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R. J. HOWARTH, C. KILBURN AND
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THE BOULDER BED SUCCESSION AT GLENCOLUMBKILLE, COUNTY DONEGAL

By R. J. HOWARTH, C. KILBURN* AND
B. E. LEAKE

Department of Geology, Bristol University and
*Department of Geology, University of Leeds

(Communicated by J. C. Brindley, M.R.I.A.)

(PLATE 1)

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ABSTRACT

A DETAILED description is given of the sedimentary sequence at Glencolumbkille, County Donegal. This is an important area for the study of the Dalradian Boulder Bed because of the completeness of the succession and the excellency of the exposures. Graphitic schists are followed successively by limestone, striped schist, dolomite, dolomitic boulder bed, semipelitic boulder bed, increasingly psammitic boulder bed with interbedded quartzites, a distinctive thin calcareous dark schist, and finally a thick quartzite, that of Slieve Tooey. The Boulder Bed is described in special detail. Four important periods of movement are recognised in the area, of which the third and fourth are the most important as they produced the Glencolumbkille anticline and syncline. Four lesser phases of late movement followed. The metamorphism reached a peak during the declining stages of the second folding in the almandine amphibolite facies when the assemblage muscovite-biotite-garnet-quartz-oligoclase-ilmenite-graphite was stable in the pelites. The presence of dolomite marble is believed to indicate that the partial pressure of carbon dioxide was high. A late period of re-heating with static growth of biotite and hornblende in the upper greenschist facies was followed by a retrogressive, lower, greenschist facies metamorphism and the production of chlorite. The metadolerites are described.

CONTENTS

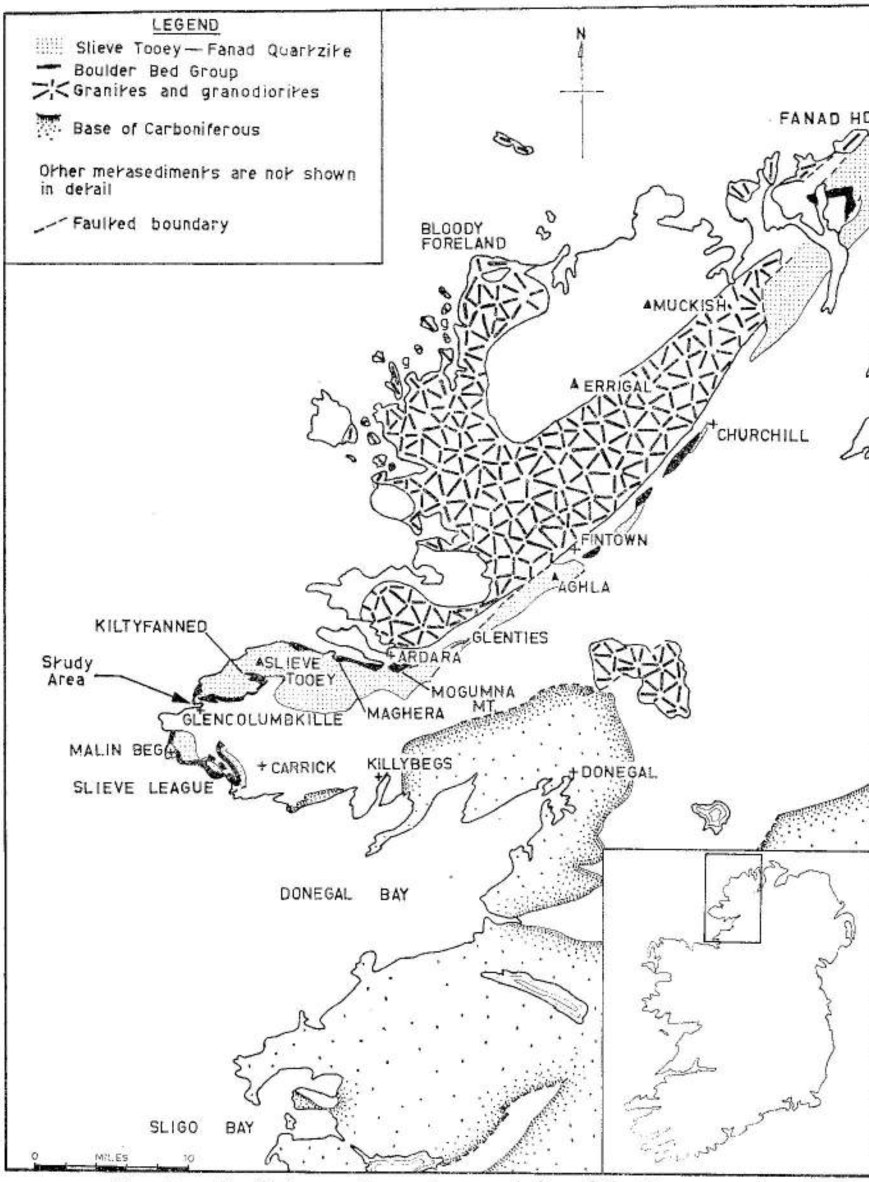
1. INTRODUCTION

Glencolumbkille lies in the southwest corner of County Donegal, near the western extremity of the Slieve League promontory (Fig. 1). The rocks are Dalradian metasediments together with metadolerites and within this sequence the important Boulder Bed Group occurs, which enables the Dalradian rocks of Connemara, Mayo and Donegal to be correlated with each other and with the Dalradian of Scotland. For Ireland, the Boulder Bed succession is particularly well exposed, thick and complete at Glencolumbkille and four miles NE of Glencolumbkille, at Kiltyfanned. The main purpose of this paper is to describe this succession and to give a general account of the geology of the area. A subsequent paper by one of us (R.J.H.) will describe the Kiltyfanned area and it is hoped that these studies will not only enable more precise stratigraphical correlations to be made but will also throw light on the origin of the Boulder Bed and the conditions of its deposition.

It is surprising that the geology of this excellently exposed and stratigraphically most important area, has received practically no attention since the Geological Survey of Ireland published the memoir on the district (Hull *et al.*, 1891) which is to be found on 6 in. sheets County Donegal Nos. 80 and 81 and on the 1 in. geological map No. 22 of the Geological Survey of Ireland. This is all the more remarkable as Kilroe, who largely wrote the memoir, fully appreciated the outstanding importance of the Boulder Bed and emphasized the excellency of the Glencolumbkille succession and exposures and subsequently (Kilroe, 1907) went on to correlate this horizon through to Connemara. One must warmly acknowledge this outstanding early work, with its careful account not only of the Boulder Bed but also of the structures in the metamorphic rocks. Nevertheless, the precise sedimentary succession, which requires detailed mapping for its elucidation, is still unknown and Kilroe (*in* Hull *et al.*, 1891), Anderson (1953, p. 399), and Kilburn, Pitcher and Shackleton (1965, p. 348) have given only very generalised successions.

The Glencolumbkille valley, ending at Glen Bay, is dominated by an east-west, Boulder Bed and quartzite ridge, reaching 800–1,000 feet, which limits the northern edge of the valley. The low ground is formed of softer pelitic schists, limestones and dolomites. The ridge ends abruptly at Glen Head in a precipitous 700 feet sea-cliff in which practically the whole succession is exposed although the Boulder Bed is probably tectonically thinned here. After Glen Head, the best exposed part of the ridge is the crag of Craigbeefan, which, apart from some screes, especially in the west, is excellently exposed.

In the present work, the first author has been responsible for the stratigraphy, sedimentation and the measurement of the type sections; the third author for the mapping of the area, the study of the metadolerites and, together with the second author, for the metamorphism and structure.



(Based on the Ordnance Survey by permission of the Government)

FIG. 1—Geological sketch map of north-west Donegal to show location of area studied.

2.0. Stratigraphic Succession

The succession passes from a pelitic base into an increasingly carbonate dominated sequence of schists until this gives way to a thick carbonate unit underlying the Boulder Bed. This group is characterised by the presence of randomly distributed, unsorted, clasts (ranging in size from granules to boulders) of igneous and sedimentary composition in a finer grained matrix. The matrix of succeeding boulder beds is calcareous at the base becoming increasingly psammitic towards the top until it passes into a blanket quartzite —feldspathic quartzite formation. Quartzites are also intercalated between the individual boulder beds. The top of the Boulder Bed appears to be marked by a thin pelite formation.

The majority of the outcrops north of Kiltyfanned, four miles north-east of Cashel, have a tectonic break at the base of the upper part of the Boulder Bed (see Fig. 1 for Boulder Bed localities). The transition from the underlying schists into this group is best seen in the Glencolumbkille area.

The stratigraphic nomenclature used in this paper is compared with previous stratigraphical terms used in Table 1.

The generalised thickness ranges for the area are:

		Top not seen.
Slieve Tooey Quartzite	......	620 — 800 + m.
Kiltyfanned Schist	......	10 — 19 m.
Boulder Bed	......	101 — 118 m.
Glencolumbkille Dolomite	......	7 — 34 m.
Glen Head Schist	......	9 — 34 m.
Skelpoonagh Bay Limestone	......	8 — 30 m.
Glencolumbkille Schist	......	200 + m.
		Base not seen.

The units forming this succession are right way up, and have sharp, conformable, contacts of sedimentary origin. Way-up criteria observed are: rare graded bedding, cross-stratification, scouring of tops of quartzites, erosion at the tops of boulder beds and in rare instances depression of thin quartzites beneath large clasts in the Boulder Bed.

2.1. Glencolumbkille Schist

The type area for these schists is south of the marshy ground between Skelpoonagh Bay and the northernmost church in Cashel village, extending to the east on either side of the Murlin River (Plate I). This type locality lies north of the river from 400 to 1,000 yds. N. 40° E. of Glen Lodge. The complexity of the structure prevents accurate measurement of a type section especially as the base of the formation is not seen within the area mapped. The schists are either dull black, weakly foliated, graphitic pelites or else they are prominently striped with light grey psammite layers (0·3–4 cm.) alternating with darker grey pelite bands (0·1–0·8 cm.) as for example, 5,000 yds. N. 59° E. of Glen Lodge. Partly chloritised garnets are often present. Intercalcated grey saccharoidal limestones which weather grey or brownish are common while thin psammites are rare.

TABLE I.

Rock-stratigraphic units described in Glencolumbkille.

<table>
<tr><th>Hull et al. 1891</th><th>Anderson 1953</th><th colspan="2">This Paper</th></tr>
<tr><td rowspan="4">Quartzitic Group</td><td rowspan="2">Slieve Tooey Quartzitic Group</td><td colspan="2">Slieve Tooey Quartzite</td></tr>
<tr><td colspan="2">Kiltyfanned Schist</td></tr>
<tr><td rowspan="2">Boulder Bed</td><td rowspan="2">Boulder Bed</td><td>Upper Boulder Bed</td></tr>
<tr><td>Lower Boulder Bed</td></tr>
<tr><td rowspan="4">Lough Gartan Group</td><td rowspan="3">Glencolumbkille Limestone Group</td><td rowspan="2">Glencolumbkille Dolomite</td><td>Cream Dolomite</td></tr>
<tr><td>Grey Dolomite</td></tr>
<tr><td colspan="2">Glen Head Schist</td></tr>
<tr><td rowspan="2">Killybegs Pelitic and Calcareous Group</td><td colspan="2">Skelpoonagh Bay Limestone</td></tr>
<tr><td>Glencolumbkille Schists</td><td colspan="2">Glencolumbkille Schist</td></tr>
</table>

The pelites and semipelites are typical fine, black schists, in section consisting of variable amounts of muscovite, considerably chloritised biotite, quartz, sericitised and kaolinised oligoclase, calcite, or dolomite with rare garnet and ubiquitous graphite, which makes up to 5% of some of the pelites. Irregular folia of lepidoblastic micas are interposed between quartz-oligoclase-carbonate layers which may, in themselves, have either a mosaic or a schistose texture. The graphite tends to be concentrated in the micaceous layers while calcite or dolomite are usually associated with quartz and feldspar though they often occur in both folia and also in late cross-cutting veins. Spongy garnets (0·1–2 mm. in diameter), full of quartz and ilmenite inclusions have grown along the schistosity and preserve relics of folded foliation and schistosity within them, as also do some of the oligoclase porphyroblasts. Late, randomly arranged, biotite is common while K feldspar, zircon, zoned blue-green tourmaline, pyrite, apatite and up to 5% of leucoxenised ilmenite may also occur.

The composition of the limestones was checked by obtaining the infrared spectra of samples selected from many outcrops and referring to the data given by Huang and Kerr (1960). In all samples dolomite, when present, varied from less than 2% to trace quantities and the stratigraphically lowest samples tend to be the purest limestones.

Correlation with schists lying at this stratigraphical position is possible at Slieve League and Kiltyfanned.

2.2. Skelpoonagh Bay Limestone

The type locality for this formation lies immediately east of Skelpoonagh Bay in the cliff and shore exposures. Elsewhere the limestone is well exposed on the cliffs below Glen Head and west of the Glencolumbkille Schist from 500 to 900 yds. N. 32° E. of Glen Lodge. The exposure along the edge of Skelpoonagh Bay shows 30 m. of limestone and although thickened by folding the upper and lower contacts appear conformable. This is also true north-east of Glen Lodge and at Slieve League.

The limestone is a fairly pure calcite marble with a little quartz, muscovite, phlogopite, pyrite, hematite and feldspar. Quantitative determinations of dolomite, calcite and quartz were carried out using a diffractometer calibrated with standard mixtures of calcite, dolomite and quartz containing silicon powder as an internal standard. The highest dolomite content detected was 7·2% and the maximum quartz content 5·0%. Infrared spectrograms showed that dolomite was either absent or below 0·7% in most of the samples examined.

This unit may be correlated with beds at Maghera to the north-east and Slieve League.

A measured section at Skelpoonagh Bay shows the nature of the transition into Glen Head Schist.

14 Glen Head Schist	c. 39·6 m.
13 Buff feldspathic quartzite	2·17 m.
12 Grey semipelitic schist	0·07 m.
11 White feldspathic quartzite	0·22 m.
10 Grey semipelitic schist	0·20 m.
9 White feldspathic quartzite	0·17 m.
8 Grey semipelitic schist	0·05 m.
7 White feldspathic quartzite	1·77 m.
6 Grey dolomitic semipelitic schist	1·10 m.
5 Grey limestone with psammitic and pelitic lenses	0·15–0·25 m.
4 Grey limestone, weathering buff	0·07–0·50 m.
3 Grey limestone with psammitic partings, weathering buff	1·00 m.
2 Grey sugary limestone; thin semipelitic or psammitic partings. Laminated. Rarely shows quartz segregations	c. 30·0 m.
1 Glencolumbkille Schist	200 +
	277·10 + m.

2.3. **Glen Head Schist**

These schists, which are characteristically striped, form most of Glen Head. Other areas where they are well exposed are : the north side of Skelpoonagh Bay; outcrops 330 yds. north-east of the southernmost church, Cashel; outcrops extending for 65 yds. north of a group of cottages 1,000 yds. N. 31° E. of the northernmost church in Cashel; and outcrops for 330 yds. north of cottages 430 yds. N. 16° E. of Glen Lodge.

From Skelpoonagh Bay to Glen Lodge the lower contact with the Skelpoonagh Bay Limestone is unexposed and there may be a faulted relation with the Glencolumbkille Schist. The base of the formation crops out north-east of Glen Lodge in the core of the Glencolumbkille anticline (see Fig 2 and Plate I).

