THE STRUCTURE, METAMORPHISM, AND STRATIGRAPHY OF THE SKIDDAW SLATES, EAST OF CRUMMOCK WATER, CUMBERLAND.

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Thesis submitted for the degree of Doctor of Philosophy (Ph.D.) under the Special Regulations of the University of Birmingham.

December, 1973.

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SYNOPSIS

The Skiddaw Slates succession of this area is 7,500' thick, and is sub-divided into five lithological groups:-

Buttermere Slates

Whiteless Pike Slates

Mosser-Kirkstile Slates

Loweswater Flags

Hope Beck Slates

Intruded into these sediments are two distinct groups of basic igneous bodies. Those of the south are very tholeiitic in composition, whereas those of the north are more silica-rich and show close geochemical affinities to the northern - Eycott - outcrop of the Borrowdale Volcanic Group.

Both the sediments and the basic igneous rocks were subjected to very low (chlorite) grade regional metamorphism towards the end of the Silurian period.

In addition, the central part of the area suffered a low-grade contact metamorphism, generating ?cordierite, andalusite, and biotite in suitable lithologies and effecting an overall baking of the aureole rocks.

The aureole is the result of the postulated intrusion beneath the area of a granitic pluton; the granite is thought to be related to the Skiddaw and Eskdale Granites (dated radiometrically at 395 $\pm$ 14 m.y.).

Four phases of deformation have been recognized within this area. F1 structures are represented by N - S striking overturned folds, with no cleavage, and are of pre-Borrowdale Volcanic age. F2 - F4 are end-Silurian events; F2 deformation produced ENE - WSW trending folds, with an associated fracture cleavage, and refolded the F1 folds into their present, generally steep-plunging, attitude. F3 folds, characterized by horizontal axial planes, are open warps with an associated fracture or crenulation

cleavage; F4 folds are open structures, with NW - SE axial planes. These last two events, F3 and F4 are thought to be due, respectively, to vertical and axial compression consequent upon the flattening of the rocks induced by F2.

DEDICATION

"The unravelling of the structure of the Skiddaw Slates would be a task of immense difficulty and would probably not repay the expenditure of the necessary time." Prof. R.H. Rastall (1910).

The thesis is respectfully dedicated to the late Professor R.H. Rastall.

ACKNOWLEDGEMENTS

My first, and largest, debt of gratitude is owed to my supervisor, Dr. Frank Moseley, who suggested the project initially and who has been, subsequently, an unfailing source of encouragement and guidance in all stages of the production of this thesis.

The thesis would never have been finished on time, however, were it not for the help of my wife, Anne, whose invaluable assistance, given at the expense of her honeymoon, is most gratefully acknowledged.

The work also benefitted greatly from discussions in the field, (and in certain other, more liquid, environments) with Barry Webb, who was working on the ground to the south of Buttermere.

The advice and assistance concerning the X.R.F. work given by Dr. G.L. Hendry are acknowledged.

My long sojourns in the Lakes would not have been nearly so productive were it not for the friendship and warm hospitality I received from the people of Lorton and Buttermere, especially Mr. and Mrs. J. Tyson; Mr. and Mrs. C. Allison; Mr. R. Twitchin; and Mr. R. Dixon.

Financial assistance from N.E.R.C. is gratefully acknowledged, as are the provisions made for me in the Geology Department, University of Birmingham, by Prof. F.W. Shotton, F.R.S.

Lastly, grateful thanks are due to my parents for their kindness, help and encouragement given to me over the years.

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CHAPTER ONE

Section 1.1 Introduction.

The aim of the present study is the structural and stratigraphic elucidation of an area of Skiddaw Slates in the N.W. Lake District, together with an inquiry into the effects upon these rocks of various metamorphic processes. The area concerned is bounded to the west by Crummock Water and the intake wall; and to the north and south by G.R. Easting 526 and Sail Beck, respectively. The eastern boundary is more tenuous, roughly following the ridge from Whiteless Pike across to Coledale Hause, thence up to Hopegill Head, and down to Easting 26 along the Ladyside Pike ridge. The area thus delimited covers about 21 sq. km., and is shown on Fig. 1:1:1.

The reasons for selecting this area were two-fold:-

a) preliminary investigation, both in the field and from published work, showed it likely to contain a reasonable variety of rock type, which it was hoped would make stratigraphic work more straightforward; and

b) the two earlier accounts of this ground by Rose (1955) and Simpson (1967) are in total disagreement on both structural and stratigraphic matters, so it was hoped that by mapping this area in detail, the obvious difference between the two theories might be resolved. (Details of Rose's (1955) and Simpson's (1967) work are discussed more fully in Section 1.3).

The area is one of high relief (Fig. 1:1:1), rising from 300' O.D. on the shores of Crummock Water to 2,791' at the summit of Grasmoor (Fig. 1:1:2). The Buttermere-Crummock trough valley (Fig. 1:1:3) is the dominant feature of the area, and all drainage therefrom leads down to this valley. Post-glacial rejuvenation has left these streams with very youthful profiles and deeply incised valleys (Fig. 1:1:4). Above the valleys rise the fellsides and ridges of the area; there are three major ridges and they can be visualised as radiating from Coledale Hause (Fig. 1:1:1), one of the three

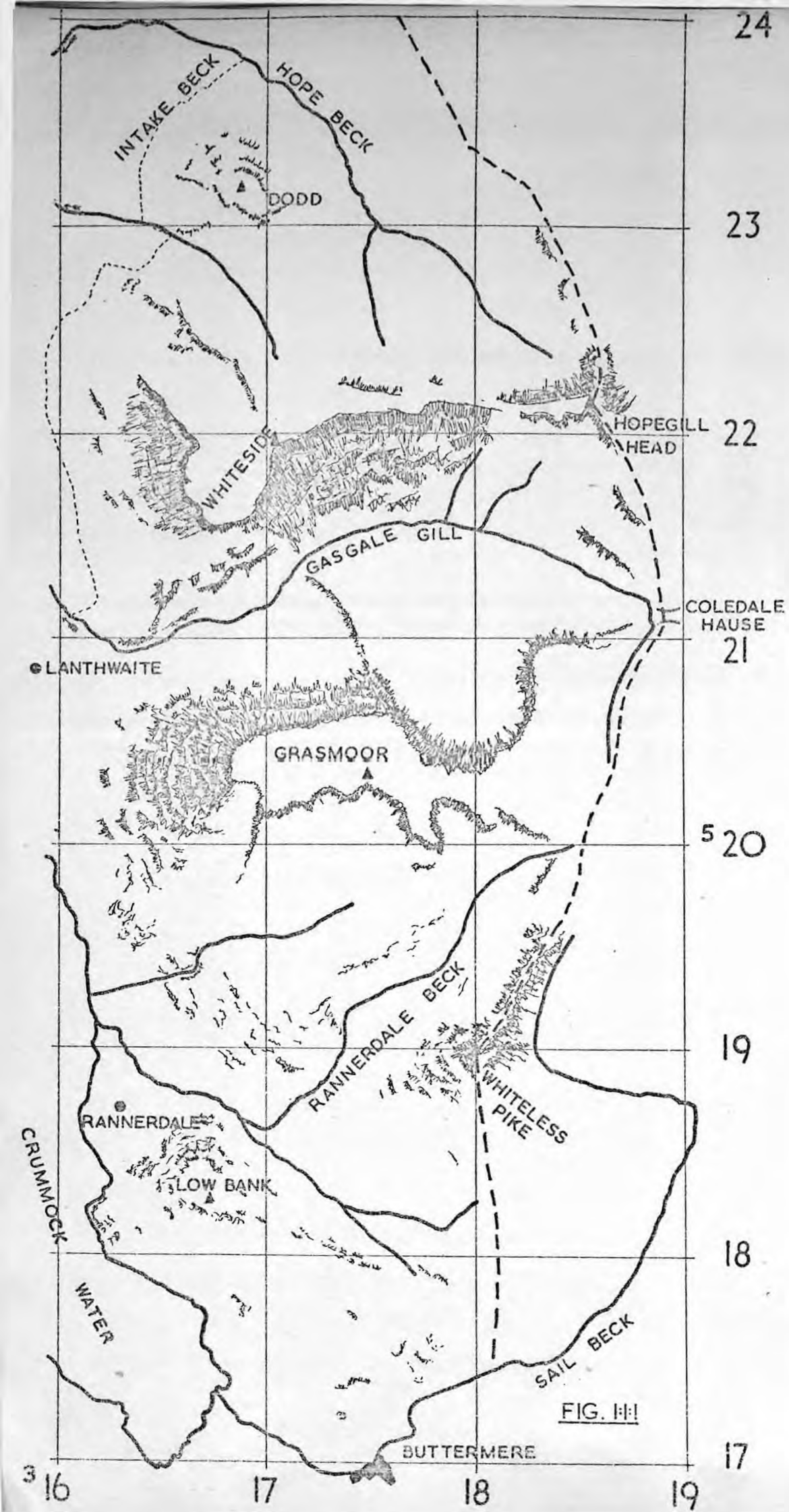

INTAKE BECK
HOPE BECK
DODD
WHITESIDE
HOPEGILL HEAD
GASGALE GILL
COLEDALE HAUSE
LANTHWAITE
GRASMOOR
RANNERDALE BECK
WHITELESS PIKE
RANNERDALE
LOW BANK
CRUMMOCK WATER
SAIL BECK
BUTTERMERE
24
23
22
21
20
19
18
17
16
17
18
19

FIG. III

Fig. 1:1:1. Map of thesis area, showing the boundary of the ground mapped (heavy broken line), the intake wall (fine broken line), and the main relief features.

cols that alone give low-level E-W access across the mountains.

The most northerly ridge, comprising Hopegill Head (2,525') and Whiteside (2,317'), runs E-W and is markedly assymetric, having steep, well exposed, much gullied crags, south to Gasgale Gill (Fig. 1:1:4), and gently dipping, unexposed, northern slopes down to Whinlatter Pass. This, more gentle, northern slope has been incised by consequent pattern streams, the most notable of which is Hope Beck. To the west of Hope Beck lies the small conical hill of Dodd, which is connected to the main mass of Whiteside by a gently inclined whaleback ridge (Fig. 1:1:5).

Grasmoor makes up the middle ridge and runs E-W on the southern side of Gasgale Gill. It is bounded on three sides by steep crags and ascent of the face of the mountain is very difficult. The lower slopes are completely obscured by scree and exposure at this level is very poor. It is remarkable that only a single stream flows off the mountain, although this is probably a function of the high permeability of the skirting scree slopes.

The southerly ridge comprises Wanlope (2,533') and Whiteless Pike (2,159'), with a lower grassy spur running down Whiteless Breast into Buttermere Village. The ridge runs S.S.W. - N.N.E. and again the lower slopes of the ridge are swathed in scree. The end of the ridge, however, which forms the western face of Whiteless Pike is almost clear of scree and is reasonably well exposed.

All but one of the major streams - Gasgale Gill; Cinderdale Beck, Rannerdale Beck and Sail Beck, flow down to the Crummock Valley sub-parallel to the three mountain ridges. The only exception is Squatt Beck, which flows N.W. and separates Whiteless Pike from the N.W. - S.E. striking ridge of Low Bank. This latter ridge has quite steep, reasonably exposed flanks, both S.W. to the Lake and N.W. to Squatt Beck, but the glacially-steepened crag end of the ridge (Rannerdale Knotts) is rocky,

Fig. 1:1:2. View south-easterly across the head of Crummock Water towards the steep western face of Grasmoor.

Fig. 1.1.3. View south from Loweswater Fells showing the Buttermere-Crummock trough valley, with Grasmoor (extreme left), Mellbreak (extreme right), Rannerdale Knotts (centre left) and the Red Pike Group (centre).

Fig. 1:1:4. View across Gasgale Gill from Grasmoor towards Whiteside and Gasgale Crags.
The deeply incised nature of the terrain is evident, and also seen is the trace of the Gasgale Fault - the marked linear feature running across the crags about half way up.







Fig. 1:1:5. View from Lorton south towards Dodd and Hope Beck. In the background is the Whiteside - Hopegill Head summit ridge.

Fig. 1:1:6 A. Looking north from Hause Point towards Grasmoor across the flat delta fan deposits of Rannerdale Beck.

B. View in the reverse direction, from Grasmoor towards Rannerdale Knotts, again showing the flat pastures of the Rannerdale Delta.







dangerously steep, and well exposed.

The only flat ground in the valley is the result of scree-fan or delta deposition, as at Lanthwaite, Rannerdale (Fig. 1:1:6), and Buttermere, and these are the only settlement sites in the area. Settlement here is of great antiquity, there being the remains of an Old English camp at Lanthwaite, and many of the present farm buildings date back to the 17th century. Often these represent only the latest structures to be built upon sites of considerably greater antiquity.

Intensive dairy farming and the growing of hay for winter feed provides the principal land use on the good alluvial soils of the valley bottoms. This gives way to free range sheep herding on the fellsides and plateaux tops, leaving the high crags and summits as the domain of raven and buzzard.

Section 1:2 History of Previous Research.

This review of the literature falls into two sections; the first deals briefly with the early history of geology in the Lake District, and then discusses in some detail the development of the present controversy concerning the Skiddaw Slates as a whole, and their relationships to the overlying Borrowdale Volcanics. The second section reviews the published work specifically related to the area of study.

The preliminary sub-division of the Lake District into two great sedimentary groups, separated by a thick sequence of volcanic rocks, the whole being disposed in a large open anticline, was first established in the early nineteenth century by Jonathan Otley. Individual investigation by many notable geologists, including Harkness, Phillips and Sedgwick, followed and this period of research was culminated in 1876 by the publication of the 1" and 6" Geological Survey maps of the Northern Lake District, and their attendant memoir, (Ward, 1876). In this memoir, Ward

estimated a thickness of about 12,000' for the Skiddaw Slates, and he thought them to be conformable with the Borrowdale Volcanics, thus rejecting the earlier suggestion (from Dakyns and Nicholson) of an unconformable junction.

More individual work followed, notably that of Marr, whose innumerable researches were summarized in his Lake District guide of 1916 in which he advocated a tectonic junction (lag plane) between the Slates and the Volcanics. Also dating from roughly this period was the pioneer work of Elles (1921 and 1933) in which she established a zonal succession for the Skiddaw Slates on the basis of their graptolite faunas. These, and other publications up to 1955, are discussed in the excellent reviews of Hollingworth (1955) and Mitchell (1956), but because they have no bearing on the present Lake District controversy, they will not be discussed further in this account. The present controversy in the Lake District dates from 1967 and the publication of Simpson's paper, since when many other papers have been published, often with extensive discussion, in which the many conflicting views have been aired.

Simpson (1967) was the first to deduce the polyphase nature of the Skiddaw Slate folding, and he described three phases of deformation - F1, F2, and F3, as follows: F1 was an intense deformation characterized by tight, upright folds trending E.N.E. - W.S.W., with accompanying penetrative cleavage (S1). F2 was intimately related to F1, and comprised open recumbent folds, with an associated horizontal fracture or crenulation cleavage, S2. F2 fold axes also trended E.N.E. - W.S.W., being coaxial with F1. Simpson interpreted both these deformation phases as pre-Borrowdale Volcanic in age since he thought the structures were truncated at the Slate-Volcanic junction which he consequently believed to be a major unconformity. The only end-Silurian deformation which he recognized was his third, F3, phase, characterized by open, generally steeply plunging,

folds and associated fracture cleavage. The folds varied considerably in orientation due to their being overprinted on a previously folded surface; around Buttermere the general plunge is to the N.E. or S.W. but elsewhere they plunge to the north or N.W.

Soper (1970) after a detailed examination of the Slate-Volcanic junction at several localities showed that the F1 and F2 fold phases could not be pre-Borrowdale in age. For example, field investigations at Hollows Farm indicated that the junction was roughly conformable, with minor disconformities only, whilst at Greenscoe Vent he found that the slate fragments incorporated into the Volcanics were uncleaved before inclusion, and that the only cleavage present passed through volcanic matrix and slate fragments alike. This was later confirmed by Knipe and Greave (1970). He also pointed out that it was most unlikely that a deformation which folded and impressed a strong penetrative cleavage on the Borrowdale Volcanics (in part) and much of the Silurian succession above (a thickness in excess of 20,000 feet), would produce such an insignificant deformation in the ductile Skiddaw Slates as Simpson's F3 was found to be. Also, it was strange that fold phases very similar in style and orientation to F1 and F2 were found in the Borrowdales and the Silurian rocks, and yet were equated with totally different styles and orientations of folding in the Skiddaw Slates.

Subsequently Helm (1970) published his account of the Skiddaw Slates of Black Combe, where he recorded six phases of deformation as follows:

D1

D2 pre-Borrowdale in age but N. - S. in trend.

D3

D4

D5 end-Silurian in age and Caledonoid (ENE-WSW) in trend.

D6

Helm, like Simpson, claimed that a major unconformity separated the Slates from the Volcanics. Helm's D4 - D6 structures were very similar to Simpson's F1 - F3 both in style and orientation, but they could not be correlated until the age of Simpson's structures was known. At Hollows Farm and on Saddleback an age has been worked out for structures that are assumed to be the lateral equivalents of Simpson's F1 and F2 on the basis of like style and orientation. This assumption is, in my opinion, a valid one because F1 and F2 remain constant over the whole of Simpson's very large area. At Hollows Farm, Simpson's S1 cleavage passes without deviation from Skiddaw Slate into Borrowdale Volcanics (Jeans, 1972) indicating at least a post-Borrowdale Volcanic age. On Saddleback, Soper and Roberts (1971) studied the relationships of the slate fabrics in the aureole of the Skiddaw Granite (dated at 399 $\pm$ 14 m.y. by Miller, 1962), where it could be seen that the andulusites in the inner hornfels zone were deformed by both the S1 and S2 cleavages, hence dating these structures as end-Silurian. This date for Simpson's F1 - F3 structures enables them to be correlated with Helm's D4 - D6, with which they are identical in style and orientation.

Simpson, however, still disputes this (in discussion to Moseley, 1972) and claims that he will not accept an age other than pre-Borrowdale for his F1 - F3 sequence until it can be proven within his area. The present work describes that evidence. For a detailed description see Chapters 3, 4, and 6, but in brief, the evidence rests upon deformed spots from the metamorphic aureole of the Dodd Dolerite, and a deformed tuff pipe on Whiteless Pike, both of which indicate a post-Borrowdale Volcanic, i.e. end-Silurian, age for Simpson's F1 and hence F2 and F3.

There is now the problem of Helm's D1 - D3 folds of pre-Borrowdale age: do they have any counterparts in the main outcrop, and hence is there an unconformity at the Slate/Volcanic junction, but of pre-F1 (Simpson) age? Roberts (1971) describes N. - S. folds in the Skiddaw

Slates of the Caldew Beck which he interpreted as pre-Bala in age, corresponding to the pre-Bala folds which Mitchell (1956) describes from the south of the Lake District. There is now some evidence (Newman, pers. comm.) that these northerly folds may be of much less importance than previously supposed, though this is disputed by Moseley (pers. comm.) from evidence in the Ullswater area. If, however, these pre-Bala folds are found to be of minor significance, then Roberts' N. - S. may well be pre-Borrowdale in age.

The present author (Jeans, 1971) has described a N. - S. pre-Borrowdale deformation, with no associated cleavage, from Newlands Beck; the Volcanics overlie the north-striking Skiddaw Slates unconformably and the end-Silurian cleavage, Simpson S1, is common to both. Further evidence of pre-Borrowdale earth movements is described by Nutt (1966) from his work in the Bampton Inlier, where he found andesite lavas resting unconformably upon upturned Skiddaw Slates, and the coarse Bampton Conglomerate, from low in the Borrowdale sequence, contained uncleaved mudstone clasts probably derived from the underlying Skiddaw Slates. The fact that the bulk of the clasts, though, were of volcanic material probably indicates that the conglomerate was just intra-Volcanic in age i.e. the deformation in the east slightly post-dated the initiation of the Volcanic episode.

More recently, Wadge (1972), has described an angular unconformity from Matterdale Beck and Causeway Foot, and Webb (1972) has published an account of N. - S. striking pre-F1 (Simpson) folding and overturning from the Buttermere area.

Thus we would seem to be left with a standard F1, F2, F3, sequence as defined by Simpson (1967) but of end-Silurian age, and then in certain areas (Newland Beck, Black Combe, Caldew Valley and Buttermere) there are found N. - S. folds, some of which can certainly be demonstrated as being

of pre-Borrowdale age. These folds, apart from those of Black Combe, do not have any associated cleavage developed.

The structural evidence, as summarized above, would seem to indicate that the junction between the Borrowdale Volcanics and the Skiddaw Slates is an unconformity, on the grounds of a pre-Borrowdale deformational episode. The distribution of the graptolite zones within the Skiddaw Slate outcrop (Wadge, 1972) supports the view of a regional unconformity, in that the entire thickness of D. bifidus and D. murchisoni zones which are exposed in the Eastern Lakes (Skevington, 1970, and Wadge, Nutt, Lister, and Skevington, 1969) is absent, or at least unrecorded, from the central Skiddaw outcrop, where no faunas younger than D. extensus have been recorded, (see Fig.1:2:1).

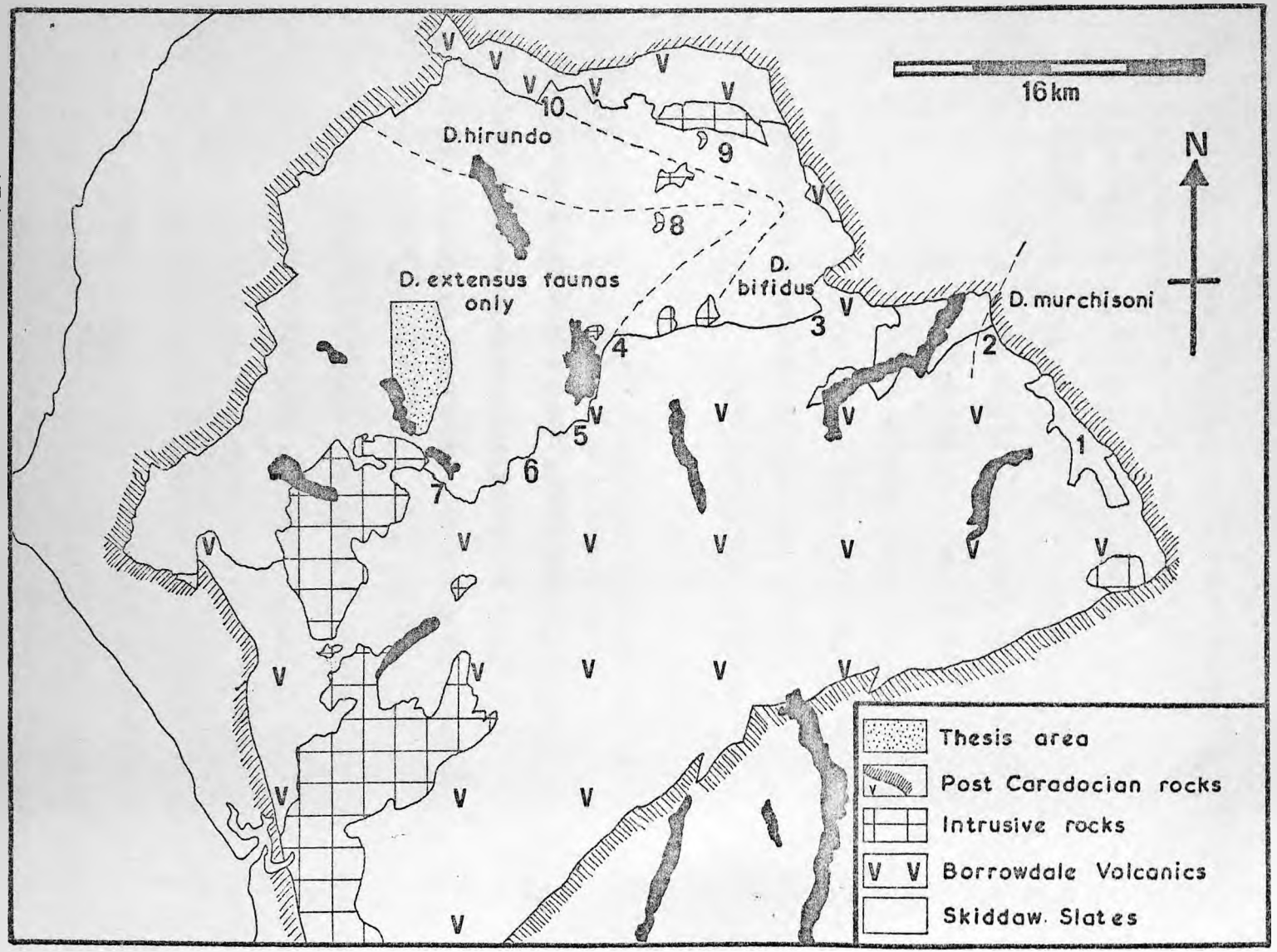
Recent investigations, however, indicate that not in every case are we dealing with a Skiddaw Slate/Borrowdale Volcanic unconformity; in places it may be intra-Volcanic and elsewhere intra-Slate. Downie and Soper (1972) describe an interbedded, conformable junction on Binsey, between the northern Eycott Group of volcanic rocks and the Skiddaw Slates, and the micro-palaeontological faunas obtained from slate horizons interbedded with the Lower Eycott Lavas indicated a basal D. bifidus age. Hence this northern volcanic succession, some 7,500' (2,500m.) thick, is likely to have been erupted prior to the folding, uplift and erosion which generated the angular unconformity and conglomerates described from the southern junction, which must be post-D. murchisoni in age. The converse situation, that of an intra-Slate unconformity, has been postulated to explain the Hollows Farm exposures (Jeans, 1970, 1972). These exposures reveal a completely conformable passage upward from dark blue slate, through pale grey-green ashy slate, into massive andesitic tuff, the whole sequence being cut only by the end-Silurian S1 cleavage. No unconformity is apparent; but field evidence from the surrounding areas (High Stile-Moseley, 1972); Newlands-Jeans, 1971; Causeway Foot-Wadge,

Fig. 1:2:1. Distribution of recorded graptolite faunas from the Skiddaw Slates, and the location of the critical areas mentioned in the text.

Key.

1. Bampton Inlier
2. Tarn Moor
3. Materdale
4. Causeway Foot
5. Hollows Farm
6. Newlands
7. High Stile
8. Skiddaw Granite
9. Caldew Beck
10. Binsey.

