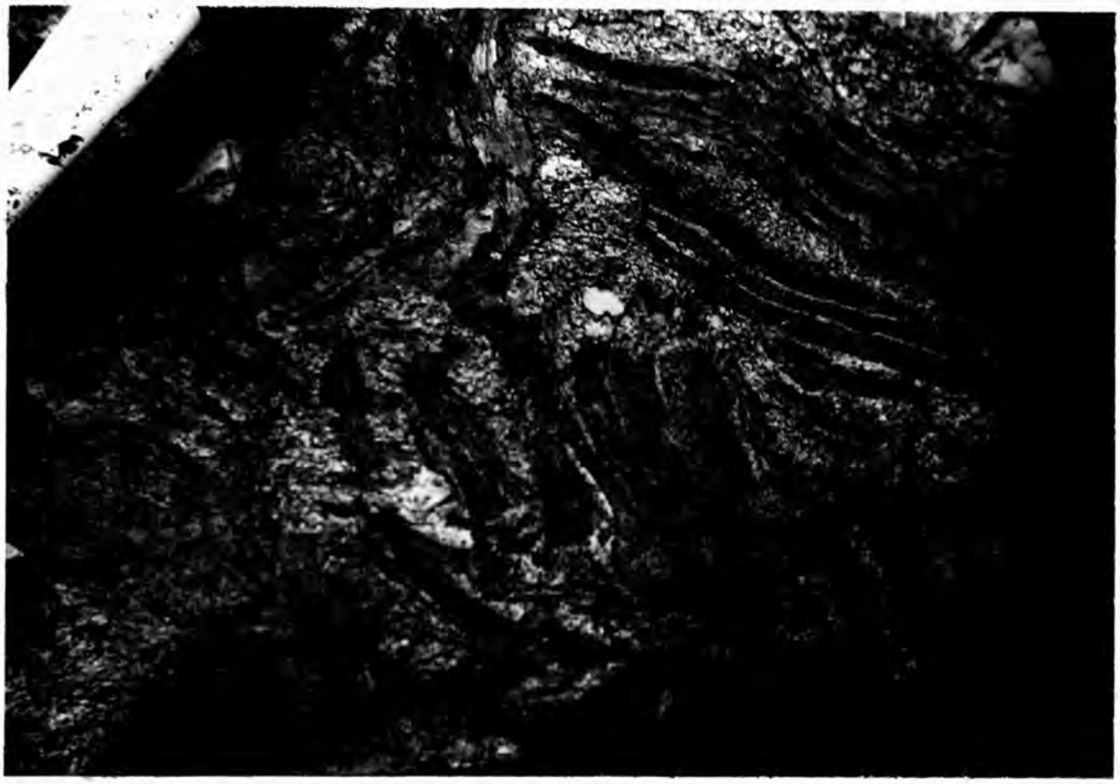


Plate 100. Disturbed inclusions of quartzose schist in the granite.

Plate 101. Contaminated granite where it has invaded and replaced pelitic country rock (see also plate 113).

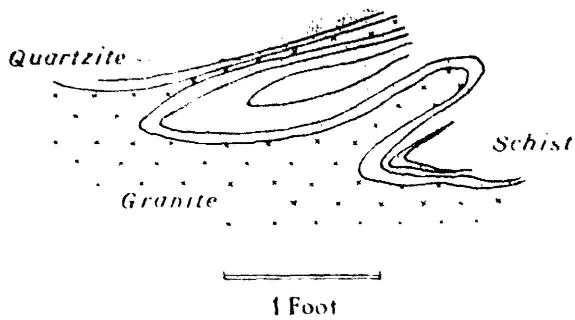
of quartzite inclusions have been noted. Sometimes, as at the point (030,087) on Creag Choinnich, such inclusions can be seen from their present orientation to have been rotated in the granite (pl. 100). At another locality (028,088) 50% of the quartzite inclusions have been little moved, while the remainder are randomly orientated after forcible detachment from a regularly-dipping quartzite mass. The orientation of elongated rafts of quartzose schist leaves no doubt as to their disturbed nature viz:- strike of elongation parallel to banding, at 80°, 310°, 285°, 170°, 165°, 340°, 325°, 325°.

There is also such evidence of the magmatic nature of the granite as drusy cavities, though these are extremely rare.

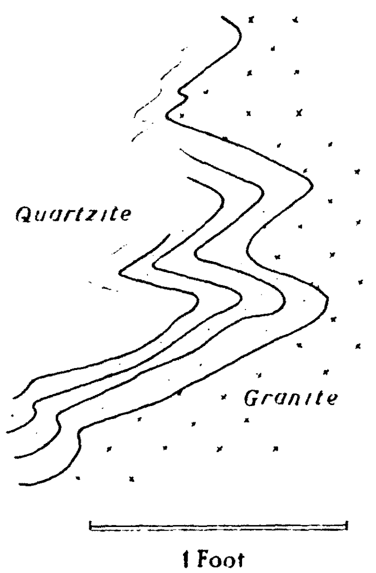
On the other hand some junctions between quartzite and granite appear initially to be gradational, in consequence of the granite becoming finer and more evenly pink in colour, to match that of the elongated felspars in the quartzite. This is a result of chilling of the granite on the one hand, and (sometimes) by felspathisation of the quartzite on the other. Shearing can produce an apparent gradation between the two.

As a general rule chilling is seldom to be recognised, though the very existence of microgranite in the same environment as granite must be admitted as a possible instance of chilling. Notwithstanding, the granite must have entered a pre-heated environment in most parts of Creag Choinnich, as elsewhere in the area.

The felspathisation mentioned above is best seen in specimen CB 5 (quartzite). Within six inches of the contact zone the two rocks are distinct, as granite and quartzite, but towards the contact from here the size and frequency of areas of quartz in the granite increase. At the same time both potash felspar and plagioclase crystals decrease in number but not in size. Ultimately the quartz areas take on the true identity of quartzite. Within this part of the quartzite the large felspars appear as euhedral crystals, but under the microscope they are seen to be more lenticular,



(i)



(ii)

Fig. 50. Diagram of granite contacts following fold profiles in pelitic rocks (i) and quartzite (ii).

Plate 102. Excellent evidence of the disturbance of schist bands caused by the incoming of granite.

having suffered shearing, and are thus of porphyroblastic origin.

Around the summit of Creag Choinnich, too, some exposures of quartzite are visibly enriched in felspar, while the adjacent granite is enriched in quartz. The selectively replacive nature of the granite can be neatly demonstrated at locality (014,082) where the granite contact follows precisely the profile of minor folds in the metasediments (fig. 50 i), though this situation can also be found in quartzite (fig. 50ii). At other points in this area of the lower nappe the granite has replaced the country rocks on a grander scale: one whole cliff-section is of a relict stratigraphy in miniature, yet there is no evidence as to the precise mechanism of replacement. Here one entire end of the feature is occupied by country rocks, but towards the other end they give way to an increasing proportion of granite sheets, sills and lenses, until the whole cliff is of homogeneous granite.

In larger areas of true schist, such as Càrn nan Sgliat, the granite shows similar variations in its mode of occurrence, though sill-like masses are less common, but vein networks more abundant. In the hornfelses thin granite veins or sheets may be difficult to distinguish: indeed, in foliated hornfelses it may be difficult to separate them at all, or to decide whether the granite-like bands are of igneous or metamorphic origin. At many localities discrete off-shoots of granite cross-cut the structures in the host only to be "lost" where finer branches grade out into the country rock.

Contamination of granite in proximity to such pelitic rocks bears further witness to its replacive nature (pl.101) and must be considered as an effective influence on the petrology of the granite (see "Petrology", section (a)).