The type locality is the cliff face 570 yds. S. 14° W. of Glen Tower. Owing to inaccessibility the type section is the base of Section A (Fig. 4) being :—

5	Glencolumbkille Dolomite	
4	Pale grey semipelitic schist with alternations of quartzite and grey semipelite	10·5 m.
3	Thinly laminated grey quartzite with semipelitic partings	4·4 m.
2	Grey dolomitic limestone with siliceous laminae 5–10 mm. apart. Weathering buff	4·0 m.
1	Glencolumbkille Schist	
		18·9 m.

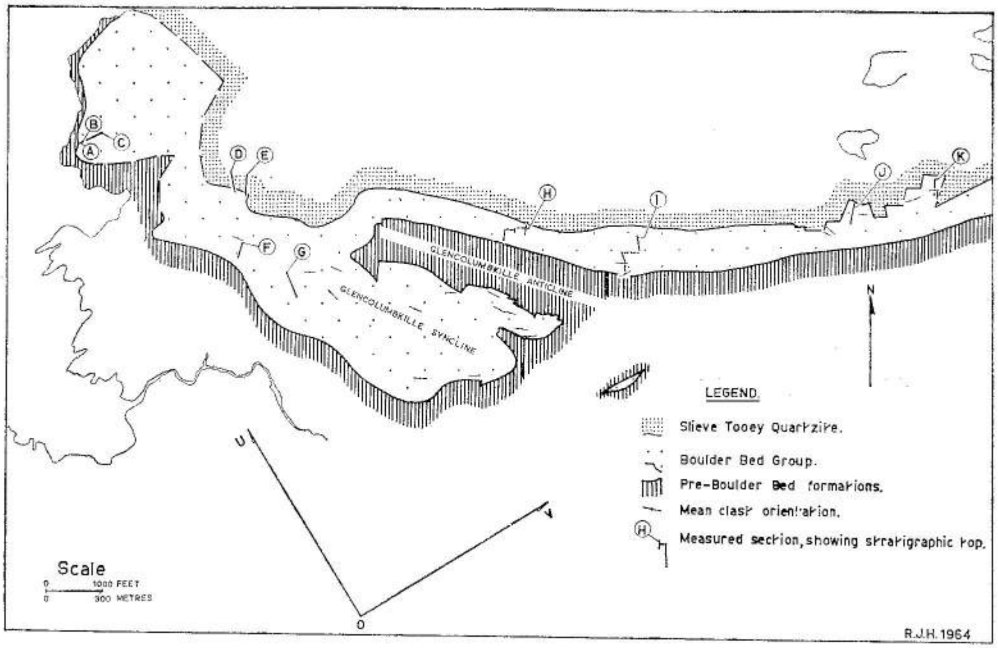


FIG. 2—Sketch map of the Glencolumbkille area showing the ground occupied by the Boulder Bed with locations of the measured sections A–K shown in Fig. 4, and the vector mean orientations of the long axes of inequidimensional clasts in the Boulder Bed. The localities of the base of the measured sections are:— A, 566 yds. S. 13° W. of Glen Tower; B, 566 yds. S. 13° W. of Glen Tower; C, 533 yds. S. 10° W. of Glen Tower; D, 1,166 yds. S. 43° E. of Glen Tower; E, 1,233 yds. S. 45° E. of Glen Tower; F, 1,250 yds. N. 42° W. of northernmost church in Cashel; G (type section), 833 yds. N. 34° W. of northernmost church in Cashel; H, 1,313 yds. N. 38° E. of northernmost church in Cashel; I, 793 yds. N. 7° E. of Glen Lodge; J, 1,813 yds. N. 54° E. of Glen Lodge; K, 2,666 yds. N. 59° E. of Glen Lodge.

The succession appears to be attenuated here, but in the lower part of the cliff at Glen Head all the contacts appear to be conformable. The succession is as follows:

4	Grey quartz-muscovite-biotite schist with alternations (1–15 mm. thick) of fine feldspathic quartzite and semipelitic. Occasional quartzites, calcareous quartzites, and thin dolomitic limestones occur towards the top	41 m.
3	Quartzite	2 m.
2	Skelpoonagh Bay Limestone	8 m.
1	Glencolumbkille Schist (base not seen)	38 m.
		89 m.

Thicknesses in this section are only approximate owing to obscuring by landslips and faulting.

The most accessible exposure of the schists lies 1,030 yds. N. 25° E. from the northernmost church in Cashel, i.e. where the road crosses the Glencolumbkille anticline.

The schists are fine grained (0·05–0·1 mm.) quartz-oligoclase-muscovite-biotite schists showing characteristic alternations of feldspathic quartzite and semipelite normally 1–15 mm. thick (although quartzites up to 8 m. occur rarely) and spaced 5–10 mm. apart. They are normally parallel bedded.

In section, the schists are largely composed of varying proportions of quartz, oligoclase, muscovite, biotite, chlorite after biotite and a little chloritised garnet, calcite, dolomite (confirmed in trace quantities by infrared spectra), zoned blue-green tourmaline, pyrite and graphite. There is, on the whole, much less graphite and more quartz and feldspar than in the Glencolumbkille Schist. Late biotite porphyroblasts are ubiquitous.

The most diagnostic feature of this unit is the prominent striping of the rocks and firm correlation is possible with the same horizon at Kiltyfanned, Maghera and on Slieve League.

2.4. Glencolumbkille Dolomite

The outcrops of this dolomite extend eastwards from Glen Head and the principal ones are located: 1,600 yds. and 1,700 yds. N. 33° W. of the R.C. church, Cashel; 1,030 yds. N. 57° E. of this church; outcrops extending round the nose of the Glencolumbkille syncline (from 1,440 yds. N. 21° E. to 1,400 yds. N. 53° E. of this church) which can be followed as an almost continuous outcrop. Scattered outcrops occur over a fairly large area for 400 yds. north of Glen Lodge on the southern limb of the Glencolumbkille anticline, and 750 yds. north of Glen Lodge on the northern limb. The type locality is 570 yds. S. 14° W. of Glen Tower. The type section (Figs. 2 and 4) is described below and appears to be conformable with the underlying Glen Head Schist.

7 Siliceous dolomite; thinly laminated; white to cream colour, weathering pale brown, with "gritty" surface	0·27 m.
6 Quartzite; pale buff	0·05 m.
5 Siliceous dolomite; creamy colour	0·04 m.
4 Quartzite; buff	0·05 m.
3 Dolomite; thinly laminated, laminae < 0·5 mm. thick, 1–5 mm. apart. Accessory phlogopite and quartz, traces of chlorite, sphene, and magnetite are present. Network of late quartz filled microfractures cause irregular weathering surface. Creamy-white colour, weathering brown	3·00 m.
2 Siliceous dolomite. Psammitic laminae 1–2 mm. apart. Accessory phlogopite, traces of orthoclase and (rare) plagioclase. Pale grey colour, weathering to a brownish crust	2·50 m.
1 Siliceous dolomite; thin psammitic laminae < 1 mm. thick, 1–2 mm. apart. Becomes increasingly psammitic towards the base. Accessory phlogopite with rare inclusions of zircon, traces of sphene and pyrite. Yellowish grey colour, weathering yellowish brown	6·17 m.
	11·98 m.

Units 1 and 2 may be grouped together as the Grey Dolomite member, the rest of the formation comprising the Cream Dolomite member. The former is not well exposed in this area except on the Glen Head cliffs, on the southern limb of the Glencolumbkille Anticline, 1,030 yds. N. 45° W. of Glen Lodge, and 80 yds. north of Glen Lodge. The majority of outcrops are of the upper member. In both the lower and upper members, laminations are frequent, marked by thin pelite, heavy mineral, dolomitic quartzite, or quartzite partings, normally 1–2 mm. thick and 1–20 mm. apart. The matrix frequently contains minor quantities of muscovite and phlogopite and is occasionally semipelitic. Quartzite and pelite intercalations are uncommon.

In section, granoblastic dolomite, frequently twinned, is dominant, with varying quantities of quartz, although dolomitic quartzites are known. The quartz is scattered through the matrix as single crystals, small porphyroblastic clots, or less commonly as laminae less than 5 mm. thick and approximately 1–2 cm. apart. Irregular subrounded patches of carbonate porphyroblasts may also be present. Lepidoblastic folia of phlogopite, muscovite (? talc) and chloritised statically grown biotite are sparse, plagioclase and orthoclase are rare and appear in the more psammatic units. The minor accessories include zircon, apatite, magnetite, pyrite and hematite. In a few instances statically grown tremolite is present as a product of low-grade thermal breakdown of the dolomite associated with sphene, carbonate, chlorite and pyrite. The tremolite occasionally contains carbonate inclusions.

Quartzite intercalations are uncommon. In section, 80–95% granoblastic quartz crystals are clear and unstrained, associated with lepidoblastic folia of muscovite. Late static biotite may be present. The micas frequently show chloritisation. Accessory minerals include orthoclase and plagioclase, frequently showing sericitised cores; carbonate; zircon, frequently as well-rounded or broken (detrital) grains; blue-green prismatic tourmaline, common only in the more pelitic units; magnetite, leucoxenised ilmenite, pyrite, and red-yellow hematite, occasionally after pyrite.

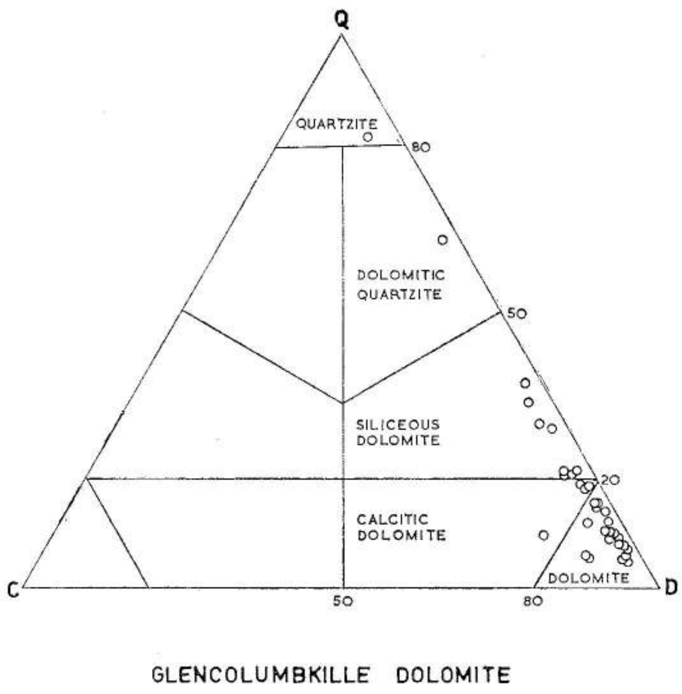


FIG. 3—Composition of the Glencolumbkille Dolomite determined by quantitative X-ray techniques.

Quantitative X-ray determinations of the dolomite compositions are shown in Fig. 3, the majority of the samples falling into the dolomite field. The detection limit using our X-ray technique is 1·3% for calcite and samples near to this limit have been checked on an infrared spectrophotometer. The molecular percentage of $MgCO_3$ in the calcite in equilibrium with the dolomite was also obtained by X-ray analysis. If the data of Graf and Goldsmith (1958) are extrapolated to low temperatures then the minimum equilibrium temperature of 340–365° C. is obtained.

The Glencolumbkille Dolomite may be correlated with units in the same stratigraphic position on Slieve League and at Kiltyfanned and Maghera, and in Scotland and Connemara (Kilburn, Pitcher and Shackleton, 1965).

2.5. Boulder Bed

The name has been retained, although it is a purely descriptive term, because it has become ingrained into the literature describing the group. The Boulder Bed is a group in which units containing clasts ranging in size from single feldspar crystals up to large boulders, randomly dispersed throughout the matrix, alternate with layers of semipelite and quartzite.

In the Kiltyfanned area it is clear that in the Boulder Bed there is a trend of increasing thickness of quartzite intercalations with increasing stratigraphic height ending with the blanket quartzite association of the Slive Tooey Quartzite. In contrast, the boulder beds show random thickness fluctuations. The Boulder Bed matrix shows a sympathetic change from dolomitic quartz wacke → quartz wacke → quartzite-feldspatic quartzite → Slieve Tooey Quartzite. There is a similar sympathetic change in clast content with dominantly dolomitic clasts at the base which, followed by the incoming of much thicker quartzite intercalations accompanied by the change to a more siliceous matrix at the top of the Boulder Bed, allows division into three units which may be recognized in most of the localities (Kilburn, Pitcher, and Shackleton 1965, p. 348). In Glencolumbkille, this threefold division is not so easily applicable, and only two subunits, the Lower (dolomitic) and Upper Boulder Bed, are recognisable. Quartzite intercalations occur throughout the whole of the sequence and impersistent semipelites and pelites occur in the Upper Boulder Bed. The lower unit is best exposed in the cliff below Glen Tower but more convenient exposures extend from 230 yds. N. 122° S. of Skelpoonagh Bay to 770 yds. N. 40° W. of the northernmost of the two churches in Cashel, reappearing 600 yds. N. 71° E. of the same church; and on the southerly limb of the anticline 1,000 yds. N. 56° E. of the above church. On the northerly limb the only outcrops occur 815 yds. N. 5° E. of Glen Lodge. The centre of the Glencolumbkille syncline is entirely occupied by the upper part of the Boulder Bed. Excellent exposures may be examined on the lower slopes of Craigbeefen and for 1,340 yds. south-east of the crag.

A small outlier of upper Lower Boulder Bed occurs 65 yds. north of Glen Lodge. The northern junction with the Glencolumbkille Dolomite is faulted while the southern boundary is unexposed but the topography suggests the presence of a fault. This Boulder Bed probably originally overlay the Glencolumbkille Dolomite in an easterly extension of the Glencolumbkille syncline, now eroded away and the Boulder Bed is preserved in a downthrown fault slice.