FIG. 1:2:1


16km
N
D.hirundo
D. extensus faunas only
D. bifidus
D. murchisoni
1
2
3
4
5
6
7
8
9
10
Thesis area
Post Caradocian rocks
Intrusive rocks
Borrowdale Volcanics
Skiddaw. Slates

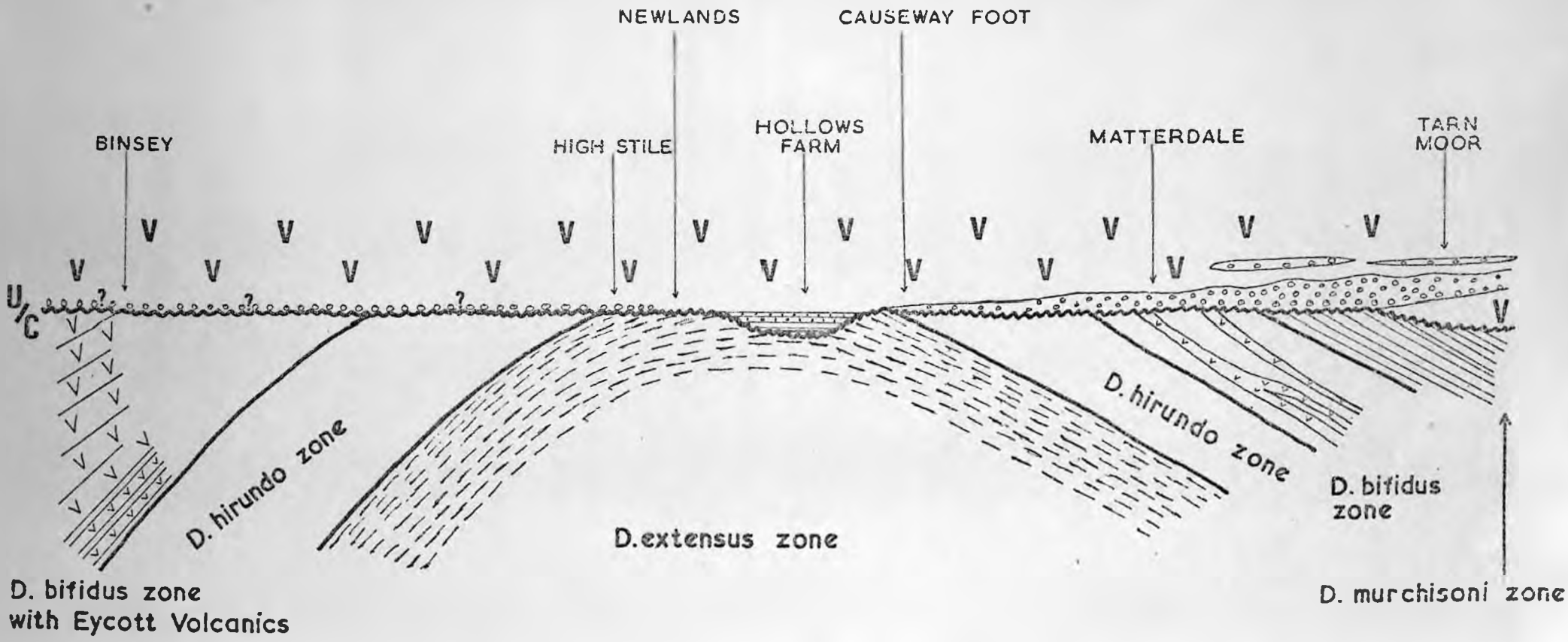
1972) suggests that it should be there. The problem is resolved if one accepts that the initiation of the Borrowdale Volcanic episode could be marked by a pelitic phase prior to the advent of volcanic material, in which case the unconformity would lie wholly in pelite. In view of the intense deformation and contact strain suffered by the pelites due to their proximity to the massive Volcanic horizons, such an unconformity would be very difficult to discover, though geochemistry and micro-palaeontology would probably be most effective in its determination.

It will be apparent that the junction problem has still not been solved completely, and that the controversy is still in a very involved state. During the last five years there has been a great wealth of literature on the subject, especially in discussions to papers, and it has been most recently summarized by Moseley (1972). However, with the ever-increasing data at our disposal, the possible solutions are being slowly refined, and Fig. 1:2:2 is offered as my interpretation of the situation at the present state of play.

Section 1:3 History of Research: Buttermere area.

The first details of the present area to appear in the literature were in Ward's (1876) memoir on the Northern Lake District. He described the area in cross-section from north to south and mentioned the "intrusive trap" of Dodd; "gritty beds" of Whiteside, highly inclined, dipping under "black splintery shales" in Gasgale Gill, and then reappearing as the "sandy beds" of Grasmoor. He observed that these sandy beds became more contorted to the south and passed under more shaly beds once again in the vicinity of Buttermere. He also recognized that the intersections between bedding and cleavage were highly complex, there being "no apparent uniformity". This single point is a good measure of the skill and quality of observation in Ward's work; the complexity of the bedding-cleavage intersections is the clue which belies the presence of at least

FIG 122



DIAGRAMMATIC SECTION ACROSS N. LAKE DISTRICT IN PRE-CARADOC TIMES, SHEWING RELATIONS AT THE JUNCTION BETWEEN SKIDDAW SLATES AND BORROWDALE VOLCANICS.

two phases of folding in these rocks, yet the significance of this conclusion was not to be recognized for nearly one hundred years, until Simpson published his analysis of the Skiddaw Slates in 1967.

Between 1876 and 1955 there was no work of any importance published on this area, a fact evidenced by Smith's (1965) excellent bibliography of Lake District geology. Then, in 1955, Rose published his account of the Keswick-Buttermere area, in which he reduced the earlier successions of Dixon (1925) and Eastwood et al. (1931), (Fig. 1:3:1), both erected on ground to the north and west of the present area, into a simple bipartite division having a lower, psammitic unit, the Loweswater Flags, and an upper, pelitic unit, the Mosser-Kirkstile Slates. The rocks were disposed in open, anticlinoria-type folds with the often-faulted outcrops of the Loweswater Flags marking the anticlinal cores (See Figs. 1:3:2; 1:3:3). He recognized the presence of a contact-metamorphic aureole in the area, which caused bleaching and spotting of both the Loweswater Flags and the Mosser-Kirkstile Slates. This transgression of the lithological boundaries by the aureole was most important because it challenged the earlier account of Dixon (1925), who thought the light-coloured, spotted rocks constituted a separate lithological unit, which he called the Blake Fell Mudstone.

Jackson (1961) only covered the northern half of the area, but increased Rose's succession with the addition of a lower, pelitic unit (Hope Beck Slates) and an upper psammitic unit (Latterbarrow Sandstone). He also reported the presence of contact metamorphic, Blake Fell Mudstone facies, with the finding of tourmaline needles and sericite-andalusite-quartz assemblages in these rocks. Also reported but not shown on his map were facies variations and minor disconformities within the two middle groups. Rose (1969) maintains that the Hope Beck Slates do not exist as a separate group underlying the Loweswater Flags, but that the slates of

EVOLUTION OF THE STRATIGRAPHIC SUCCESSION IN THE SKIDDAW SLATES

SIMPSON (1967)	JACKSON (1961)	ROSE (1955)	DIXON (1925)
BORROWDALE VOLCANICS			
LATTERBARROW SANDSTONE			
SUNDERLAND SLATES			
WATCH HILL GRITS	LATTERBARROW SANDSTONE		
MOSSER SLATES	MOSSER-KIRKSTILE SLATES	MOSSER-KIRKSTILE SLATES	MOSSER SLATES
LOWESWATER FLAGS	LOWESWATER FLAGS	LOWESWATER FLAGS	LOWESWATER FLAGS
KIRKSTILE SLATES	HOPE BECK SLATES		KIRKSTILE SLATES
BLAKE FELL MUDSTONES			BLAKEFELL MUDSTONE
BUTTERMERE FLAGS			
BUTTERMERE SLATES			
TOTAL 30,000'	TOTAL 6,000'		

FIG. 1:3:1

Hope Beck are part of the Mosser-Kirkstile succession, and overlie the Flags.

A second paper by Jackson (1962) examined the zonal succession of the Skiddaw Slates first established by Elles (1921 and 1935) on the basis of the graptolite faunas. He confirmed Elles' downward sequence of D. bifidus Zone, D. hirundo Zone, and the upper three sub-zones of the D. extensus Zone, but below the D. deflexus sub-zone he found no graptolites at all. The localities which Elles quoted for her sub-deflexus zones (Dichograptus and Bryograptus kjerulfi), when checked by Jackson, were found to be characteristic of D. deflexus and higher, so there is no evidence that the Skiddaw Slates are older than Arenig in age.

This earlier work was basically lithological in approach, and it was not until 1967 with the publication of Simpson's work that the Skiddaw Slates received a modern structural treatment. As mentioned in Section 1:2 Simpson described three phases of deformation in this area, all three being represented by major and minor structures. This sequence of minor structures was substantiated by the present work, but because of the absence of Borrowdale Volcanic rocks from the area it was impossible to date these structures directly. By analogy with other areas (Jeans (1971); Soper and Roberts (1971)), I regard Simpson's F1-F3 structures as being wholly end-Silurian in age, and there has hitherto been no descriptions of any earlier structures from this area, though Webb (1972) describes N. - S. folds which pre-date Simpson F1 structures from the area to the S.E. around Hassness. During the present study I found much evidence of earlier, pre-F1 deformation and this will be described in full in Chapter 5.

The major structures described by Simpson (1967) from this area are shown on Figs. 1:3:2 and 1:3:3, and comprise the Braithwaite Synform (F1 age), the Gasgale Slide (F1), the Crummock Recumbent Synform (F2), and

Fig. 1:3:2. Map of the area showing the conflicting interpretations of Rose (1955) and Simpson (1967).

Key.

B.S. = Braithwaite Syncline

G.S. = Gasgale Slide

C.S. = Crummock Recumbent Synform

B.A. = Buttermere Anticline

Wh = Whiteside

H.H. = Hopegill Head

C.H. = Coledale Hause

G = Grasmoor

W.P. = Whiteless Pike

L.B. = Low Bank

INSET. Location map of N.W. England.

Lakes in black. Thesis area shaded.

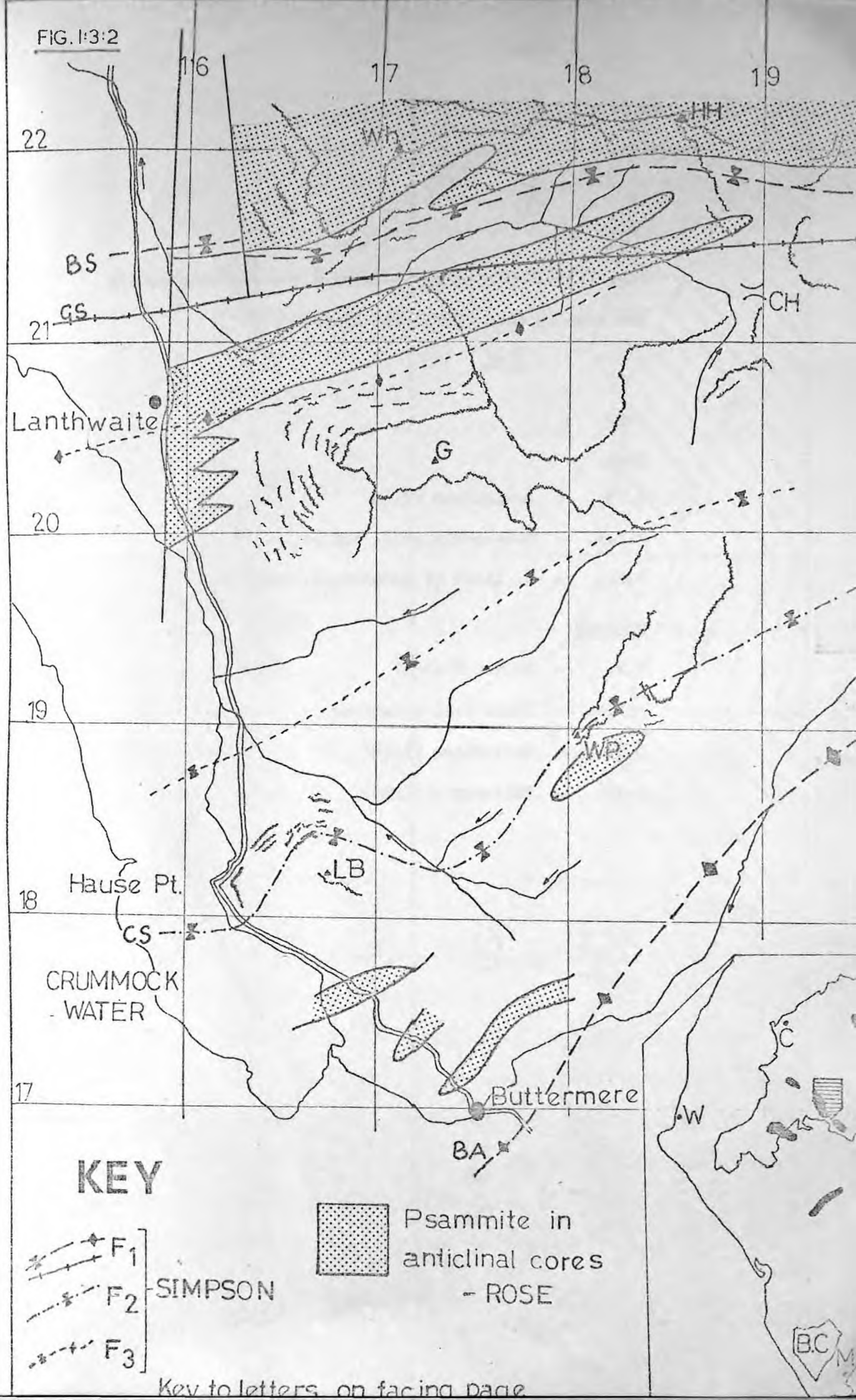
W = Whitehaven

C = Cockermouth

M = Millom

B.C. = Black Combe

FIG. 1:3:2



Key to letters on facing page

Fig. 1:3:3. Comparison between N – S sections across the area by Rose (1955) and Simpson (1967).

Key.

Rose.

L.F. = Loweswater Flags

M.K.S. = Mosser-Kirkstile Slates

N.M. = N. limit of metamorphic aureole.

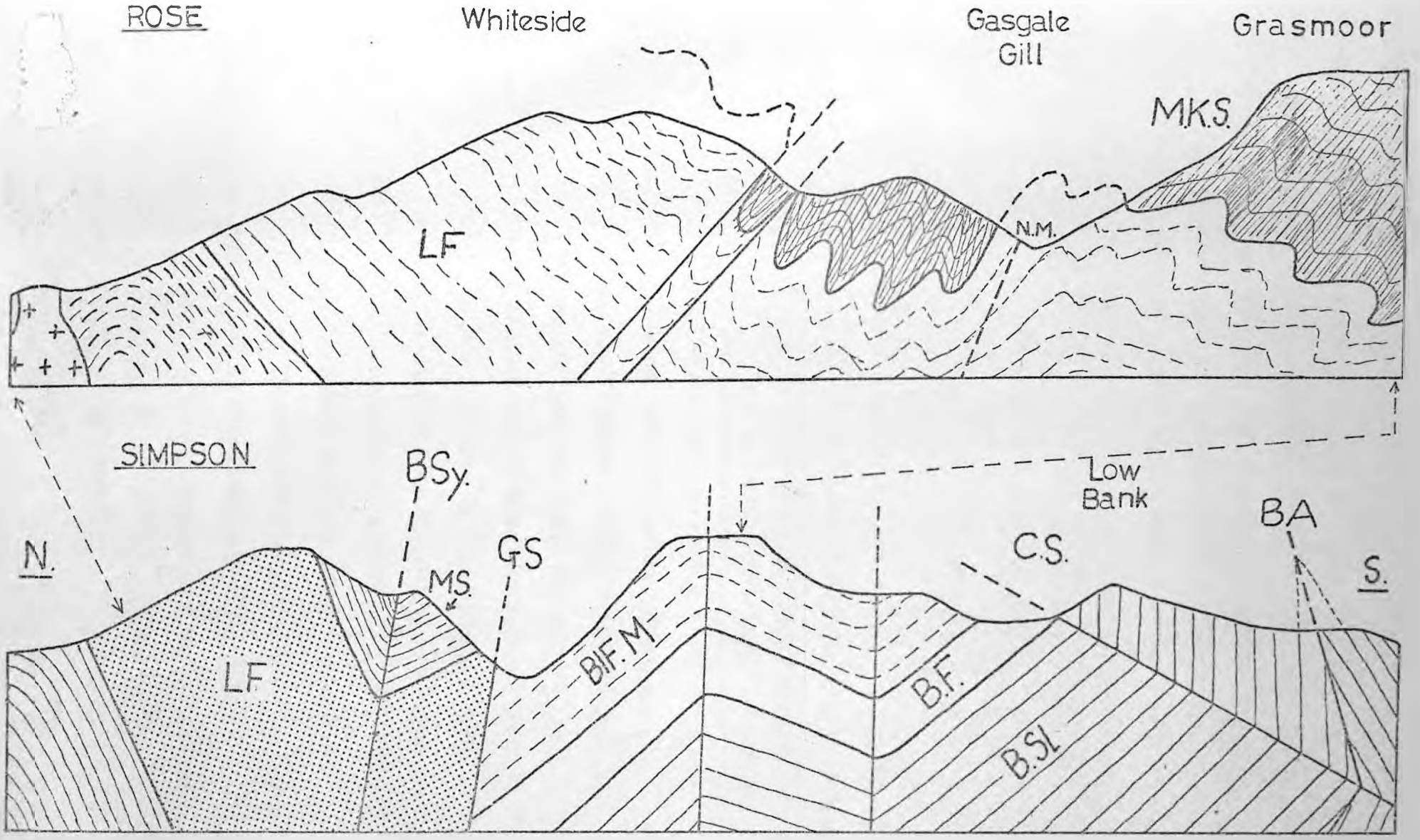
Simpson.

M.S. = Mosser Slates

BlF.M. = Blake Fell Mudstones

B.F. = Buttermere Flags

B.Sl. = Buttermere Slates.



Key to letters on facing page

the Buttermere Antiform (F1). The Braithwaite Synform occupies a position corresponding to one of Rose's pelitic infolds, and the presence of a synformal structure was also substantiated by myself. The existence of the Gasgale Slide is difficult to confirm; much of the ground along its trace is unexposed, but where exposure is good, as at the mouth of Gasgale Hill, little evidence of a major dislocation is seen.

It is possible to be more definite about the Crummock Recumbent Synform, for it does not exist, the rocks around Hause Point being disposed in an upright anticline, much complicated by the presence of earlier, pre-Simpson F1, folding.

The Buttermere Antiform is also rejected for it is possible, on the basis of fold styles and younging directions (see Section 2:4), to prove the presence of a major synclinal structure running E.N.E. just south of Buttermere Village. This agrees with the evidence to the south of my area (B.C. Webb, personal communication) which indicates that the psammite bands at Moss Force (193175) are younging to the north as they disappear downwards, suggesting that a synclinal area lies in that direction.

It is interesting to note that Simpson's (1967) stratigraphic succession is by far the most complex ever published for this area and shows a return to the earlier work by such as Dixon and Eastwood (both op. cit.) in resurrecting the Blake Fell Mudstones as a lithological unit. This is in contrast to the earlier work of Rose and Jackson who thought the Mudstones to be the thermally metamorphosed equivalents of the Mosser-Kirkstile Slates. Since Simpson's major structures are now in doubt as discussed above, his stratigraphic subdivisions must also now be less credible, and this aspect will be discussed in greater detail in Chapter 2. However, the basic structure of Simpson's succession with two psammitic units separated by three pelitic units does accord well with the lithological succession erected in the present study.

As mentioned above, there have been found within this area folds which pre-date Simpson's F1 folds, which are here taken to be of end-Silurian age. Henceforth in this thesis, therefore, these early folds will be styled F1, thus relegating Simpson's three phases of folding to F2, F3, and F4, respectively, and this usage will be adhered to for the remainder of this work.

CHAPTER TWO

STRATIGRAPHIC SUCCESSION.

The area mapped lies within the main northern outcrop of the Skiddaw Slates and this is the only rock group present, excluding the minor igneous intrusions (Chapter 3) and Quaternary deposits. In Section 1:3, the conflicting interpretations of the succession were outlined (Fig. 1:3:1) and this, in a way, pinpoints the difficulties of working in these rocks. These are, the lack of distinct marker horizons, the lack of palaeontological control, and the intense polyphase deformation. The latter can render the pelitic units almost wholly intractable from the stratigraphic standpoint. Lithologically, the Skiddaw Slates consist of a great thickness of shales, mudstones, flags, and psammites, with each rock type passing imperceptively into the next, and this is the characteristic assemblage of the turbidite-greywacke suite, indicating deposition under geosynclinal conditions. The palaeogeographic implications of this will be discussed further in Section 2:5.

When mapping initially, an attempt was made to distinguish four discrete lithologies - slate, mudstone, flags, and psammites - but this was later abandoned when it became clear that all the lithostratigraphic divisions could be delimited solely by the use of pelitic and psammitic divisions. This removed the great difficulties involved in trying to make consistent distinctions in the field between, for example, slates and mudstones, or flags and psammites, where only tenous, gradational boundaries exist.

On the basis, therefore, of this bipartite division into psammitic and pelitic lithologies, a stratigraphic succession for the area was worked out, as follows:

Buttermere Slates	1,000'+ *
Whiteless Pike Flags	800 - 1,000'
Mosser-Kirkstile Slates	1,500'
Loweswater Flags	3,500'
Hope Beck Slates	400'+
	7,400'+

The succession therefore consists of three pelitic divisions separated from each other by two intervening psammitic units. Existing lithostratigraphic names have been used wherever possible to avoid further encumbering an already weighty literature, but in one case (Whiteless Pike Flags) this proved impossible because no earlier name was applicable, and in another, the case of the Buttermere Slates, the existing name was used only after careful consideration. The problem was that Simpson (1967) had introduced Buttermere Slates as the lowest unit in his succession (see Fig. 1:3:1) whereas, in fact, they represent the highest unit in the succession, thus creating room for confusion. However, since the Buttermere Slates can be proven to overlie the Whiteless Pike Flags both by superposition and way-upness, the objections were overruled and the existing name preferred, especially since it accurately describes the type locally, and, in fact, refers to those rocks described by Simpson as Buttermere Slates, his difficulty being that of faulty structural interpretation.

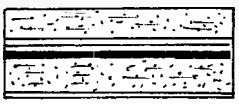
The position of these rocks within the faunal succession of the Arenig-Llanvirn stages is only partially known; this from the work of Jackson (1962) previously mentioned in Section 1:3. Jackson described the zonal position of those stratigraphic units up to the Mosser-Kirkstile Slates, but no further; unfortunately, no fossils were found during this study, so the exact zonal position of the higher units can not be assessed, but I would expect the Buttermere Slates to approach a high D. hirundo or a low D. bifidus age. The known correlation of the litho-

* Thicknesses shown take no account of tectonic thickening; i.e. are uncorrected.

Fig. 2:1:1. Stratigraphic succession as erected by the present study, equated with the known occurrences of the graptolite zones (mainly after Jackson, 1961).



Massive graded psammite units.



Interbedded flaggy lithology.



Pelites.

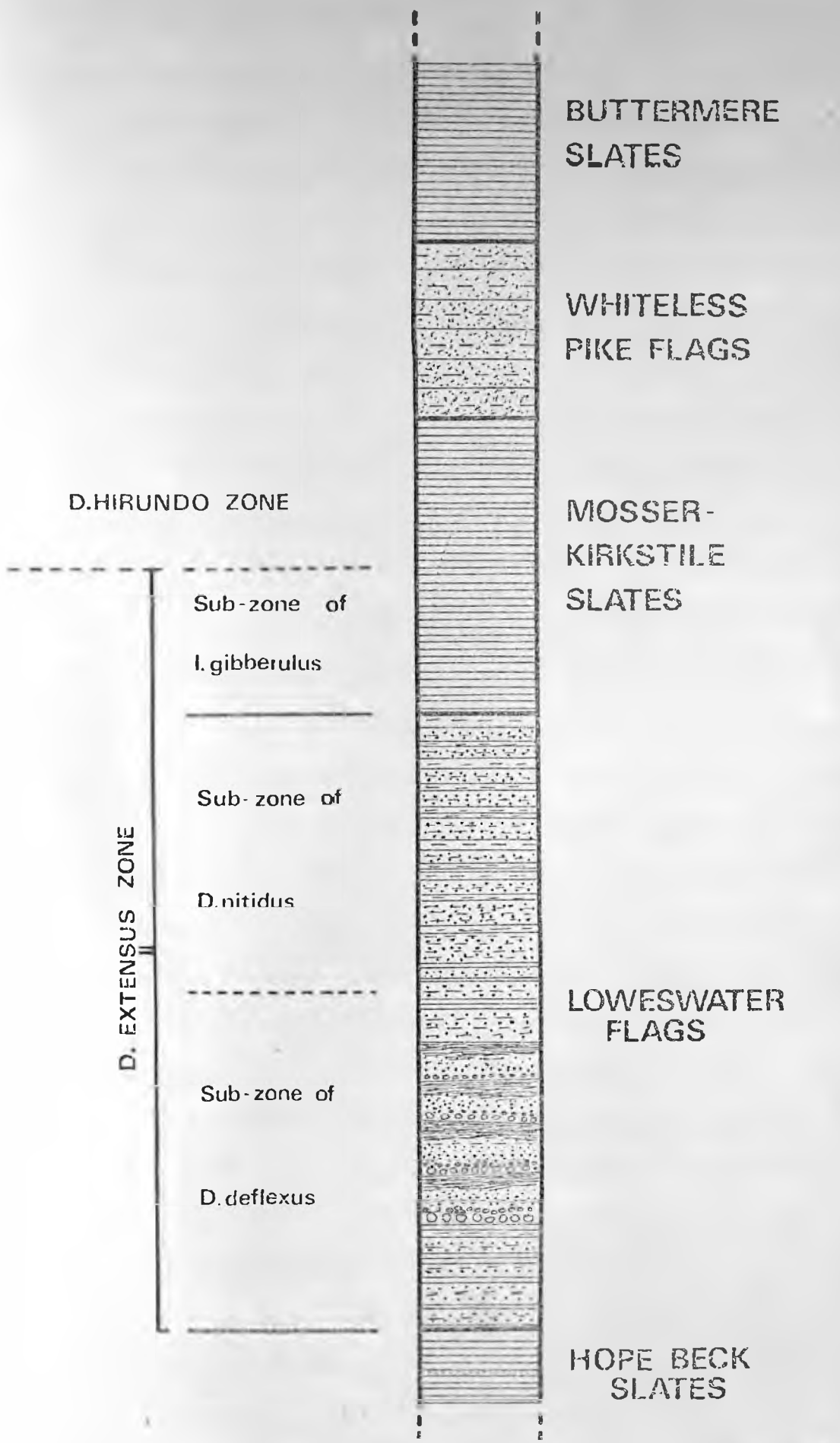

BUTTERMERE SLATES
WHITELESS PIKE FLAGS
MOSSER-KIRKSTILE SLATES
LOWESWATER FLAGS
HOPE BECK SLATES
D.HIRUNDO ZONE
D. EXTENSUS ZONE
Sub-zone of
I. gibberulus
Sub-zone of
D. nitidus
Sub-zone of
D. deflexus

FIG. 21

stratigraphic units with the graptolite zones is shewn in Fig. 2:1:1.

The remainder of this chapter will be made up as follows: Sections 2:2 and 2:3 will respectively describe the lithological characteristics of the pelitic and psammitic rock units outlined in the succession above. This is considered to be the most practical method because if described completely as individual formations, then the great similarity of all the psammitic and pelitic units would result in much undue repetition. The penultimate section (2:4) considers the outcrop distribution of the formations, whilst in Section 2:5 an outline of the inferred depositional environment is suggested and this is related to a palaeogeographic reconstruction of the Southern Caledonides. This last section draws heavily upon results of the geochemical investigation of the sediments of the Grasmoor area, which are presented in full in Appendix C.

Section 2:2 The Pelitic Groups.

Within the present area three pelitic groups have been distinguished; the Hope Beck Slates; the Mosser-Kirkstile Slates; and the Buttermere Slates. The term "slates" requires some comment because it is rarely that a true slaty condition is developed. "Slates" is used in the context of Hope Beck Slates or Buttermere Slates, for example, on the basis of historical precedent, and merely describes a pelitic succession of rocks. These are made up of varying proportions of unbanded dark shales (Kirkstile facies of the Geological Survey, 1968) and banded mudstones and silts (Mosser facies, op. cit.), with the latter generally in preponderance. There is a complete gradation between the Kirkstile and Mosser facies, based on the increasing proportion of silt-sized material in the latter and, with further increase in the silt fraction, the Mosser facies shews a complete gradation into the flaggy lithology.

The three major pelitic units are impossible to distinguish on lithological grounds because of their great similarity, the only difference

between them being a direct result of either metamorphism or deformation. This view is reflected by earlier work, for Jackson admits (1961) that the Hope Beck Slates can not be distinguished from the Mosser-Kirkstile Slates unless their relationship to the intervening Loweswater Flags are known, and Simpson (1967) also admits the similarity between his three slate divisions. Thus it is only by structural mapping and on such criteria as way-upness and super-position that the succession can be erected, and the evidence upon which this has been done is presented in Section 2:4.