Evidence of a more contradictory nature can be found at locality (098,122) and locality (122,158), north of the Dee. Granite sheets cross-cut folds of all ages in the schist and disturb certain bands (pl.102), yet in other bands, gently folded, the granite appears to be replacive (pl. 103),

Plate 103. Evidence of the replacive nature of the granite.

Plate 104. Discordant sheet of granite preserving the structure of the schist.

Plate 105. Apparent displacement of schist-bands across a thin sheet of granite.

and where discordant the original structure in the schist is perfectly preserved (pl. 104) in the granite.

Even at the only locality where bands of schist are displaced across thin intrusive sheets of granite the displacements are greater than the width of the sheets would allow (pl. 105). The most likely explanation is that the granite occupies a minor fault-zone.

Consequently the field relations of the granite to the quartzo-felspathic and pelitic rocks indicate that the granite has come into place by both forceful intrusion and by replacement. It is envisaged, from the evidence presented, that the forceful nature of the granite alternated with that of replacement due to the build-up of pressure from below. During periods of reduced pressure the presence of the granite allowed the country rocks to heat up, and replacement commenced. Then when the pressure reached a high level the outer portions were forcefully injected into the country rocks, in some cases reaching colder levels.

A special study of granite/limestone contacts has not been made, but the evidence found is much as presented above. The granite shows no special affinity for invading or replacing calcareous rocks. Net-veining is unknown. However, at the summit of Creag Choinnich the granite exhibits a mineralogical departure from normal in consequence of its proximity to the limestone there. Very close to the contact the granite is converted to an apparently monomineralic white rock ("anorthosite") further described under "Petrology". Intrusive sheets of this rock have also been found (pl.106).

The field relations of the granite to the diorite leave no doubt that the diorite was the earlier phase, as found by Oldershaw. Diorite may be veined by the acid rock or sharply transected by it. Some localities reveal inclusions of diorite in granite: in others the diorite is unmistakeably being granitised at the junction, providing an apparent transition between the two. This may arise by the development in the diorite of patches enriched

Plate 106. Intrusive sheet of "anorthosite" (A) with two thinner ones to the right.

Plate 107. An epidiorite sheet (E) transected by granite (G).

in quartz and potash felspar, or by the formation of a zone of "mixed" rock.

At locality (087,138) the "mixed" rock can be studied. The two rock-types were emplaced as separate sheets but towards their mutual boundary they become atypical and variable. Some exposures resemble contaminated granite, others granitised diorite. Detailed investigation reveals "contamination" on all scales, portions of diorite having been found in all stages of digestion (see "Petrology").

Porphyritic microdiorites and "dark porphyries" vary considerably in age from pre- to post-granite.

By their very nature epidiorites must be considered to be pre-granite, and their field relations leave no doubt that this is so(pl. 107).

(b)"Felsites"etc.

Although distinguished for convenience on the maps, "felsites" and porphyritic microgranites are not entirely distinct from the granite(s), but they occur only as relatively thin sheets, except perhaps on Càrn nan Sgliat. They are mostly intruded along or near the planes of nappes and slides, and sometimes may be the main indicators of such structures.

The map covering Càrn nan Sgliat (map 2) shows that swarms of concordant "felsites", as lenses and sheets, sometimes multiple or branching, coincide with a level of probable folding or thrusting within the upper nappe quartzite. One large sheet transects the granite, it will be noticed, according to the map, but field evidence from the contacts is not so positive (see p. 84).

In the area north of the Dee few minor acid sheets have been encountered. Those mapped proceed undeflected in the granite and are believed to be later than it. Their scarcity confirms the observation that the "felsites" diminish in size and frequency towards the northeast in the Braemar area.

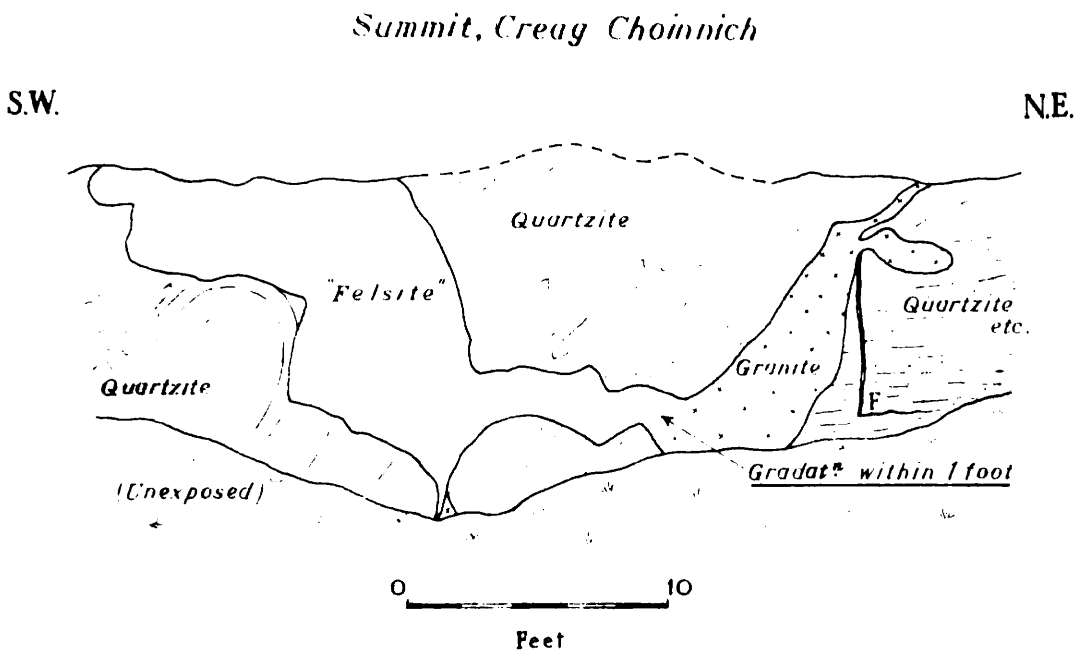


Fig. 51. Sheet of porphyritic "felsite" grading quickly to granite.

An important distinction between them and true granite is that they have seldom been found in the hornfelsed schist of the area, and never in the Ben Eagach Schist.

(c) Granites to "Felsites" etc.

Field evidence proves that some of the hypabyssal acid rocks are pre-granite and some post-granite. There are those, too, which must be broadly contemporaneous, some of which may well be but locally chilled phases of the granite itself. The majority cannot be relatively dated with confidence: of those that can most are pre-granite.

On the southeastern slopes of Creag Choinnich several "felsites" are terminated by granite and one of these is invaded by tongues of the granite. Near the summit of the same hill some "felsites" appear to cut across granite, but are ultimately cut out by the same. This proves that, as with the metasedimentary rocks, the granite may preserve parts of earlier acid intrusions almost intact.

This part of the area is not without contradictory evidence. For instance near the summit of Creag Choinnich one porphyritic "felsite" which can be traced as such for many yards is cut out by granite at one end but grades into microgranite and then to granite within two feet at the other end (fig. 51). Also, the junctions of "felsites" preserved within granite may not be sharp, with joints crossing from one rock into the other (pl. 108).

Sometimes indirect evidence has to be used where contact relations are unhelpful. On the north-west side of Creag Choinnich the main "felsite" in the granite is proved to be the earlier by the presence in the granite of quartzite inclusions veined by identical "felsite".