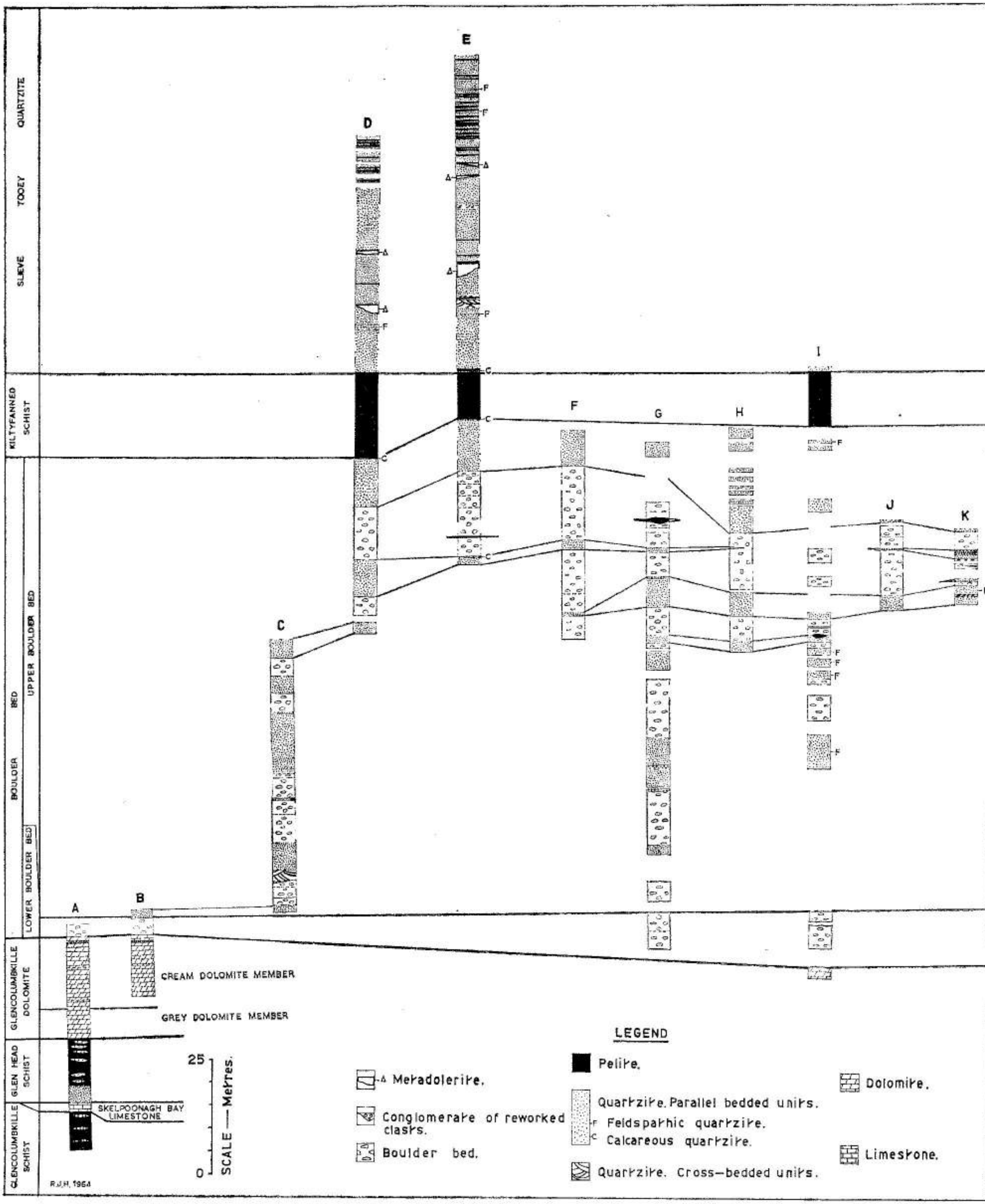


FIG. 4—Stratigraphic sections of the Boulder Bed hung from the top of the overlying Kiltyfanned Schist. Locations of the sections are shown in Fig. 2.

The type succession (Figs. 2 and 4) is as follows :—

UNIT	DESCRIPTION	THICKNESS (metres)
27	White quartzite	3·0
	UNEXPOSED	10·0
26	Boulder Bed : Pale grey psammitic matrix containing subrounded to rounded pebbles and cobbles of white granite and rare pink granite. Massive. Weathers buff. Distinct planar contact with unit 25	2·0
25	Boulder bed : Greenish-grey psammitic matrix containing subangular to subrounded pebbles and cobbles of white and red granite. Bedding planes (10–32 cm. apart) show very sharp to sharp planar contacts. Euhedral pyrite in the lowest 10 cm. Weathers to a darker grey surface	1·2
24	Semipelite with psammitic layers showing white, alternating with grey, semipelite bands 1–2 mm. thick. Some bands may contain carbonate. Euhedral pyrite in top 30 cm. Massive. Sharp planar contact with unit 25	1·6
23	Boulder bed : Grey psammitic matrix with subangular to subrounded pebbles, cobbles, and (rare) boulders of white and pink granite (17%), quartzite (46%), feldspar (33%) and dark schist (4%) granules and pebbles. Euhedral pyrite in top 20 cm. Sharp to distinct planar bedding plane contacts (10–40 cm. apart). Sharp planar contact with unit 22	0·8
22	Semipelite with white psammitic layers alternating with grey semipelite bands 1–2 mm. thick	0·4
	UNEXPOSED	1·0
21	Boulder bed : pale grey psammitic matrix with rare subrounded granules and pebbles of white granite, pink granite may also occur as cobbles	3·3
20	White quartzite. Distinct planar top and bottom contacts.	0·7
19	Boulder bed : greenish-grey psammitic matrix with angular to subrounded cobbles and boulders of white and pink granite. One angular boulder (35 × 22 × 20 cm.) showed a crack in the top surface c. 2 cm. wide and 12 cm. deep filled with matrix during sedimentation. Several distinct bedding planes (20–160 cm. apart). Bottom contact gradational over 0·5–1 cm.	5·0

18	White quartzite containing subrounded rare pebbles and granules of white granite. Distinct planar contact with unit 17	0·6
17	Feldspathic quartzite; white; sharp to distinct planar bedding contacts (20–180 cm. apart)	5·5
16	Boulder bed : Pale grey psammitic matrix with rare subrounded to rounded granules of white granite and boulders of pink granite. Distinct planar bedding plane contacts (30–120 cm. apart)	6·0
15	Boulder bed : Grey psammitic matrix with subrounded white granite pebbles and cobbles (7%); quartzite (84%), and feldspar (9%) granules and pebbles. Bedding plane contacts planar and distinct (20–180 cm. apart). Distinct planar contact with unit 14	3·4
14	Quartzite : white, with distinct planar bedding plane contacts (20–60 cm. apart). Sharp irregular contact with unit 13	1·6
13	Boulder bed : Grey psammitic matrix with pebbles and boulders of white granite and quartzite. Small diffuse quartzite intercalations near top of unit (suggest some erosion and winnowing prior to deposition of unit 14). Massive	1·5
	UNEXPOSED	0·5
12	White quartzite. Sharp planar bedding plane contacts (10–90 cm. apart)	4·0
	UNEXPOSED	2·0
11	Boulder bed : Grey-green psammitic matrix with abundant subrounded white and pink granite pebbles and cobbles (17%), quartzite (57%), feldspar (21%), and dark schist (5%) pebbles. Massive; sharp planar contact with unit 10	12·5
10	Feldspathic quartzite : white, weathering buff. Distinct planar bedding plane contacts (100–180 m. apart). Base is transitional (1 cm.) into unit 9	5·3
9	Pebbly feldspathic quartzite : white with white granite pebbles scattered sparsely through matrix. (Probably intraformational conglomerate of derived winnowed-out clasts.) Transitional contact (> 1 cm.) with unit 8	0·8
8	Feldspathic quartzite : white with bedding plane contacts sharp (1–5 cm. apart)	3·0
	UNEXPOSED	0·1
7	Feldspathic quartzite (as above). Basal 6 cm. transitional into underlying boulder bed (unit 6) and contains subrounded clasts of white granite (3 cm. max. diam.) in quartzite matrix. (Probably result of winnowing of boulder bed matrix.)	0·3

6	Boulder bed : grey, becoming lighter towards top. Contains white and pink granite pebbles and cobbles (10%), quartzite pebbles (77%), and feldspar pebbles and granules (13%). Distinct planar bedding planes (50–400 cm. apart). Basal 15 cm. contains pyrite and abundant pebbles of white granite and vein quartz. Contact with underlying quartzite irregular	11·7
5	Feldspathic quartzite : weathering buff. Large quantities of hematite and limonite and segregations of euhedral pyrite in the top 5 cm. Rare granules and pebbles of white vein quartz. Sharp planar lower contact	0·1
4	Feldspathic quartzite : white to cream. Bedding plane contacts flat and sharp (2–30 cm. apart). Thicker beds show parallel lamination in which the coarse grained laminae equal the fine in number. One graded bed 12 mm. thick with a coarser base of quartzite granules observed	2·2
	UNEXPOSED	6·0
3	Boulder bed : pale greenish-grey psammitic matrix with subrounded granules and pebbles of white and pink granite (8%), quartzite (85%), and feldspar (7%). Distinct planar bedding plane contacts	4·2
	UNEXPOSED	2·0
2	Boulder bed : Pale grey to dark grey dolomitic-semipelite matrix (in parts) with subrounded pebbles and cobbles of cream dolomite (65%), quartzite (34%) and white granite (1%). The dolomite clasts may become deficient in some places towards the base of the unit. The bedding plane contacts are planar (90–100 cm. apart), the basal contact with unit 1 is transitional (1 cm.)	3·9
1	Boulder bed : cream coloured dolomitic-semipelitic matrix, containing cream dolomite (96%), white granite (1%) and quartzite (3%) pebbles and granules. The matrix forms a bedded complex with coarse and fine layers, approximately equal, 5–20 cm. thick, with semipelitic and psammitic intercalations	4·0
	Total thickness	112·2

The Lower Boulder Bed is characterised by a dolomitic matrix and by the presence of cream dolomite clasts, similar in composition, as determined by quantitative X-ray determinations, to the Glencolumbkille Dolomite. The proportions of dolomite, quartz, feldspar and granite clasts, as shown in Fig 5A, fall within the field of more than 30% dolomite and less than 10% of

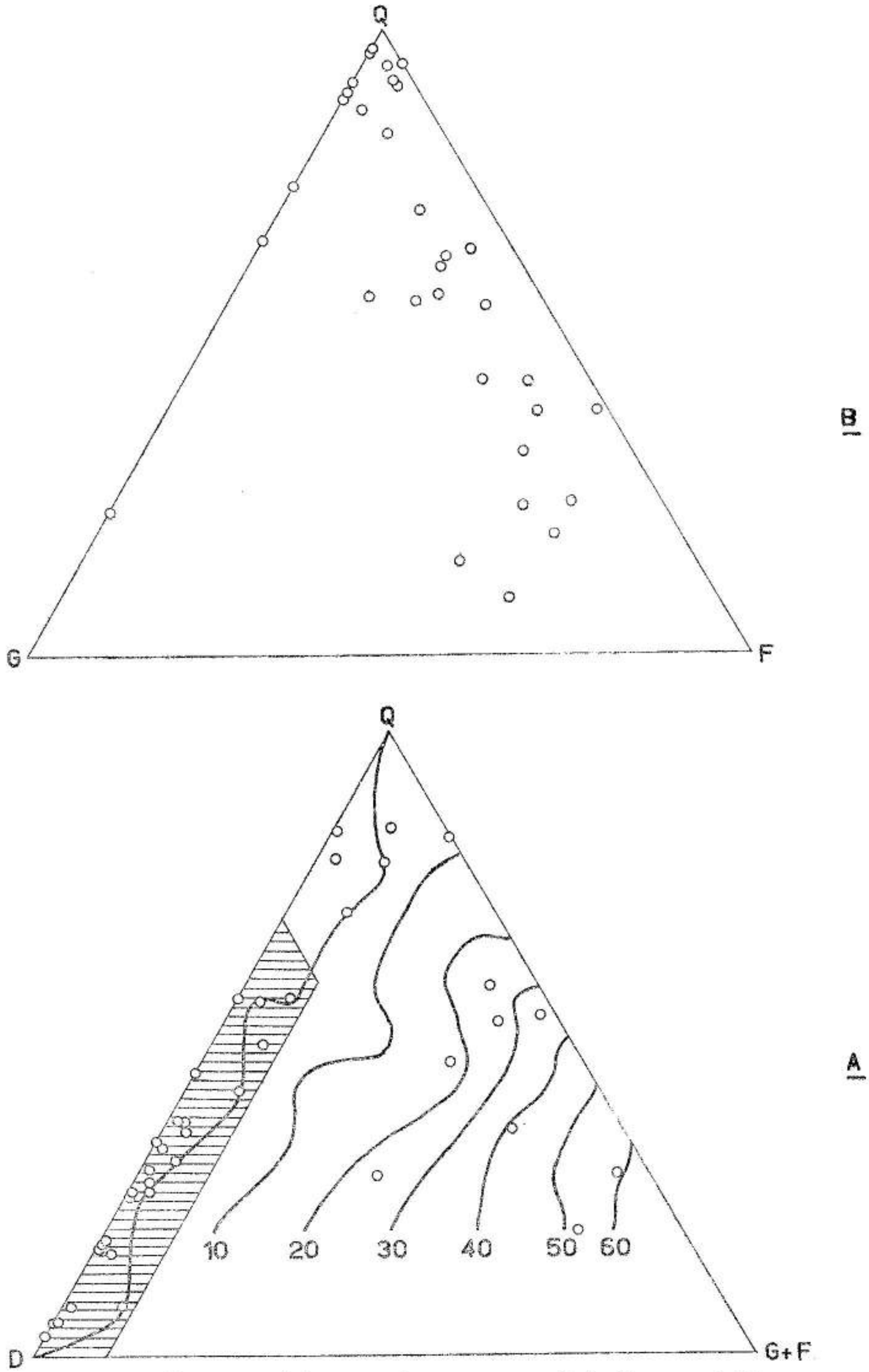


FIG. 5—Clast composition at various exposures of the Lower and Upper Boulder Bed. A—Quartz + quartzite (Q), dolomite (D) and granite + feldspar (G + F) clasts as a percentage. Minor quantities of other clasts are neglected. Contours are percentage of feldspar clasts. All plots within the hatched area are from the Lower Boulder Bed; the other plots are from the Upper Boulder Bed. B—Upper Boulder Bed clasts on the basis of quartz + quartzite (Q), granite (G), and feldspar (F). Minor quantities of other clasts neglected.

granite plus feldspar, the feldspar alone being normally less than 5%. In outcrop the dolomite clasts are conspicuous, weathering with a brown crust or weathering out completely leaving the rock face pitted with holes.. The smaller clasts are variably flattened and are often surrounded by thin layers of biotite. More pelitic or psammitic laminations stand out in the brownish weathered matrix and rarely parallel bedded quartzite intercalations (0·5–2 m.) occur.

The matrix of the Lower Boulder Bed is lighter than that of the Upper Boulder Bed and is a quartz-dolomite (20–60%)-muscovite-biotite schist in which calcite rarely exceeds 2%. Lepidoblastic folia contain muscovite, phlogopite and partly chloritised dark green to brown biotite, often with zircon inclusions. Patches of partly sericitised orthoclase and albite-oligoclase are sometimes common, both angular and rounded zircons occur and zoned tourmaline, apatite, magnetite, pyrite, hematite and limonite are accessory minerals. The association quartz-epidote-sphene-calcite with statically grown actinolite fibres exists in one thin section.

The top of the Lower Boulder Bed is defined by the change from a dolomitic matrix to a semipelitic matrix and by a change in the clast composition to less than 30% dolomite and more than 10% granite plus feldspar (Fig. 5). A marker for the upper part of the Lower Boulder Bed appears to be the occurrence of rare boulders of yellow vein quartz conglomerate at approximately this level at Kiltyfanned, Glencolumbkille and Malin Beg, three miles south-west of Glencolumbkille.

The Upper Boulder Bed forms the greater part of the group and has a quartz-muscovite-partly chloritised biotite-microcline-albite oligloclase matrix. Muscovite folia are set in a granular or schistose quartz matrix. Late porphyroblasts of green biotite are ubiquitous and some microcline-quartz and microcline-plagioclase intergrowths are present but feldspar is rare in the more pelitic layers in which blue-green tourmaline is entirely concentrated, suggesting original enrichment of boron in these layers. Carbonate, epidote, zircon, apatite, variably leucoxenised ilmenite, pyrite and hematite are accessories. The assemblage poikiloblastic hornblende with biotite, quartz and euhedral epidote occurs in one thin section. The clast composition in the Upper Boulder Bed is shown in Fig. 5B.