The characteristics of the three slate units are now briefly described:

i) Hope Beck Slates. The Hope Beck Slates were first defined by Jackson (1961) from the area of Hope Beck itself. They formed the lowest unit of his succession, as they do in mine, and they have been found on each occasion to be completely devoid of macro-fossils. However, work is currently in progress on the micro-fossils in an attempt to fix their zonal position (Downie, personal communication).

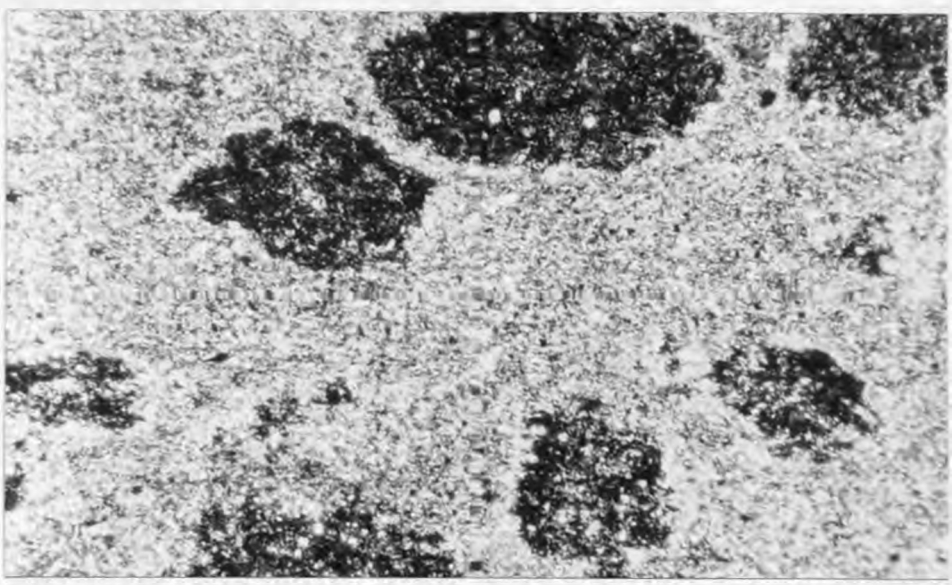
The Hope Beck Slates are dark blue-black in colour, weathering to pale brown, and are generally fissile parallel to bedding. They are dominantly composed of finely banded mudstone, though with very occasional, thin (2"), bands of silty psammite. Bedding is usually easy to distinguish, being marked by alternations of lighter, silty bands and darker, more pelitic, bands (Fig. 2:2:1). In the field these micro-beds were measured at roughly 2-3mm in thickness, but under the microscope some are seen to be much finer than this, averaging 0.7-1.0mm in thickness. The mudstones, when seen microscopically, are found to consist almost entirely of detrital quartz, sericite, disseminated pyrite, chlorite and an irresolvable matrix. The sericite and chlorite also show considerable palimspastic recrystallization parallel to the bedding, (Fig. 2:2:2), thus

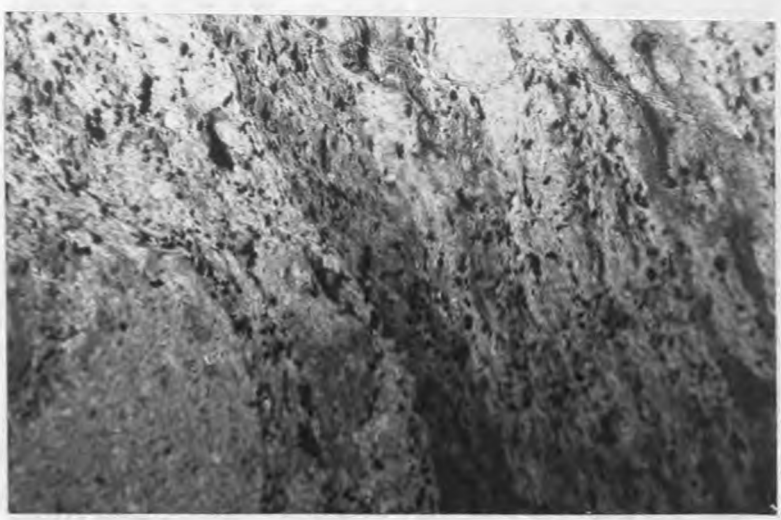
Fig. 2:2:1. Field appearance of the Hope Beck Slates (G.R. 172.236).

Fig. 2:2:3. Anhedral spots (?chloritic) in the Mosser-Kirkstile Slates of the Grasmoor area. Photo taken under crossed nicols (X Nicols) shows their cloudy opaque-appearance (x25).

Fig. 2:2:2. Mimetic recrystallization of sericite and chlorite parallel to bedding. (X40; Plane - Polarized Light (PPL)).







explaining the strong tendency of these rocks to be fissile along this direction. As mentioned above, the banding is of the order of 1mm in thickness, but it shows poor lateral continuity, with both pelitic and silty bands tending to pinch out and merge. There are no signs of eroded tops to the pelitic layers, to suggest some form of cyclic deposition with grading as recently described by Piper (1972).

ii) Mosser-Kirkstile Slates. The first description of the Mosser-Kirkstile Slates from this area was that of Rose (1955) who combined the Mosser Slates and the Kirkstile Slates of Dixon (1925) into a single lithological unit lying above the Loweswater Flags. Unfortunately the names "Mosser" and "Kirkstile" have facies connotations, as outlined in the introduction above, and these are not equally represented in the combined unit, it being dominated by the banded muds and silts of the Mosser facies.

In outcrop, the Mosser-Kirkstile Slates vary quite widely in colour from dark grey and dark blue-black to light grey, and grey-brown. The lighter coloured rocks are restricted to the centre of the area, being found from Gasgale Gill in the north, to Rannerdale Knotts and Whiteless Pike in the south, (see Fig. 4:2:1). Outside this area, both to the north and south, the rocks are wholly dark in colour, with the bedding forming the principal plane of parting, and the rocks are very similar to the Hope Beck Slates already described, both in the field and under thin section.

The mudstones of the central area are quite different from those to the north and south. They are light greyish in colour and freshly broken surfaces often show a marked dark spotting (Fig. 2:2:3); they have a brittle siliceous appearance and more often break across the bedding than along it. Cleavage is present as a weak fracture cleavage, often sufficiently widely separated to be termed joints. The features described

above are indicative of quite a brittle rock, yet on some joint faces very complex fold patterns can be seen, suggesting that the rocks were once much more ductile. ~~than they are now.~~

As far as major constituents are concerned, there is no noticeable mineralogical difference between the two types of Mosser-Kirkstile Slates. Within the silty beds the quartz grains are sub-angular and range in size from 100u - 50u, with a mean size of roughly 70u. Crystals of sericite up to 1.5mm in length have grown in the silty bands; they have grown with little interference and do not show the same degree of parallel orientation as the much smaller sericites in the pelitic bands. This is probably a function of the lower ductility of the silty bands. Chlorite, too, formed as comparatively large (500u) elongate crystals.

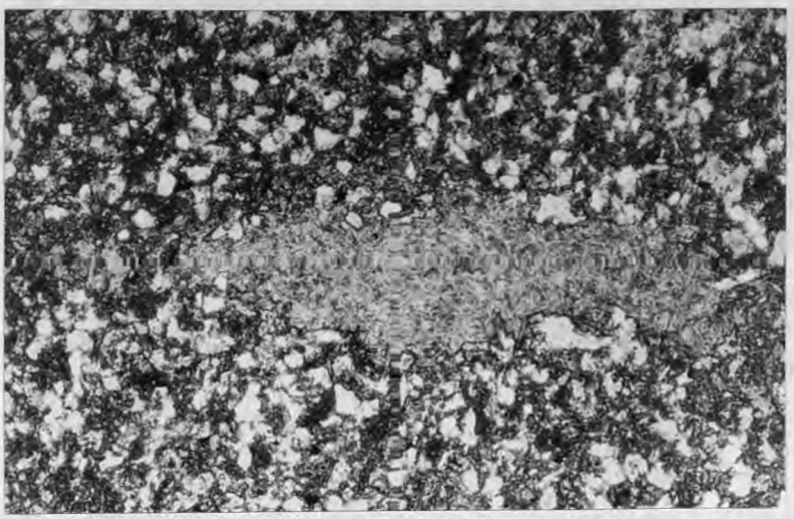
Within the pelitic bands the major component is sericite, presented as a finely crystallized lepidoblastic felt with a high degree of orientation parallel to the bedding. The quartz grains are smaller (30-50u) and there are no discrete crystals of chlorite as in the silty bands. Instead, tabular chlorite (variety penninite) is present in veins up to 800u in width and 10mm long; most veins parallel the bedding but the larger ones are transgressive (Fig. 2:2:4). Penninite also occurs in elongate aggregates of small interfering crystals of tabular habit (Fig. 2:2:4). Disseminated pyrite blebs are common to both silty and pelitic bands and they are evenly distributed throughout. They are only 20 - 50u in diameter and are too small for their structure to be determined optically.

Within the mudstones of the central area there are found small elliptical spots, averaging 500u in length, and they appear almost irresolvable in thin section, with very little reaction to transmitted light, due to their high content of inclusions. It is these spots that form the basis of the evidence for the proposed granitic body thought to

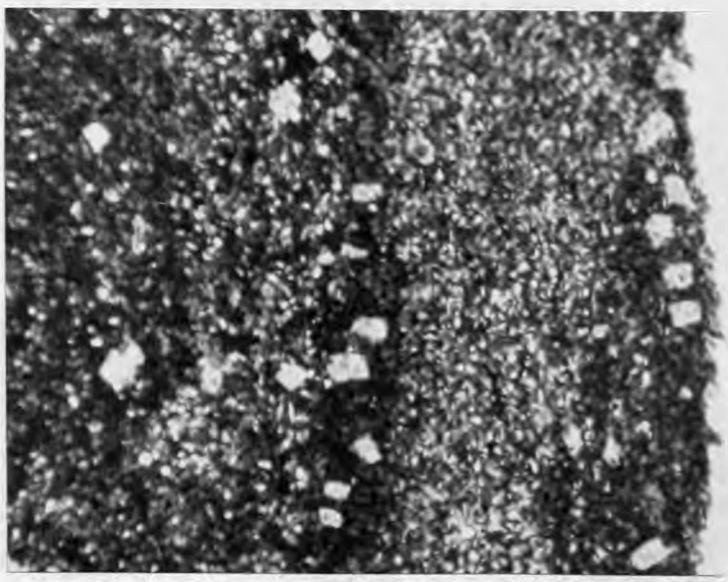
Fig. 2:2:4 A. Elongate aggregate of small interfering sheafs of chlorite. (x40. PPL)

B. Quartz-chlorite vein cutting bedding. Note mimetic recrystallization of sericite parallel to bedding. (x35 . PPL)

Fig. 2:2:5. Example showing the restriction of the metamorphic spotting to pelitic horizons, being absent where layers are silty or psammitic. (x25 . PPL)







underlie Grasmoor, and a more detailed description of them will be given when the contact metamorphism due to this granite is discussed in Section 4:2. The spots are confined to the finer pelitic bands, being absent when the rock is at all gritty, as in the coarser silty bands (Fig. 2:2:5).

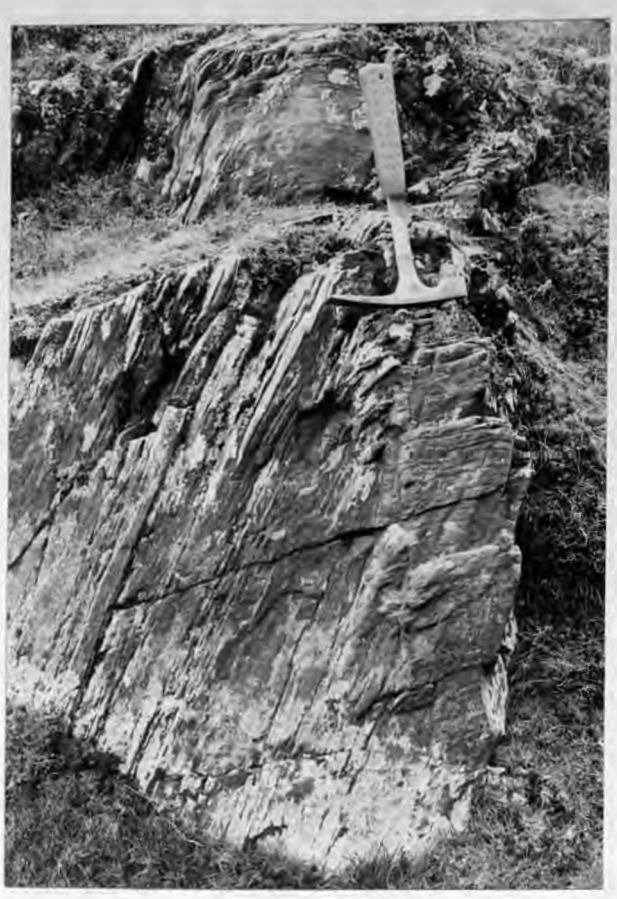
iii) Buttermere Slates. The Buttermere Slates represent the highest unit of the succession erected, and they, of the three pelitic units most closely approximate to the dark blue slates of the Kirkstile facies, especially in the vicinity of Buttermere itself. The slaty rocks of Buttermere pass downwards into banded mudstones and silts of the Mosser facies, which make up the lower part of the Buttermere Slates unit. These mudstones and silts are very similar to the non-metamorphosed rocks of the Mosser-Kirkstile and Hope Beck Slates, and it would be impossible to distinguish them in hand specimen unless their relationships to either of the psammites were known.

Bedding in the slaty facies is sometimes marked by slight colour banding but unless the cleavage is parallel to it, the bedding is often impossible to discern (Fig.2:2:6). The slates are highly deformed, being tightly folded, well cleaved and showing one or more penetrative lineations; assymetric minor folds are extensively sheared parallel to the longer limbs.

In thin section, the rocks are seen to be composed of thin alternating laminae of silt and clay-grade sediment. The clay-grade bands are largely composed of sericitic mica showing a high degree of bedding - parallel orientation. The silty bands consist of very fine sub-angular quartz grains (25u) set in an irresolvable matrix which is now almost completely made over into sericite and chlorite. Also present within the groundmass are very small anhedral aggregates of a light brown-green mineral of quite high relief which is either tourmaline or epidote. The slaty bands are almost wholly recrystallized into a lepidoblastic felt of

Fig. 2:2:6. Field appearance of Buttermere Slates, note strong "bedding cleavage".

Fig. 2:2:7. A and B. Siliceous concretions (A. illustrates the shape of the "hamburger bombs").







sericite mimetic along, and in parallel orientation with, the bedding. Very fine quartz grains and larger chlorite aggregates make up the matrix of the slates. Finely disseminated throughout the whole rock are tiny crystals of pyrite, often elongate, and always with a random orientation, i.e. the mimetic texture of the sericite is not reflected by the pyrite crystals.

iv) Nodular Structures. Nodular structures were found to be quite common within the pelites of the Mosser-Kirkstile and Buttermere Slates. As defined by their shape, there were three different types of nodule although two of these were each represented by only a single example. The usual type was of ellipsoidal shape and ranged in size from 6" - 24" in maximum dimension, which was usually parallel to bedding. The short axis (bedding-normal) was more constant at 4" - 8" and the nodules had an oblate rather than prolate three-dimensional form. Those which most clearly approximated to the spherical shape were characterized by equatorial cracks running parallel to the bedding, hence my field name for them of "hamburger bombs" (Fig. 2:2:7). All but two of the concretions were of this type, the exceptions being a kidney-shaped block, 12" x 9", found above Buttermere cross-roads, and a spindle-shaped body (Fig. 2:2:8) measuring 12" x 6", found above the mouth of Gasgill Gill. These looked quite volcanic in origin; both were found to have a pattern of concentric cracks around their margin, and both lay with their long axes oblique to the bedding, this in marked contrast to the equatorial cracks and the bedding-parallel orientation of the "hamburger" nodules.

In thin section, the two types of nodule (hamburger and volcaniform) are also quite different. The "hamburger" type is largely composed of very fine crystals of quartz in an interlocking, granophyric texture (Fig. 2:2:9); zircon, tourmaline and pyrite occur and there are occasional plates of secondary calcite. In some, though not in all, there is

Fig. 2:2:8. Volcani-form "bombs".

B. In fact the only "bomb" to have a composition compatible with a volcanic origin. (G.R. 178.170).

A. Though spindle shaped this "bomb" was analysed as a siliceous concretion (G.R. 165.211).





evidence of what is taken to be remnant bedding. The accessory minerals and the remnant bedding imply a sedimentary origin for these nodules.

The volcaniform types are much darker and more difficult to resolve in thin section. The Gasgale specimen has a fine granophyric texture and consists dominantly of quartz, with accessory iron ore, sericite and oxydized chlorite. The groundmass is dusty-black, isotropic, and completely irresolvable. More variation in grain size is shown by the Buttermere specimen. The groundmass is again very fine-grained, dark and difficult to resolve. Secondary plates of calcite are widely developed, often intergrown with pools of clear quartz showing well developed strain shadows. Bright blue tabular chlorite is found growing in cavities between the quartz and calcite plates, (Fig. 2:2:10) and very small, highly altered crystals of plagioclase may be present.

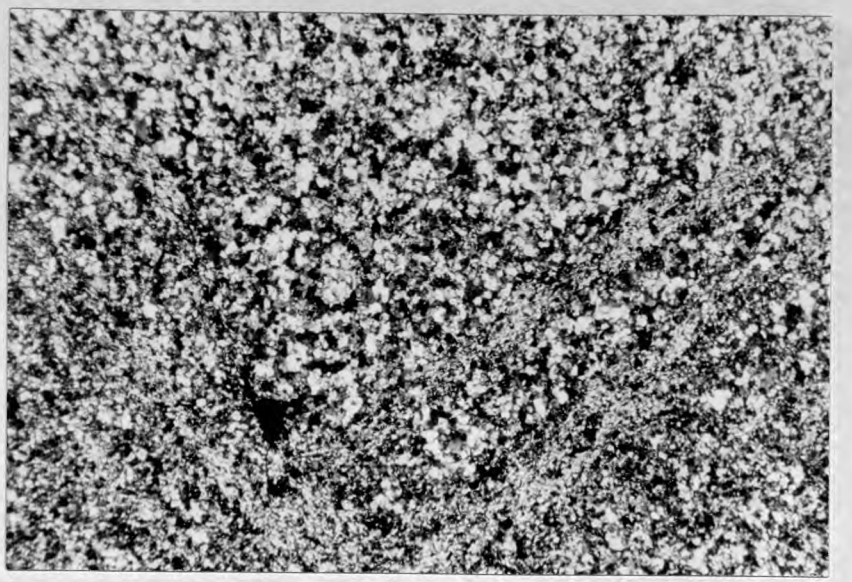
A further point of difference between the volcaniform and the hamburger nodules is the presence within the latter of cone-in-cone structures. This structure is seen in the field where picked out by differential weathering from the surrounding nodule, and in thin section the cone shapes are seen to be outlined by either geniculate traceries of white mica (Fig. 2:2:11), or by a system of fine cracks, often carrying limonitic mineralization. Nothing of this was seen in the volcaniform nodules, but neither was there anything else within them to suggest their likely origin. To this end, specimens of both volcaniform nodules were analysed by X.R.F., together with the two specimens of the "hamburger" nodules to act as controls. The results, recalculated to 100% anhydrous, are tabulated in Table 2:2:12, and shown diagrammatically on Fig. 3:1:10 where comparison of these nodules with the igneous rocks of the area is made.

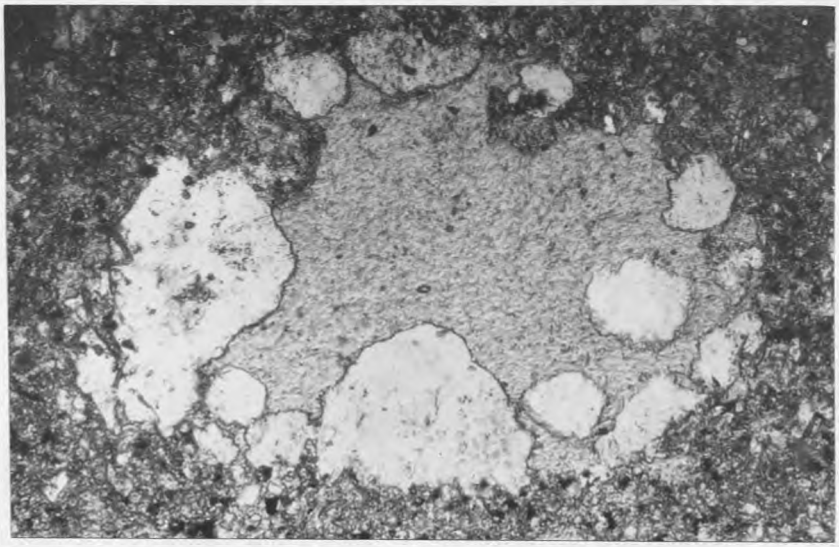
The main conclusion to be drawn from the analyses is that the two volcaniform nodules are completely different in composition, and therefore

Fig. 2:2:9. Granophyric quartz groundmass typical of the siliceous concretions.
Mica traceries visible (x25: x Nicols).

Fig. 2:2:10. Vesicular infilling of chlorite (chl) with quartz (q) and calcite (c) lobate outgrowths. From the single example found of a volcanic "bomb". (x40 . PPL).

Fig. 2:2:11. Geniculate traceries of sericite outlining the cone boundaries. (x25 . PPL).





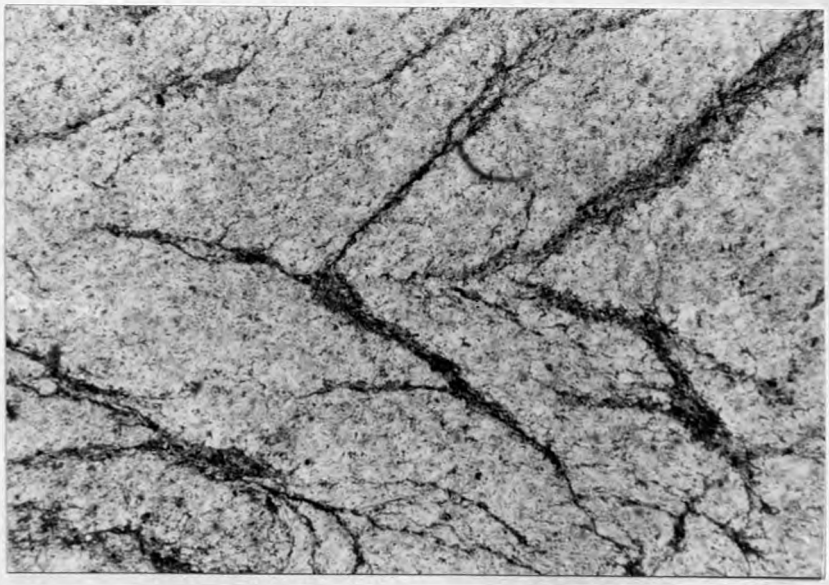


TABLE 2:2:12

CHEMICAL COMPOSITION OF THE NODULES
(recalculated to 100% anhydrous)

	9005	9004	9001	9003
SiO_2	87.58	82.84	45.66	85.98
TiO_2	0.29	0.31	1.38	0.41
Al_2O_3	8.36	6.43	17.17	7.55
Fe_2O_3	1.84	6.61	22.79	4.05
MnO	0.02	0.03	0.28	0.02
MgO	0.36	1.30	7.00	0.72
CaO	0.26	2.03	4.13	0.13
Na_2O	0.39	0.14	0.25	0.72
K_2O	0.82	0.25	1.22	0.24
P_2O_5	0.08	0.08	0.12	0.18
	100.00	100.00	100.00	100.00
	C/C	C/C	"BOMB"	SPINDLE C/C

by extrapolation in origin. The Buttermere specimen, 9001, is basic in composition and as Fig. 3:1:10 shows, consistently falls within the compositional field of the southern igneous rocks, strongly suggesting its origin as a volcanic bomb. The Gasgale specimen, 9003, although with its spindle shape superficially more volcanic than 9001 in appearance, is in fact seen to be very similar to the two hamburger nodules in composition and hence is of sedimentary (diagenetic) origin. The analyses for 9003, 9004, 9005, are very similar with only Fe_2O_3 showing marked variation, being especially high in specimen 9004. This is due to secondary introduction of limonitic material along the cracks which, as explained above, outline the cone-in-cone structures of this nodule, (Fig. 2:2:13).

Such concretions as these are commonly associated with sub-greywacke sands and muds having a simple clay matrix as the aggregating agent for small grains of silica. Strakhov et. al. (1954) considered concretion formation to be characteristic of their third stage of diagenesis, following the initial stages of bacterial activity and the formation of diagenetic minerals. Under these conditions colloidal silica within the sea-water can replace all or part of the clay matrix, probably by direct precipitation in the openings between the individual clay mineral crystals, (Dapples, 1967). In as much as the clay-sized interstitial material is known to contain small particles of quartz, these could act as nuclei for the precipitation of additional silica from solution. Penecontemporaneous re-solution is probably inhibited by relatively rapid deposition, (Strakhov et. al., 1954). The colloidal silica would be derived from several sources; devitrification of opaline silica; dissolution of detrital quartz grains; and decomposion of unstable volcanic material, probably being the major contributors.

The presence of cone-in-cone structures within most of the siliceous concretions is thought to be due to increasing load pressure on the

nodule during compaction and burial of the sedimentary pile. The clay minerals would become segregated from the silica and would be concentrated along the margins of the developing cone interfaces, in much the same way as chiastolite crosses are thought to form. With metamorphic recrystallization, the clay minerals were converted into white mica and the silica recrystallized into a granophyric mass of fine crystals, obscuring any trace of the cone-in-cone structure other than that preserved by the sericite stringers, (Fig. 2:2:9 and 2:2:11), or the mineralized fracture surfaces.

The major cone boundaries are often the site of slight overthrusting between two adjacent layers of the cone, and the double thickness of material thus resulting makes the cone-in-cone pattern quite distinct (Fig. 2:2:14). This distinction of major cone units is aided still further by secondary limonitic mineralization (Fig. 2:2:13). The direction of translation along these small overthrusts is difficult to deduce with certainty, but their cuspate form would seem to indicate movement parallel to the cone axis and not perpendicular to the cone margins. The axial angle of the major cones is quite constant, ranging from 40° - 50°, with 48° the median value. Within the boundaries of the major cone units however, a whole network of minor cones is preserved and these have somewhat sharper axial angles, ranging from 30° - 40°, suggesting that they were more highly appressed.