Relationships proving the finer acid rocks to be post-granite are scarce. North of the Dee one sizeable "felsite" cuts cleanly through the mass of mixed granite/diorite rocks already referred to. At locality

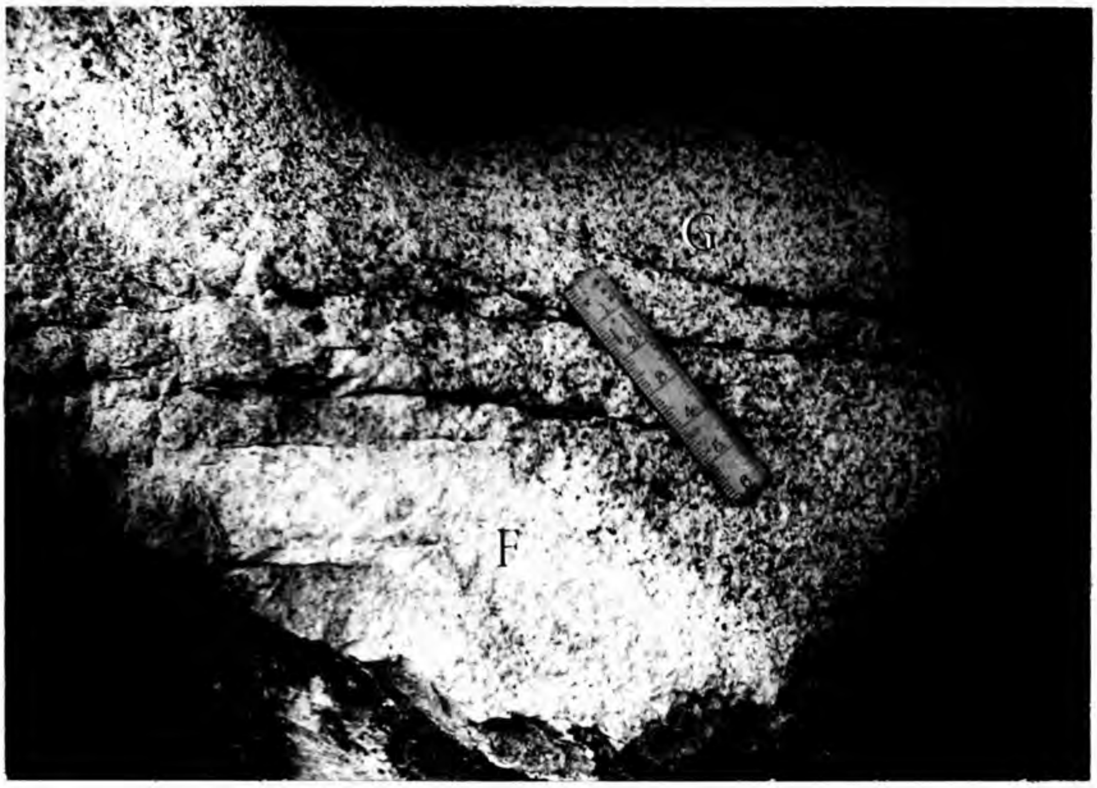


Plate 108. Junction of "felsite" (F) and granite (G) crossed by a set of joints.

Plate 109. "Felsite"(F) and granite (G) intruding quartzite (Q).

(040,082), south of the Dee, typical fine "felsites" appear to bear the same relationship to the granite, yet not far away the granite is of variable character, becoming much more like a porphyritic felsite. In fact much of the evidence of granites/"felsites" relationships involves the variation of one towards the other (such as that of figure 51). This seems to be particularly true of porphyritic "felsites" or microgranites which are frequently to be found in association with granite. They may intrude granite at one point yet grade perfectly into it at another.

Composite sheets are not unknown. One sheet of dark reddish "felsite", only inches thick, has been traced for over 20 feet in quartzite and granite on the southeastern slopes of Creag Choinnich. At one point it lies wholly within a 6-inch sheet of granite (pl. 109). The igneous boundaries are sharp for the most part though at two points the two acid rocks have merged. This evidence shows that two fluids were in coexistence, for the position of the finer <u>within</u> the coarser proves that it is not simply a chilled phase of the other produced in situ.

One wonders, however, whether in the case of larger sheets grading from granite to "felsite" etc. along their length this is evidence of granite magma "giving birth" to felsitic material. Such instances have been observed on Ant-Sron where a gradual change along acid sheets has been confirmed. It can be seen from the map, too, that at the granitic "end" the sheet becomes an irregular and much larger body in contrast to the felsitic end which is significantly much more sheet-like, concordant and thinner.

Examination of the large sheet of true felsite on Càrn nan Sgliat has established that on a bigger scale, too, fine- and coarse-grained acid igneous rocks may be broadly contemporaneous, the two phases blending together locally. Sudden variations in texture and/or colour exist in both rocks. Contact portions tend to have a deeper red colour than either parent,

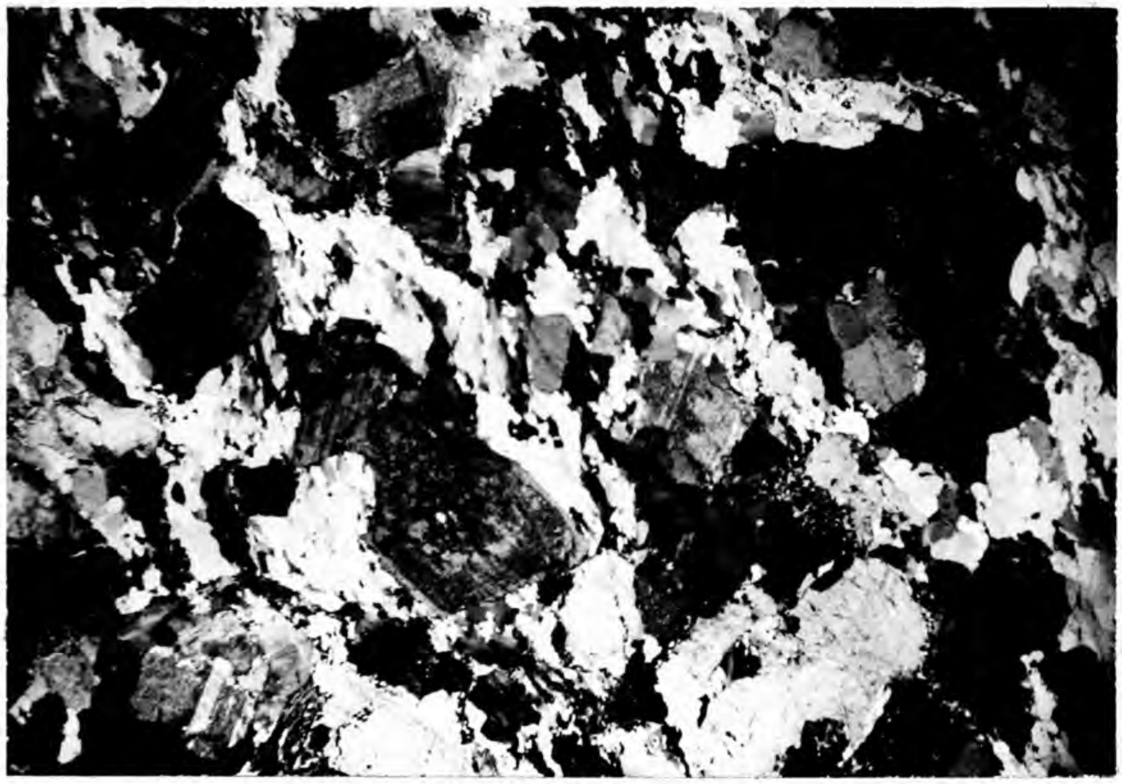


Plate 110. A typical specimen of granite showing straining of quartz crystals. CB 20/1; x-nicols; x 15.

while one exposure constitutes a highly porphyritic acid rock unlike any other found, though obviously produced locally at the time of intrusion.

Petrology

(a) Granites

Hand specimens of granite are grey to cream to distinctly pink, speckled rocks, varying in grain size from 5 mm downwards. Even in the hand specimen granites may be divided into biotite-rich and biotite-poor types, but while two separate granites can thus be mapped over short distances (see p. 87), no overall distinction on this basis is feasible. Any foliation present is due largely to the common alignment of micaceous minerals.