The Upper Boulder Bed may show a general transition from greenish semipelites at the base into grey, psammitic and more feldspathic beds towards the top but this is not evident everywhere, being more pronounced at Kiltyfanned than in Glencolumbkille.

Within individual beds it is possible to discern bedding planes often coinciding with abrupt changes in clast composition or associated with thin (1–2 cm.) quartzite layers. This suggests that diastemic breaks in the sedimentation took place together with slight reworking of the topmost sediment. Contacts between individual boulder beds and their intercalated quartzites are normally sharp and not transitional and erosion at the base of a boulder bed is uncommon though occasional scouring of the underlying quartzite has given local transitional junctions with clasts in a quartzite matrix. More commonly erosion of the top surface yielded pockets of winnowed-out clasts lying on the top of the boulder bed. Intraformational conglomerates may occur intercalated in quartzite, e.g. Fig. 4, Section G, unit 9 and in individual boulder beds, e.g. 1,660 yds. N. 52° E. of Glen Lodge where a lens (175 × 75 × 85+ cm.) of subangular to subrounded pebbles of granite, gneiss, vein quartz, schist and heavy minerals are concentrated.

Normally the boulder beds are completely unsorted. Rarely, where reworking of the top few cm. of a bed has taken place crude grading may occur and micro-grading may be present at the top contact of intraformational conglomerates. There is no evidence of a turbidite sequence. Clasts do not usually penetrate the bedded layers and lode structures are rare.

Pelitic layers associated with the boulder beds are quartz-muscovite-biotite schists with chlorite after biotite and accessory tourmaline, carbonate, zircon, sphene, apatite, oligoclase, epidote, pyrite, magnetite, leucoxenised ilmenite and hematite.

Quartzites in the boulder beds usually consist of 90–95% quartz with minor muscovite, sericitised orthoclase and albite-oligoclase, carbonate and biotite while magnetite, leucoxenised ilmenite, zircon and hematite are accessories. The quartzites are normally parallel bedded (2–180 cm.) with sharp planar contacts. Cross-stratification is rare, graded bedding extremely rare and never exceeding 5 cm. in thickness. The cross-strata have sets of the planar type (McKee and Weir, 1953) and are small scale solitary or grouped, non-erosional or gradational with a planar lower surface and are of homogeneous composition (terminology after Allen, 1963). A few sets of cross-strata have been faulted prior to lithification of the sediment. Current ripples, distinguished from tectonic ripples by the criteria outlined by Spry (1963), are occasionally present.

The clasts vary from a lower limit of about 3 mm., below which extensive recrystallisation of the matrix prohibits unequivocal identification, up to boulders commonly 30–50 cm. in diameter and exceptionally more than a metre. If the granite has a mean density of 2·63 grm. per c.c. then the largest boulders must weigh 2–3 tons. In general, the smaller clasts are often quartzite and feldspar at all levels in the Boulder Bed.

It is usually difficult to extract whole clasts and only the cross-sections of the clasts can be readily measured. These show a roundness range of 0·3–0·9 and a sphericity range of 0·3–0·7 on the visual chart of Krumbein and Sloss (1963, p. 111) if extremely distorted clasts are excluded. Eighty pebbles and cobbles were extracted from one boulder bed, measured and plotted on a Zingg diagram (Fig. 6). This shows a moderate scatter, probably partly due to slight deformation, and suggests a minimum sphericity of 0·5 (Krumbein and Sloss, 1963, p. 107).

The clast compositions are shown in Fig. 5 and they can be divided into extrabasinal igneous and metamorphic fragments and intrabasinal sedimentary rocks. The former includes vein quartz, quartzite, quartz conglomerate, gneiss and granite clasts, the latter the dolomites, apparently derived from the Glencolumbkille Dolomite, and perhaps quartzite. Only three rock types are really common as clasts, quartzite, granite and dolomite.

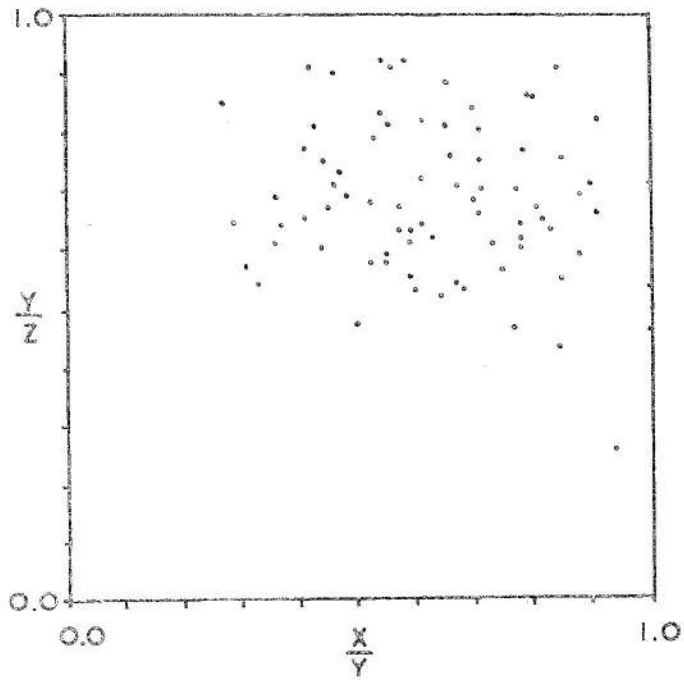


FIG. 6—Zingg diagram of clast sample. Three axes $X > Y > Z$ cm. measured. Location: six inch sheet 81 Co. Donegal, 1,266 yds. N. 75° W. of Lougheraherk School.

The granites vary in colour from white or pale yellow to pink and red. Their textures and mineralogy lack distinctive traits. Quartz, only rarely showing graphic intergrowth, albite-oligoclase, muscovite, clots of partly chloritised biotite, often with zircon inclusions, orthoclase, frequently perthitic, and microcline. The feldspars are patchily sericitised. Carbonate sometimes fills late cracks. Accessory minerals are zircon, apatite, blue-green and yellow tourmaline, rare rutile, magnetite, leucoxenised ilmenite and pyrite often partly hematised. The quartzites are white and composed of 85–97% granoblastic quartz with muscovite and biotite. Calcareous quartzites (10–15% carbonate) are uncommon but porphyroblastic (introduced?) carbonate is a minor accessory of a few clasts together with zircon, apatite, magnetite, and hematite. The dolomites are largely granoblastic dolomite with a little quartz, muscovite (talc?), biotite, tourmaline with an earlier fabric shown by carbonate inclusions, pyrite, rare magnetite and hematite.

The deformation of the clasts depends both upon their size and their composition, the smaller clasts of dolomite being the most deformed. The style of deformation is generally the flattening type of Flinn (1962) and the mixed clast types in the Lower Boulder Bed (Fig. 7) have a mean axial ratio of 16 for dolomite, 3 for quartzite and 1 for granite at one locality. This relative deformation appears to be similar in all localities.

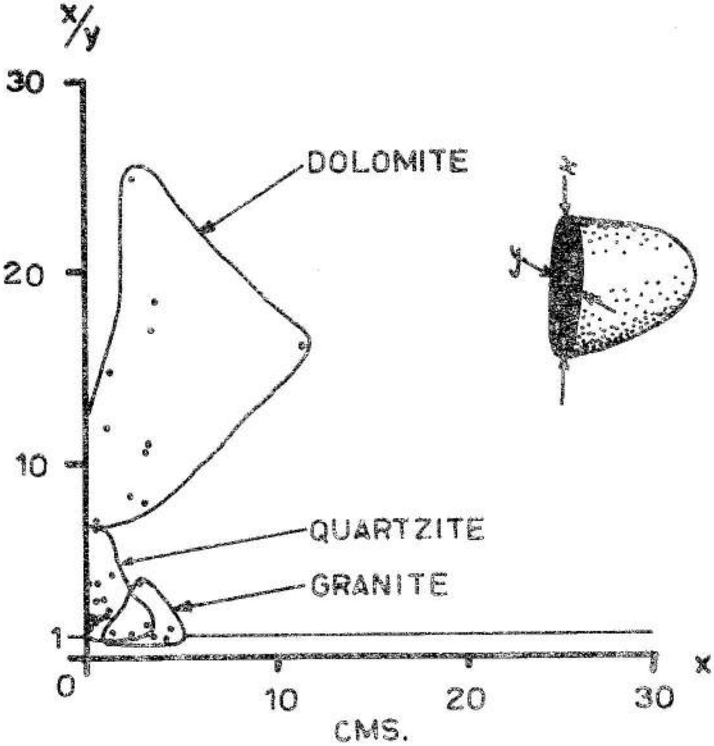


FIG. 7—Deformation of Lower Boulder Bed clasts. Plot of X/Y versus X cm. for clasts measured perpendicular to maximum flattening. Location: six inch sheet 81 Co. Donegal, 1,400 yds. N. 76° W. of Lougheraherk School.

It has been suggested by Shackleton (1961) that a till fabric is recognisable in the Boulder Bed at Fanad, County Donegal and the orientations of clasts in the Glencolumbkille area were measured to investigate this. At some localities only clasts smaller than 2 cm. maximum diameter were noticeably sheared out. The orientations of the deformed and "apparently undeformed" inequidimensional clasts were measured separately and plotted (Fig. 8, A_{1-2}, B_{1-2}), At both localities there was no statistical difference between the vector mean orientations of the two groups, and in all the diagrams (Fig. 8, A_1, B_1, C and D_1) the means are within 15° of the F_2 schistosity (p. 145). At locality D the orientation of clasts in two mutually perpendicular faces, at right angles to the bedding plane, as shown by a thin quartzite, could be measured. The clasts lie sub-parallel to the bedding suggesting an original fabric in this orientation especially as there was no apparent deformation of small clasts at this locality. A two-dimensional moving average of clast orientations for the area (Fig. 8) shows a consistent pattern over the whole region, aligned parallel with the axial planes of the Glencolumbkille anticline and syncline. It is concluded that the fabric of the clasts has been aligned by the F_2 movements but remains subparallel to the bedding in some outcrops. A till fabric cannot be recognised.

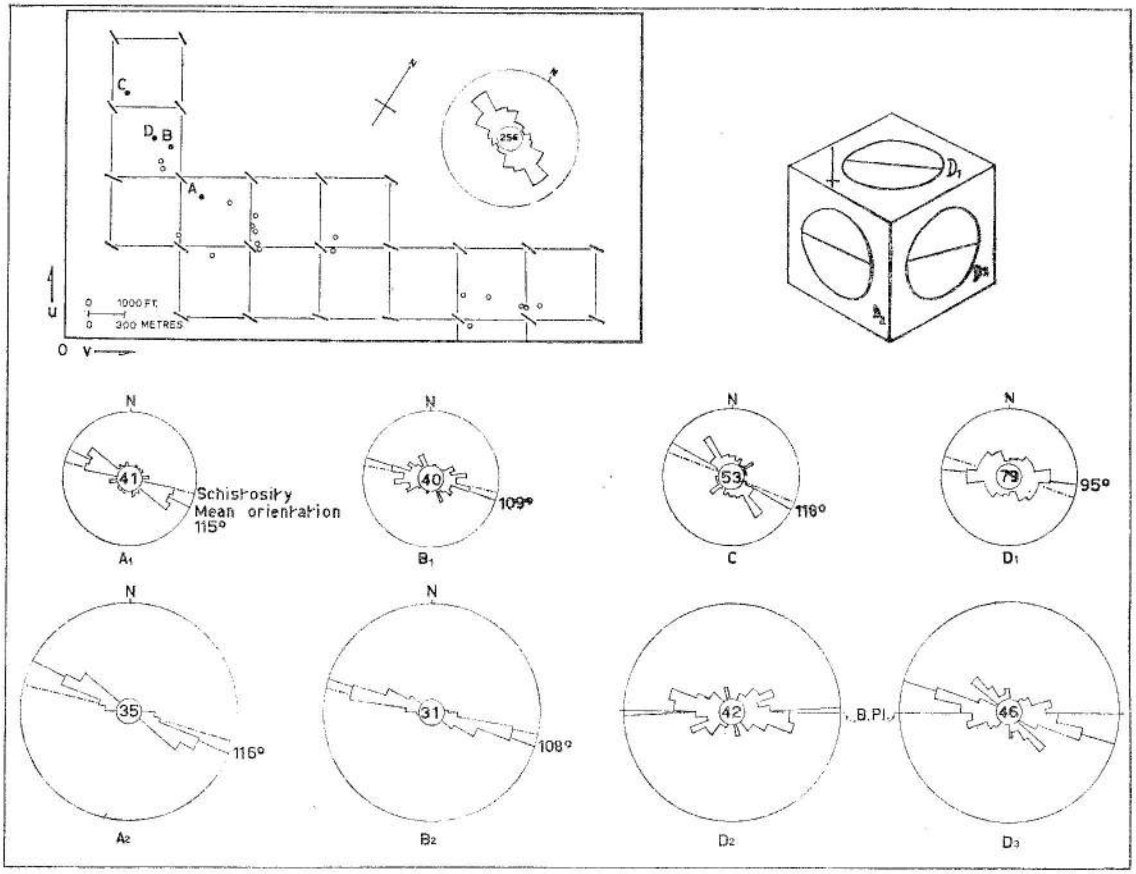


FIG. 8—Orientation of clasts in Boulder Bed at four localities, *A*, *B*, *C*, and *D*. A_1, B_1, *C* and D_1 are rose diagrams of "apparently unsheared" clast elongations. A_2, B_2 are "sheared" clasts. D_2 and D_3 are "unsheared" clasts measured on two faces perpendicular to the bedding plane. Class interval in all diagrams except D_1 is 10°, class interval is 20° in D_1. Figure at centre of rose diagram is the number of observations.

Map at top left shows the regional two-dimensional moving average of clast orientation directions and regional rose diagram, class interval 20°. For origin of *U*, *V* coordinates see Fig. 2.

2.6. Kiltyfanned Schist

Kiltyfanned, four miles north-east of Glencolumbkille, contains the type section of this schist, 66 yds. north of the east end of Kiltyfanned Lough. Details of this section will be described elsewhere. In Glencolumbkille, the exposures lie on the north side of the valley, on the crags of Craigbeefan and for 2,000 yds. east of a point 900 yds. north of Glen Lodge. There is also a small exposure, probably of this formation, 566 yds. N. 26° E. of the northernmost church in Cashel.

The schist is dark green to black with quartz segregations up to 4 cm. long. Sometimes it is so compact that it is difficult to distinguish from fine grained metadolerite without recourse to thin sections. The base of the formation is frequently associated with more calcareous dark green schist.

In thin section lepidoblastic folia of heavily chloritised biotite, muscovite and magnetite are interposed between quartz-epidote-carbonate layers which may show a schistose or granular texture. Carbonate crystals with exsolved iron along twin and cleavage planes suggest an original ferroan composition. Epidote, sphene, zircon, apatite. oligoclase, tourmaline, magnetite, leucoxenised ilmenite and hematite are accessory minerals.