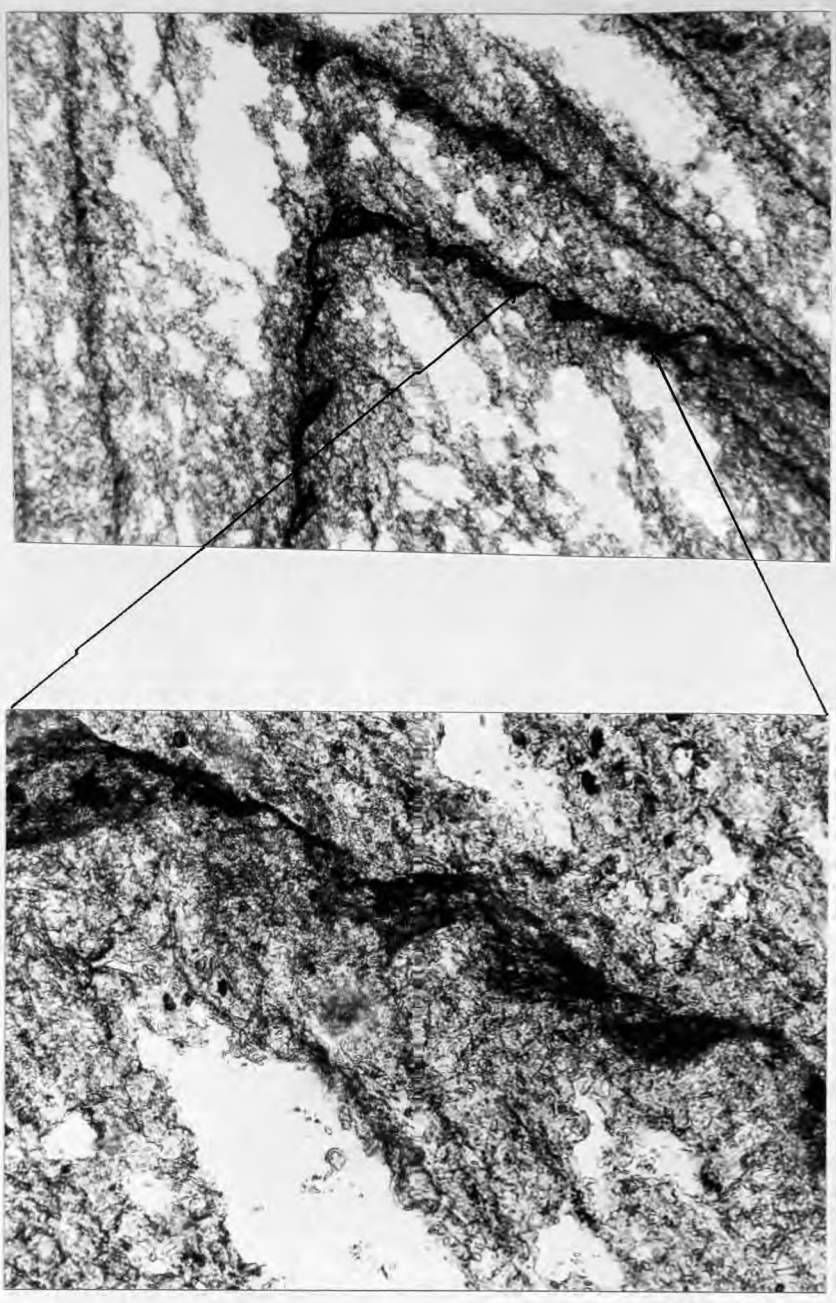
The nodules have diametrically opposed cone-in-cone patterns about a central axial zone, hence the common "hamburger" form, and continued growth would appear to have occurred outwards from this central zone along diverging cracks (dihedral angle 40° - 50°), (Fig. 2:2:15). This results in concentric, conical patterns of quartz crystallization parallel to the developing cone margin.

Thus the formation of these concretions would seem to be a function

Fig. 2:2:13. Limonitic staining along the cracks which outline the cone margins. (x25 PPL). Circle refers to enlarged section shown in the next figure.

Fig. 2:2:14. Enlargement of the cone-margin, showing the overthrust appearance of the junction. The direction of growth is indicated by the arrow (x40 PPL).

Fig. 2:2:15. The zone about one of the equatonal cracks, showing the way in which accretionary growth of the concretions from the centre outwards is thought to occur. A newly-formed cone structure is seen at the top of the photo. (x40 PPL).



of high overburden pressure, sufficient to induce triaxial shear failure in possibly unconsolidated sediments, yet not so high as to inhibit the outward growth of the structure normal to the bedding (i.e. parallel to the cone axis). This was due to steady access of quartz along the bedding and its deposition in the central zone of the concretion.

Section 2:3 The Psammitic Units.

There are two psammitic units within the proposed succession and they together account for well over half the total exposed thickness (Fig. 2:1:1), with the Loweswater Flags, in the order of 3,500' thick, by far the thickest unit in the succession.

The sequence is entirely a gradational one and hence any boundaries drawn between psammitic and pelitic units tend to be highly arbitrary and subjective in nature. The boundaries eventually selected are not based on any single factor but on a combination of field criteria which together were found to mark the passage from pelitic to psammitic (and vice versa) with sufficient uniformity throughout the whole area. These criteria are:

i) the first incoming of distinct psammitic bands, usually of 2" - 3" thickness,

ii) an increase in the psammite : pelite ratio to approximately 1:2 or 1:3, from the much higher values in the pelitic units, and

iii) the change in ductility contrast; for whereas the ductility contrast between the silty and muddy bands in the Mosser facies is low, it is always much higher between the psammitic ribs and pelitic interbeds of the flags leading to well developed buckle folding, boudinage and mullion structures, (Fig. 2:3:1).

The same changes occur in a reverse sense when the transition is from psammite into pelite.

Fig. 2:3:1. Typical field appearance of the psammitic lithologies.

A. Gasgale Crags (G.R. 173.228)

B. Penn (G.R. 166.225)

C. Dove Crags (G.R. 165.208)

The ductility contrast between the psammitic and pelitic bands of this lithology is well seen.







(i) Loweswater Flags.

The Loweswater Flags comprises a massive sequence of psammites, flags and interbedded pelites, and make up nearly half the total exposed sedimentary column of the area. The group can be sub-divided into upper and lower units of interbedded 'flag' lithology, separated by a massive middle psammite unit (see Fig. 2:1:1). The ratio of psammite to pelite varies widely from 1:3 at the top and bottom of the group, where transition into pelite occurs, to 20:1 in the massive central part of the sequence.

The same kind of colour variation across the area is seen in the Loweswater Flags as in the Mosser-Kirkstile Slates described previously, being pale grey on the flanks of Grasmoor; at the head of Gasgale Gill and around Rannerdale Knots. Elsewhere, to the north, the colour is darker, this variation being best observed in the pelitic interbeds, wherein spotting can also be seen in the central area. Another effect of the contact metamorphism is the increase in resistance to weathering which it affords the pelitic rocks; this means that they are not etched out so easily, and hence the alternation of psammite and pelite is much more difficult to distinguish in the central part of the area. For this reason the Loweswater Flag/Mosser-Kirkstile Slate boundary is less well defined in the Grasmoor - Lad Hows area than elsewhere to the north.

The pelitic interbeds are very similar, both in hand specimen and thin section, to the pelitic lithologies described previously. Lepidoblastic felts of sericite are interbanded with very fine silty bands containing larger sericite and chlorite crystals elongated parallel to the bedding. Pyrite is very finely disseminated throughout the rock as tiny blebs and needles.

The psammite bands vary widely in thickness and in coarseness of grain, but they must be divided into two groups for descriptive purposes. The first are thinner, finer grained psammitic bands in the upper and

lower parts of the succession. They range in thickness from 1" - 4", and have a restricted suite of sedimentary structures, with only cross bedding, convolutions, and parallel laminations being recorded. The second group comprises the massive psammites of the central part of the sequence. They range in thickness from 6" - 36" and show virtually the full sequence of sedimentary structures associated with turbidite deposits, namely sole marks and other bottom structures, graded bedding, convolutions, cross bedding, and parallel lamination (Fig. 2:3:2). Comparison of the contained sedimentary structures of these two divisions with the analysis of Allen (1970), (Fig. 2:5:1), would suggest the emplacement of the massive psammites as the proximal deposits of turbidity currents, with the thinner flaggy horizons representing their distal equivalents, although in a vertical sense rather than a horizontal one.

Other sedimentary structures present in these rocks are all the result of penecontemporaneous deformation. They include slump folding (discussed in Section 5:2), slump faulting and other, more complex, loading and shear structures, (Fig. 2:3:3). Indicators of possible flow direction in these rocks were rare. The cross-laminated division of the psammitic units is always too thin or too poorly exposed to measure foreset directions, and this leaves only sole marks on the base of the graded units; and the sense of over-folding shown by the slump folds, as possible flow direction indicators. Unfortunately, the evidence from these two sources is mutually contradictory. The slump folds (see Chapter 5) are orientated E - W and generally overturned to the north, indicating downslope flow in that direction. The sole marks measured, however, all had their long axes directed E - W, indicating flow in a direction at right angles to that shown by the slump folds. The significance of this is discussed in Section 2:5.

The thin psammite bands from the flaggy units are seen to be quite

Fig. 2:3:2. Examples of sedimentary structures from the Loweswater Flags.

A. Overturned cross-lamination

B. Slump and convolute bedding

C. & D. Lode carts. (following page).









Fig. 2:3:3. Isoclinal slump-folds from the Loweswater Flags.

A. N. end of Rannerdale Knotts. (G.R. 163.184) (See also Fig. 5:2:1c).

B. Dove Crags; fold axis plunges at 12° to 014°. (G.R. 176.206).





a heterogeneous group of rocks under the microscope. Some are quite well sorted, with sub-angular quartz grains consistently between 40u - 60u; others are very poorly sorted, the quartz grains ranging from 70u - 10u in diameter. Some show a well defined bedding, others are quite massive. All, however, are characterized by the very fine grained matrix, now more-or-less altered to chlorite and a little sericite, and by the presence of pyrite. This is present both as small crystals (20u - 30u) which are not uniformly scattered through the rock but restricted to diffuse zones parallel to the bedding (Fig. 2:3:4), and also as large euhedral aggregates (Fig. 2:3:5). These aggregates range in size from 300u to 3.2mm in diameter, are occasionally elongated parallel to the bedding, and are totally different from anything I have found described in the literature. Their origin will be discussed in more detail below.

The bedding in these thin sections is most clearly defined by the thin pelitic bands. In these, sericite is the dominant phyllosilicate, but yields to chlorite in the silty or psammitic bands. The chlorite makes up virtually the entire matrix of these coarser units and often appears to have corroded the quartz grains around which it lies (Fig. 4:1:2). Chlorite is also found as well-formed lamellar crystals mantling the large pyrite aggregates (Fig. 2:3:6). This latter is possibly due to the concentration of iron available for incorporation in the chlorite - which is of the penninite variety.

Also found in the vicinity of these pyrites, or around cavities in the slide, which to judge from the iron-staining once held pyrite, are clusters of very small granular, yellow-brown, crystals of high relief, which I think are epidotes. Detrital heavy minerals identified were zircon, rutile, and tourmaline; they were evenly distributed throughout the rocks, and no concentration in heavy mineral bands was noted.

According to Pettijohn et. al. (1972), the fine-grained, flaggy

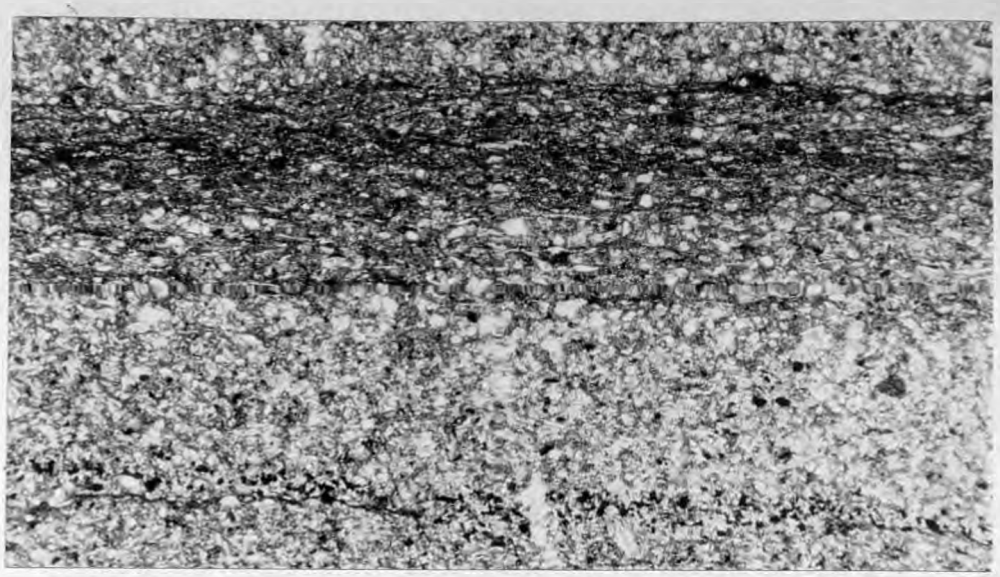
Fig. 2:3:4. Concentration of pyrite parallel to bedding.

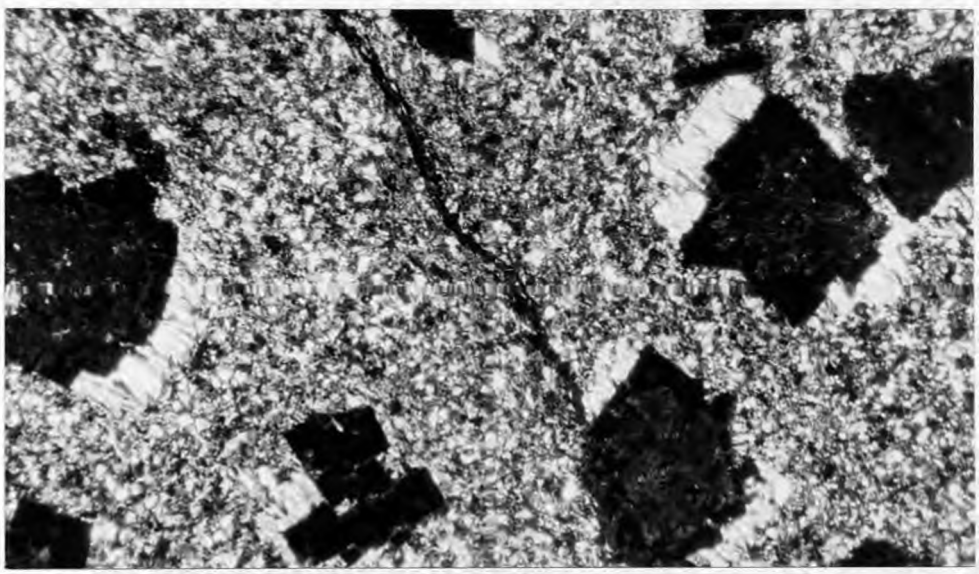
(x25 . PPL)

Fig. 2:3:5. Euhedral pyrite megacrysts with fibrous quartz pressure shadows growing parallel to the bedding.

(x25 x Nicols)

Fig. 2:3:6. Large chlorite crystal growing upon euhedral pyrite megacrysts, having its cleavage orientated perpendicular to the common face. (x40 . PPL)







psammites would be classified as quartz-wackes, on the basis of the low modal values of feldspar and lithic fragments (both less than 10%) and the high percentage of chloritic matrix.

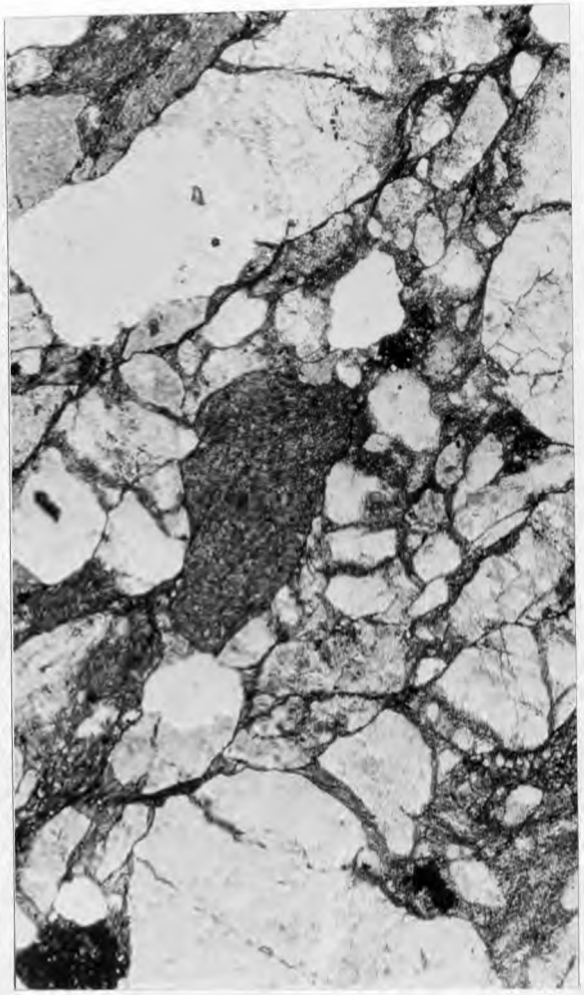
Following the same authors, the coarse grained, graded psammites from the centre of the succession, would be classified as true greywackes, since they contain more than 15% of matrix and roughly equal amounts of feldspar and lithic fragments. Modal analyses of specimens from the base of two psammite units gave figures of: 48% and 52% quartz; 31.5% and 22% matrix; 13% and 13.5% lithics; and 13% and 13.5% plagioclase. Much of the quartz is in the form of large, highly-strained and coarsely-polycrystalline, grains, (Fig. 2:3:7), up to 2 - 3mm in diameter; the lithic fragments also contain much siliceous material, in the form of micro-crystalline grains, probably representing chert fragments. Inclusions of pelitic material, occasionally difficult to distinguish from the chloritic matrix, (Fig. 2:3:8), make up the remainder of the lithic constituents. The feldspar is all plagioclase, of albite-oligoclase composition; it is very finely twinned (Fig. 2:3:9) and only partially sericitized, being quite fresh in appearance. The matrix is of chloritic composition and generally of very fine grain, though occasionally larger crystalline felts of chlorite do develop. Several of the rocks show a very irregular, hackly, fracture cleavage, and this cleavage cuts both grains and matrix alike, being undeviated even by quite large fragments of quartz. Secondary limonitic mineralization has permeated down many of these fractures and thence around grain boundaries, and has converted the chlorite of the matrix in to a bright yellow-brown, non-pleochroic, ferro-oxydized chlorite. Also present within these rocks are pyrite crystals, averaging 200-300u in diameter, but they have not been seen to attain the large sizes observed in the finer psammitic bands of the flaggy lithology.

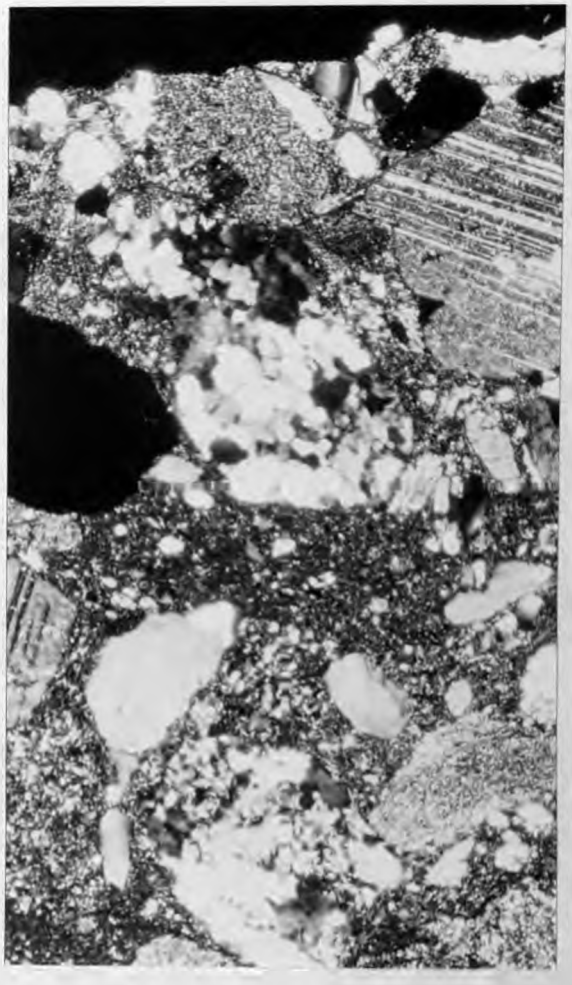
Fig. 2:3:7. Highly strained quartz fragments in a basal greywacke unit from the Loweswater Flags. (x25; x Nicols)

Fig. 2:3:8. Pelitic fragment from greywacke basal unit. (x25 . PPL)

Fig. 2:3:9. Plagioclase fragments from greywacke basal unit. (x25; x Nicols)







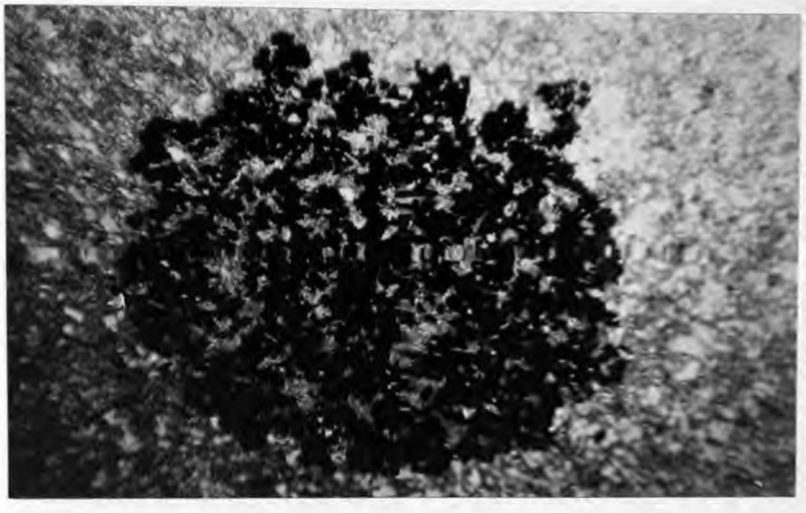
(iii) Pyrite Generation.

Within the pelitic fraction of the Loweswater Flags the pyrite is present as microscopic blebs and needles scattered uniformly throughout the rock. In the psammitic fraction there is a bi-modal size distribution of pyrite, it being present as very fine crystals (30u - 50u) concentrated along the bedding, and also as large irregular cubic aggregates occasionally framboidal in form up to 3.2mm in size (Fig. 2:3:10). These pyrite megacrysts are not seen, however, in either the massive graded psammites of the middle Loweswater Flag succession, or in the Whiteless Pike Flags.

Why, then, does the greatest development of pyrite occur in the, theoretically, most unlikely lithology, namely the deposits of distal turbidites; and why are they restricted to this lithology and not found in the more massive proximal deposits?

Before analysing this phenomena in detail, a brief review of the current theories of pyrite formation will be given. This is based largely on Berner's (1971) account, though data from Love (1964) is included. The basis of the formation of pyrite is given in Fig. 2:3:11. Pyrite is the stable iron mineral in the environment characterized by fine grained marine sediments enriched in organic material, and where anaerobic bacterial sulphate reductions are active. Pyrite is always authigenic because it is very unstable under conditions of transport and deposition. The chief source of iron is detrital iron minerals and the finer grained they are, the more reactive they are. The iron becomes soluble as the ferrous ion, this being a product of the bacterial and inorganic reduction of the ferric oxides. The major source of the sulphide radical is H_2S, which is derived dominantly from bacterial sulphate reduction, though some contribution is made by the decomposition of organic sulphur compounds. The Fe^{++} combines with the sulphide radical to form

Fig. 2:3:10. Framboidal pyrite crystal, with chlorite filling the interstices. (x35; PPL).



FeS, the initial, highly unstable, black iron ores, e.g. Mackinawite, that are almost immediately made over to pyrite (FeS_2) by reaction with elemental sulphur. Studies of recent sediments show that about 70% of the total pyrite content is formed from FeS within the top two inches of the sediment - water interface (Fig. 2:3:11). The pyrite formed as tiny crystals and spheres. The spheres show a framboidal texture formed by the close packing of tiny cubic crystals. The mean size of both crystals and spheres lies between 4u - 20u, though the larger spheres reach 40u in diameter.

This mechanism is alright for the small pyrite grains found throughout these rocks, but it does not explain the presence of the large megacrysts within the thin distal psammites. A summary of 3 possible mechanisms for the development of these megacrysts is given below, though each has quite effective arguments against it:

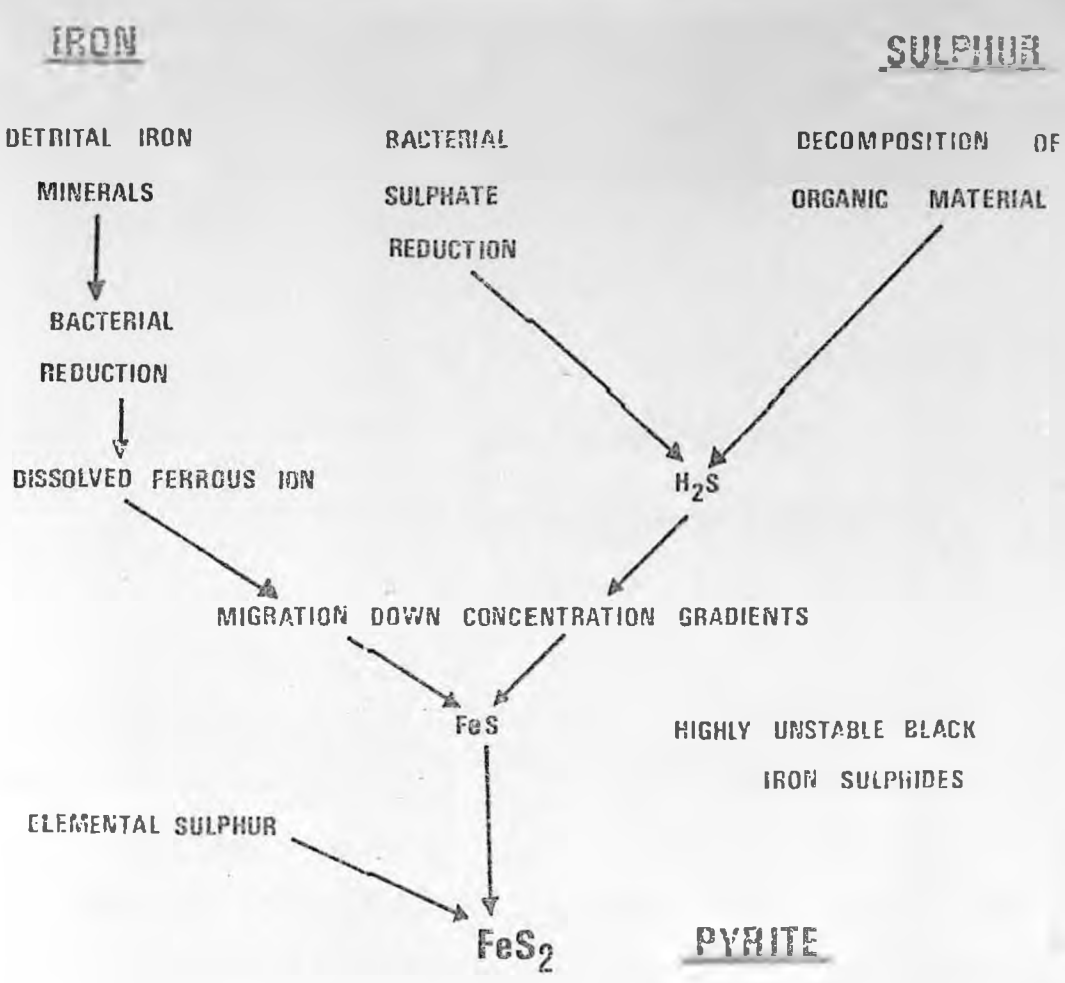
Possibility 1: Formation of pyrite in situ by the standard diagenetic processes (see above) but because the rock was coarser and hence more porous than the pelitic material in which pyrite usually forms, the iron and sulphide radicals were able to migrate more freely along concentration gradients and hence develop these larger crystals.