A microscopic study of thin sections of over 70 specimens reveals that most granite samples conform to the following pattern:-

Quartz varies in abundance from 25 to 35% of the total volume of the rock. Invariably recrystallised, it is highly strained, forming lenticular growths (pl.110). Its relative purity compared to quartzite speaks for an igneous origin, as already implied on page 45 .

The combined felspar percentage is around 60, but weathering and sometimes obscurity of the twinning prevents details of the felspar ratios from being obtained. Generally the potash and plagioclase felspars are about equal, plagioclase perhaps slightly more abundant save in finer granites or certain biotite-rich types where the reverse is true.

The potash felspar tends to have crystallised earlier, as can be deduced from the inclusion of some of it in the plagioclase (this is not an exsolution effect). The potash felspar is also the more strained, undulose extinction being the rule. Under crossed nicols strain shadows closely resemble bands of exsolved but untwinned albite which are present in other crystals. Strain shadows are also confusingly like microcline twinning. Incipient conversion of orthoclase to microcline has been found in every thin section examined. This, incidentally, is in direct contradiction to the evidence of Oldershaw, and it must be concluded that this change is undertaken in a sharply-defined zone right at the edge of the Lochnagar Granite, and in all outer granites. It is believed that the straining, the development of microcline twinning and the exsolution of albite are concomitant.

Universal stage measurements have established that the 2V angle for the potash felspar is consistently within the narrow range of that for microcline in some granites. In other granites orthoclase is also present, the angle 2V approaching 50°. Good magmatic zoning has been observed in the potash felspar of but one specimen, CB 177.

The interfacial boundaries between the potash and plagioclase felspars tends to be marked by complex intergrowths and sometimes "swapped rims". The An content of the plagioclase is usually reduced where it adjoins the potash felspar.

In certain specimens the plagioclase displays a fair crystal outline. This, and characteristic cloudy brown alteration, help in its recognition when other features are obscured. Zoning is ill-defined in most cases, but in conjunction with albite and sometimes Carlsbad twinning the variation in composition across the zones has been ascertained in a few granites (see p. 88). Optical determination of the composition shows that the plagioclase of the area, excluding the edge of the Lochnagar mass, is variable. Variation has been found even within the limits of one thin section. The unzoned plagioclase of most granites lies within the range of An_1 to An_{15}, the limits being An_{5-6} and An_{17}. Zoned crystals may have cores as high as An_{28}.

Biotite, present in most sections, is always associated with replacive chlorite. Crystals, though small, tend to be bent. There may be small amounts of muscovite intergrown with the biotite, but the total micas plus

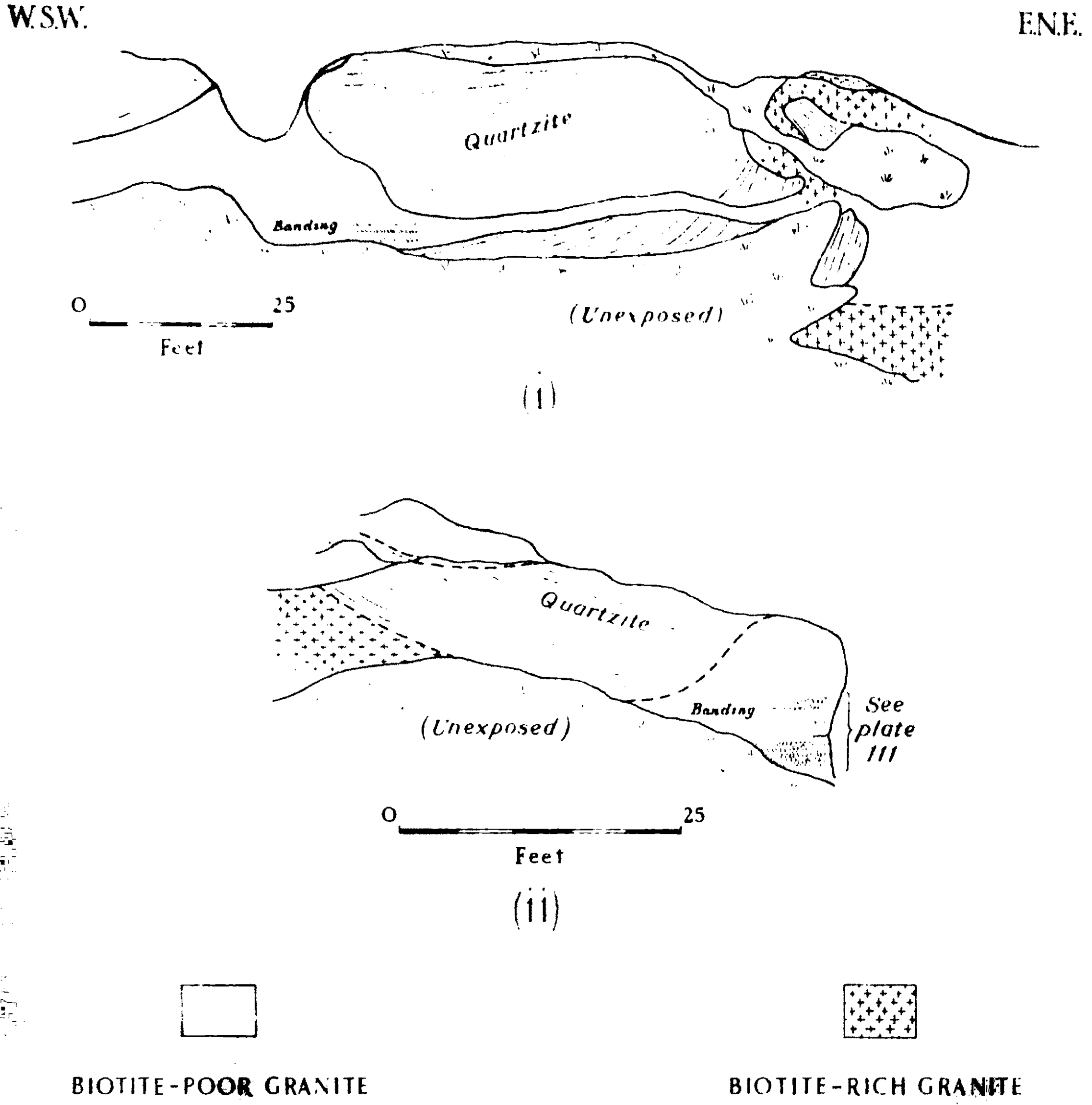


Fig. 52. Diagrammatic sections containing biotite-poor and biotite-rich granite. Section (ii) lies just to the E.N.E. of (i).

Plate 111. Smooth-weathering (s) and rough-weathering granites (see also plate 112).

Plate 112. Close-up of part of plate 111, showing junction of the two types of granite.

Plate 113. Local mineralogical variation in the granite, the banding suggesting a derivation from the granitisation of banded schist etc.

chlorite do not exceed 10% of the whole-rock volume except in the biotite-rich granites.

Zircon, with pleochroic haloes against biotite, is a minor accessory. Small amounts of magnetite may be present. Apatite is absent.

Variations in granite are best developed in minor bodies. For instance at locality (017,076) a biotite-poor and a biotite-rich granite can readily be distinguished (fig. 52). There is no evidence of their relationship, though within this area a granite of intermediate nature has been found. Another exposure is of thin bands that alternate between light and dark featureless granite, but their regularity and close proximity point to a genesis by local contamination.

Sometimes no petrological distinction can be made between two apparently different granite bands (pl.111,2). In the instance illustrated the granites cannot be distinguished on the cliff-face at right angles to that shown. Their only difference is that one of them is smooth-weathering on surfaces parallel to a weak micro-jointing.