The unit follows the top of the highest quartzite in the Boulder Bed Group and below the uniform Slieve Tooey Quartzite. The base of the schist seems to be a marker horizon for the top of the Boulder Bed, a conclusion in agreement with the stratigraphical sequence of the Boulder Bed proposed by Kilburn, Pitcher and Shackleton (1965).

2.7. Slieve Tooey Quartzite

Anderson (1954) named this quartzite after Slieve Tooey mountain six miles north-east of Glencolumbkille. The reference section in the present area is Section E. of Figs. 2 and 4.

The quartzite has parallel bedding, commonly 0·5 to 42 cm. (mode ca. 2·5 cm.) which is usually distinct and non-gradational, marked by darker layers between white or light grey flags. Most beds are well-sorted but pebbly layers and graded beds of quartz or feldspar granules, 1–5 cm. thick, are not uncommon. Cross-stratification is of the planar type (McKee & Weir, 1953) with most commonly single sets which range up to about 2 m. and are homogeneous in lithology. The lower bounding surfaces are non-erosional or gradational planar or possibly trough shaped, and concordant. Occasional cross-strata are graded. Current ripples are present. The regional transport pattern is consistent and dominantly towards the north.

Calcareous quartzites are uncommon but feldspathic quartzites (20–30% feldspar) are prevalent. Partings of semipelite and muscovite rich layers (up to 10 cm.) are customary and in thin section 93–95% quartz is typical of the non-feldspathic quartzites. Muscovite is always present while sericitised plagioclase, microcline, apatite, zircon, leucoxenised ilmenite and magnetite are less common minerals.

A strong quartz rodding is almost ubiquitous while some tectonic ripples occur parallel to this lineation.

This unit can be correlated over a wide area of Donegal, Connemara and Scotland.

2.8. Sedimentation

Following the deposition of black shales and some limestones came a dolomite and then a thick series of boulder beds whose matrix became progressively more siliceous until, after a thin calcareous pelite, a thick quartose sandstone was deposited. The Boulder Bed was originally ascribed (M'Henry, 1891) to ice rafting during an ancient ice age. Kilburn, Pitcher and Shackleton (1965) suggest it as a subaqueous till and the present work supports a glacial origin but this will be discussed in detail elsewhere when all the evidence from Donegal has been assembled It seems that at first ice action tore off pieces of the dolomite from the underlying sediment and deposited them elsewhere in the earliest boulder beds. Major problems include the origin of the extrabasinal clasts and also whether the sedimentation was merely interrupted in a progression to quartzose sandstone or whether the Slieve Tooey Quartzite is dominantly a much worked sand produced by extensive erosion of a granitic and metamorphic land area which may have been the source of the extrabasinal clasts. It appears less likely to have been produced by winnowing of extensive boulder bed deposits elsewhere. If the tectono-environmental classification of Krumbein and Sloss (1963, p. 510) is followed the sequence may be interpreted as a transition from a deep water quiet offshore "circalittoral" environment into an infralittoral stable or slightly unstable shelf environment.

3.1. Igneous Intrusions

Metadolerites

The metadolerites generally form sills and are typically made of about 60% green hornblende, 25% partly zoisitised plagioclase, 5% sphene, 5% quartz, 3% ilmenite and 2% biotite. They are generally massive, non-schistose and unfoliated rocks, except at their edges where a few centimetres or a few metres of schistose amphibolite may be found, e.g. in a quarry 270 yards N.E. of Glen Lodge. The granoblastic texture, formed by stumpy hornblendes (1–2 mm. av. diameter) and plagioclase is disturbed by later re-crystallisation of needles and prisms of hornblende which randomly cross the earlier hornblende and penetrate and replace the plagioclase. The stumpy hornblende presumably crystallised during the peak of the metamorphism while the latter needles belong to a period of late re-heating. The plagioclase contains many small, but distinctly formed β zoisite and clinozoisite prisms, randomly arranged, which have replaced the feldspar during this late re-heating, leaving a relatively sodic plagioclase whose composition ranges from albite to middle oligoclase. This late replacement of what must have been labradorite or calcic andesine probably took place during the crystallisation of the amphibole needles and was accompanied by a little late biotite and sphene. Sphene usually forms rims around, and within, the skeletal ilmenite, has variably replaced the ilmenite and is itself generally partially altered to leucoxene. Small (1 mm. diameter) garnets are less common while accessory euhedral apatite needles, calcite, often in porphyroblasts but occasionally in veins, and pyrite with hematite replacement rims are ubiquitous.

In a few small outcrops of schistose amphibolite, 340 yds. and 600 yds. north of the road junction, which is 500 yds. east of Glen Lodge and also at a point 560 yds. north of Spot Height 152, east of Glen Lodge, there are unusually large poikiloblasts of garnet, up to 1 cm. in average diameter set in about 55% prismatic blue-green hornblende, 10% quartz, 15% acid plagioclase, 10% biotite and 5% ilmenite. Carbonates may be present while some specimens are quartz rich. These garnets are completely filled with minute (0·03 mm.) quartz and ilmenite inclusions, such that the guest crystals almost equal the volume of host. Occasionally carbonate inclusions exist. The textures show that the garnets grew statically after the S_2 foliation and schistosity. Their euhedral shape and the existence of a clear, inclusion-free rim in some of the garnets, suggests that they may have enlarged themselves during the late period of re-heating.

On Craigbeefan, above the Kiltyfanned Schist several thin (0·3–2 m.) metadolerites have been appreciably replaced by biotite, such that an assemblage consisting of 35% very dark brown biotite, 30% albite-oligoclase with pistacite (not zoisite), 20% hornblende, 15% quartz, a few carbonate porphyroblasts and sphene, has been formed. The biotite has clearly replaced hornblende and has grown during the late thermal metamorphism. Such rocks are not easily distinguished in the field from the nearby Kiltyfanned Schist and thin sections have been used to confirm the field identifications. While a little late biotite is common in the metadolerites it is only on Craigbeefan that such extensive replacement occurs.

It is probable that all the main outcrops of metadolerite belong to one discordant intrusion which has been broken into now separated pieces. This intrusion can be traced southeastwards from Craigbeefan, where it is at the top of the Boulder Bed, through the Boulder Bed down to the Glencolumbkille Dolomite, 750 yds. N.W. of Glen Lodge where it is squeezed out, only to reappear, much thicker and within the Glencolumbkille Dolomite immediately N.W. of Glen Lodge. Here, it contains a thin hornfelsed felspathic quartzite inclusion.* The metadolerite which occurs ½ mile N.E. of Glen Bay and which is even further down the succession, being within or just below the Glencolumbkille Dolomite, is probably the same intrusion, having trangressed to lower levels. This also includes a thin quartzite band. This may be the same metadolerite as intruded the top of the Glen Head Schist S.E. of Glen Head but this correlation is less sure.

Thin sections of several series of specimens collected across the thickness of this principal metadolerite suggest that garnet only appears in the upper part of the intrusion. This indicates that a little differentiation may have taken

* This might be a slightly sheared granite vein.

place in the original dolerite and that the top may be a little more iron-rich, thus favouring the crystallisation of some garnet at the expense of the hornblende, which is less ferrophile.

3.2. Dykes

Late quartz porphyry and biotite-hornblende lamprophyre dykes have been mapped. Although both these sets are displaced by the north-easterly striking faults their precise age is unknown as the number of movement periods on these faults is unknown.

A thin garnetiferous micro-granite vein crosses the metadolerite by the *n* of Craigbeefan on 6 in. sheet Donegal 80. This interesting rock is almost entirely made of quartz, microcline and albite-oligoclase with epidote and clinozoisite. A little biotite, chlorite and euhedral garnet are also present. The edge of the metadolerite against the vein has been altered; the hornblende replaced by biotite or epidote and chlorite and the plagioclase saussuritised. The presence of distinct epidote and clinozoisite crystals in the plagioclase of the vein suggests that the vein pre-dates the late metamorphism.

4. History of Deformation

The structure of the Glencolumbkille area has been produced by a complex sequence of deformation which is tentatively divided into four important phases and four late, brittle subordinate phases. Of these the second was the strongest, and only the second, third and fourth produced any major structures. The first four phases were the result of a strong north-south compression, whilst the last two were the result of a weaker compression of uncertain direction. Although the deformational history is complex it is found that the main structure of the Glencolumbkille area is relatively simple. It must be emphasised that the structural and metamorphic sequences suggested are for a relatively small area, and as such will only serve as a guide to more extensive work in the Slieve League peninsula.

A distinct schistosity, either at a low angle to, or coplanar with, the bedding, is the most obvious structure formed by the first deformation F_1. Although it commonly appears to be a bedding schistosity it has also been seen in the axial planes of first folds, e.g. S.W. of Skelpoonagh Bay. Accentuation of this schistosity which, on the evidence of relict internal foliations was originally phyllitic, has been due to later mimetic crystallisation. Metamorphic segregation of quartz occurred with this deformation and produced veins within or discordant to the schistosity. Apart from the schistosity few records of the first deformation remain in the Glencolumbkille area, and no major structures or large scale stratigraphic inversions can be related to it. Minor folding of the earliest phase is rare and only observed in distinctively banded lithologies, e.g. banded schists. It is isoclinal with angular hinges and axial planes whose orientation is solely dependant on the effects of later deformations.

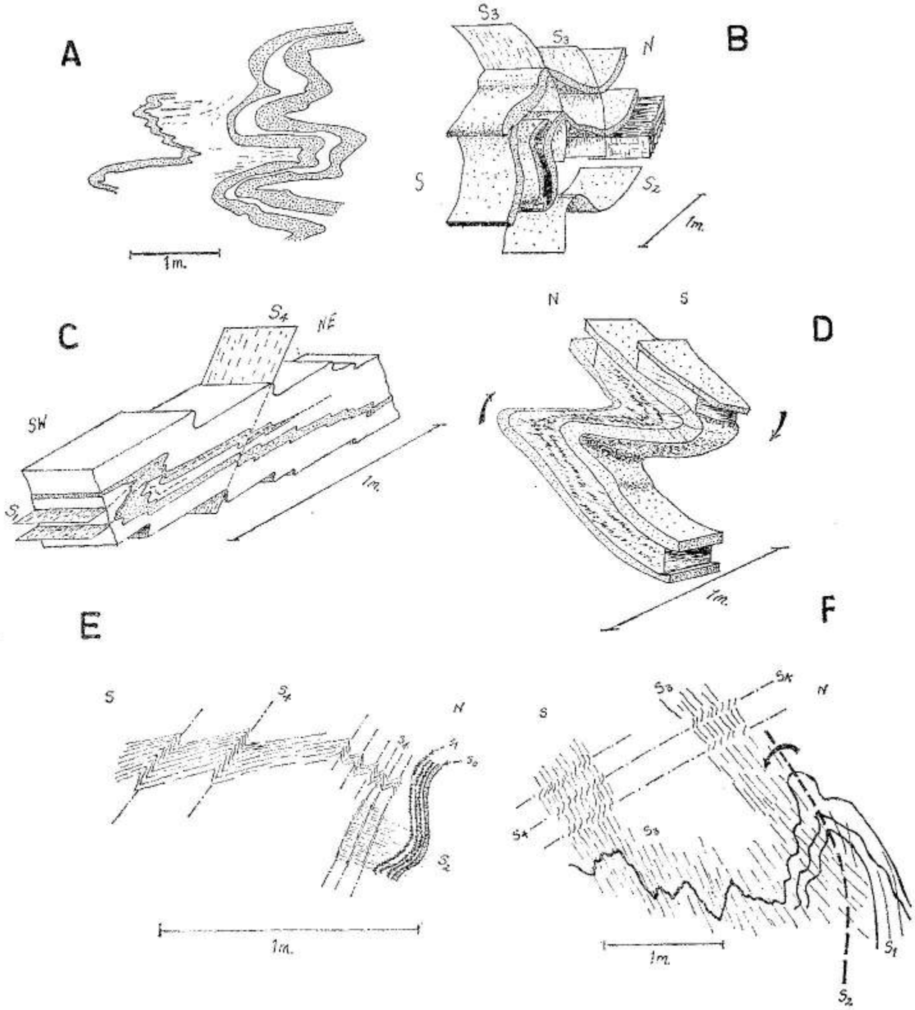


FIG. 9 A—F_2 minor folds in banded psammites at the head of Skelpoonagh Bay.

B—Diagrammatic representation of quartz vein in schist 800 yds. north of Glen Lodge. S_1 schistosity folded by F_2 and F_3.

C—Diagrammatic representation of F_4 and F_1 fold styles in the ground north of Glen Bay.

D—F_4 fold in banded psammites with strong feature cleavage fans in fold hinges: sense of rotation is the reverse of that normal in the area at the head of Skelpoonagh Bay.

E—Relationship of S_0, S_1, S_2, and S_4 diagrammatically. Semipelites at the bridge at Gaveross.

F—Relationship of S_1, S_2, S_3 and S_4 in the central part of the Glencolumbkille Anticline; c.f. with 9B where S_2 is almost horizontal. 550 yds. north of Glen Lodge.

The second phase of deformation (F_2) was intense. It produced a pronounced widespread schistosity or strain-slip cleavage, with tight similar folds. It also compressed pre-existing early folds, flattened the dolomite clasts of the Boulder Bed and folded the quartz veins produced during the previous deformation. Inception of the Glencolumbkille anticline may be associated with this phase. The second schistosity varied with the lithology of the rocks; in pelites and semi-pelites a true schistosity was formed, while in psammites a strain-slip cleavage with partial reorientation of the micas was produced. The width of the strain-slip lamellae was less than 5 mm. Although the second schistosity often became dominant the first was partially preserved especially as traverse micas in S_2 strain-slip bands and in S_2 folia and quartz augen. Second phase minor folds are either curvilinear or angular (Fig. 9A). They are tight and in pelites can be isoclinal while their axes are widely divergent in trend, a scattering which could be partly original but is mainly due to the effects of later movements; their axial plane schistosity generally dips from 30° to 50° northwards. Quartz veins produced during the previous deformation were folded and various trends of axes have resulted from the folding of the originally discordant veins. An intersection lineation and commonly a mineral lineation, especially in the quartzites, were developed parallel with the F_2 fold axes. A particularly distinct quartz rodding at 100–110° was formed in the Slieve Tooey Quartzite. Near Skelpoonagh Bay, later twisting of the F_2 structures has resulted in this quartz rodding striking almost north-south. Major F_2 folds are prevalent in the west, particularly at the head of Skelpoonagh Bay. They are tight curvilinear folds variable in both direction and plunge. At Skelpoonagh Bay folds up to 10 m. in amplitude occur in banded psammitic rocks. They plunge steeply down dip and contrast with the F_4 folds which have a sub-horizontal plunge. Associated minor folds (Fig. 9A) occur in the more finely-banded psammites and in the dolomites which lie below.

Deformation became weaker after the second phase, and phase three structures (F_3) are open or moderately tight folds associated with a puckering or a strain-slip cleavage. This cleavage consists of strain-slip bands 5 mm. or less in width; S_3 planes usually lie within 30° of earlier schistosities, dipping northwards between 20° and 45° and striking between W.S.W. and W. F_3 minor folds are rare and show, like the strain-slip bands, a vergence towards the south; their axial trend is about W.N.W. dipping at 10° to 30° except, in the region of the Glencolumbkille Anticline where due to late modification a wide divergence of planes and fold axes is evident: Figs. 9B and 9F show the styles of F_3 minor folds and their relationship to S_1 and S_2. During this deformation the Glencolumbkille anticline, together with the complimentary syncline on its southern side, were initiated.