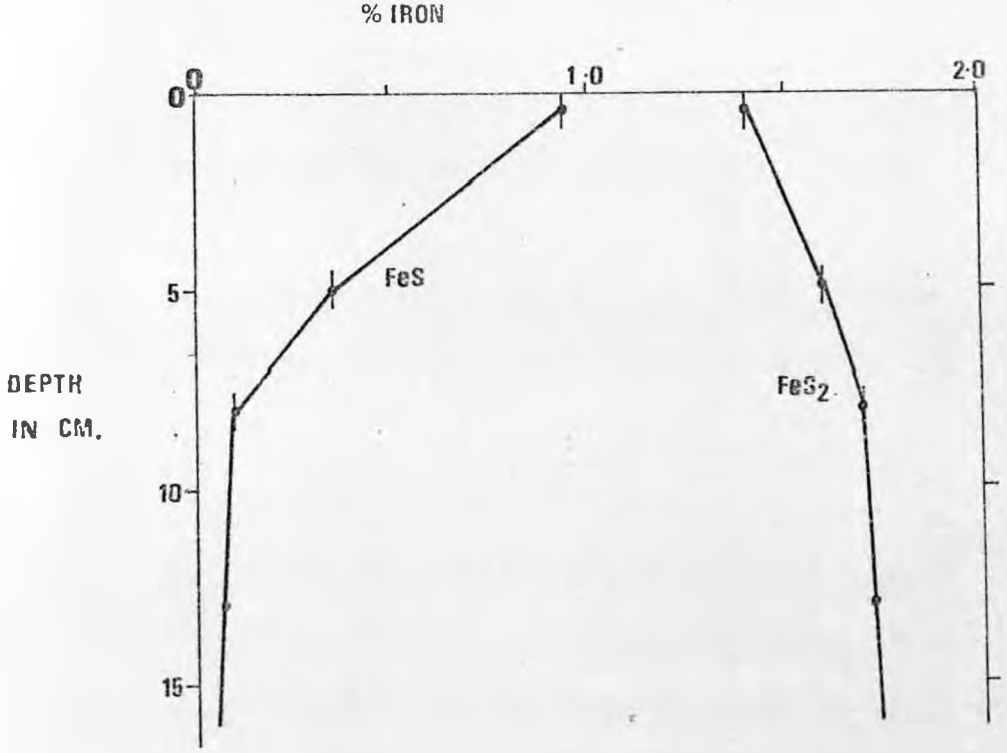
Against: a) The psammites are much coarser than the usual pyrite bearing strata and hence the required fine grained detritus would not be available.

b) It does not explain why a good proportion of the pyrite remains fine grained-and bedded.

c) The flags, I think, are deposits of turbidity currents, this judged on the presence of many of the diagnostic features, and turbidites are characterized by the relatively rapid deposition and coarse grain size.

DIAGRAMMATIC FORMATION OF PYRITE





CONCENTRATION OF FeS AND PYRITE VERSUS DEPTH FOR OFF-SHORE SEDIMENTS (AFTER BERNER; 1971)

FIG. 2:3:II

d) No evidence of organic detritus.

Possibility 2: The pyrite was formed as above, and the large megacrysts are the result of metamorphic recrystallization.

Against: a) Ditto b) above.

b) All the rocks showing these pyrites come from outside the zone delimited by the incipient cordierites (see Section 2:3).

Possibility 3: The large pyrites grew in pelites under conditions of slow fine-grained deposition, with a good supply of organic detritus, on a hypothetical shelf or slope removed from the present area. These were then disrupted, incorporated into turbidity currents flowing downslope, and the pyrites therefore winnowed free of their native pelites. The pyrites were deposited together with the sub-angular quartz grains to make up the present lithology. This possibility is endorsed by the similarity of the psammitic bands to turbidity deposits and the somewhat moth-eaten appearance of the large pyrites. The bedded concentrates of fine-grained pyrite resemble graded deposits.

Against: a) Pyrite is very unstable under conditions of transport and deposition (Berner, 1971); though this possibly applies only to terrigenous conditions.

b) Such a flaggy succession as this from which the pyrites come is characteristic of distal turbidites, and the large crystals are unlikely to be carried this far.

The most likely explanation, I think, lies somewhere between 1) and 3) above, namely that the pyrite formed as the standard small grains in a shelf or slope environment, as in 3), which was then incorporated into a turbidity current. Winnowing of the grains occurred, but most were still light enough to be deposited with the thin distal psammites. The coarser, proximal deposits also contained pyrite crystals, probably a

little larger, but the more rapid burial of these deposits, due to their closeness to the source of the sediment, effectively removed these pyrite crystals from the near surface zones of solution and re-deposition. The distal turbidites, though, were sufficiently removed from the terrigenous source to remain within these surface zones for a much longer time, and the higher permeability of these rocks as compared with their pelitic equivalents enabled the more rapid migration and recrystallization of the pyritous solutions, along the inevitable concentration gradients. This will explain:

a) The restriction of the megacrysts to the distal psammites,

b) their occurrence within turbidites, and

c) it still permits much of the pyrite to remain fine grained and bedded.

The absence of these megacrysts from the Whiteless Pike Flags is possibly a reflection of their much more rapid accumulation, thus inhibiting the redistribution of their contained pyritic material.

(ii) Whiteless Pike Flags

The Whiteless Pike Flags are made up of about 1,000 feet of thin-bedded psammites and pelites; the outcrop runs from Whiteless Pike across Squatt Beck, and over Low Bank, before disappearing in the unexposed ground around Bush House and the shores of Crummock Water.

The psammite bands are generally thin (1" - 3"), but do show good sedimentary structures, which are of great use in structural interpretation. The sedimentary structures are those typical of Walker's (1965) C division and above, namely cross bedding, convolutions and parallel lamination, and hence the Whiteless Pike Flags, like their equivalents in the upper and lower parts of the Loweswater Flags, are interpreted as the deposits of distal turbidites.

In thin section they are similar in all respects to the flags of

the Loweswater group, with the exception that no pyrite megacrysts were recorded.

The flags become somewhat lighter in colour towards the top of Whiteless Pike but the changes due to contact metamorphism are not clearly defined. Very diffuse spotting is shown in the pelitic interbeds on the N.W. flank of the fell, and this disappears by the head of Rowantree Beck, but between these two extremes the actual metamorphic limit is very difficult to define.

The ductility contrast between the psammitic and pelitic interbeds within these flags is not so well marked as in the Loweswater Flags, and on many occasions differential weathering has been found to etch out distinct bands which, when studied in thin section, appear identical in grain size and composition to those more resistant beds around them. This serves to illustrate the fact that weathering is a much finer judge than the eye in assessing the relative positions of these rocks in the psammite-pelite spectrum.

Section 2:4 Geological Description.

The area mapped during this study extends from Hope Beck in the north to Buttermere in the south, and within this area it has been found possible to subdivide the Skiddaw Slates into five lithological groups:

Buttermere Slates	1,000'+
Whiteless Pike Flags	800' - 1,000'
Mosser-Kirkstile Slates	1,500'
Loweswater Flags	3,500'
Hope Beck Slates	400'+
	7,400' approx.

The lithological characteristics of these formations have been discussed in the preceding sections; the present section will describe

Fig. 2:4:1. Reduction of final map to 2½" scale to show main outcrop and structural details.

Key.

1. Dodd Monocline
2. Braithwaite Syncline
3. Braithwaite Anticline
4. Whiteside Syncline
5. Sand Hill Anticline
6. Sand Hill Syncline
7. Gasgale Anticline
8. Dove Crags Syncline
9. Dove Crags Anticline
10. Coledale Anticline
11. Whiteless Pike Syncline
12. Whiteless Pike Anticline
13. Hause Point Anticline
14. Hause Point Syncline
15. High Buttermere Folds
16. Buttermere Syncline
17. Rannerdale Beck F1 Anticline Inlier
18. Hause Point F1 fold axes
19. Dodd F1 syncline.

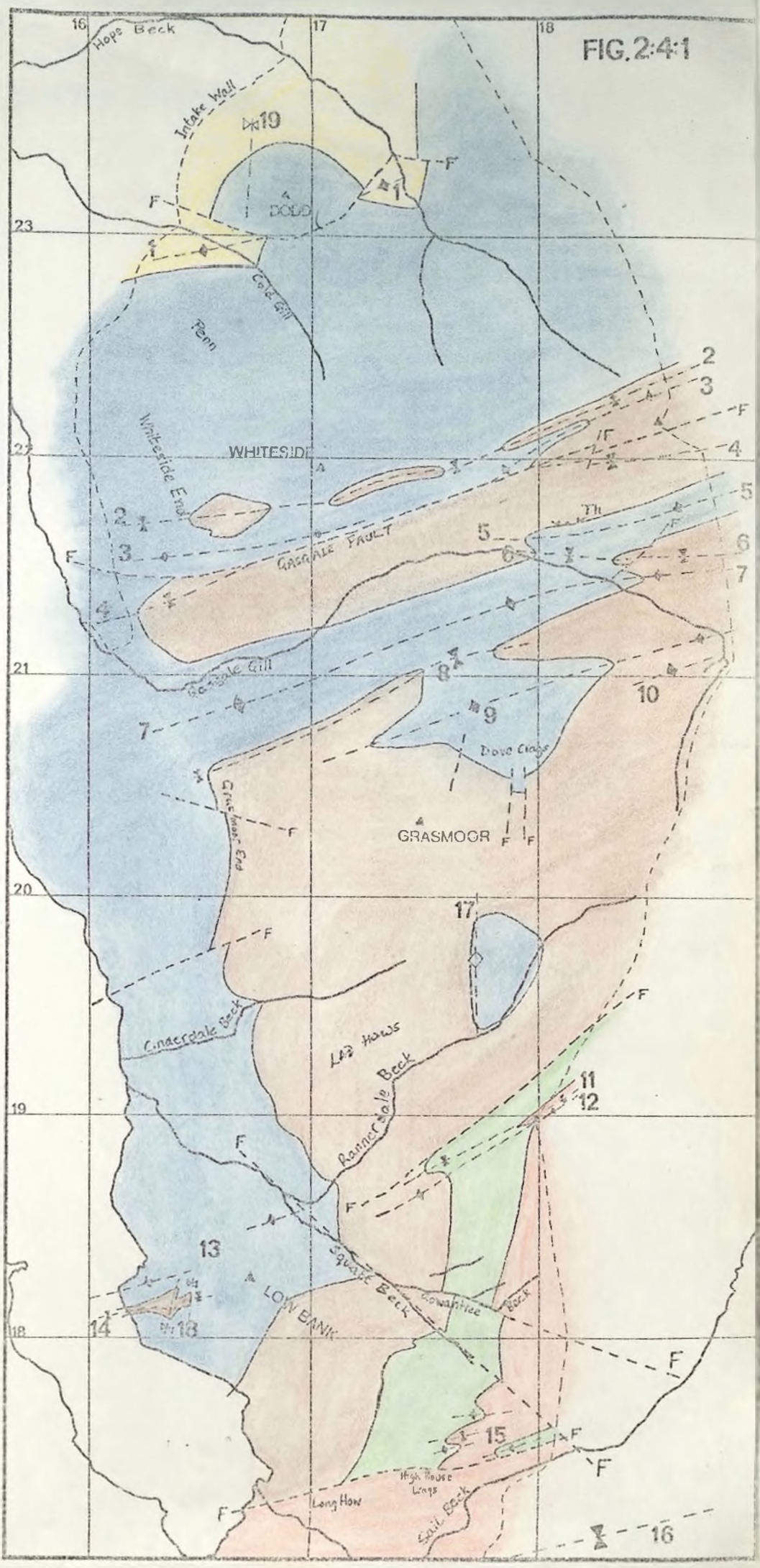

FIG. 2:4:1
Hope Beck
Intake Wall
DOLO
Cole Gill
Penn
Whiteside End
WHITESIDE
GASCALE FAULT
Gascale Gill
Dove Crags
GRASMOOR
Grasmoor End
Cinderdale Beck
Lad Hows
Rannerdale Beck
Square Beck
LOW BANK
Gowan Hee
Long How
High House Crags
Sail Beck

Fig. 2:4:2. A. Indication of the confidence levels with which results are to be regarded.

B. Sketch-section to illustrate major structural features of the area.

Buttermere Slates.

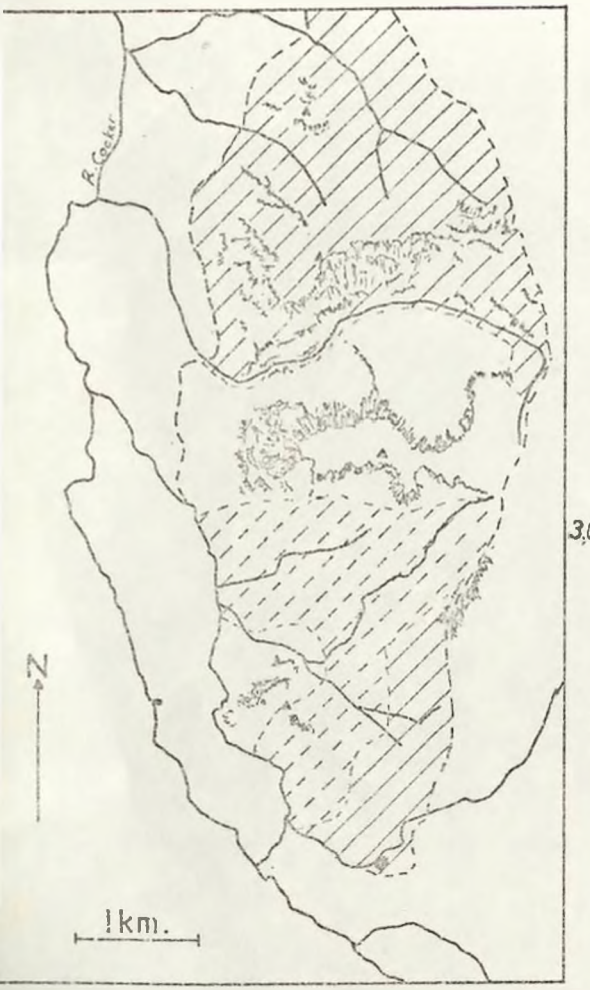
Whiteless Pike Flags.

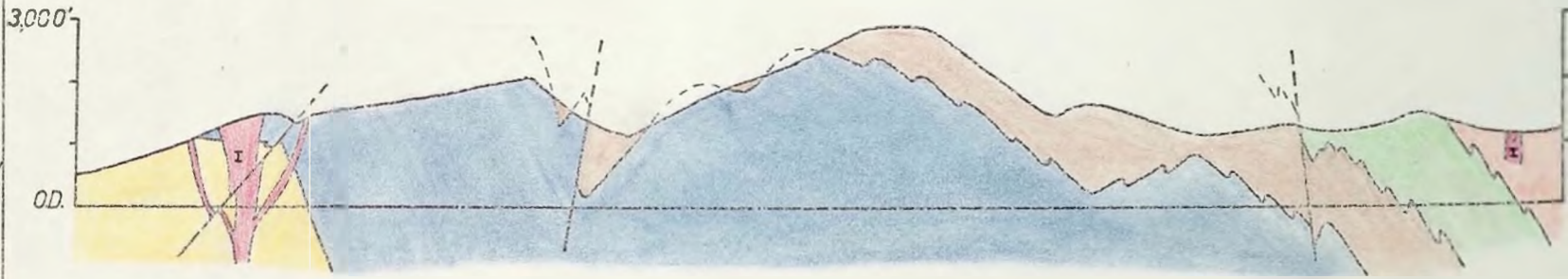
Mosser-Kirkstile Slates.

Loweswater Flags.

Hope Beck Slates.

I Igneous intrusions.





HORIZONTAL SCALE: 2": 1 mile.

VERTICAL EXAGGERATION: 0·88x.

HIGH MEDIUM LOW

CONFIDENCE LEVELS

FIG. 2:4:2

their distribution across the area.

Several problems had to be overcome before it was possible to map these lithologies with the required degree of confidence. The first was the difficulty of making consistent distinction between them, due to the gradational nature of their boundaries and the effects of contact metamorphism and heterogenous deformation. The second difficulty was the discontinuous nature of much of the exposure especially in the centre of the area, where the southern flanks of Grasmoor and the sides of Cinderdale and Rannerdale Becks were almost entirely covered with scree and heather.

With this in mind, it is possible, and valuable I think, to record the degree of confidence with which the results are presented. This is shown graphically in Fig. 2:4:2; the correlation of the area of limited confidence with areas of both contact metamorphism and poor exposure is not coincidental. The limited confidence in the Rannerdale Knots area, however, is due solely to the extreme complexity of the structures found there.

With these factors in mind, the geology of the area will now be described. The area will be covered from north to south and although detailed reference is intended to the 6" map and sections (Plates 1 and 2) in the rear pocket, a small scale map and section are included in the text (Figs. 2:4:1, 2) for more rapid general reference.

In the north of the area the emergence of solid rock from beneath the mantling glacial drift of Lorton Vale, roughly coincides with the line of the intake wall. It was from this area that Jackson (1961) first described the Hope Beck Slates, and they are exposed within the valley of Hope Beck itself for a distance of roughly 800 yds. upstream from their first emergence at the intake wall. The Slates are dark and well bedded, with occasional silty bands (Fig. 2:2:1), and the bedding is flexed into gentle, open warps, about sub-horizontal, E-W axes, giving low dips (30°)

to the north and south. The beds are cut by a fracture cleavage which dips at 20° - 30° to the south, post-dating the gentle E-W folding.

At Grid Reference 173.233, the Hope Beck Slates disappear under a cover of peat and lacustrine silts which mark the site of an old glacial lake basin, the flat, marshy floor of which extends for roughly 400 yds. upstream of this point. The lake was probably dammed by a rock lip rather than a morainic mound, because there is quite a deep outflow gorge cut into the Hope Beck Slates immediately below the mouth of the basin. The junction between the Hope Beck Slates and the overlying Loweswater Flags is obscured by the drift, since above the head of the basin, the first solid outcrops encountered are of massive Loweswater Flags dipping steeply to the south.

South-west of Hope Beck lies the low, rounded cone of Dodd, and here the junction of the Hope Beck Slates and the Loweswater Flags can be demonstrated, with the Slates passing conformably upwards into the Flags at roughly 1,000' O.D. on the northern face of the fell. It follows from this that Rose's (1969) interpretation of the Hope Beck Slates being younger than the Loweswater Flags can not be upheld, and that Jackson's (1961) earlier interpretation was correct.

Passing westwards around the fell, bedding dips in the Slates and Flags decrease from 30° - 40° S.W. to horizontal, and then incline again to the S.E. at 20° - 30°, thereby indicating the presence of an open N - S syncline, probably of F1 age, striking across the western flanks of the fell. The intensity of F2 folding increases from north to south across the fell; the Hope Beck Slate outcrop in the north is virtually devoid of F2 folding, but in the overlying Loweswater Flags open F2 folds (Fig. 2:4:3) begin to appear and by the time the crags immediately west and east of the summit are reached, the rocks are folded into large scale south

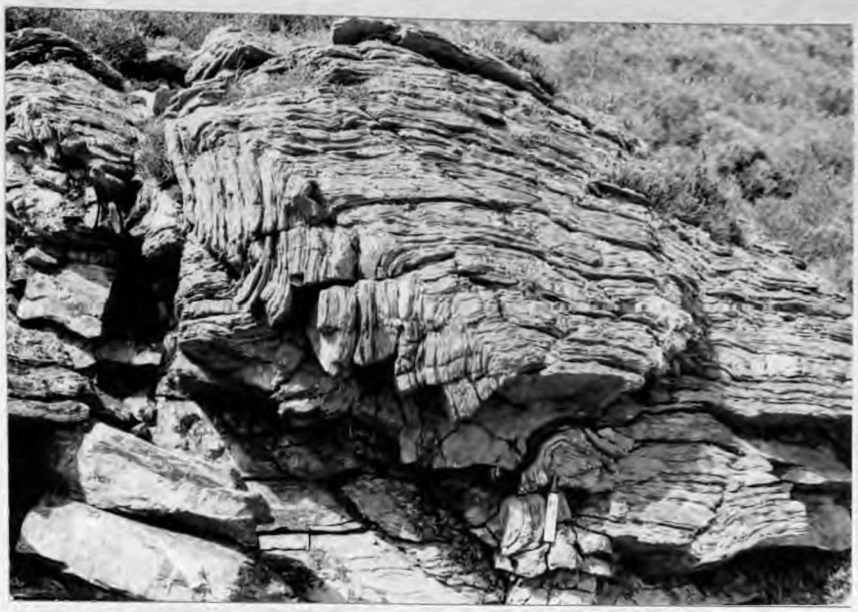
facing F2 overfolds of several tens of feet amplitude (G.R. 166233 and 169233). The hinge of the major Dodd monocline lies further south and is nowhere seen, having been faulted out and removed by erosion, but this fold was an order of magnitude larger than the overfolds just referred to, having an estimated amplitude of thousands of feet. Thus on Dodd, the intensity and amplitude of the folding decreases northwards away from the major structure. The overall S.W. and westerly dips of the Loweswater Flags capping Dodd are a function of the F1 syncline, mentioned above, lying to the west of the summit. A low-angled, north-dipping cleavage, roughly axial planar to the main Dodd monocline, is found throughout the exposed succession, and is also correlated with the low-angle cleavage in the Slates of Hope Beck. At locality G.R. 166233, this low-angled cleavage has been refolded by the minor monoclinal fold mentioned above, and in the steep middle limb of the fold, the refolded cleavage dips at 40° - 50° south. However, the corresponding fold on the eastern side of the fell (G.R. 169233) is cut by this same low north-dipping cleavage, and does <u>not</u> refold it; clearly, therefore, this congruous minor folding must have occurred both before and after the formation of the major Dodd monocline. Further discussion of this difficult problem is given in Chapters 5 and 6.

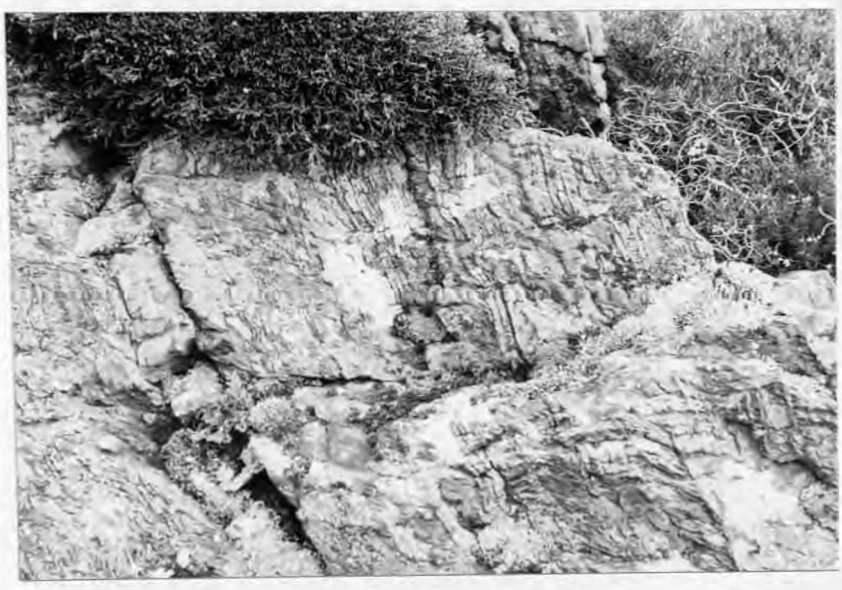
The Dodd Dolerite is best exposed on the northern face of the fell (Fig. 3:1:8), where it first appears as two inclined sheets dipping steeply S.E., which then merge into a single unit at 920' O.D. which may be then followed up the face to the summit plateau, where it disappears under a carpet of bleaberry and heather. Evidence for its continuation is found in the S.E. corner of the summit, where a train of dolerite boulders descends a gentle grassy gully. On the northern face, the bedding dips in the Slates steepen considerably as they approach the intrusion, even becoming overturned on the western margin. It is thought

Fig. 2:4:3. Overturned style of F2 minor folds from the Loweswater Flags of Dodd. (G.R. 170.232)

Fig. 2:4:4. Anomalous "reversed" cleavage in the steep limb of Dodd minor fold. (G.R. 167.232).

Fig. 2:4:5. View from Lanthwaite Green across to Whin Ben and Whiteside. The line of the Gasgale Fault (A-A') is clearly indicated.







that these features are due to contact strain during the F2 (end-Silurian) deformation, the pre-F2 age of the intrusion being evidenced by chloritic spots of contact metamorphic origin in the adjacent Slates, which have been deformed and sheared by F2 (Fig. 4:2:8).

The summit of Dodd is separated from the main ridge rising towards Whiteside by the fault gully termed Dodd Pass (G.R. 170230). The trace of the fault is not well marked to the west, but to the east a marked, marshy gully descends to Hope Beck, intersecting the Beck at the head of the glacial overspill gorge described above.

On the ridge south of this fault there is a further outcrop of the Dodd Dolerite; it is poorly exposed but seems to extend westwards and form the slight positive feature at the "lip" of Dodd Pass. The hinge of the Dodd monocline is nowhere exposed and everywhere south of the fault steep south-dipping strata are encountered - of Hope Beck Slates in the valley of Cold Gill, passing up into Loweswater Flags towards the top of the ridge. The pelitic interbeds of these steep-dipping Loweswater Flags show the same anomalous, reversed cleavage (Fig. 2:4:4), dipping 30° - 50° south, as described from the minor fold west of Dodd (G.R. 166233). In the bottom of Cold Gill, the Hope Beck Slates are almost vertical, and they are cut by a strong horizontal fracture cleavage; so strong, in fact, that at first glance the rocks appear to be flat-bedded. The origin of this cleavage is discussed in Chapter 6, but it is clearly unrelated to the folding that brought the Slates to their vertical position. The rocks exposed on the northern slopes above Cold Gill show very complex structures which were not fully understood, but the slope is very unstable and highly eroded due to excessive heather burning, so soil creep may have contributed to the structural confusion by disrupting and re-orientating the outcrops.

The Loweswater Flags of the steep, south-dipping limb of the Dodd monocline outcrop across the complete width of the area, from Ladyside Pike (G.R. 185225) to Whiteside End. Individual bedding dips very widely throughout this belt, due to abundant minor buckle folding, but a mean dip for the whole Flag unit would be about 60°, steepening as the Braithwaite Syncline is approached to 60° - 70° in the west, and 90° in the east, where the structure seems generally more tightly appressed. The beds strike consistently E. - W. across the whole area, and F2 fold plunges are horizontal, except on Whiteside End where gentle (20° - 30°) plunges to the east occur. Other than this, there is no evidence for the southwards continuation of the Dodd F1 syncline.

The Loweswater Flags are best exposed between Cold Gill and the Braithwaite Syncline, on the two sub-parallel crags of Penn and Whiteside End. The lowest 500' of the succession is composed of thinly interbedded psammites and pelites, marking the transition from the pelitic regime of the Hope Beck Slates, into the massive, turbidite-emplaced psammites, which are over 2,000' thick, and form the main part of the Flags division. As the Braithwaite Syncline is approached, the incoming of the Mosser-Kirkstile Slates is presaged by a decrease in the psammite-pelite ratio and by the decreasing thickness of the psammitic interbeds. This transitional, flaggy, lithology occupies the uppermost 500' - 700' of the Flags succession.

The Braithwaite Syncline is a well established structure, having been described by Rose (1955), Jackson (1961) and Simpson (1967) in turn. It trends E. - W. from Whiteside End to Gasgale Crags, whence it swings to ENE - WSW across towards Hopegill Head. The Syncline, well seen at G.R. 165218, where a thick quartz vein has been injected along the axial surface of the fold, is asymmetric and faces south, with the northern limb dipping steeply south at 60° - 90°, and the southern limb dipping more

gently northwards at 30° - 60°. Five hundred feet further south, the axis of the complementary Braithwaite Anticline is crossed, whereupon the bedding turns over steeply to the south once more. This steep southerly limb of the Braithwaite Anticline is well exposed in the crags to the west of the marked scree gully descending S.W. from Whiteside End, (Fig. 2:4:5, and G.R. 166.215).

The Mosser-Kirkstile Slates are exposed in the axis of the Braithwaite Syncline, but only in three separated periclinal structures. The separation of these infolds may be due to one of two causes:

i) heterogeneous strain during the F2 formation of the Syncline,

ii) subsequent warping of the Syncline along N. - S. axes by F4 folds.

It is not a reflection of a refolded F1 surface because, unlike Dodd, where the F1 minor folds possess the right symmetry for the proposed F1 syncline, on Whiteside the symmetry of the F1 minor folds is not related to the positions of the F2 culminations and depressions.

The Mosser-Kirkstile Slates in the synclinal axis are highly folded and cleaved, as are the Loweswater Flags which re-emerge from beneath them on the southward-rising middle limb of the Braithwaite fold pair. The trace of the Braithwaite Anticline is not continuous across the area, being faulted out for part of the length of Gasgale Crags, but it re-emerges in the anticlinal crest striking ENE towards Hopegill Head, and can be seen in the flags exposed on the path up Whiteside from Whin Ben, around G.R. 168.216. Stratigraphic and structural control on this face of Whiteside is again good, with abundant way-up indicators in the Loweswater Flags.