Mostly, however, field differences reflect genuine mineralogical ones. Distinctly biotite-rich granites tend to be very slightly coarser than normal and there may be direct evidence of very local "contamination".* In CB 28 there are, in addition to coarse flakes of biotite, clots of finer biotite showing two directions of preferred orientation. Still smaller trails of opaque material may be matched in a schist inclusion. A few of the trails are also enclosed by the felspars, proving that the digestion of schist by the granite commenced at an early stage in its cooling history. One part of the inclusion mentioned has been partly "detached" from its parent. This must have involved a measure of mechanical action on the part of the granite.

Distinctly mica-deficient granites, generally slightly finer than average, are characterised by intergrowths between the potash felspar and

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* This may also be true of the "anorthosite" previously referred to (p.81). It comprises andesine (An_{36}) with a little anhedral diopside and sphene, probably xenocrysts derived from skarn rocks.

Plate 114. Intergrowths between alkali felspar and quartz (graphic texture).
CB 24; x-nicols; x 20.

the quartz (pl. 114). The felspar is otherwise similar to normal but makes up to 62% of the rock by volume, while plagioclase may be as low as 3.4%. Chlorite, often the only dark mineral, is around 2% of the rock.

Graphic intergrowth tectures are best developed in the intrusions on Creag Choinnich, especially on the southeastern slopes. As with biotite-rich and biotite-poor types, there are both sharp and gradational contacts between normal granite on the one hand, and apparently finer pink granites and red acid rocks on the other. The latter are seen in section to be mostly felsitic (and constitute further evidence of granite/felsite associations). The pink types may be almost entirely of felspar, which is seen in thin section to be potash felspar of coarse grain-size. In CB 24 these crystals are entirely dominated by a sieve texture caused by graphic intergrowth with quartz. Blebs of the latter range from 0.1 mm in maximum dimension downwards. In any one felspar there are most often two regular patterns of different size adopted by the quartz.

When plagioclase is more abundant certain crystals have a similar graphic texture with quartz. It is noteworthy that under such conditions the quartz crystals have been protected from shearing effects and show no strain extinction whatever.

In some pegmatitic granites (which are never extensively developed) plagioclase with quartz inclusions may be the only felspar represented.

All the above-mentioned rocks can be distinguished from normal in the field, but other variants can only be established as such under the microscope, mainly by the nature of their plagioclase felspars. When Carlsbad/albite twinning is present zoning is common, a feature shared with diorites. In unzoned crystals the composition is An_{16} or less, otherwise they range from An_{24} to An_{31} in the core to An_{17} or less peripherally.

Specimens of granites close to the main Lochnagar mass show no significant variation in their plagioclase anorthite content (circa An_{12}) as compared

Plate 115. Porphyritic microgranite with intergrowths of quartz in the felspar crystals. This phenomenon is found in the groundmass also. CB 202; x-nicols; x 35.

Plate 116. Weakly porphyritic "felsite" with graphic texture. CB 42B; x-nicols; x 35.

to that of granites from Creag Choinnich and Craig Leek (lower in value). It would thus appear that the anorthite content of the plagioclase rises as one crosses into the main granite mass, where conditions were more stable during crystallisation.

(b) "Felsites" etc.

Microgranites associated with granites have the same general petrology. This, in conjunction with their local extent, suggests a close relationship in origin. Existing differences lie in their lesser amounts of mica and greater amounts of quartz. There are no equivalents of the biotite-rich granites.

The main masses, particularly sheets, of the biotite-rich granites further contrast with granites in being porphyritic in varying degrees, and it has been observed that deformation textures are best developed in those most strongly porphyritic.

The sheet nearest to being non-porphyritic (represented by CB 89) can only with difficulty be distinguished from quartzite in the field. Very rare phenocrysts of quartz, less than 2 mm in diameter, have been found after an extensive examination of the available exposures. The rock appears granular in thin section and, apart from iron ore, all crystals visible are anhedral. Potash and plagioclase felspars are equally represented. Flakes of mica are sporadic and tiny, being partly altered to chlorite, though fine new muscovite has been created from the alteration of plagioclase.

CB 73 is essentially the same, though deformed phenocrysts of quartz can be seen under the microscope, while particularly obvious are larger "spots" of partially altered biotites showing a common elongation caused by deformation. In yet another similar sheet there are anhedral crystals and smaller aggregates of yellowish garnet.

In the more porphyritic types the order of abundance of phenocrysts tends to be quartz, potash felspar and then plagioclase, their size being

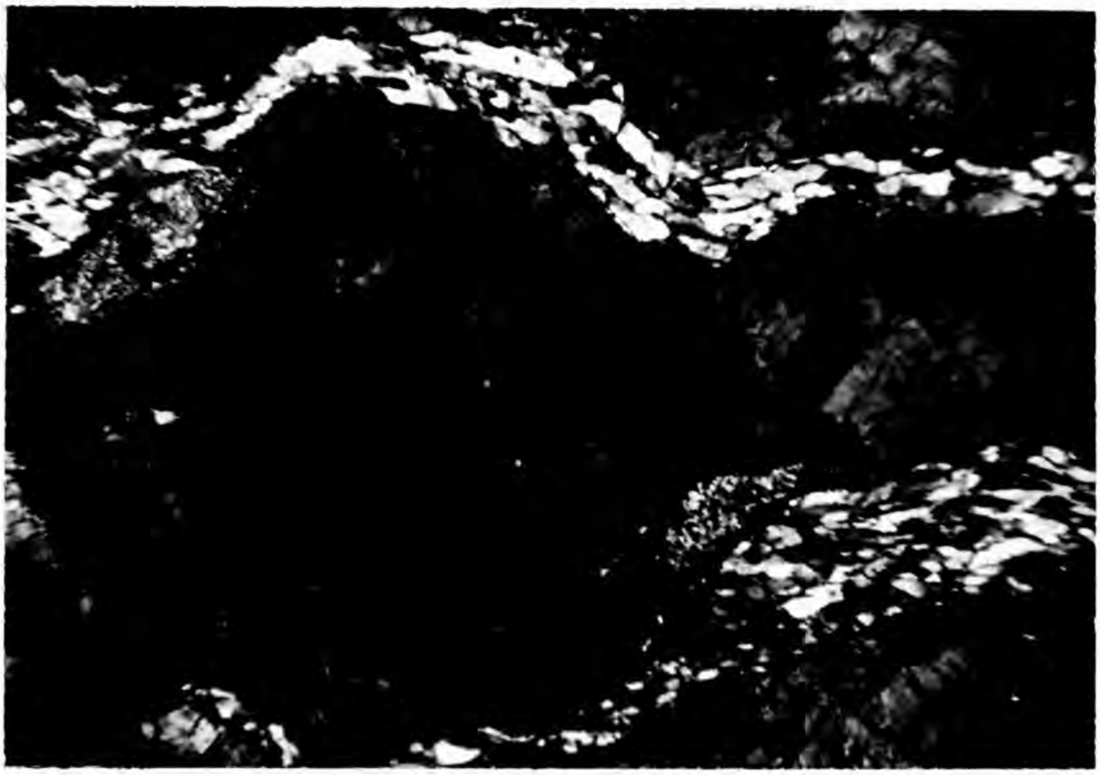

Plate 117. Highly deformed and sheared granite(?) in which the felspars show bending of the cleavage (centre) and even peripheral granulation. CB 37; x-nicols; x 35.