In the Glencolumbkille area the most conspicuous minor structures, i.e. similar folds of open concertina form and kink bands associated with strong fracture and strain-slip cleavages, were formed during the fourth phase of deformation (F_4). These structures, are ubiquitous, but they are exceptionally well seen in the Glen Head Schist 360 yds. S.S.W. of the C of Cloghan (on sheet 81) and E.S.E. from there to about 800 yds. N.N.E. of Glen Lodge. In finely banded lithologies the fold hinges are angular, and in coarsely banded rocks curvilinear: in both cases limbs are straight (Fig. 9C). They show a vergence northwards, with short limbs dipping steeply northwards and long limbs gently southwards; their attitude and that of their associated strain slip cleavage (Fig. 9D) which dips to the south, is in the west of folds with axial plane strikes of 60° and axial plane dips of 20°–40° south eastwards. This is somewhat different from that in the eastern and central areas where the axial planes strike approximately 120° and dip south-westwards between 20° and 75°. This swing in strike of axial planes and cleavage is probably due to late flexuring. Variation in fold plunges is slight, with most axes being sub-horizontal. Figs. 9E and 9F show the relationship between S_4, S_2, S_1 and S_0. Major structures, such as the Glencolumbkille anticline and syncline, which originated in the previous phase were further developed by this phase. What was probably a fairly open anticline formed by F_3 with a northward dipping and plane, became accentuated by F_4 folding into a tight fold with a southward dipping plane, a gentle westward plunge at about 20°, and a general style and trend normally associated with F_4. A gently southward dipping fault which crosses the feldspathic quartzite nearly parallel to the bedding 750 yds. E.S.E. of the N.E. corner of Skelpoonagh Bay appears to be an F_4 structure. Similarly the main face of the crag of Craigbeefan, which is a smooth face dipping to the S.S.W. at 40°, is probably also of F_4 age although it might be F_8. Also, the hollow which strikes towards Skelpoonagh Bay from immediately north of the northernmost church in Cashel, and is parallel to Craigbeefan, presumably contains either an F_4 or an F_8 fault.

Brittle conditions prevailed during the fifth phase of deformation (F_5) and open monoclines or shear zones both with a coarse fracture cleavage were produced. These structures are discordant to the general schistosity and strike N.E.–S.W. with axial planes which dip steeply to the N.W. Their plunge is variable and dependant on the attitude of bedding and schistosity. In shear zones the associated minor folds are of a chevron type and vary in intensity along the axial plane. S_5 strain-slip cleavage is strong, but irregular, with planes that bifurcate and die out. Semipelites and psammites readily break along these planes revealing an uneven and often puckered surface. In fold hinges these planes form a cleavage fan, but even away from the major structures they possess a variable dip. No major F_5 folds have been mapped but many of the northeasterly striking faults, particularly those on Craigbeefan belong to this deformation phase and the widely spaced (1–10 yds.) fracture cleavage in the quartzites and Boulder Bed, though some of this is also of F_6 age.

The sixth phase of deformation also affected brittle rocks. The F_6 folds have a similar strike of their axial planes to those of F_5, being 40–50° but the planes dip to the south-east between 20° and 65°; the plunge being gently either to the north-east or to the south-west. Many of the northeasterly trending faults, especially south of Skelpoonagh Bay, belong to F_6 and the main fault which limits the lower exposure of Boulder Bed is F_6 and dips S.E. at

65°. F_6 folds are best seen between Skelpoonagh Bay and Glen Bay in the Glencolumbkille Schist.

Two complimentary sets of steeply dipping strain slip cleavage with axial planes striking 340° and 20° form late conjugate folds (F_7) and knick bands while the latest knick bands (F_8) strike 120–130° and form small faults.

All the movements after F_4 took place under brittle conditions and are not so important as the previous periods of deformation.

5. Structure

Within the Glencolumbkille area the two most obvious structures are the Glencolumbkille Anticline and the Glencolumbkille Syncline. These east-west trending folds with a steeply dipping axial plane are impressed on a northerly dipping sequence of metasediments, and are responsible for the excellent development of the Boulder Bed in this area. In the Kiltyfanned Anticline, three miles to the north of this area, a similar structure again exposes Boulder Bed amidst large areas of Slieve Tooey Quartzite.

In the central area, north-east of the village of Cashel the anticline has a short steep, northerly-dipping northern limb and a long gently southward-dipping southern limb. Its axial plane, which strikes at 115° with a dip southwards of 60°, and its axis which plunges westwards at 10° are both close to the average measurements of the F_4 minor folds. In the central area, the anticline is tight; eastwards it dies out, and westwards it opens out but it can still be traced to the Glen Head-Skelpoonagh Bay area.

Although in general shape and attitude it is more akin to the F_4 folds, the Glencolumbkille anticline is not of F_4 age. It was initiated in the third phase of deformation and was only modified by the fourth.

Evidence for the age of the fold is seen north-east of Cashel. There, on the southern limb of the anticline the S_4 cleavage dips at a lower angle than the bedding and graded bedding showing that the cleavage youngs downward in an uninverted succession. This cleavage-bedding relationship, which is consistently observed in this area, indicates the superposition of the cleavage upon an earlier structure. In the centre of the anticline the interference of cleavage and schistosites of different ages is uncommon but sufficient examples have been seen to ascribe the inception of the anticline to the third phase of deformation.

A north-south section across the Glencolumbkille area is shown in Fig. 10A. On this the traces of the most probable axial planes for the Glencolumbkille Anticline and the Glencolumbkille Syncline are indicated. Two possible planes are suggested for the anticline. One is the bisectrix of the fold angle and its dip is approx. 55° to the south. The other is a curved plane which dips southwards in its upper region but dips steeply northwards in its lower part. This latter suggestion is based on the attitude of schistosity and S_3 cleavage along the length of the fold; as mentioned before the fold becomes less important both eastwards and westwards and this is reflected in the style, attitude and intensity of the minor folds. From the attitude and relationship of F_2, F_3 and F_4 minor folds the curved plane is considered most probable

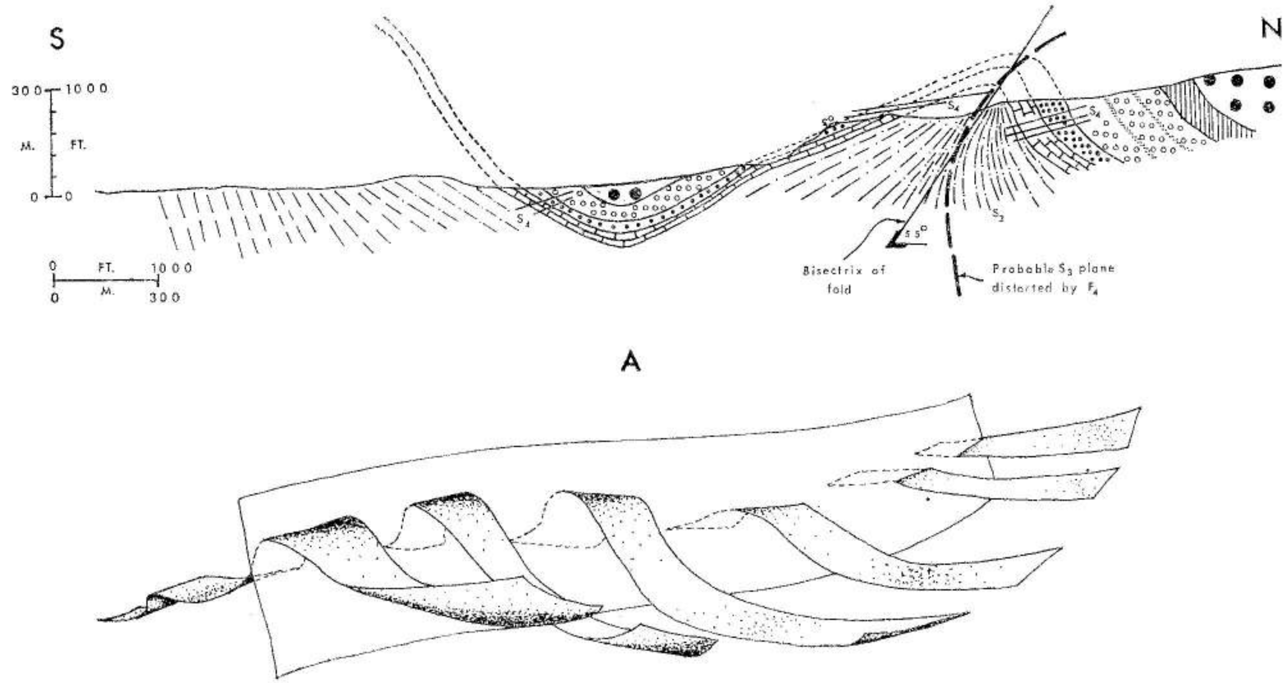


FIG. 10 A—North-south section across the Glencolumbkille Anticline and Syncline. Vertical scale and horizontal scales are equal. Section line 800 yds. west of Glen Lodge. Key to ornament as in Plate A.

B—Representation of the form of the Glencolumbkille Anticline and Syncline. Viewed from south west.

even though a bisecting plane is of almost identical attitude to the average plane for the F_4 folds.

This being so it appears that fairly open folds with steeply northward dipping axial planes were formed during the third deformation. These were then tightened and partially overturned to the north by the following fourth deformation. The result is a complex fold produced by two consecutive phases of deformation.

In the west the main anticline is not very obvious because of faulting and lack of exposure away from the coast. Some large folds are observed, these are usually F_2 or F_4 in age. On the cliffs south of Glen Head tower there are excellent F_4 folds in the Boulder Bed and the underlying dolomite. At Skelpoonagh Bay, in a series of banded psammites and marbles there is complex refolding of F_2 folds by F_4 folds. Two major tight east-west trending F_2 folds are bounded to the north and to the south by fairly open F_4 folds. The F_2 folds have steep southward dipping axial planes with axes that plunge down-dip while the F_4 folds have subhorizontal axes with moderately steep southward dipping axial planes. Minor folds of both periods abound; these produce, especially in the marbles, abundant and often chaotic refolds.

South of the main anticline is the basin-form Glencolumbkille syncline whose form is indicated by the dips of the strata involved (Fig. 10A).

6. Metamorphism

Two episodes of metamorphism have affected the Glencolumbkille rocks. The earliest and most intense reached its climax shortly after the peak of the second deformation, and was then of almandine amphibolite facies. The second metamorphism largely post-dated the movements and reached its climax in the upper greenschist facies (quartz-albite-epidote-almandine subfacies of Fyfe, Turner and Verhoogen, 1958, p. 224). It is not certain whether or not continuous metamorphism occurred between these two metamorphic events.

During F_1, foliation and schistosity were formed with the assemblage muscovite-biotite-ilmenite-quartz-oligoclase-graphite-calcite or dolomite in the pelites and semipelites. A fine grained phyllitic fabric was produced, which was practically destroyed by later mineral growth. This is suggested by the relict internal foliations sometimes found in later garnets which indicates that the micas were originally fine grained and that subsequently larger mica grew in the S_1 foliation by recrystallisation. Similarly, in the calcareous and dolomitic rocks, tremolites contain a relict fine grained S_1 foliation. The degree of recrystallisation and granulation varied from very little in many of the psammites, which may retain delicate original sedimentary features, to complete recrystallisation the pelites.

The apogee of the initial metamorphism probably post-dated F_2 and gave the following mineral assemblages: quartz-muscovite-biotite-oligoclase-garnet-ilmenite-graphite in the pelites and semipelites; quartz-oligoclase-K feldspar-muscovite-biotite in the psammites, tremolite-calcite-dolomite-phlogopite-sphene-oligoclase in calc-silicate rocks and marbles, hornblende-basic andesine or labradorite-quartz-garnet-ilmenite-sphene (with epidote?) usually in a non-schistose texture, in the metadolerites. This non-schistose, or fels texture, suggests that strong movements did not prevail during the peak of the metamorphism which is supported by the observation that some of the garnets in the pelites enclose small folds which fold the first schistosity. On the other hand some garnets contain a planar internal schistosity derived from S_1 which now lies at a discordant angle to the second schistosity. This indicates that the garnets were either rotated slightly in the second schistosity or the second schistosity has been flattened about the garnets. Thus garnet growth may have started during F_2 and continued during the decline of this folding. The common occurrence of quartz, calcite and ilmenite inclusions and the general absence of sphene in the garnet suggests that calcite did not decompose before the growth of the garnet. Although the second deformation usually formed a schistosity, some semipelites contain a strain-slip cleavage in which segregation of mica and quartz occurred while more pelitic bands may have "strain-slip eyes"—augen structures of quartz grains with transverse micas lying parallel to the relict S_1 plane. The new schistosity (S_2) contained much coarser muscovite and biotite compared with S_1, no doubt reflecting the greater ease of diffusion under the higher temperature conditions. It is notable that the pelites richest in graphite contain fine grained muscovite.

As partly chloritised garnet occurs throughout the Glencolumbkille schists, and in parts of the metadolerites, the line drawn on Anderson's (1953, p. 419) map showing the present area to be largely outside the limit of garnet, is clearly erroneous. The presence of garnet in parts of the metadolerites and the possibility that epidote may have coexisted with hornblende and andesine or labradorite, suggests that the lithostatic pressure during the peak of the metamorphism was fairly high (Shido, 1958, p. 210; Fyfe, Turner and Verhoogen, 1958, p. 229). This is supported by the presence of both dolomite and quartz in many slides of the Glencolumbkille Dolomite, which indicates that substantial breakdown of the dolomite was prevented by a high partial pressure of carbon dioxide, for the temperature must have been fairly high if garnet could crystallise. The rare presence of diopside in Glencolumbkille Dolomite adjoining metadolerite suggests early, localised hornfelsing by the dolerite magma.

The metamorphism progressively declined from its peak after F_2 and during F_3 a strain-slip cleavage with bending of micas was more commonly produced than a schistosity though considerable recrystallisation of muscovite and biotite occurred in the Glen Head schists in the core of the Glencolumbkille Anticline. Garnet, however, does not seem to have crystallised. Plagioclase seems to have contined to grow for some time after F_2 and some plagioclase augen which have grown in the strain-slip cleavage of some semipelites have formed a crude foliation of ilmenite-epidote-muscovite-plagioclase bands alternating with bands of quartz.

Between F_3 and F_4 chlorite started to form by alteration from biotite. Chlorite growth continued during F_4 but the metamorphism was now too low for any other mineral growth, with the possible exception of some epidote and acid plagioclase in the metadolerites. Chlorite that had formed between F_3 and F_4 was both deformed and recrystallised during F_4, giving some development of chlorite in S_4 planes. The remaining minerals were mainly fractured and deformed by progressively more brittle conditions.

After F_5 a static metamorphism of upper greenschist facies resulted in the development, in the pelites and semipelites, of relatively large (1–5 mm.), randomly oriented, biotite and chlorite porphyroblasts. This metamorphism was therefore the result of a re-heating of the area after a period, during the fifth defomation and probably later, of either no metamorphism or metamorphism so low that chlorite porphyroblasts did not form at the expense of the biotite. Little or no growth of muscovite occurred in this late metamorphism but plagioclase sometimes overgrew a fabric deformed by F_5.