When Gasgale Crags are viewed from the summit of Grasmoor, the line of the Gasgale fault is clearly seen striking across the crags and

up towards Hopegill Head (Fig. 1:1:4). The fault is of reversed style, hading steeply north, with an apparent downthrow to the south of roughly 2,000'. To the west, its continuation is probably reflected in the deep gully on Hobcarton Crags just west of Hopegill Head; to the east, the fault is thought to pass along the southern termination of Whiteside End.

South of the fault, in the mouth of Gasgale Gill and again near its head, there are exposed northerly-dipping Loweswater Flags. The Flags pass conformably upwards into Mosser-Kirkstile Slates, and over the entire exposed length of Gasgale Crags these Slates remain steeply dipping to the north. To connect these north-dipping outcrops with the steep, southerly limb of the Braithwaite Anticline requires the presence of a large moderately tight, synclinal structure - the postulated Whiteside Syncline. However, the fact that north-dipping Slates are in juxtaposition across the Gasgale Fault with south-dipping Loweswater Flags, indicates that the axis of the Whiteside Syncline and the Mosser-Kirkstile Slates of its northern limb have been downthrown along the fault and are hidden from view, at least along the exposed length of Gasgale Crags. In the poorly-exposed ground to the west (N. of Whin Ben, G.R. 165213) and to the east (between Hopegill Head and Sand Hill), the Whiteside Syncline may emerge from beneath the fault but this is not certain and is only tentatively indicated on the map (Plate 1).

In the west of the area, around the lower reaches of Gasgale Gill, the Loweswater Flags rise steadily to the south (with one exception to be mentioned later), but when they next appear, above the scree apron of the northern crags of Grasmoor, the Flags are dipping to the south, indicating the presence of the Gasgale Anticline in the intervening ground. In the west of the area, however, from Sand Hill (G.R. 188218) down to the upper reaches of Gasgale Gill, the presence of an intervening fold pair is indicated. The evidence is well seen in the small

tributary gully containing the isolated doleritic body (G.R. 182217); coming south down the gully the north-dipping, much corrugated, Mosser-Kirkstile Slates are succeeded by Loweswater Flags which emerge from beneath the Slates just above a small thrust plane. Younging evidence here clearly indicates that the Slates do overlie the Loweswater Flags. Immediately below the overthrust lies the small doleritic body (described in Chapter 3) which is intruded into north-dipping Loweswater Flags. Just below the dolerite, the Flags then pass over an anticlinal hinge and descend in trains of minor F2 crumples to the south. This anticline, when traced westwards along the strike, coincides with the outcrops of flags around Sand Hill; these flags young to the south, towards outcrops of Mosser-Kirkstile Slates, and so are taken to be Loweswater Flags equivalent to those exposed in the dolerite gully. The complimentary syncline to the south is not well indicated, but must be present to separate the Sand Hill Anticline from the easterly extension of the Gasgale Anticline.

This latter structure is well marked in the vicinity of the waterfall in Gasgale Gill (G.R. 185214) where north-dipping right-way up flags emerge from beneath the scree on the north side of the Gill and turn over, with some axial thrusting, to dip steeply S.E. and disappear under Mosser-Kirkstile Slates immediately above the waterfall. From this point up to Coledale Hause the valley lies entirely in steep dipping, N. - S. striking Slates. There is no evidence of F2 folding in this distance but the exposures in the south side of the Gill reveal the presence of an asymmetric F1 overfold, facing ESE, and plunging gently (10°) to S.30.W. The gentle plunge of this fold indicates that the area must lie in a zone of relatively little F2 disturbance, otherwise the F1 plunge (down F2 dip) would be much greater.

The 10° southerly dip of the F2 surface indicated by this F1 fold

plunge, returns to steeper (30° - 45°) northerly dips by the sharp southwards bend at the head of the Gill (G.R. 189212). At this point is exposed an open, upright anticline with limbs dipping north at 30° - 50° and south at 40°, which may represent an extension of the Dove Crags Anticline from the east. This Anticline and its corresponding Syncline, (probably obscured in Gasgale Gill by the F1 folding) have, in combination with the glacial erosion of the Dove Crags corrie, produced a re-entrant pericline of Loweswater Flags to the south of the line of general incoming of the Mosser-Kirkstile Slates.

Decreasing limb-dips and amplitude make it difficult to trace this fold-pair WSW to Grasmoor End, but there is evidence for its presence as a broad trough-arch structure in the crags of the NW corner of the fell. Although everywhere along its length the Dove Crags Anticline is only a broad, quite open, structure, it is nonetheless one of the more important structural elements of the area, separating the basically south-dipping succession to the south of Grasmoor from the north-dipping succession down to the Gasgale Fault. To the north of this broad arch of Grasmoor, the dominant structure is the south-facing Dodd Monocline-Whiteside Syncline; to the south of the arch the north-facing Buttermere Syncline is the dominant structure, hence the unusual situation arises in which the amplitude of the congrous "parasitic" structures dwarfs that of the arch on whose flanks they occur, and across which they face each other.

The reason for this relative lack of deformation across the crest of the Grasmoor Arch is thought to be due to the presence, at depth, of the granitic body which formed a resistant "buffer" against which the rocks to the north and south were deformed, leaving the intervening arch relatively undisturbed. The decreasing width of the contact metamorphic aureole in the east of the area probably indicates the increasing depth, or decreasing size, of the granite below; in either case the likely

result would be an increase in the intensity of deformation in this part of the area. This is indeed the case; the Braithwaite fold-pair is much tighter (vertical limb-dips) in the east than the west; there is a similar increase in limb-dip and amplitude of the Dove Crags Anticline, and, in addition, there is the presence of the Sand Hill fold pair which is completely unrepresented in the west of the area.

Evidence of F1 folding in the Whiteside-Gasgale Gill area is not common, and the effect of F1 on the outcrop pattern of the area is certainly small. The form of the F1 folded surface is indicated by the plunge of F2 folds, and these are everywhere low to the ENE or WSW quadrants, thus suggesting a low relief for the F1 surface, though this may be a function of F2 flattening.

Further evidence of the pre-F2 shape of the F1 surface comes from the form and symmetry of the F1 minor folds. In the south of the area, around Low Bank and Buttermere, this is a useful tool, but in this northern area the evidence is inconclusive, there being no uniformity of facing the direction or style.

F1 structures become dominant in only three isolated areas:

i) at the head of Gasgale Gill where, as described previously, a large, east-facing F1 fold-pair occurs, and plunges gently SSW;

ii) on the south bank of Gasgale Gill near its mouth, around G.R. 162209, where steep westerly dipping flags occur, separated from north-dipping flags on the opposite bank by a WNW-striking fault;

iii) right on the intake wall just below Whiteside End (G.R. 161215) where an F1 anticlinal overfold is refolded by an F2 anticline, exposing pelites in the core of the structure, (see Fig. 5:3:6). What these pelites are, and how they come to be exposed in this situation, is a mystery, for the outcrop is isolated from adjacent exposures (of Loweswater Flags) by grass and heather.

Fig. 2:4:6. Massive, blocky nature of the Mosser-Kirkstile Slates within the Grasmoor aureole. (G.R. 167.205)

Fig. 2:4:7. Tight, angular. S-shaped F2 folds above Buttermere (G.R. 177.176). Looking east, down-plunge, the minor fold style suggests the shape of the major Buttermere F2 Syncline to the south (right).





The N - S bedding strikes of exposures ii) and iii) above may be related if a dextral movement along the WNW-striking fault in the mouth of Gasgale Gill is removed, in which case exposure ii) would represent the middle limb of the F1 overfold as it turns over from a gentle easterly dip to a steep westerly dip, (see Fig. 5:3:12).

There is no evidence anywhere along the valley of Gasgale for the presence of Simpson's (1967) Gasgale Slide (see Fig. 1:1:1). He may possibly have been misled by the edge of the contact metamorphic aureole which, as defined by the appearance of spots of andalusite (and possibly cordierite) in the pelites, follows the line of his Gasgale Slide quite closely. The change in appearance of the rocks across this line is quite marked, with shaly pelites becoming lighter in colour and more brittle, and silty mudstones becoming more massive and often very difficult to distinguish from more psammitic rocks (Fig. 2:4:6). The metamorphic gradient here must have been quite steep, because the limits of the aureole are quite sharp, with often less than 10 yards separating spotted from unspotted rocks. It is this indurating effect of the contact metamorphism, together with the generally poorer exposure, that is responsible for the lower levels of confidence with which the results from the Grasmoor-Lad Hows area are presented.

South of the crest of the Grasmoor Arch, down to Rannerdale Beck, the contact metamorphism has made the ground very difficult to interpret, for both Loweswater Flags and Mosser-Kirkstile Slates appear light grey and massive, with even bedding difficult to discern, a situation not aided by the broken nature of the exposure. The structure down to Cinderdale Beck is thought to be similar to that of Whiteless Pike, with trains of north-facing F2 overfolds rising to the north, but being downthrown in that direction by numerous strike faults. These strike faults are represented by the numerous parallel gullies, striking ENE, often with a

spring at their head, which cut the low crags above the lake. This conclusion is drawn from the fact that the dip of the beds as recorded (40° - 50°) is nearly always steeper than the outcrop of the enveloping surface (taken as the Loweswater Flags - Mosser-Kirkstile Slate junction) will permit.

Cinderdale Beck roughly corresponds to the line of an F2 synclinal axis and south of this the beds rise towards the low spur of Lad Hows before turning south once more to dip steeply down to Rannerdale Beck. Tourmaline veining is marked on the NW slopes of Lad Hows, and the effects of F1 folding are also important in this area. South from, and within, the middle reaches of Cinterdale Beck, there is a marked occurrence of N-S bedding strikes; they represent the sharp infolds of F1 synclines, the gentle northerly plunges of which reflect the limb-dips of the F2 folds which deform them. To the east, a second F1 fold is responsible for a major feature of the outcrop pattern of the area, namely the inlier of Loweswater Flags around the head of Rannerdale Beck. Steep westerly dips in the Mosser-Kirkstile Slates towards the east of Lad Hows presage the appearance of Loweswater Flags along the steep middle limb of a large F1 fold. Sedimentary structures here confirm the succession indicating a younging direction towards the Slates in the west. The F2 anticline of Lad Hows must be translated northwards across the F1 steep belt because all the F1 plunges in the Flags indicate an F2 surface dipping to the south; F2 fold plunges are few, but indicate a generally low, westerly dip for the F1 surface across the inlier, steepening as the Flags disappear beneath Mosser-Kirkstile Slates in the steep, north-striking belt.

South of Cinderdale Beck, the Mosser-Kirkstile Slates outcrop on the lower slopes of Whiteless Pike. They are overlain conformably by

the Whiteless Pike Flags, an 800' - 1,000' thick unit of interbedded psammites and pelites very similar to the upper and lower divisions of the Loweswater Flags. These Whiteless Pike Flags are in turn overlain, on the summit and eastern slopes of the fell, by the highest unit in the succession, the Buttermere Slates. Sedimentary structures are well displayed in the Whiteless Pike Flags and the way-up evidence from these confirms the succession established by superposition.

In the immediate vicinity of the summit of Whiteless Pike there occurs a major, ENE-trending, F2 fold-pair, of S shape and facing northwards, having long limbs dipping steeply (60^{o} - 90^{o}) southwards and a middle limb dipping more gently (45^{o}) to the north. The anticline is the better marked of the two folds and can be traced WSW down the fellside virtually to the intake wall (G.R. 173186). The syncline, to the north, can be traced down the face in a more westerly direction until exposure disappears at G.R. 174188. The axial traces of both folds follow sinuous courses across the fellside due to the topography of the pre-existing F1 surface, an easterly dip for which is suggested by the dominantly easterly plunges of the F2 minor folds.

South from the anticlinal axis the rocks dip steeply to the south, in trains of asymmetric, north-facing, S shaped overfolds, (Fig. 6:2:4). The steeper (usually vertical-overturned) middle limb of these folds are often down-faulted along strike-faults, the site of which are often suggested by the numerous ENE-WSW gullies of this face (Fig. 3:1:2, background). In this way, the situation arises in which the individual bedding dips as measured in outcrop are steeper than the mean dip of the enveloping surface of the minor folds.

The Whiteless Pike fold-pair increases in steepness towards the summit, and the northerly limb of the syncline is extensively sheared parallel to the (sub-vertical) axial planes of the fold-pair. Further-

more, the distance between the outcrop of the Whiteless Pike Flags near the top of the ridge and the exposures of Loweswater Flags in the bottom of the beck, is insufficient to accommodate the full thickness of the intervening Mosser-Kirkstile Slates. These two features suggests that the steep northerly limb of the syncline is down-faulted in the same manner as described for the minor folds to the south (see Plate 2). It is stressed that this solution is purely empirical, the trace of the proposed fault running through unexposed ground for its entire length. There is, however, no doubt that the two flag outcrops are of different stratigraphic groups, the Loweswater Flags younging towards, and the Whiteless Pike Flags away from, the intervening Mosser-Kirkstile Slates.

Also exposed on the face of Whiteless Pike are extensive systems of tourmaline veining and, at G.R. 179187, there is the proposed vent-breccia pipe of Borrowdale Volcanic age discussed in Chapter 3.

Below Whiteless Pike lie the valleys of Squatt Beck and its tributary, Rowantree Beck, both of which have been excavated along major fault lines, a fact which prevents direct correlation of structures across the valley. In the sides of Rowantree Beck F1-F2 interference structures of considerable complexity are exposed but these seem to have little effect on the outcrop pattern, which is wholly determined by the F2 structure.

Across Squatt Beck from Whiteless Pike lies the NW - SE striking ridge of Low Bank, with the crags of Rannerdale Knotts at the NW end. This NW end of the ridge is composed of massive Loweswater Flags which occupy the core of a major F2 anticline. This anticline is quite an open structure down by Hause Point and is well exposed on the road at G.R. 162182, but becomes increasingly tight as the summit of Low Bank is approached. The exact location of the axis of the anticline (termed the Hause Point Anticline) near the crest of the ridge is difficult to

fix, but it must lie between the steep southerly dipping strata of the summit and the steep north-dipping beds, with steep-plunging F2 Z folds, found at G.R. 167184, one hundred yards to the north. On the eastern side of Rannerdale Knotts (G.R. 168185) the hinge of the Hause Point Anticline is again well exposed, but along a line considerably displaced from that followed by the axial trace of the fold to the west of the ridge. Both an F1 steep zone and a marked N-S fault gully occur in the vicinity of the summit and either, or both, of these may be responsible for the observed displacement of the Anticline. The Hause Point Anticline is thought to be equivalent to the Whiteless Pike Anticline on the opposite side of Squatt Beck, where steep F1 down-folding reduces considerably the size of the Loweswater Flag outcrop. There is no indication of a complimentary syncline, as on Whiteless Pike, in the rocks of Hause Point.

On the crags immediately above Hause Point (G.R. 163181) there is an outcrop of pelitic rocks which are interpreted as an infold of Mosser-Kirkstile Slates into Loweswater Flags along the axis of a minor F2 syncline. The presence of two N-S steep belts, however, within the outcrop suggests that the extent of the inlier is largely controlled by the form of the pre-existing F1 surface. The outcrop therefore represents a Type 1 interference pattern as described by Ramsay (1967, page 522).

Over the remainder of Rannerdale Knotts the extreme complexity of the F1/F2 interference patterns defied detailed description, though this was not aided by the precipitous nature of the terrain and the frequent dissection of the outcrop by joint or fault gullies.

The descending limb of the Hause Point Anticline is exposed in a traverse SE from the summit, down the ridge of Low Bank. Successively younger horizons are exposed in this direction, and these are continuous (intervening faults omitted) with the SE-dipping strata of Whiteless Pike. Younging evidence, both in the Loweswater Flags and the Whiteless Pike

Flags, though scanty, does confirm that this succession is right-way-up. There is nothing in these rocks to support the conclusions published by Simpson (1967); the Crummock Recumbent Syncline does not exist and the rocks of the ridge are south-dipping, not north-dipping as shown on his cross-section (Fig. 1:1:2).

The rocks of Whiteless Pike and Low Bank maintain a steep, SE dip down towards Buttermere, interrupted only by minor basic intrusions to be described in Chapter 3, and a proposed E-W fault running north of Long How and High House Crags. To the north of this fault, the boundary between the Whiteless Pike Flags and the Buttermere Slates is quite complex, reflecting the presence of F2 fold pairs which expose Flags in the cores of the anticlines and Slates along the corresponding synclinal axes. Along the fault, which everywhere follows a marked marshy gully or a distinct negative feature, the rocks are downthrown to the south and only the highest unit of the proposed succession (the Buttermere Slates) is exposed south of this line.

The dominant structure of this southern area is the Buttermere Syncline, of F2 age, which trends ENE-WSW just south of Buttermere village. The rocks to the north of the village lie on the northern, SSE-dipping limb of the syncline, and this interpretation is confirmed by way-up evidence (load-casts in Buttermere Quarry) and by the S shaped, N-facing style of the F2 minor folds (Fig. 2:4:7). The southern limb of the Buttermere Syncline outcrops on Robinson and Moss Force (B,C. Webb, personal communication) and is inverted, suggesting that the overturned style of the minor folds closely approximates to that of the major structure. There is no evidence to support Simpson's (1967) contention that Buttermere is the site of a major refolded anticline,

The dominant F1 minor folds in this area are Z shaped and plunge steeply to the SE quadrant, down the limb of F2 Buttermere Syncline.

This, coupled with the presence of zones of steep, N-S striking bedding, indicates that the general attitude of the F1 surface prior to the F2 deformation would have been gently rising to the west, flexed at intervals into asymmetric west-facing folds having a steep to vertical middle limb. Complex F1/F2 interference patterns occur in the vicinity of Buttermere village, and these will be described further in Chapters 5 and 6.

In conclusion, therefore, the overall structure of the area is a function almost exclusively of the F2 deformation. It comprises two major, asymmetric fold pairs (the Dodd Monocline-Whiteside Syncline in the north, and the Buttermere Syncline in the south) and these face towards each other across the gently arched central area of Grasmoor. The lack of deformation here is thought to be due to the buffering effect of the underlying granitic pluton.

Section 2:5 Environment of Deposition and a Palaeogeographic Reconstruction.

The Skiddaw Slates of this area, as described in the preceding sections of this chapter, consist of a great thickness of monotous, marine psammites, flags and pelites. These are typical lithologies associated with the flysch facies, and comparison of the Skiddaw Slates with the characteristics of flysch as outlined by Dzulinski and Walton (1965) confirm this view. The features in common are:

1) marked alternation of immature psammite and pelite,
2) slump deposits and pebbly mudstones present, (reported by Jackson, 1961),
3) very little contemporaneous vulcanism,
4) absence of large scale cross bedding and ripple marks.

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The absence of the last two features, and the constancy of the pelitic intercalations, implies a deep-water origin, and the sequence of sedimentary structures found within such flysch deposits as these is compatible with their emplacement by turbidity currents. There are those who would dispute a turbidite origin for such deposits (see Selley, 1972), on three basic issues:

1) Nobody has actually seen a turbidite, even in the deep-sea canyons, despite the use of explosives in an attempt to trigger them.
2) Recent deep-sea sands may well be traction-deposited and furthermore, they are clean and well-sorted, not like the matrix-loaded greywackes.
3) Some ancient flysch deposits show features imcompatible with their presumed emplacement as turbidites - namely that they were deposited by currents that were not slope controlled, again suggesting the action of traction currents.

The first objection may be just a function of insufficient sampling, and the second objection has been effectively countered by Cummins (1962). He argued that the greywacke matrix is produced by the secondary, diagenetic alteration of initially unstable detrital fragments, such as feldspars and mafics. This view has recently been affirmed by Butler and McManus (1973), who have shown statistically that the grain size distribution of modern deep-sea sands will correspond to that of ancient greywackes if one assumes that 20% - 40% of the sand-sized particles in the former are lost by diagenetic disintegration and the products incorporated in the matrix. Thirdly, the occurrence of non-slope-controlled turbidites is very rare, and until a mechanism for the formation of the flysch sequence of sedimentary structures by traction

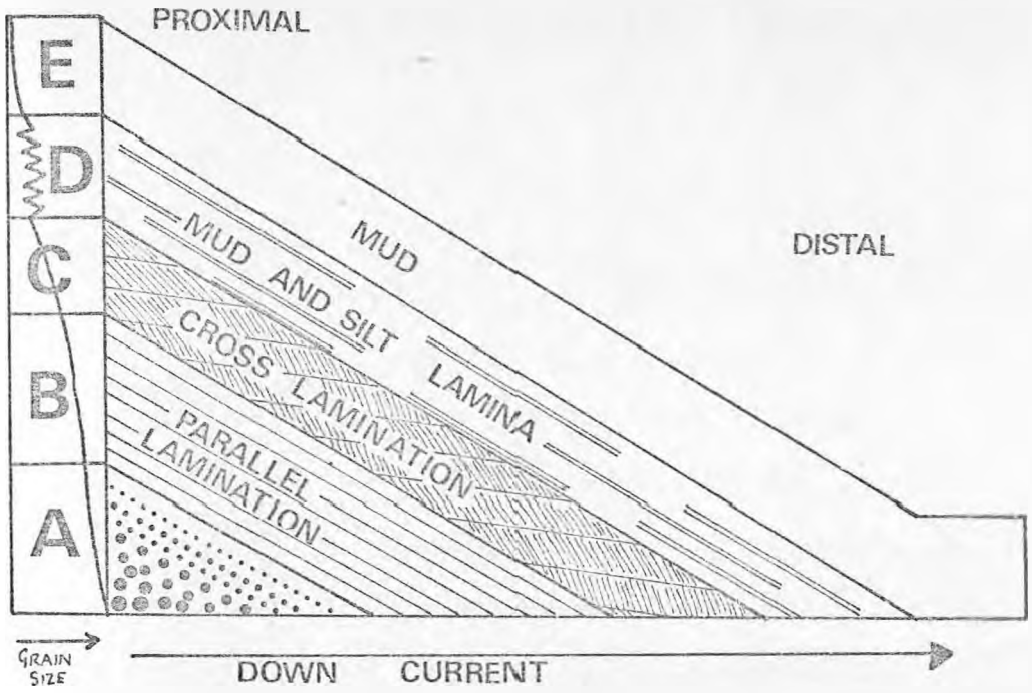
currents is proven, the benefit of the doubt must remain with turbidity currents as the only plausible mechanism for the emplacement of these greywacke type rocks.

The hydrodynamic properties of turbidity currents and the relationships between flow regimes and sedimentary structures have been discussed by Allen (1970) and Walker (1967); the basis behind this study is the premise that the further a turbidity current flows from its source, the lower its energy, and hence the sediment load becomes finer and the conditions of deposition more tranquil. It therefore follows that the deposits from proximal (near source) and distal (source-remote) turbidity currents may be distinguished on the basis of their petrography and structure, (Fig. 2:5:1). The psammites of Whiteside fit Walker's (1967) description of proximal turbidites: the beds are thick; medium to coarse grained, with individual sandstone bands amalgamated into thicker units from which pelitic intercalations may be largely absent. Sedimentary structures are present in profusion, and the bounding surfaces of individual beds are sharp.

The flags of the Whiteless Pike and Loweswater divisions represent distal turbidites, and are usually found to contain only divisions D and E of Walker's (1965) classification, though the convoluted and micro-cross-laminated C division is occasionally present. Such distal turbidites as these are characterized by rapid alternations of fine grained silty sandstones and pelite, with individual beds thin and regular. The sandstones may be well graded, and laminations are common, though scours and washout structures are absent, recording deposition from a turbidity current of waning power.

The Hope Beck, Mosser-Kirkstile, and Buttermere Slates represent pelagic accumulations of fine silt and clay grade material either during periods of quiesence, i.e. no turbidity current flow due to greater

FIG 25:1



HORIZONTAL & VERTICAL VARIATION IN TURBIDITES (AFTER ALLEN 1970) SHEWING DISTINCTION BETWEEN PROXIMAL & DISTAL DEPOSITS

stability of the continental slope, or a decrease in supply of terrigenous material; or alternatively, because the outer limit of turbidite deposits at these times did not reach this far.

The palaeogeographic conditions that prevailed in this area during the Lower Ordovician are very difficult to deduce because of the wide extent of later rocks which mask the relationships between the isolated Lower Palaeozoic deposits. However, using as much independent evidence as possible, an intelligent guess may be made, though even then it will still remain what Rast (in discussion) has derisively called a "geo-cartoon".

The Skiddaw Slates were deposited during Arenig and Llanvirn times, in a subsiding basin or trench, that lay marginal to, or within, the continental shelf. Within the Slates there is an almost total lack of any fossils but planktonic forms (graptolites and micro-fossils); more neritic forms are virtually absent, the only exceptions being scattered fragments of brachiopod and crustacea. This would imply a depositional environment considerably removed from the shoreline, which from somewhat equivocal slump fold orientations, would seem to have lain to the south. The margins of this trench or basin were areas of rapid sedimentary accumulation, and this sediment was transported into the axial regions of the trough by turbidity currents. The "trigger mechanism" for these flows need not necessarily have been earthquake shocks (Helwig, 1970), but could simply have been the increasing shear stress induced by the weight of rapidly accumulating sedimentary material (Terzaghi, 1956). The alternations of pelagic, distal and proximal deposits that are recorded in the stratigraphical column of the area are most likely to be the result of fluctuations in the distance of the land source from the axis of the trough, due either to changes in sea-level, or to eustatic rise and fall of the land surface itself.

TABLE 2:5:2

COMPARISON OF MEAN SKIDDAW SLATE COMPOSITION WITH SELECTED VALUES FROM THE LITERATURE

OXIDE %	SKIDDAW SLATES	AVERAGE SHALE	AVERAGE GREYWACKE	CRUSTAL ABUNDANCE
ELEMENT ppm	JEANS	SHAW (1956)	PETTIJOHN (1957)	TAYLOR (1964)
SiO_2	53.43	58.9	66.7	63.1
TiO_2	1.31	0.78	0.6	0.8
Al_2O_3	24.11	16.7	13.5	15.1
Fe_2O_3	9.70	6.1	4.5	6.0
MnO	0.182	0.09	0.1	-
MgO	1.85	2.6	2.1	3.0
CaO	0.89	2.2	2.5	4.1
Na_2O	1.18	1.6	2.9	3.4
K_2O	3.56	3.6	2.0	3.0
P_2O_5	0.113	0.16	0.2	-
		TAYLOR (1965)	TAYLOR (1965)	
S	227	-	-	-
Cr	127	100	140	100
Ni	48	70	50	80
Cu	104	50	40	55
Zn	88	100	50	70
Ga	21	20	15	15
Rb	211	140	120	90
Sr	136	300	450	375
Y	25	25	10	33
Zr	219	140	160	165
Ba	890	700	500	425
La	45	-	-	30
Ce	101	-	-	60
Pb	61	20	15	13
Th	17	12	10	10
$K_2O:Na_2O$	2.98	2.25		
$Al_2O_3:Na_2O$	20.4	10.4		
$P_2O:CaO$	0.127	0.074		
$Ga:Al_2O_3$	0.88×10^{-4}	1.2×10^{-4}		

Further information on the problem of the depositional environment is afforded by the geochemical analyses made of 33 rocks from the Grasmoor and Rannerdale areas, despite the fact that these analyses were initially made as part of the metamorphic investigation. The full analysis lists are given in Appendix C but the relevant details will be abstracted here. Table 2:5:2 lists the mean composition of all the analyses made, and compares them with values for the "world average" shale (Shaw, 1956; Taylor, 1965) and greywacke (Pettijohn, 1963; and Taylor, op. cit.), and also with the standard crustal abundance figures (Taylor, 1964). The analyses span the upper flaggy group of the Loweswater Flags and most of the Mosser-Kirkstile Slates, so the rocks are basically pelitic in aspect, and this is borne out in their similarity to the "Average Shale" figures of Shaw (1956) and Taylor (1965). The only significant difference lies in the very high Al_2O_3 values of the Skiddaw Slates, and this is attributed (Ernst, 1969), to the presence of hot, humid climatic conditions in the sediment source area, with the intense tropical weathering releasing abundant alumina as solution and suspension load to the rivers. Good corroborative evidence for this assumption has been provided by Dr. J. Briden (pers. comm.) who has obtained palaeomagnetic latitudes for the Lake District in early Ordovician times of 20° - 25° south.