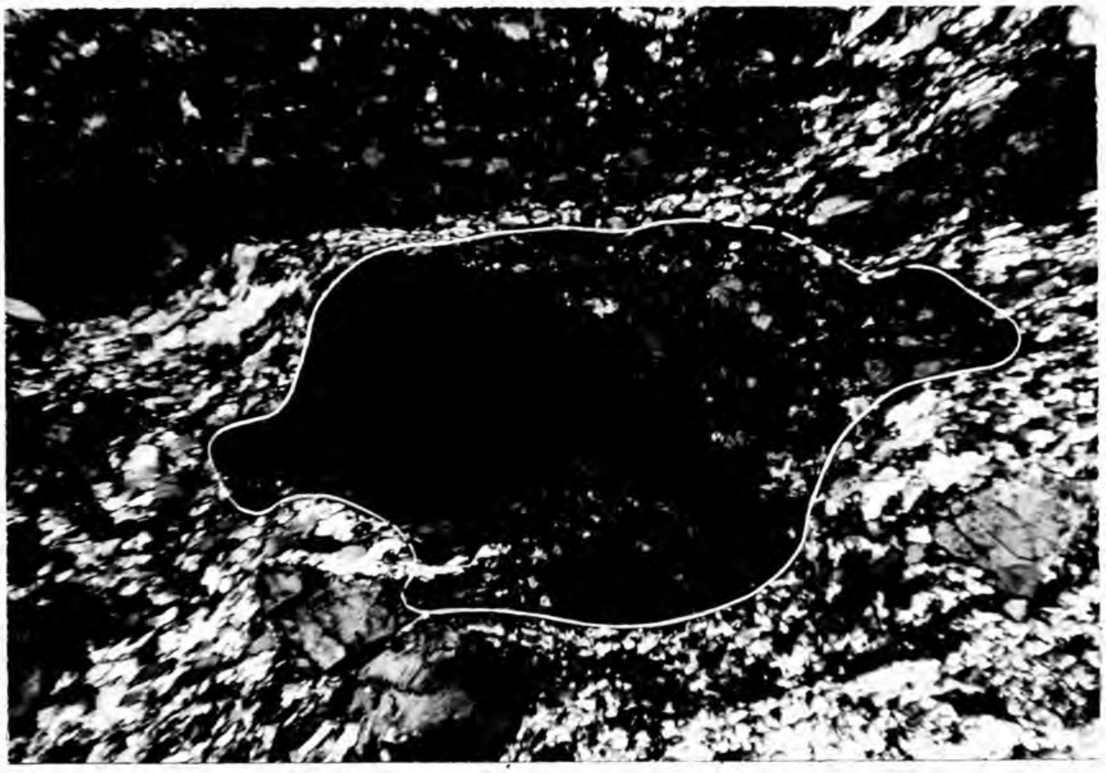

Plate 118. More porphyritic aspect, possibly caused by local extremely intense deformation. The central felspar crystal is surrounded by a large area of granulated felspar particles (within white line). Adjacent to CB 37; x-nicols; x 35.

comparable with those of typical granites. In some examples quartz intergrowths may be seen at the edges of potash felspar crystals*.

Where deformation of the minor sheets has been intense, as in CB 54, it becomes more difficult to recognise the phenocrysts of quartz, though they are doubtless always present. Even where deformation has been more acute, e.g. CB 59, the thin section permits one to see the extent to which the felspars have resisted it. The structure produced is much like an incipient strain-slip in a mica-schist, and cannot be confused with magmatic flow structures as can be the case with microdiorites.

Lastly CB 37 illustrates the possibility of a purely mechanical gradation between acid porphyry and granite, brought about by extreme deformation. Granulation of even the felspar in this flaser-type rock has rendered it difficult to decide whether this was originally a granite or a felsitic porphyry. In the field it looks distinctly granitoid, while under the microscope certain porphyritic features come to light (pl.117,8).

B. INTERMEDIATE & OTHER ROCKS

Field Relations

There is a gradation from diorites into microdiorites and thence into epidiorites. As with the acid rocks the finer types tend to be porphyritic. All of these will be considered together.

None of these rocks forms large bodies, though there are several sizeable sheets of fine diorite at the northern end of the area. These follow a Caledonoid trend and diminish in size southwards as they grade into true microdiorites. No true diorites have been found south of the Dee.

The dioritic sheets are less regular than their microdioritic counter-

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* Subsequently two samples have been found with well-developed intergrowths between potash felspar and quartz (pl.115,116).

parts and exhibit bifurcations and sudden terminations. The most easterly sheet on Craig Leek cuts straight up the steep slope at the northern end of the feature. At the top, where the limestone is flat-lying and unde - formed, the vertical sheet turns over to form a sill.

The microdiorites at the southern end of Craig Leek illustrate another feature - that dykes tend to be more porphyritic and of regular form, while the "sills" (or semi-concordant bodies) are thicker, less regular and less porphyritic. Some of the thin porphyritic dykes post-date the granite. This relationship has been confirmed outside the area mapped.

From the maps it would appear that nearly all thin dioritic sheets are post-granite. This is not so. Ample evidence of relative age from the contacts of the long dioritic sheet in the centre of Craig Leek shows that at its northern end it has withstood the invading granite which has simply penetrated down either margin into the quartzite.

An anomalous situation is apparent at a locality a few hundred yards to the northwest where a thin porphyry dyke branches. One branch is clearly terminated by a granite sheet, while the other branch itself terminates the same or an even earlier granite sheet.

On Creag Choinnich several "dark porphyries" lie within the granite. The largest of these is discontinuous, and there can be little doubt that its various outcrops are but remnants or rafts. Other evidence, dealt with later, also points to an earlier age for these porphyries.

All the intermediate rocks form sharp boundaries with the metasedimentary rocks. Apart from the late porphyries (probably tension dykes) there is no substantial evidence for or against displacement across the bodies. Dioritisation of schist has been found only once, and this on a small scale.

True epidiorites and amphibolites are the sparsest of the igneous rocks. As thin dykes or highly inclined sheets, seldom more than a few feet wide, they are of limited length, due in some cases to post-intrusive deformation.

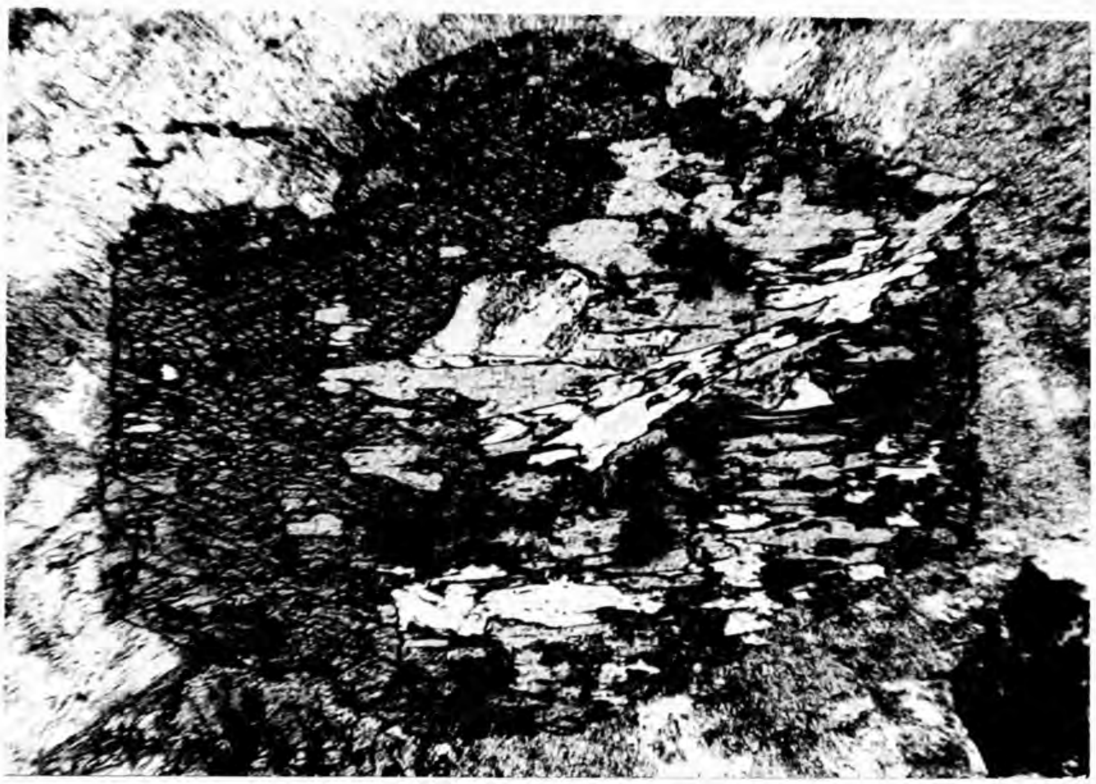

Plate 119. Hornblende being replaced by biotite, chlorite and felspar (white). CB 203; x 80.