In the metadolerites hornblende partly recrystallised into randomly arranged needles and the calcic plagioclose broke down into albite or acid oligoclase and zoisite or epidote. New biotite and sphene also crystallised in these rocks. The general reaction seems to be labradorite + hornblende + ilmenite $\rightarrow$ albite + zoisite + sphene + new hornblende, the calcium needed for the formation of sphene, zoisite and an increased volume of hornblende coming from the calcic plagioclase; the old and new hornblendes will have different compositions. During this important late metamorphism clots of disarranged tourmalines grew locally, e.g. in the Upper Boulder Bed crystals up to 3 cm. in length occur in a quartzite intercalation 1,910 yards S.S.W. of 916 triangulation point (75 yds. S.E. of B.M. 179·2). Statically grown stellate tremolite porphyroblasts which have carbonate inclusions in their centres and an inclusion-free outer zone were formed (e.g. quarry, 240 yds. N.E. of Glen Lodge). Late ilmenite and also porphyroblasts of pyrite (up to 2 × 2 × 2 cm. near to Glen Lodge) also crystallised at this time though it is apparent that both pyrite and ilmenite have crystallised at several different times. The crystallisation of these relatively large crystals of pyrite and tourmaline implies appreciable movement of elements such as B, S, Si, O and Fe even if only to segregate the B and S in the host rock. But the replacement of substantial amounts of hornblende by biotite in the metadolerites near to the top of Craigbeefan can only be accounted for by K metasomatism from outside the metadolerites. The especially prominent late biotite porphyroblasts in the Boulder Bed in the upper parts of Craigbeefan supports this interpretation. It is not certain whether garnet crystallised during this late metamorphism. Euhedral garnets, with inclusion-filled cores and inclusion-free rims occur 720 yds. from Glen Lodge on a bearing of 50° east of north. The inclusion-free rims may well have crystallised at this time but definite proof is, at present not available.

The late thermal metamorphism, with its resulting quartz grain growth in the pelites and semipelites, welded together the shear planes and cold-worked surfaces along which movement unaccompanied by recrystallisation had occurred. Thus the pelites and semipelites, despite their many strain slip and fracture cleavages, do not disintegrate when hammered.

During the latter part of the late metamorphism there was widespread chloritisation and epidotisation. Biotite, hornblende and garnet were wholly or partly altered to chlorite and porphyroblastic chlorite developed. There appear to be differences in colour, interference colour and form between the chlorites formed between the fourth and fifth deformations and those formed during the last metamorphism. Thus the early chlorites derived from biotite often contain rutile needles while the latest ones do not. There are also differences between the late, light-green chlorite developed by replacement of early biotite and the darker chlorite formed from late porphyroblastic biotite. All these chlorites differ from the porphyroblastic, late-grown, chlorites which are spindle or barrel shaped and have a rounded, smooth form, with poor cleavage and a well-defined polysynthetic twinning parallel to (001). They overgrow the earlier fabric, including the S_5 fracture cleavage and include internal schistosities and foliations, as shown by inclusions such as graphite and quartz. Undoubtedly these chlorites and late epidote replacing plagioclase are the last significant minerals to have formed in the metamorphisms of the area, post-dating all the principal movements although their distortion by rejuvenated strain-slip planes shows that some movements occurred after their growth. Contemporaneous with the chloritisation was partial sericitisation of the garnets and plagioclases and the development of leucoxene from ilmenite and sphene.

This late metamorphism has not been precisely dated but it is after F_5 and before F_7 and was not accompanied by movement. Part of the difficulty in dating this metamorphism lies in the problem of whether the F_5 and F_6 faults had posthumous movement. These faults clearly displace both quartz porphyry and lamprophyre dykes (Plate A) and similar faults are reported by Pitcher *et al.* (1964, p. 252) to be either Devonian or early Carboniferous and to have a complex history of repeated movement. At Glencolumbkille some movement on these faults is later than the thermal metamorphism but it is possible that this movement is not F_5 or F_6 but is some much later movement. Consequently this metamorphism has not been precisely dated.

It is evident that the rocks now contain a jumbled collection of minerals belonging to a sequence of crystallisations and it is not possible to give a mineral assemblage for each rock type in each stage of the two metamorphisms because there is often no means of identifying the different mineral ages. Too much earlier work in metamorphic terraines has ignored this ubiquitous state of affairs. In summary, the metamorphism increased to a peak during the later stages of F_2 when the almandine amphibolite facies was reached. A decline followed to lower grades until after F_5 a late re-heating occurred to the upper greenschist facies which was followed by widespread chloritisation of the lower greenschist facies. Fig. 11 diagrammatically summarises the relationships between deformation and metamorphism.

7. Discussion

This study is a contribution to the detailed stratigraphy of the Portaskaig Boulder Bed Series in the Dalradian of Scotland and Ireland. In Ireland, there are only two localities in which the Boulder Bed Series seems to be largely complete, not greatly thinned tectonically and little deformed. These are at Fanad in north Donegal and at Glencolumbkille with its contiguous extension into Kiltyfanned. The present area is therefore of great importance for the correlation of the Boulder Bed Series and it is surprising that no previous detailed work has been attempted. Thus the measured thickness of the Boulder Bed (101–118 m.) is at least twice as thick as that suggested by Kilburn, Pitcher and Shackleton (1965, p. 348).

The evidence assembled by Kilburn, Pitcher and Shackleton (1965, p. 335) in support of a glacial origin for the Boulder Bed, as distinct from subaqueous slide-breccias or mud flows is impressive and will not be repeated here. The constancy of the succession over 600 km. is remarkable. An aspect not discussed by the above authors is the common association of dolomite or limestone with supposed tillites, which suggests a much more definite environmental control over the sedimentation than subaqueous slide breccias or mudflows would imply. Thus carbonate horizons associated with supposed tillites have been reported from the Pre-Cambrian of several Continents, though the glacial origin of some of these "tillites" is not clearly proven.

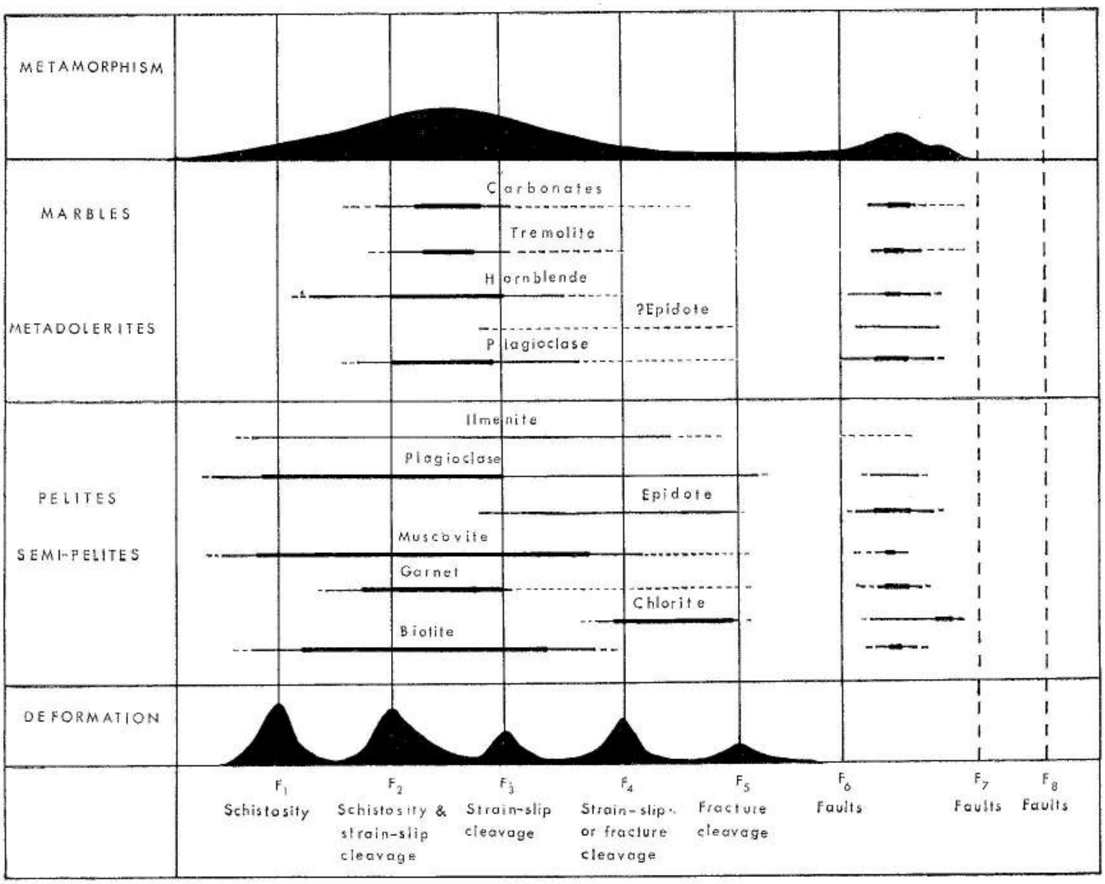


FIG. 11—Diagrammatic relationship of the history of metamorphism and deformation at Glencolumbkille.

The European localities include (1) the Tillite Formation of Kap Oswald north-east Greenland where two tillites are interbedded between dolomitic argillites, and calcareous sandstone, the upper tillite conformably overlain by a laminated dolomite; stromatolitic limestones occur some distance below the base of the Tillite Formation (Schaub, 1950, 1955), (2) the Polarisbreen Series of Ny Friesland and Olav V Land, Spitzbergen (Harland and Wilson 1956, Wilson and Harland, 1960) contains dolomite or dolomitic siltstone immediately above and below the Polarisbreen Tillite Formation. (3) The Moelv conglomerate of Southern Norway (Holtedahl, 1922) in places succeeds calcareous or dolomitic sandstones, rarely with a slight angular unconformity (Föyn, 1937).

In the Riphean of the Southern Urals the Valdaisk (Ashinsk) Series ("rich in block conglomerates of the tillite type") overlie the stromatolitic dolomites of the Karatarsk series with a "sharp discordance" (Menner, 1960).

The well described "tilloids" of the West Congo System in Angola overlie limestones and calcareous quartzites and are overlain by a thick carbonate sequence, but their glacial origin has been disputed (Schemerhorn and Stanton, 1963). In south-west Africa the Chuos tillite of Western Damaraland (Gevers, 1931) appears to have a few thin limestone intercalations in its upper part. The Waldidiala Series of Sierra Leone and Guinea overlies the stromatilitic limestones of the Medina Konta series and is reported to contain a basal tillite (Furon, 1960). In the Tasudeni Basin of the western Sahara the base of the Cambrian is defined by a tillite overlain by a dolomitic limestone, which is itself separated from underlying sediments, including stromatolitic carbonates, by a marked unconformity (Zimmermann, 1960).

The best and probably the most unequivocal examples are those from Australia. The Sturtian tillite in the North Flinders Ranges (Mawson, 1949) commonly overlies dolomite and magnesite-bearing shales, and is overlain in places by a thin limestone followed by calcareous shales. In the Kimberley region of Western Australia, Dow (1965), describes the Moonlight Valley and Fargoo tillites. The former is underlain by considerable areas of striated and grooved pavement and overlain by 6–12 ft. of laminated dolomite, the latter is underlain by up to 30 ft. lenses of dolomitic conglomerate and containing lenses of dolomite and dolomitic sandstone showing pre-consolidation slumping and in extreme cases rolled-up masses of dolomite.

The similarity between the Portaskaig Boulder Bed of Donegal and these Precambrian tillite-like rocks in their association with carbonate rocks is remarkable and probably implies a similar, glacial, origin for all these occurrences.

This work attempts the first detailed description of the metamorphism in any part of the 130 sq. miles of the Slieve League Promontory. Contrary to Anderson's (1953, p. 419) map, garnet has been found to be abundant in the area mapped. A complex sequence of crystallisations has only been partially unravelled and much work remains to be done. Two metamorphisms have been clearly distinguished; the first and main metamorphism took place, before, during and after the F_2 fold movements with its climax shortly after the F_2 movements. Metamorphism continued at least intermittently until at least the F_5 folding but at a lower grade and with less construction of the rock fabric. The main metamorphism, of garnet zone or upper greenschist facies, was probably accompanied by fairly high lithostatic pressure for dolomite and quartz occur together in the Glencolumbkille Dolomite and garnet is found in parts of the metadolerites. A late static metamorphism, also of upper greenschist facies came after F_5 and implies a re-heating of the area at a late stage which was followed by final retrogression.

The apogee of the main metamorphism after F_2 is similar to that deduced by Rast (1958, p. 427) and Sturt and Harris (1961, p. 689) for the Dalradian rocks of Central Perthshire, and to that described by Dalziel and Brown (1965, p. 309) for part of the Moine sequence, while Zwart (1962, p. 38) has obtained a similar result in the Boost area of the central Pyrenees.

As the metamorphic grade of the Boulder Bed Series in Donegal is generally below the garnet zone the Slieve League promontory is of higher grade than elsewhere. There may be an increase in the metamorphism passing from Fanad in north Donegal to Slieve League to N.W. Mayo where kyanite and staurolite occur (Trendall and Elwell, 1963) only to be followed by metamorphic decline towards the metamorphic rocks around Clew Bay. Current evidence from the Alps and Pyrenees suggests that such metamorphic variations are not primarily related to differing depths of burial but to differing rise of the isotherms and the geothermal gradient.

The structural history of four important and four late subordinate brittle phases of deformation has nevertheless resulted in a fairly simple overall structure of the Glencolumbkille anticline and syncline. In this small area it has not been possible to integrate the important structures into the broad tectonic picture of Donegal because in the Slieve League promontory no detailed structural work, or even a more detailed geological map than that published 75 years ago by the Geological Survey of Ireland, is available. It seems probable that the area lies on the upper limb of a major recumbent anticline, which may close to the south, perhaps near Slieve League. Some of the latest brittle movements can probably be correlated with the Leannan fault system (Pitcher *et al.*, 1964, p. 241).

The existence of metadolerites of definite intrusive origin and now largely converted to amphibolites is greatly relevant to the problem of the origin of amphibolites in the Dalradian Series, especially in Connemara and Scotland. It seems highly probable that the whole suite of these amphibolites was originally dolerite sills and dykes.

8. ACKNOWLEDGEMENTS

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9. REFERENCES

ALLEN, J. R. L. 1963 *Sedimentology*, **2**, 93–114.

ANDERSON, J. G. C. 1953 *Quart. J. geol. Soc. Lond.*, **91**, 399–422 (for 1953).

DOW, D. B. 1965 *Geol. Mag.*, **102**, 407–414.

DAZIEL, I. W. D. and BROWN, R. L. 1965 *Scot. J. Geol.*, **1**, 304–312.

FLINN, D. 1962 *Quart. J. geol. Soc. Lond.*, **118**, 385–433.

FÖYN, S. 1937 *Norsk. Geol. Tidsskr.*, **17**, 65–163.

FURON, R. 1960 *Géologie de l'Afrique*. 2nd ed. Payot, Paris.

FYFE, W. S., TURNER, F. J. and VERHOOGEN, J. 1958 *Geol. Soc. Amer. Mem.* **73.**

GEVERS, T. W. 1931 *Trans. Geol. Soc. S.A.*, **34**, 1–17.