The high Rb and K_2O values, taken with the low Na_2O, indicates the lack of authigenic albitic feldspar, and hence Rb and K are probably wholly confined to detrital potash feldspar and illitic clay minerals, now recrystallized to sericite.

The high P_2O_5:CaO ratio (0.127, compared with average 0.074) indicates that only a certain proportion of the phosphate is taken up with the Ca as apatite; the remainder was probably provided from biogenous processes, these being characteristic of warm seas and shelf

environments today (Taylor, 1965), although the high MnO values do indicate deposition in rather deeper water, (Keith and Degens, 1959).

Possible indicators as to the nature of the source area of the sediment are provided by the trace element concentrations. The Slates are concentrated in Rb, Zr, Ba, La, Ce, Cu, Pb, and Th, and are depleted in MgO, CaO, Na_2O, and Ni; in other words they are depleted in those elements of basic affinity, and they are concentrated in those elements characteristic of granitic terrains. The lowish $Ga:Al_2O_3$ ratio (0.88×10^{-4}) tends to militate against a sedimentary source area, on the basis that Ga tends to be concentrated with respect to Al_2O_3 in sedimentary rocks, (Hickman, 1972).

Therefore evidence from three directions suggests the presence of a source area composed of potassic granitic rocks, either igneous or gneissic in origin. The high Cu values are a little anomalous but, in mitigation, it must be said that Cu is known to be concentrated by biogenous processes, though admittedly only in near-shore environments.

The plotting of individual analyses, or element ratios, against the height from which the sample was taken, yield a further environmental indicator (Fig. 2:5:3). The decrease of Total Alkalis and K : Rb with height is interpreted as being due to the increasing influx of basic volcanic material into the sediments, and this is further borne out by the decrease with height of Pettijohn's (1957) maturity index (Al_2O_3 : Na_2O), though admittedly one does not see the increase in MgO, Fe_2O_3, or CaO that should accompany such a volcanic influx.

In conclusion then, the Skiddaw Slates represent a flysch sequence emplaced largely by turbidity currents in a subsiding trench or basin marginal to, or just within the continental shelf. Sediment was supplied largely from the south, from a granitic source area, though there was some influx of basic material towards the top of the sequence. The zone

Fig. 2:5:3. Geochemical data plotted against height from which sample was collected.

Decrease of K/Rb ratio and (K_2O + Na_2O) with height suggests influx of basic volcanic material. This is possibly borne out by the decrease in Al_2O_3/Na_2O with height.

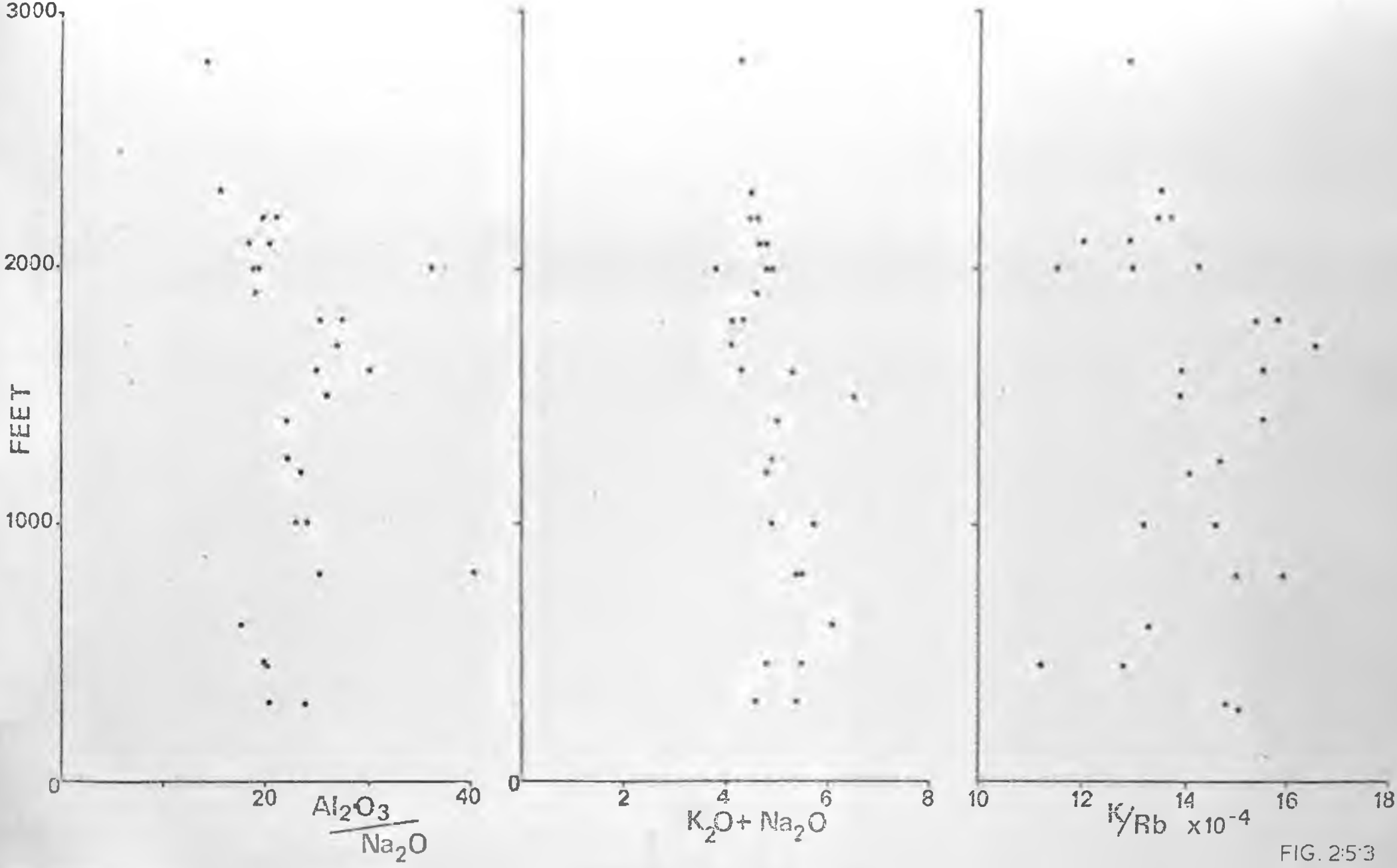

3000
2000
1000
0
FEET
20
40
$\frac{Al_2O_3}{Na_2O}$
0
2
4
6
8
$K_2O + Na_2O$
10
12
14
16
18
$K/Rb \times 10^{-4}$

FIG. 2:5:3

of deposition lay at a palaeolatitude of roughly 25°south, and the source area probably suffered a sufficiently tropical climate to promote the passage of high concentrations of alumina, both in solution and suspension into the rivers.

A PLATE TECTONIC RECONSTRUCTION OF THE SOUTHERN CALEDONIDES OF GREAT BRITAIN

PETER J. F. JEANS

(As accepted for publication in Nature).

Gunn[1] has recently published in these pages an elegant solution to the problem of accommodating two subduction zones, one dipping north under the Highlands[2], the other dipping south under the Lake District[3] between the continental crust thought to underlie the Southern Uplands[4] and that of the Highlands. Gunn[1] considers the Midland Valley of Scotland "to be the oceanic remnant of the proto-Atlantic Ocean", and he suggests that the Highland Boundary Fault and the Southern Uplands Fault "mark the positions of diverging Benioff zones". The ophiolite complex at Ballantrae and the ophiolitic rocks of the Highland Border Series, both of which may be traced aero-magnetically across into N. Ireland, are considered to be fragments of ocean floor emplaced by thrusting and welding along the zones where oceanic crust was over-ridden by the converging continental plates.

The purpose of this communication is to extend this hypothesis to account for the geological features of the Southern Caledonides of Britain (Wales, Lake District and the Southern Uplands). As mentioned above there is strong evidence[4] to indicate that the Southern Uplands are underlain by continental crust, and hence it would seem reasonable to postulate that a single continuous continental plate underlay the whole of the Southern Caledonides; certainly there is no evidence for any crustal suture line between the postulated plate margin at Ballantrae and the Welsh Borders. Two major depositional basins would therefore be located on this plate, namely the Lake District-Southern Uplands basin and the Welsh basin, which were separated by the oft-proposed Irish

Sea geanticline. Evidence for the northwards continuation of the Manx-Skiddaw sequence beneath the post-Llanvirn rocks of the Southern Uplands is possibly found in the Coldingham Beds of Berkshire, especially as they show very similar refolding patterns to those of the Skiddaw Slates[5, 6, 7]. It is necessary to postulate the presence of the Irish Sea geanticline as an actively eroded upland area to provide the detritus for the Skiddaw-Manx and Welsh sediments, and palaeoslope and sediment transport directions from the Skiddaw Slates[7, 8] do indicate the presence of a shoreline to the south and supply of sediment from that direction.

Recent radiometric dating of the Ingletonian rocks has yielded Cambro-Ordovician ages[9], which indicates their likely deposition in the Manx-Skiddaw trough, as the possible time-equivalents of the lowest Manx Slates.

It is suggested that during Arenig times (Fig.1), the south dipping oceanic plate was gently angled, giving the andesite volcanism of the Cader Idris, Rhobell Fawr and Pembroke, whilst beneath the Manx-Skiddaw trough the descending plate was insufficiently deep for partial melting and volcanism to occur. Following Rast[10], the Bala Fault is shown as the feeder to both N. and S. Wales volcanics. Eruptions in these areas continued into the Llanvirn and were characterized by either spilites and andesites, or rhyolites and keratophyres, the latter representing the products of crustal fusion and differentiation consequent upon the emplacement within the crust of large bodies of hot, mantle-derived magma.

By Llandeilo times (Fig.2) the inclination of the Benioff zone must have increased to permit the formation of andesite melts beneath the Manx-Skiddaw trough; upon rising to the surface these andesites built up an island arc, now represented by the Eycott and Borrowdale Volcanic Groups. Although dominantly composed of andesite, the more extreme

differentiation products occur towards the top of the Borrowdale sequence. The Borrowdale Volcanics were wholly sub-aerial in eruption[11] and this suggests that the compressive forces responsible for the closing of the proto-Ocean were sufficient to raise the top of the Skiddaw Slates above sea level, and this is reflected in the widely reported sub-Volcanic unconformity[12], and pre-Volcanic, albeit N-S orientated, deformation phases[6, 13, 14].

The inclinations of the downgoing slabs required for this model are compatible with present-day observations. The diagrams show the situation in which no crustal shortening is assumed; taking 150 km. as a reasonable depth for andesite genesis, this gives lithospheric dips of 30° and 52° for North Wales and the Lake District, respectively. If a 50% crustal shortening is admitted, the respective dips would be decreased to 15° and 32°. These values compare favourably with present-day observations presented by Oliver[19], who records lithospheric dips ranging from 10° - 20° beneath Peru and North Chile, to 60° for the majority of the Western Pacific Benioff zones, e.g. Kermadoc, New Hebrides, Mindano, etc.

Cessation of the Volcanic episode by the end of the Llandeilo was accompanied by deformation and subsidence of the island arc. This subsidence and marine incursion occurred earlier in the east than in the west, for the Volcanics at Cross Fell are overlain unconformably by the corona Beds which have no equivalents in the main outcrop, the Volcanics there being overlain by later members of the Coniston Limestone Group. Progressive deepening of this shelf sea would then lead to the initiation of the mio-geosynclinal trough in which the thick succession of Silurian greywackes and slates was deposited.

Meanwhile, during Caradocian times, vulcanicity in North Wales was renewed on a huge scale, starting in Conway and the Capel Curig

Fig. 2:5:4.

1. Manx Slates.
2. Skiddaw Slates.
3. Coldingham Beds, Berwick.
4. Ballantrae Ophiolite Complex.
5. Ingletonian.
6. Irish Sea Geanticline.
7. Menai Straits Fault Complex.
8. Bala Fault - acting as feeder for:-
9. Cader Idris, Rhobell Fawr and Pembroke Volcanics.
10. Eycott Volcanic Group.
11. Borrowdale Volcanic Group.
12. Snowdon and Conway Acid volcanics.
13. Subjacent granitic plutons, e.g. Tan-y-Grisian.
14. Proposed "heat plume", fault controlled, rising from Benioff zone at depth, causing melting at base of crust.
15. Highland Border Series and Boundary Fault.
16. Southern Uplands Fault and Ballantrae Ophiolites.
17. End-Silurian intrusion of granites in the Southern Uplands and Lakes due to rising "heat plume".
18. Prograding delta fan depositing the Girvan greywacke sequence.
19. Thin Moffat sequence; axial zone progressively swamped by northerly derived greywacke deposits.
20. Silurian rocks of Lake District and Cross Fell.
21. Down-pinching ocean crust.

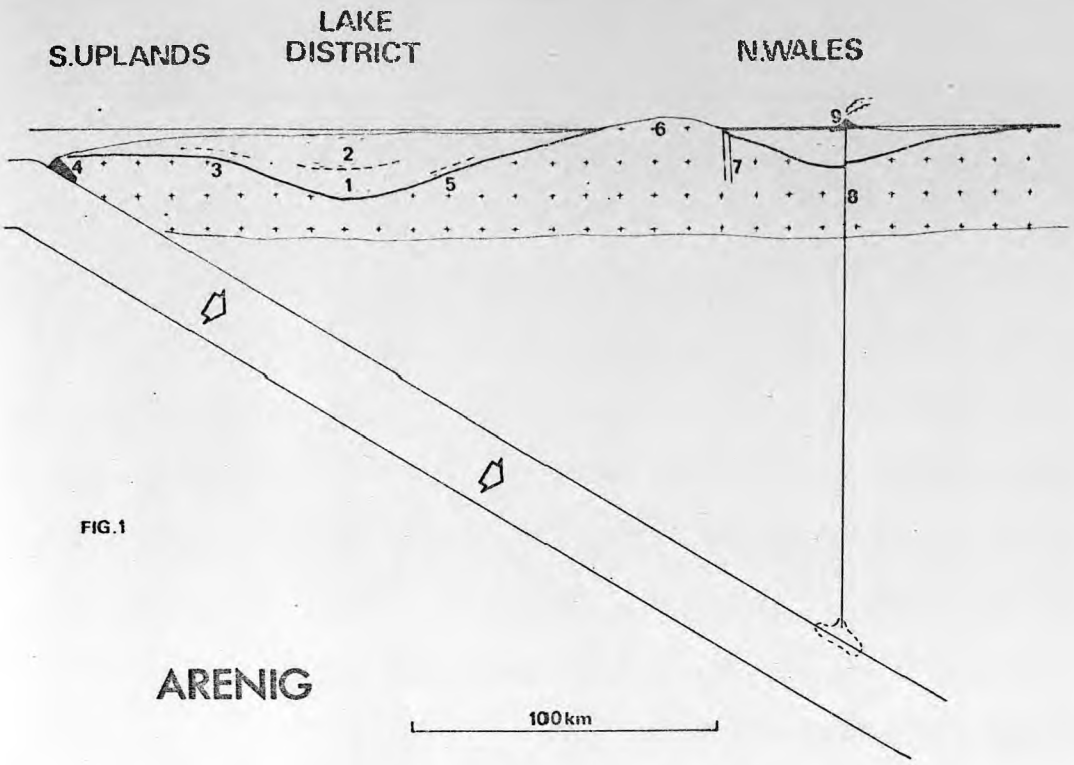


FIG.1

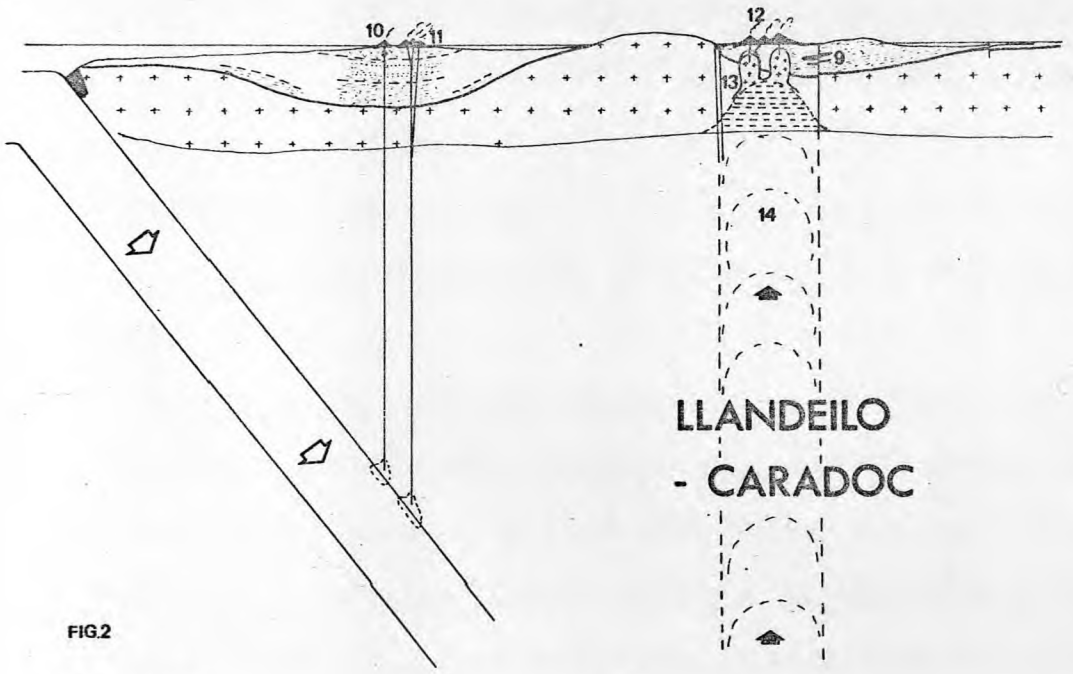


FIG.2

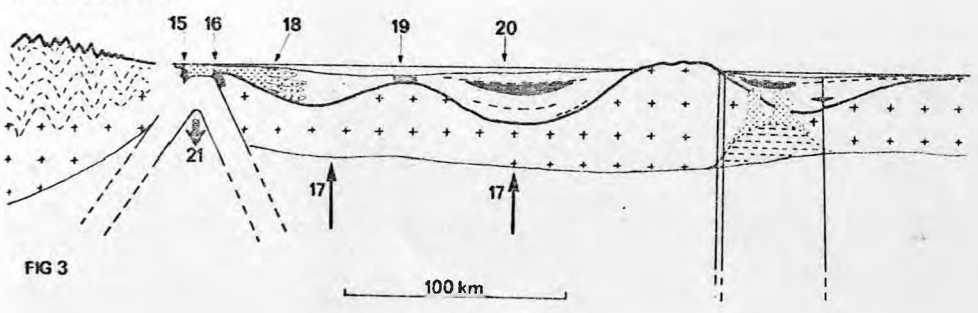


FIG 3

area east of Snowdon. Snowdon became a major centre of volcanism, with the eruption of vast thicknesses of rhyolite and ignimbrite, and the emplacement of subjacent granitic plutons such as Tan-y-Grisiau. There is evidence that the whole of Snowdonia is underlain by a large sub-volcanic granitic body[10], and geochemical analyses from the ignimbrites (Hendry, personal communication) suggest their origin by partial melting of sediments at no great depth below the surface (10 km.). The mechanism by which this crustal melting occurred is not known, but it has been suggested[10, 15] that the crust below Snowdonia was invaded by a large, mantle-derived basic pluton, and that the heat from this basic body melted the continental crust and sediments above, thus providing the acid magma for eruption and epizonal injection. Whether the basic material was derived directly from the Benioff zone at depth, or whether only heat was transferred upwards to melt mantle material nearer the base of the crust is not known, but the latter hypothesis is illustrated in Fig. 2, the rising "heat plume" being controlled by the deep faults of Menai and Bala.

By the end of Ordovician and beginning of Silurian times the situation as envisaged in Fig. 3 was reached, with the closing rafts of continental crust beginning to pinch down the remnant ocean crust of the Midland Valley. By lowest Llandovery times (acuminatus zone, Cocks and Toghill[20]) coarse grained terrigenous sediment from the rising Highlands to the north was beginning to appear in the prograding greywacke fans of the Girvan succession, whilst the central Moffat region at this time was the site of very slow pelagic deposition. This axial zone was progressively encroached upon throughout the Silurian, as the whole Southern Uplands basin was rapidly filled by greywacke turbidites of northerly derivation[16].

In the Lake District the Coniston Limestone Group was succeeded

conformably by Silurian rocks, again of greywacke lithology, and derived from the south, where the Irish Sea Geanticline remained a positive feature. The recent dating of part of the Kirkcudbright succession of greywackes and shales as Upper Llandovery by Walton[17], means that there is only 30 miles of unexposed ground between the Lower Silurian of the Cross Fell Inlier and the most southerly outcrop of the Southern Uplands succession, so the postulated continuity of the Skiddaw-Southern Uplands depositional basin at this time is not unreasonable.

The final episode in the evolution of the Southern Caledonides came at the end of the Silurian with the emplacement of high level granitic plutons into the Skiddaw and Southern Upland successions. As in North Wales, where the Arenig andesite vulcanism was followed after roughly 50 m.y. by granitic intrusion and ignimbrite eruption (Caradoc age), so in the Lakes and the Southern Uplands. The andesite vulcanism (Llandeilo) is succeeded, again after 50 m.y., by the granites of Criffel, Dalbeattie, Skiddaw, Eskdale, etc., of end-Silurian age[18]. Despite the absence of the contemporaneous ignimbritic eruptions in the Lakes or the Southern Uplands, the parallels with North Wales are still so similar as to suggest the existence of a magmatic cycle of roughly 50 m.y. duration, and the operation beneath the Lakes and the Southern Uplands of a similar mechanism to that invoked for North Wales, namely a rising "heat plume" from the down-going slab which induced partial melting at the base of, or within the continental crust.

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CHAPTER THREE

IGNEOUS GEOLOGY

This area does not possess a very diverse igneous history and only two igneous phenomena have been recorded. These are:

i) small intrusions of dolerite-diorite composition, and

ii) innumerable quartz veins, which occasionally bear tourmaline.

A third igneous effect is postulated in the following chapter, and in addition to this, a single example of what is thought to be a volcanic "bomb" was found.

Section 3:1 Minor Intrusions of Basic Composition.

These basic intrusions are restricted to the northern and southern parts of the area, being absent from the intervening ground of the Grasmoor block; their locations are shown on the accompanying map, (Fig. 3:1:1). These bodies have been studied both in the field and in the laboratory, the latter involving X.R.F. analysis of the rocks as well as conventional thin section studies.

The southern intrusions are restricted to the southern flanks of Whiteless Breast, Low Bank and Buttermere Village, and they are quite varied in form, with all gradations from tabular, dyke-like bodies to circular, pipe-like structures occurring. The Buttermere Bridge dolerite (175.170) forms a sheet, 2' thick, dipping gently northwards at the western end of the exposure, and 30 yds. eastwards there is a 3' thick vertical dyke, striking E-W, which probably connects with the sheet to the west. Associated with this sheet are occasional bodies of vein quartz, but the exposure is very overgrown and the relationships are difficult to ascertain.

At Buttermere cross-roads (177170) there are a further two dyke-

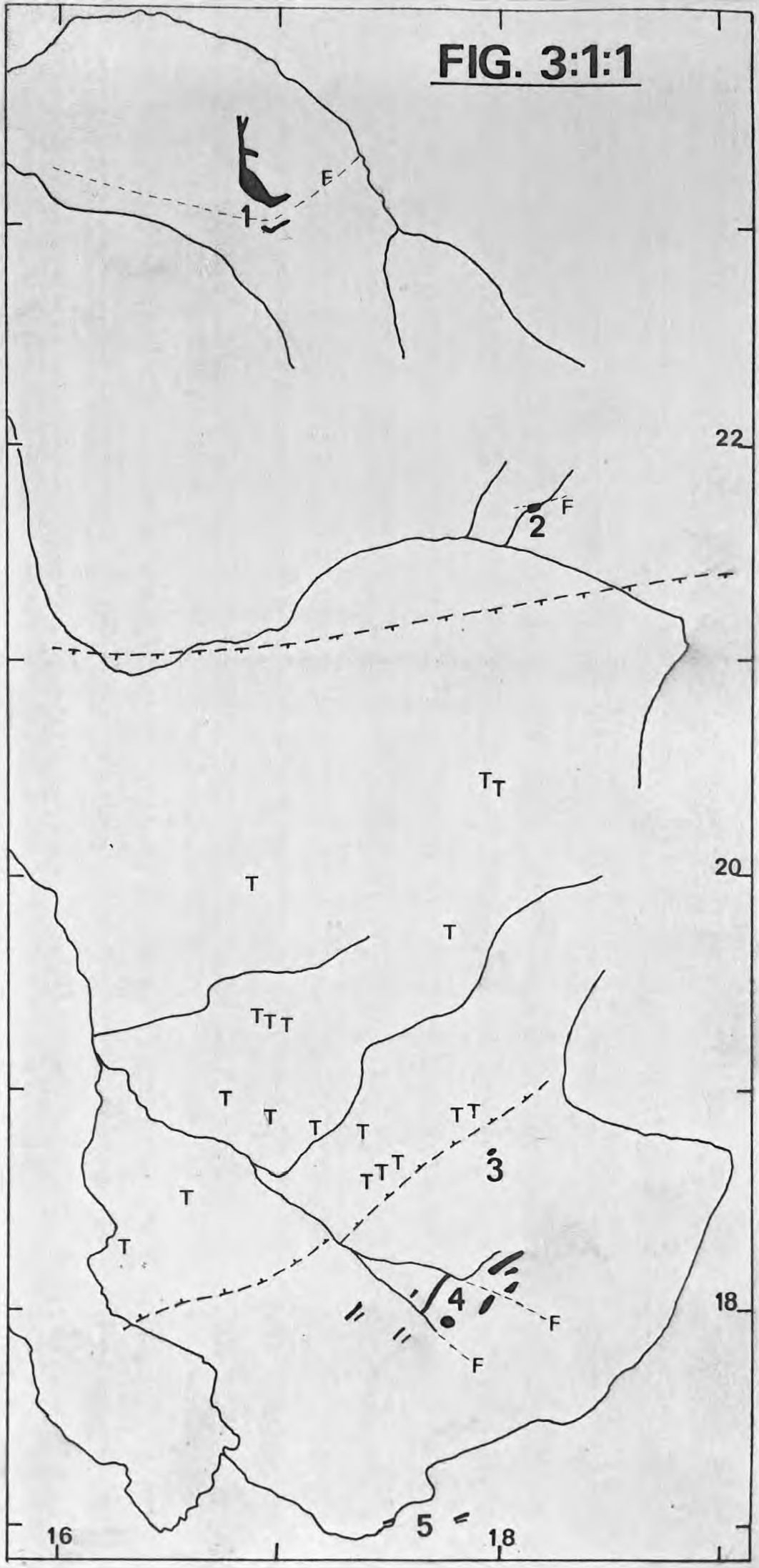
Fig. 3:1:1. Location of the basic igneous bodies.

Key.

1. Dodd dolerite.
2. Thrust - gully dolerite.
3. Whiteless Pike vent.
4. Whiteless Breast - Low Bank dolerites.
5. Buttermere dolerites.

T = presence of tourmaline veining.

= limits of metamorphic aureole.


FIG. 3:1:1
1
F
22
2
F
T T
20
T
T
T T T
T
T
T
T
T T
T T T
3
T
T
4
F
F
18
16
5
18

like bodies, both striking E-W and ranging in thickness from 2' - 3'. The two dykes are slightly divergent to the west and may be connected with the Bridge dolerite but no evidence for this was found. In outcrop the dyke rock is massive and grey in colour (Fig. 3:1:1A), and the slaty country rock shows clear indication of contact strain as the junction with the dyke is approached.