Many are lenticular in plan, too, but may still maintain an obviously transgressive relationship to the country rocks. Their igneous nature is thus not in doubt. Other bodies which exhibit no distinct igneous features may be hornblende-biotite-schists of sedimentary origin.

Petrology

(a) Diorites - Porphyritic Microdiorites

In the main diorites the felspars and amphiboles are commonly around 3 mm in length. Quartz is restricted to interstitial areas or to intergrowths with felspar and is of limited abundance though of widespread occurrence. Mostly of plagioclase, the felspar is subhedral to anhedral and variably sericitised. Zoning is common and sometimes oscillatory. Irregularities therein may result from metasomatic effects induced by later granite intrusion. The cores of some felspars reach a composition of An_{28}, but others may be as low as An_{15}. This, too, could be the result of metasomatism. Finally, small amounts of microcline, tending to corrode the plagioclase, complete the picture.

The amphibole, a pale olive green hornblende showing weak pleochroism, has better crystal outlines than the felspar. It has been partially replaced in the core by biotite, chlorite and sometimes felspar (pl.119). Pyrites, too, may be found in this situation.

Sphene is a not-too-common accessory, forming isolated blades less than 1 mm long. Smaller crystals of "epidote" are, like sphene, subhedral, but more sporadically distributed. Highly elongated crystals may be slightly warped, while the smallest have been found enclosed by sphene.

Diminutive crystals of apatite have been tentatively identified. Others, pleochroic needles of a neutral to drab olive hue, are as yet unidentified. Although much like tourmaline, $\epsilon > \omega$ and not the reverse. They cannot, therefore, be tourmaline and may be a coloured variety of apatite.

Diorites not fitting into the above picture are quartz-diorites. In them the felspars are more lath-like, often with Carlsbad/Albite twinning, and the crystal edges have a composition less than An_{20}. The hornblende is less well-formed than before, and biotite, instead of replacing the amphibole, exists as separate and equally abundant crystals. Chlorite is absent but sphene is as just described.

Some of the quartz-diorites reveal the presence of puzzling new needles situated within quartz and felspar crystals. They are six-sided in cross-section and range from colourless to very pale green. They greatly resemble apatite yet there appears to be a gradation in type to prisms of unmistakeable amphibole. This is all the more puzzling because of the discovery in CB 171 (microdiorite) of similar, and at times larger, needles which are unquestionably of apatite and which display no variation. The further discovery in typical diorites of such prisms with a coloured central zone and seemingly higher birefringence would suggest that such phenomena could in fact be apatite needles moulded on to needles of amphibole. This specimen (CB 171) is also unusual in that the plagioclase ranges in composition from An_{50} outwards to rims of An_{18} or even less.

The area of "mixed" granite and diorite, described on page 82 , partly consists of abnormal types of diorite, such as CB 158, which is weakly porphyritic. The amphibole has completely recrystallised, though preserving its initial outlines. The central zone of one amphibole in the section examined was found to contain partially altered material probably representing an earlier-formed pyroxene. In parts of the section large microclines have partially replaced the plagioclase.

Other samples may be enriched in quartz, the most obvious sign of granitisation. Usually there is a higher-than-average proportion of potash felspar, and in CB 155 large plates of microcline surround the almost unaltered hornblende crystals.

The remaining questionable diorites are situated at Sron á Bhruic (map 2) where, as before, granite is ubiquitous. Patches of doubtful diorite have been found surrounded by granite, and sometimes the latter grades into a darker type in which xenoliths of granitised schist are just discernible. Are these diorites to be regarded as partly digested inclusions of hornblende-schist?

In thin section certain characters emerge to support an origin by contamination of granite. For instance in CB 164, distinctly leucocratic and coarser than normal, biotite is well-formed; but hornblende is scarce and never associated with the mica. The percentage of quartz lies between that of a granite and that of a true diorite, while the plagioclase composition trends outwards from crystal cores of An_{38} or more to An_{20} or less.

The effects of the granite on earlier diorites etc. are slightly different. At the contact the diorite is usually enriched in quartz of a coarser grain-size than that already present. Metasomatic zoning has been induced in the felspars, which also show an increase in cloudy alteration (equivalent to a reddening in the hand specimen - also noted by Oldershaw). Chlorite may be present after hornblende.

(b) Porphyritic Microdiorites - Epidiorites

The gradation already mentioned in this group of rocks is believed to be a function of age coupled with progressive regional metamorphism. In the field the least altered sheets of such rocks display no cleavage or schistosity, and phenocrysts (usually present) are clearly visible.

The first sign of transformation is manifest in a weak marginal cleavage, parallel to the contacts. Increase in metamorphism has caused the cleavage to spread to affect the whole body. At this stage the phenocrysts are difficult to recognise, and further alteration, caused mainly by increase in temperature, has served to produce a hornfels or a schistose rock (epidiorite). Thin section examination of a number of specimens confirms these gradations.

The least deformed type of porphyry is typified by CB 71, which can be seen to have three "phases":- a fine crystalline groundmass, numerous small to medium-sized felspar phenocrysts, and rarer large felspars and amphiboles. The groundmass is of quartz-diorite composition, the quartz crystals reaching 0.1 mm in length. Potash and plagioclase felspars lie in intimate association, and are completely anhedral. The amphibole is irregular in shape and comprises green hornblende. Its associates are iron ore and small amounts of sphene, also anhedral. The amphibole ranges in size up to that of the second phase felspar (0.2-0.3 mm in length), which is subhedral, having characteristic outlines, but sutured edges. Some individuals have been deformed by flow, though there is no immediately obvious flow-texture. Twinning and zoning, though somewhat blurred, indicate that most, if not all, of the felspar is plagioclase. The large amphiboles have largely recrystallised. The third and largest phase of felspar is of plagioclase only, and exhibits good crystal outlines and a rhythmic reversal of the zoning.

The initial stages of deformation in these porphyries, prior to the appearance of a megascopic cleavage or schistosity, have involved loss of good felspar outlines and the association of biotite with the hornblende. The flow alignment of the felspar and amphibole crystals has been intensified by shearing parallel to the margins of the body.

The next stage in the gradation, such as that reached by CB 12, reveals a true cleavage, also parallel to the margins. Microscopic examination shows that the rock has definite affinities with flaser rocks. It is at this stage that interpretation of structures becomes problematical as has been found with CB 82.