GRAF, D. L. and GOLDSMITH, J. R. 1958 *Geochim et cosmoch. Acta.*, **13**, 218.

HARLAND, W. B. and WILSON, C. B. 1956 *Geol. Mag.*, **93**, 265–286.

HOLTEDAHL, O. 1922 *Amer. J. Sci.*, 4, 165–173.

HULL, E., KILROE, J. R., and MITCHELL, W. F. 1891 Explanatory memoir to accompany the maps of South-West Donegal, Sheets 22, 23, 30 and 31 (in part), of the Geological Survey of Ireland. Mem. geol. Surv. Ireland.

HUANG, C. K. and KERR, P. F. 1960 *Amer. Min.*, 45, 311–324.

KILBURN, C., PITCHER, W. S. and SHACKLETON, R. M. 1965 *Geol. J.*, **4**, 343–360.

KILROE, J. R. 1907 *Proc. R. Irish Acad.* (Sect. B), **26**, 129–160.

KRUMBEIN, W. C., and SLOSS, L. L. 1963 *Stratigraphy and Sedimentation.* 2nd Ed. Freeman. San Francisco and London.

MAWSON, D. 1949 *Trans. R. geol. Soc. S. Aust.*, 72, 244–251.

McKEE, E. D. and WEIR, G. W. 1953 *Bull. geol. Soc. Amer.*, **64**, 381–390.

MENNER, V. V. 1960 *Intern. Geol. Congr.*, 21*st Copenhagen,* 1960, *Rept. Session, Norden.* Pt. **8**, 18–23.

M'HENRY, A. 1891 *in* Explanatory memoir to accompany sheets 3, 4, 5 (in part), 9, 10, 11 (in part), 15 and 16 of the maps of the Geological Survey of Ireland, comprising North-West and central Donegal. *Mem. geol. Surv. Ireland.*

PITCHER, W. S., ELWELL, R. W. D., TOZER, C. F., and CAMBRAY, F. W. 1964 *Quart. J. geol. Soc. Lond.*, **120**, 241–273.

RAST, N. 1958 *Trans. R. geol. Soc. Edinb.*, **63**, 413–431.

SCHAUB, H. P. 1950 *Medd. Groenland,* **114**, no. 10.

——— 1955 *Medd. Groenland,* **103**, no. 10.

SCHEMERHORN, L. J. G. and STANTON, W. I. 1963 *Quart. J. geol. Soc. Lond.*, **119**, 201–241.

SHACKLETON, R. M. 1961 *Proc. geol. Soc. Lond.*, 1587, 75–76.

SHIDO, F. 1958 *J. Fac. Sci.* Tokyo (sect. 2), **11**, 131–217.

SPRY, A. 1963 *Amer. J. Sci.*, **261**, 756–766.

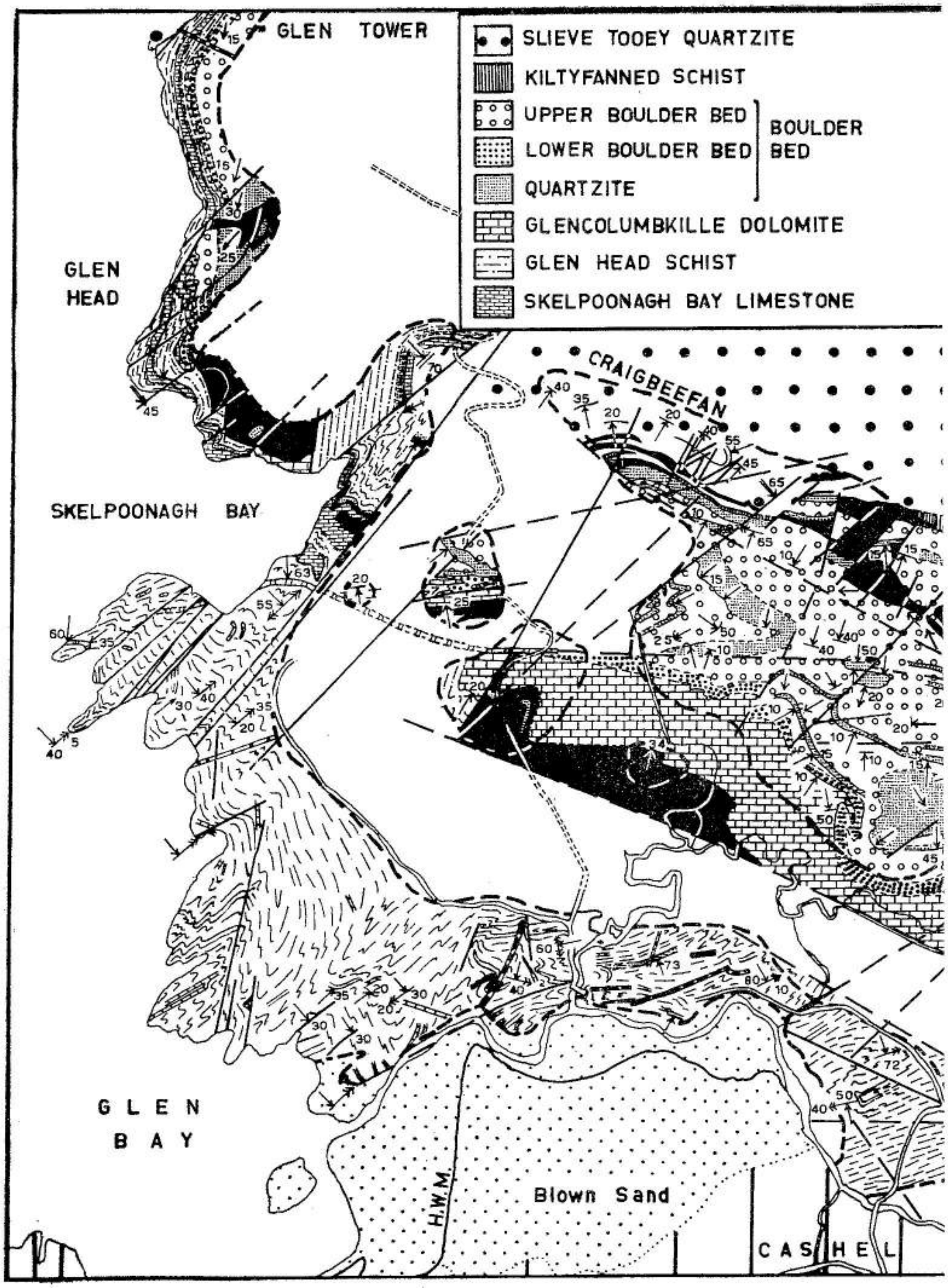
STURT, B. A. and HARRIS, A. L. *Liverpool and Manchester Geol. J.*, **2**, 689–711.

TRENDALL, A. F. and ELWELL, R. W. D. 1963 *Proc. R. Irish Acad.* (Sect. B), **62**, 217–247.

WILSON, C. B. and HARLAND, W. B. 1960 *Geol. Mag.* **101**, 198–219.

ZIMMERMANN, M. 1960 *Intern. Geol. Congr.*, 21*st, Copenhagen,* 1960, *Rept. Session, Norden.* Pt. **8**, 26–36.

ZWART, H. J. 1962 *Geol. Rund.*, 52, 38–65.


GLEN TOWER
SLIEVE TOOEY QUARTZITE
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GLENCOLUMBKILLE DOLOMITE
GLEN HEAD SCHIST
SKELPOONAGH BAY LIMESTONE
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CRAIGBEEFAN
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GLEN BAY
H.W.M.
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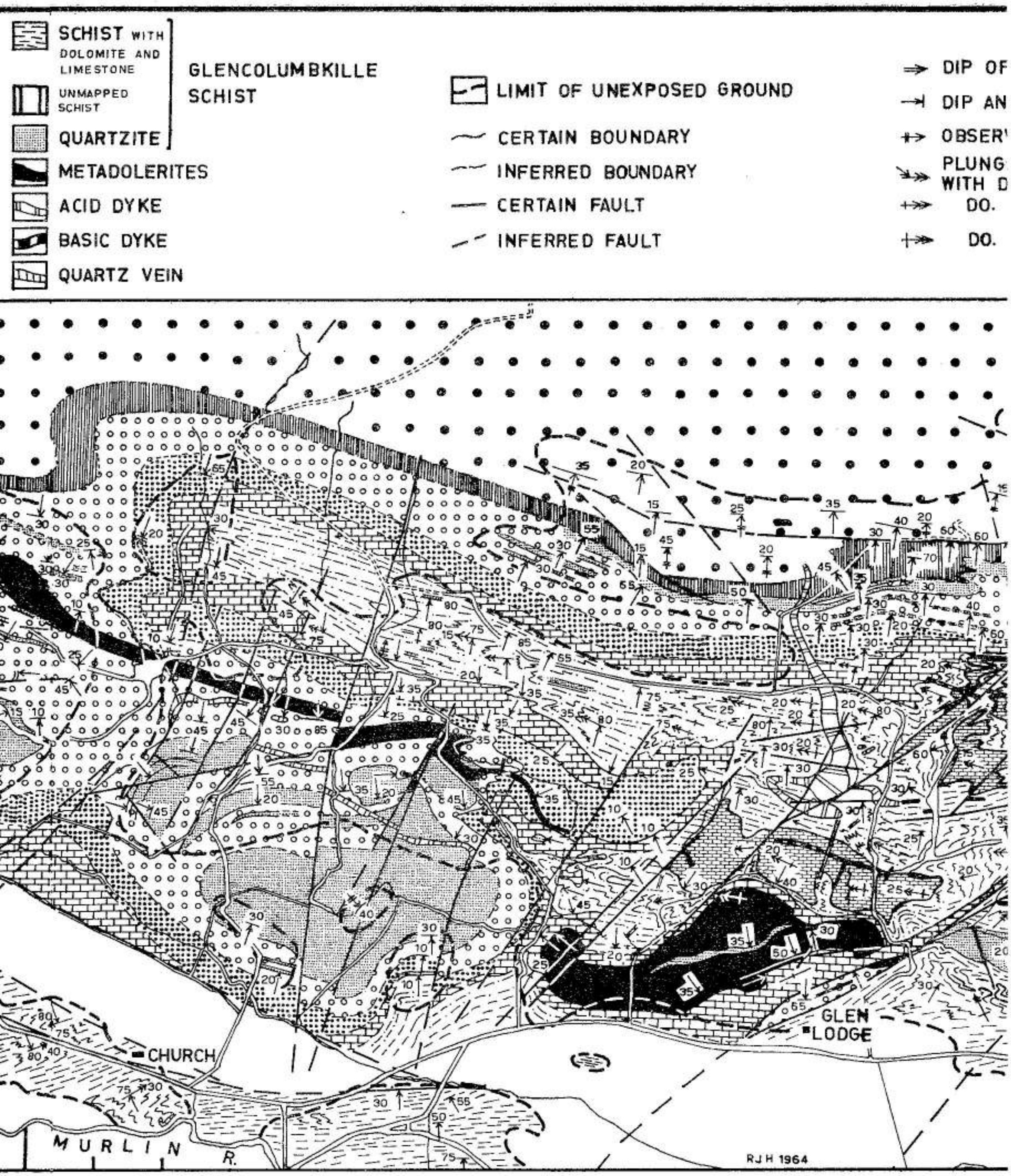
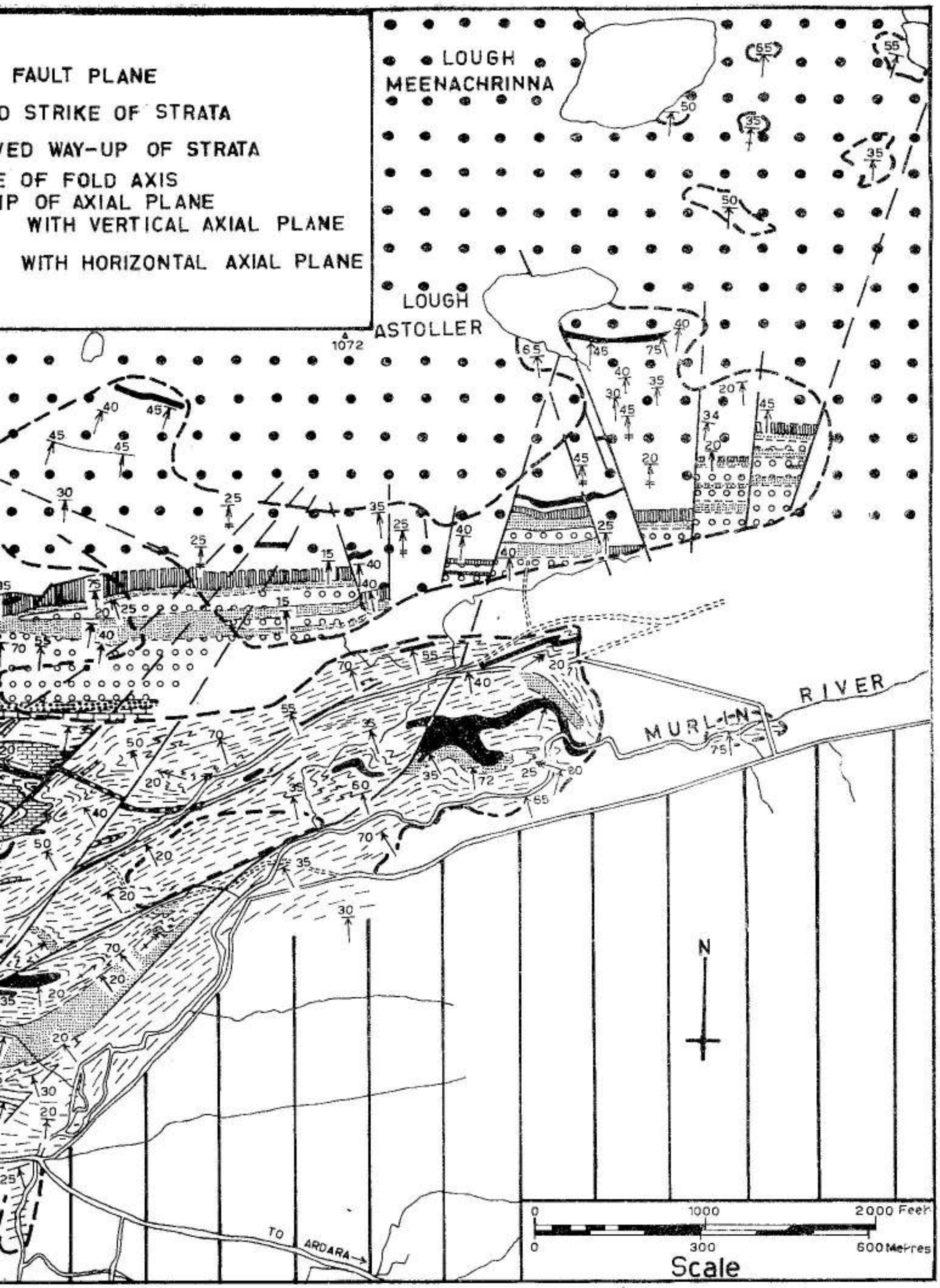


(Based on the Ordnance Survey by permission of the Government)

Geological map of Glencolumbkille. Limestones and dolomites are **distinguished**, where possible, on the basis of infı



ared spectra.