CB 82 represents the large dyke or steeply inclined sheet that runs

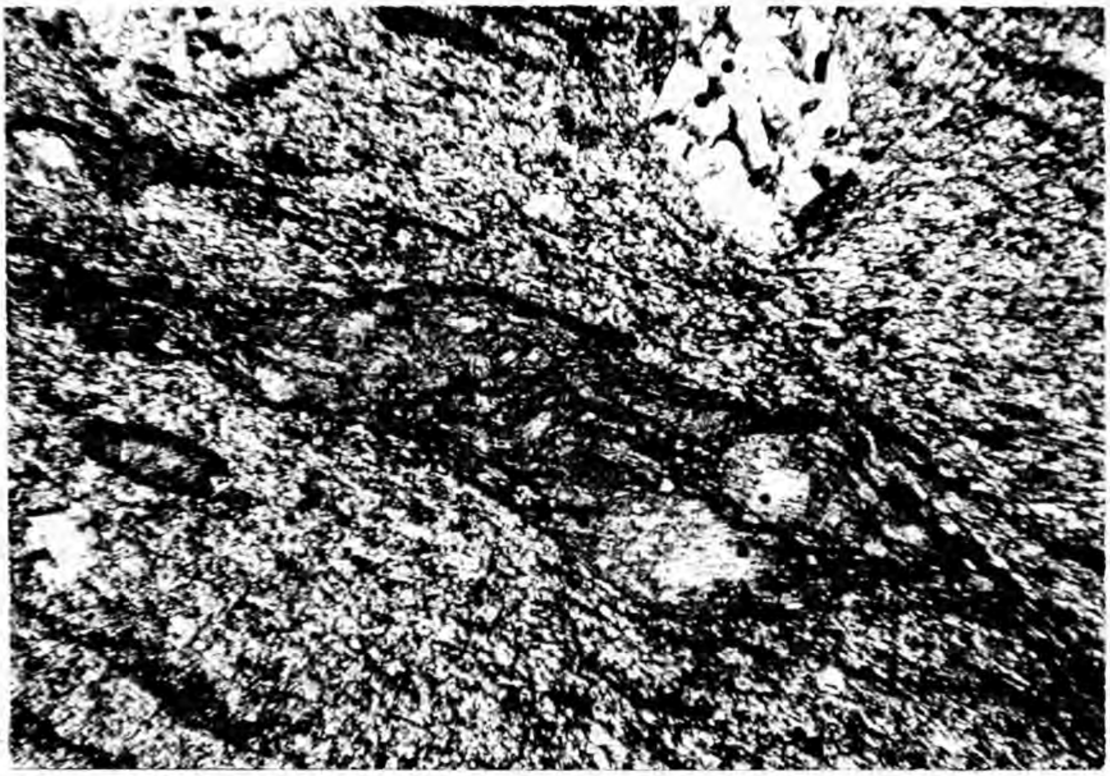

Plate 120. Augen-like recrystallised phenocrysts(?) of actinolitic amphibole. CB 82; x 35.

across the northwest side of Creag Choinnich, apparently transecting the granite. The hand specimen is of a seemingly highly cleaved dark igneous rock. The contact with the granite is straight and parallel to the cleavage, while in the granite there is a weak orientation of the biotites at approximately 90° to the junction. However, detailed examination of the contacts reveals that microstructures in the porphyry are cross-cut by those of the granite. Apophyses of the latter penetrate the former, too.

The groundmass of the microdiorite is of granular felspar and anhedral hornblende well-orientated parallel to the cleavage. Larger masses of magnetite tend to be so aligned also. The felspar phenocrysts (An_{45} - An_{53}) show a partial alignment, while the groundmass appears to have flowed around such felspars as are randomly orientated. The hornblende phenocrysts(?) are abundant but vary from the dark green common type to a lighter-coloured actinolitic type which is fibrous. Though some may be individual crystals most phenocrysts(?) have clearly recrystallised to a highly tapering, augen-like shape (pl. 120).

In this condition one wonders whether the flow-texture is of igneous or dynamic origin; and whether the amphibole "augens" are highly deformed and recrystallised phenocrysts, or aggregates of flowing crystals, or even porphyroblasts.

The almost complete lack of deformation in the felspar phenocrysts favours the igneous flow hypothesis for the texture, while the mere existence of felspar phenocrysts suggests that large crystals of hornblende are likely to have formed also. However, it is the precise shape of the amphiboles which has the most significance. The undulating nature of their tapering "tails" can only be easily explained by deformation (pl. 120). It is thus inferred that these amphiboles were centres where pressures in the rock were relieved, possibly not long after intrusion. They gave rise, in other words, to planes of selective shearing. This theory also fits well the case of similar bodies, more highly deformed, where the phenocrysts have been almost completely sheared out.

Next in the main gradational sequence is a rock like CB 18, in which there is little evidence of a porphyritic texture. It is more melanocratic than anything so far described. Crystals are mostly fine-grained and anhedral. There is little quartz, while of the felspars only plagioclase can be identified with certainty. Nevertheless there are in the section several large aggregates of interlocking plagioclases which (the aggregates) have a rough felspar outline. That some of the plagioclase individuals are also coarser than the groundmass is a good indication that this rock was originally a porphyry. Bright green hornblende is widely associated with, and being replaced by, deep brown biotite and lesser amounts of chlorite, all of which define a weak cleavage. Some of the biotite and opaque ore material tendsto be concentrated along minute shears. The rock has thus affinities with both epidiorites and microdiorites.

Thermally altered equivalents of CB 18 still retain their weakly discernible porphyritic features, but the biotite has been almost entirely upgraded to hornblende with the development of a hornfels texture. The presence of chlorite in such situations could be related to late shearing.

In other specimens of epidiorite, usually reduced to isolated exposures or tectonic inclusions, the remnant phenocrysts lie with their maximum length parallel to the schistosity, proving the latter to have developed along planes containing the direction of flow. It is therefore likely that the schistosity is parallel to the original contacts of the body.

In more highly deformed, or non-porphyritic bodies, it may be impossible to ascribe an igneous or a sedimentary origin to the epidiorite. Field evidence becomes critical, but this does not distinguish between remnant phenocrysts and porphyroblasts, a problem which is not unknown in the area.

At the end of the gradation scale lies a group of quartz-free rocks containing little or no felspar. They are better referred to as amphibolites or hornblende-(biotite)-schists or hornfelses. Of these CB 143 reveals sporadic clusters and groups of coarser-than-usual hornblende crystals, which, though indefinite, is nevertheless suggestive of an old porphyritic texture.

Conclusions

Although the succession of the Dalradian metasedimentary rocks is not obvious, the observed relationships among the formations are compatible with the sequence proposed by Bailey (1922). Evidence from the field, from thin section examination and from petrofabric analysis has shown that these rocks have suffered polyphasal metamorphism. The rocks were first regionally metamorphosed, during which period three deformation phases were effective. These were largely successive, though in parts of the area the last two may have overlapped considerably. The first was responsible for the development of the main Dalradian nappe-structures. This was followed by cross-folding, accompanied by the climax in the development of new metamorphic minerals with the ultimate appearance in the pelitic rocks of kyanite. The minerals thus formed were then deformed by the third phase of folding.

It is probable that due to local intrusions of granite there was sufficient rise in temperature to effect the growth of thermal minerals beginning with plagioclase, before the end of the third deformation phase. The higher-temperature minerals are clearly closely related to individual intrusions of granite and were formed after the period of maximum folding of the third phase. Around the Lochnagar granite this later mineral growth is seen as a contact thermal effect, the pelitic rocks having been converted to high-grade hornfelses. In the contact areas (including the local extensions of the granite) sillimanite has formed after an even later phase of deformation.

The intrusion of granite itself is believed to have come about in phases: its deformation is anything but uniform, while the thermal minerals have developed over a relatively long period of time. Petrofabric analysis of granite/quartzite specimens has indicated that parts at least of the granite have suffered two deformations though the relative ages of these have not been determined. The igneous cycle commenced with the intrusion of basic rocks followed by dioritic rocks and culminating in acid ones. The latter were emplaced in pulses of forceful intrusion alternating with periods of non-dilationary granitisation. During forceful intrusion partially crystallised granitic magma was forced past the area of heating to form partially concordant sheets of acid porphyry or felsite.

Finally it can be claimed that field work has shown that extremely complicated structures in both the igneous and metamorphic rocks can be mapped on a much smaller scale than is normally attempted. This in turn has revealed that these smaller-scale features mirror the major structures.

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