Within the Whiteless Breast - Low Bank areas there are found no less than 10 outcrops of dolerite, although many of them may represent segments of the same intrusive body, but separation by faults and incomplete exposure makes this impossible to confirm. The bodies are elongated towards the N.E. quarter with orientations ranging from 020° - 075°, and are varied in form, both dyke-like and ellipsoidal bodies being present. The ellipsoidal intrusions form quite marked positive features (Fig. 3:1:2), unlike the dykes which are quite difficult to detect. In outcrop the rock is dark, fine-grained and obviously crystalline; most of the outcrops of elliptical form show a strong cleavage fabric in continuity with that of the Slates, and contact strain phenomena occur quite commonly at dyke margins. Ragged masses of vein quartz are often found injected along the junction between the dolerite and the country rock; these quartz bodies may be quite extensive (10'-20' long, 2' thick) though much broken and sheared by subsequent deformation.

High on Whiteless Pike (180.187) is the last of the southern intrusions, a small, much-fractured, body, 20'-30' in diameter, and elliptical in plan. It shows marginal alteration effects and these render the rock pale brown in colour and quite felsitic in appearance; the unaltered central parts are dark brown-grey in colour and look distinctly brecciated. The bedding in the adjacent sediments is deformed around the body, lying parallel to virtually its entire circumference, and contacts with the country rock are everywhere steeply inclined.

Fig. 3:1:1. A. Outcrop of the Buttermere cross-roads dolerite. (G.R. 176.170).

Fig. 3:1:2. The hummocky relief of the dolerite outcrops of Whiteless Breast, seen from Low Bank, looking NE.





Thin section examination of the last intrusion reveals its fragmental character quite clearly. Set within a very fine-grained groundmass of chlorite and secondary quartz are large felts of chlorite material, crowded with tiny zircon crystals (Fig. 3:1:3); these are interpreted as xenolithic blocks of pelite. Small xenocrystic patches of quartz also occur; they are much broken and shattered, but still display a well-rounded outline, and have a fine-grained granophyric texture due to complete recrystallization. This included material is set in a fine-grained groundmass of igneous origin, which has completely recrystallized to quartz, chlorite, sericite and (?) epidote, its texture closely resembling that of the other igneous bodies described below. No felspars have survived but what are thought to be pseudomorphs of chlorite after pyroxene are present (Fig. 3:1:3A); the petrographic composition of the matrix however is impossible to ascertain. However, both field and microscopic evidence would suggest that this intrusion represents a breccia-filled volcanic pipe genetically related to the overlying Borrowdale Volcanic rocks.

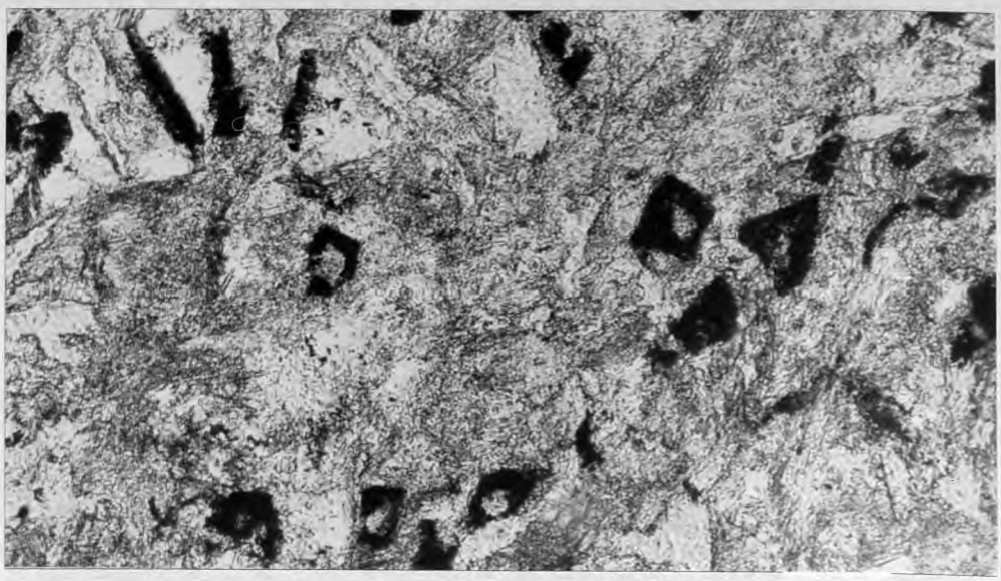
The remaining basic intrusions of Buttermere and Whiteless Breast all show similar alterations to a greater or lesser extent. The freshest specimens still show clear laths of plagioclase felspar, but these have all suffered some degree of saussuritization, breaking down to clinozoisite, sericite and albite. In some cases it is possible to test plagioclase composition by the symmetrical extinction angle method (Michel - Levy), and they fall within the oligoclase-andesine range (angles range from 0° - 25°); u Canada balsam.

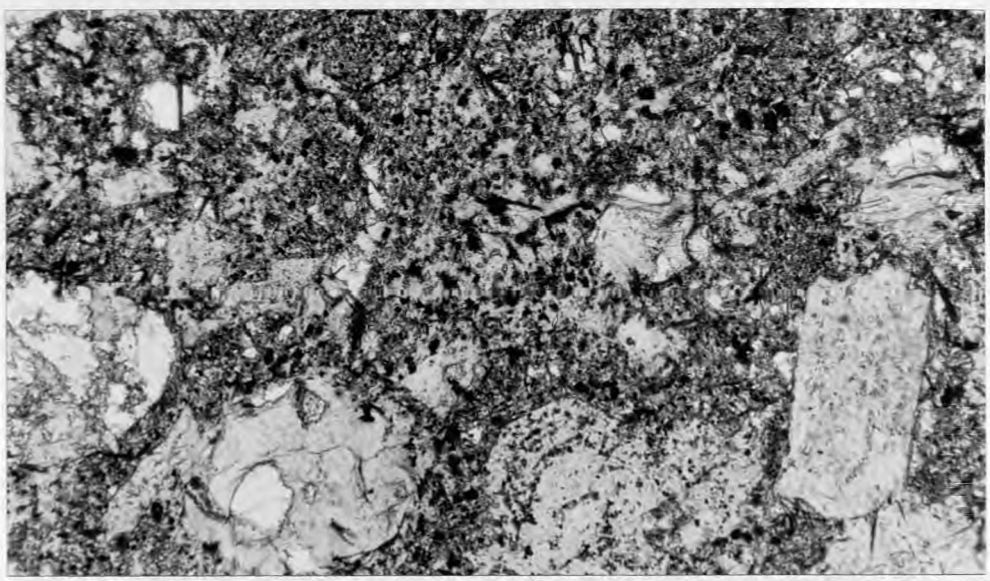
The mafic constituents, with the exception of biotite, are completely altered to a fine-grained felt of chlorite, usually the blue, penninite, variety, which is virtually irresolvable. Quartz and K-felspar also occur, usually as micro-pegmatitic intergrowths (Fig. 3:1:4), though

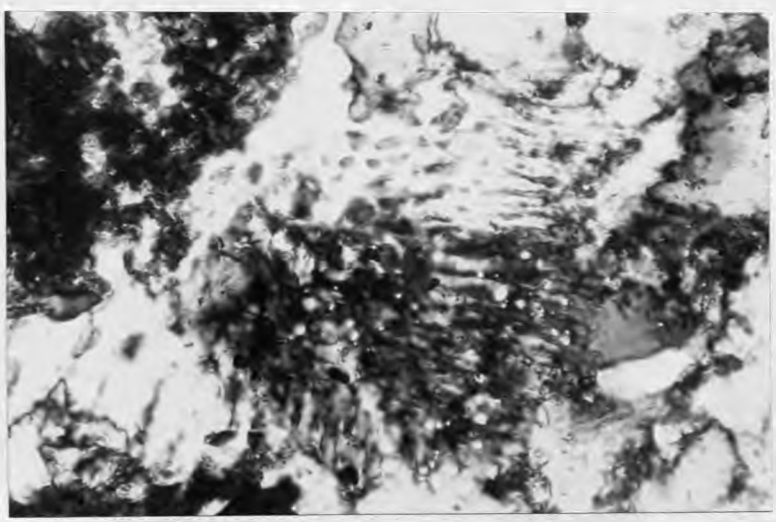
Fig. 3:1:3. Tiny zircon crystals from a pelitic xenolith included within a marginal breccia associated with the large Low Bank dyke. (x 100. PPL)

Fig. 3:1:3. A. Pseudomorphs of chlorite after pyroxene in the vent breccia of Whiteless Pike (G.R. 180.188) (x 25. PPL)

Fig. 3:1:4. Micro-pegmatitic intergrowth of quartz and K-felspar from Squatt Beck dolerite (G.R. 177.180).
(x 100; x Nicols).







whether these are primary or secondary is impossible to say. There are very small sphenes formed from ilmenite, and small quantities of calcite are also present as irregular aggregates of small anhedral crystals. Biotite still survives unaltered in some specimens, though it is very pale in colour, as if bleached, with weak pleochroism, but containing very marked, zircon-shaped, pleochroic haloes.

In the more intensely altered specimens, the secondary growth of carbonate becomes the dominant factor of the petrography, and can occupy up to 60% of the rock. Between these large aggregates; some of which are quite circular in shape and show good radical crystallization (Fig. 3:1:5), the groundmass is made up of mafic minerals completely pseudomorphed by chloritic material, together with much secondary granophyric quartz. There is some indication that the calcitization of these rocks was a relatively early event, because the radially-crystallized, (?) vesicles mentioned above are completely calcitic in composition, with the exception of marginal arcuate outgrowths of chlorite and quartz which are seen to be replacing the earlier calcite (Fig. 2:2:10). A similar phenomenon was described by Lehmann (1965) from the diabases of S.W. England.

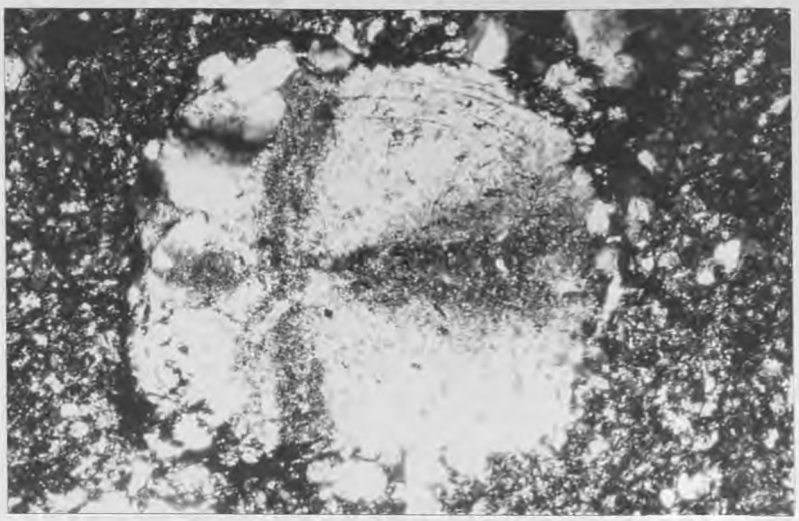
The story for the northern group of intrusions is much the same, with extensive alteration obscuring virtually all of the features diagnostic of the rocks' composition. Plagioclase has been altered to oligoclase, calcite or zoisite (Fig. 3:1:6), and quartz; K-felspar is rarely seen and has probably contributed sericite on alteration; the mafics (amphibole and pyroxene) are completely made over to chlorite, sericite, calcite and exsolved iron ore (Fig. 3:1:7); only biotite rarely survives.

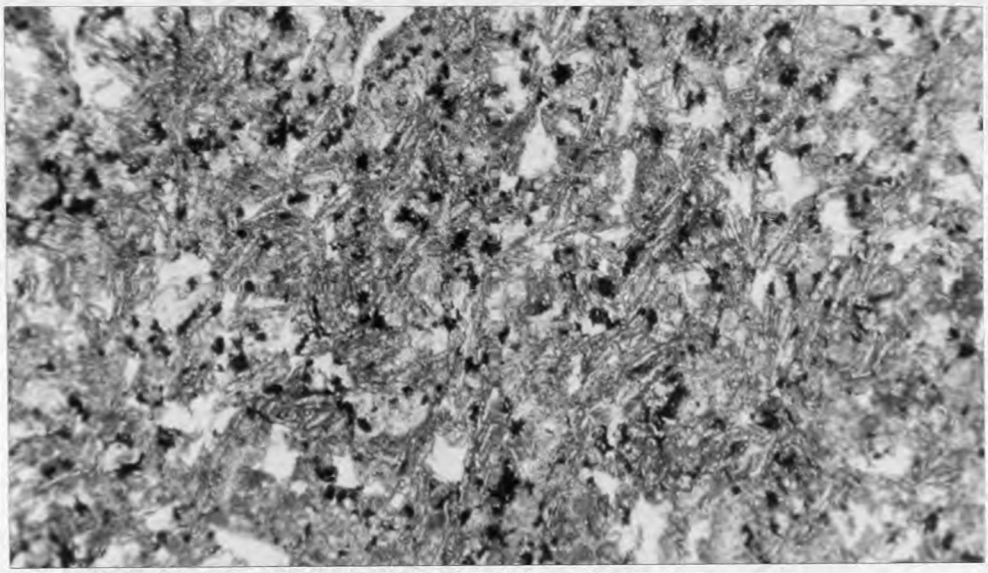
The field relations of these intrusions are more distinct than those in the south. The Gasgale Gill body appears beneath a small over-

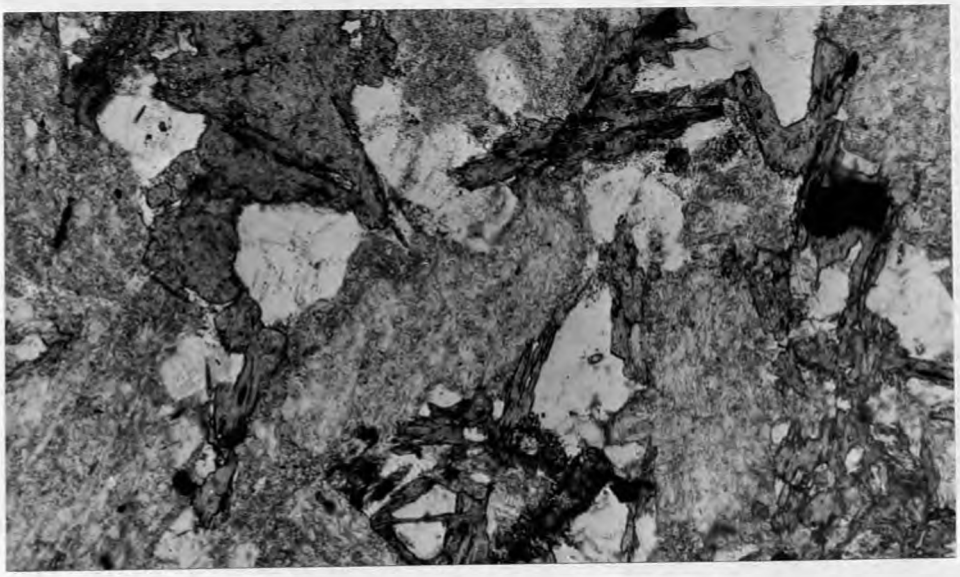
Fig. 3:1:5. Calcite infilling well-formed vesicle, with small blebs of quartz at the margin. Buttermere dolerite (G.R. 176.170). (x 40; x Nicols).

Fig. 3:1:6. Plagioclase alteration in the Dodd dolerite (x 25. PPL)

Fig. 3:1:7. General appearance of igneous rocks - fine-grained alteration products. (x 25. PPL).







thrust mass of Skiddaw Slate in a deeply-cut, much vegetated, tributary valley. Relationships at the non-tectonized junctions are difficult to decipher but they seem to be steeply dipping and cross-cutting the strike of the surrounding sediments. The rock is pale grey in colour, medium grained and very massive.

The Dodd intrusion is clearly defined on the north face (Fig. 3:1:8) of the fell, but on the summit and the S.E. face it is obscured by vegetation, though at the latter locality a trail of boulders betrays its position underground. The intrusion is sheet-like in form, dipping N.E. at 50° - 60°, and is somewhat wider in the centre, around the summit of Dodd, and narrower towards the N. and S.E. margins. Two lateral apophyses occur on the northern face, and being concordantly emplaced along the bedding, these offshoots have a gentle northerly dip. Across Dodd Pass, on the gentle slopes to the south, another offshoot occurs, again sill-like in form, but the relationships of this body to the parent are obscured by poor exposure and the intervening fault although its continuation downhill can be mapped where it produces the slight positive feature that marks the western end of Dodd Pass (G.R. 169.229). The steeply inclined junctions between the Slates and the intrusion show marked contact-strain effects (Fig. 3:1:8), with cleavage increasing in intensity, and bedding increasing markedly in dip (and even overturning) close to the junction.

Geochemistry

A geochemical study of these basic intrusions was undertaken to discover if there was any systematic variation in their composition across the area, and whether this was related to geographical distribution of the bodies, (see Fig. 3:1:1). Also investigated was the possibility of establishing the relationships of the intrusives to the succeeding

Fig. 3:1:8. Outcrop appearance of the Dodd dolerite intrusion.

A. N. face of Dodd; steeply inclined dolerite sheet forming marked positive feature. (G.R. 168.235)

B. Foot-wall contact of the Dodd dolerite with Hope Beck Slates. The Slates increase in dip towards the junction (A-A') and become quite closely cleaved parallel to the junction.





TABLE 3:1:9

COMPARISON OF THE MEAN COMPOSITIONS OF THE NORTHERN AND SOUTHERN ROCKS, AND THE PROPOSED VENT.

	MEAN NORTH	MEAN SOUTH	VENT
SiO_2	52.65	47.44	57.11
TiO_2	1.29	1.17	1.10
Al_2O_3	17.26	13.50	13.81
Fe_2O_3	13.68	14.99	15.90
MnO	0.30	0.32	0.23
MgO	5.95	8.79	5.15
CaO	1.10	4.95	0.90
Na_2O	2.28	0.99	0.30
K_2O	1.26	0.59	0.53
P_2O_5		no difference	
Cr	145	459	932
Ni	55	121	191
Cu	190	182	204
Zn	134	113	99
Ga	14	12	11
Rb	78	31	29
Sr	154	105	62
Y		no difference	
Ba	311	214	127
La	25	19	12
Ce	-	-	-
Pb	27	16	21
Th		no difference	

Borrowdale Volcanic rocks, in view of the recent discoveries of the marked differences in age and composition of the Eycott and Southern outcrops. (Fitten and Hughes, 1970; Downie and Soper, 1972).

Six specimens of the northern group and nine from the southern group were taken for analysis, one of the southern specimens being from the presumed vent on the upper slopes of Whiteless Pike (spec. no. 9012). The techniques of sample preparation and analysis are outlined in Appendix A, and the full list of results is given in Appendix B. From these analysis lists, but more easily from the mean compositions (Table 3:1:9), it can be seen that the northern and southern groups are significantly different in both major and trace element composition. The composition of the vent, 9012, has close affinities with the northern rocks, despite its occurrence in the south of the area.

Graphical representation renders the distinction between northern and southern intrusions even clearer; Fig. 3:1:10 shows plots of MgO, CaO, Na_2O+K_2O, and Al_2O_3 against silica; and Fig. 3:1:11 shows the variation of Sr against Rb, and Sr against Ba. Also plotted on the Harker diagrams are the compositions of the nodules described in the previous chapter; all but one of them are very siliceous in composition, but the exception, specimen 9001, falls consistently within the field of the southern igneous rocks and is therefore considered to be of igneous origin, i.e. a volcanic fragment.

The analyses of these rocks has enabled, therefore, their division into two compositional groups, one from the north, and one from the south of the area. Bare analyses, however, mean little in petrographic terms, and so they were re-cast in terms of C.I.P.W. norms using the departmental NORM 3 program, which also computed various common petrographic ratios and indices. These results are tabulated on Table 3:1:12.

Normative recalculation is especially useful in this case because

Fig. 3:1:10. Variation diagrams showing the distinctions between the N - S igneous groups, the concretions and the vent.

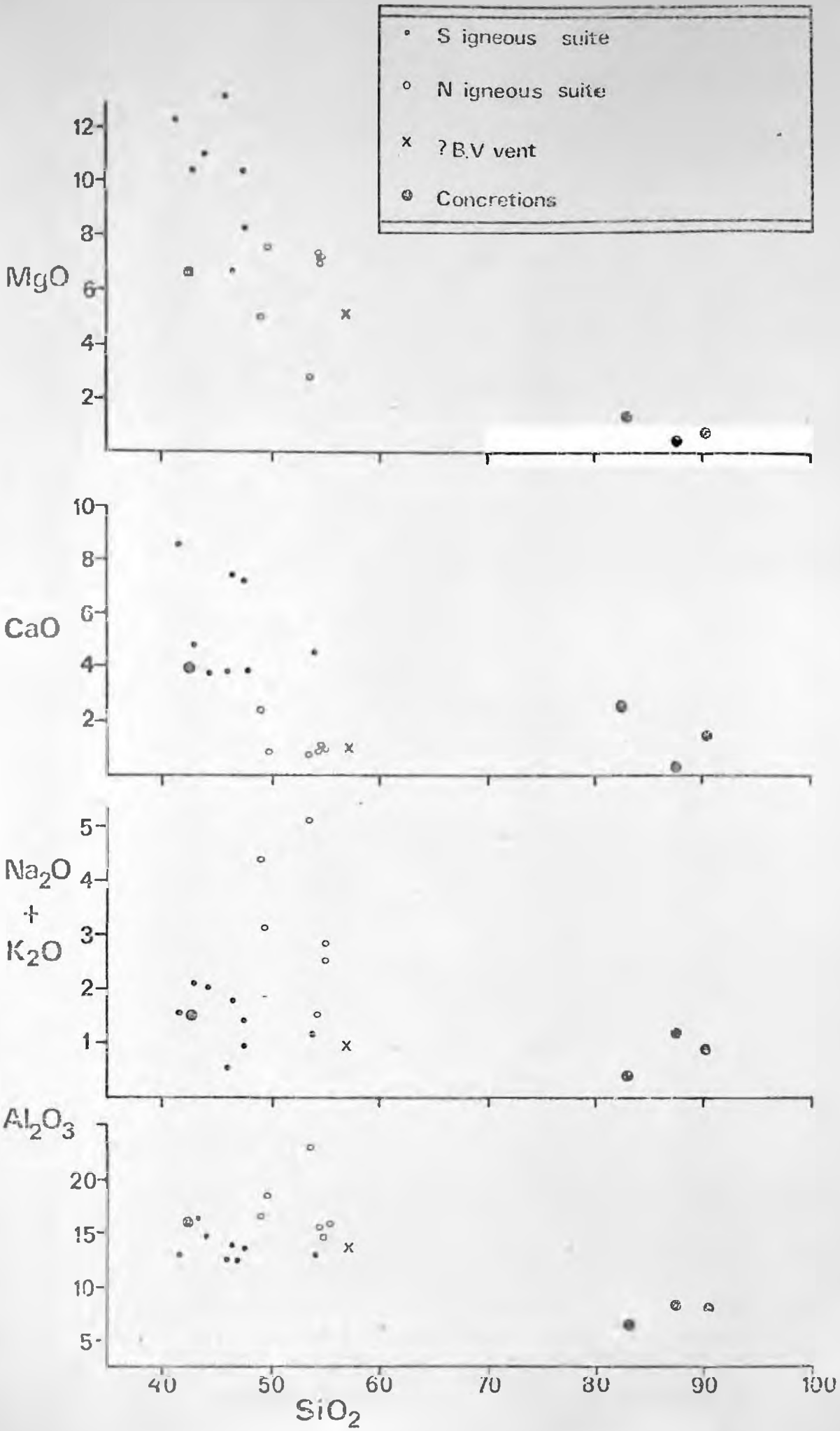
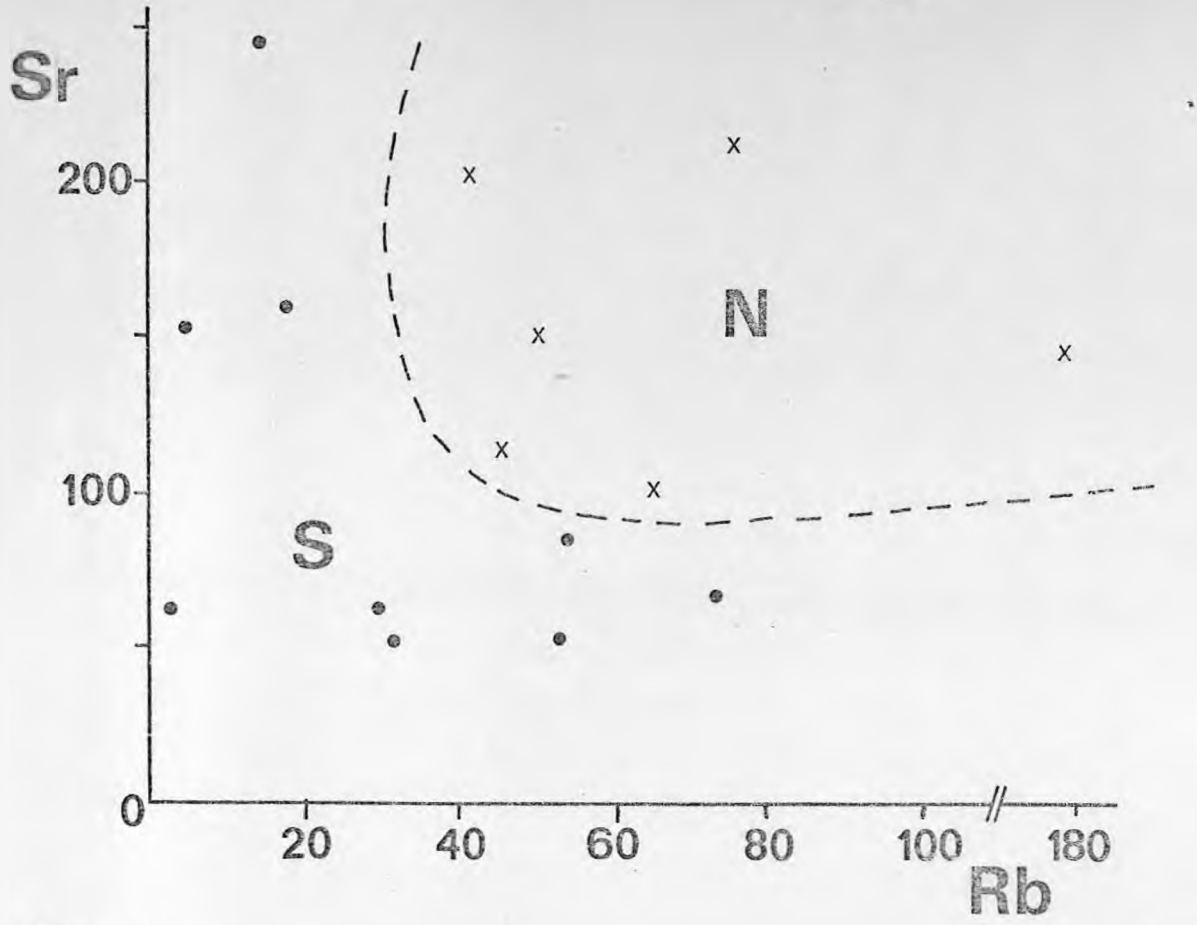

S igneous suite
N igneous suite
? B.V vent
Concretions
MgO
CaO
$Na_2O + K_2O$
Al_2O_3
SiO_2
40
50
60
70
80
90
100

FIG. 3:10

Fig. 3:1:11. Sr/Rb and Sr/Ba diagrams indicating the differences between the N. and S. igneous rocks.

FIG. 3 (II)


Sr
200
100
0
N
S
20
40
60
80
100
180
Rb

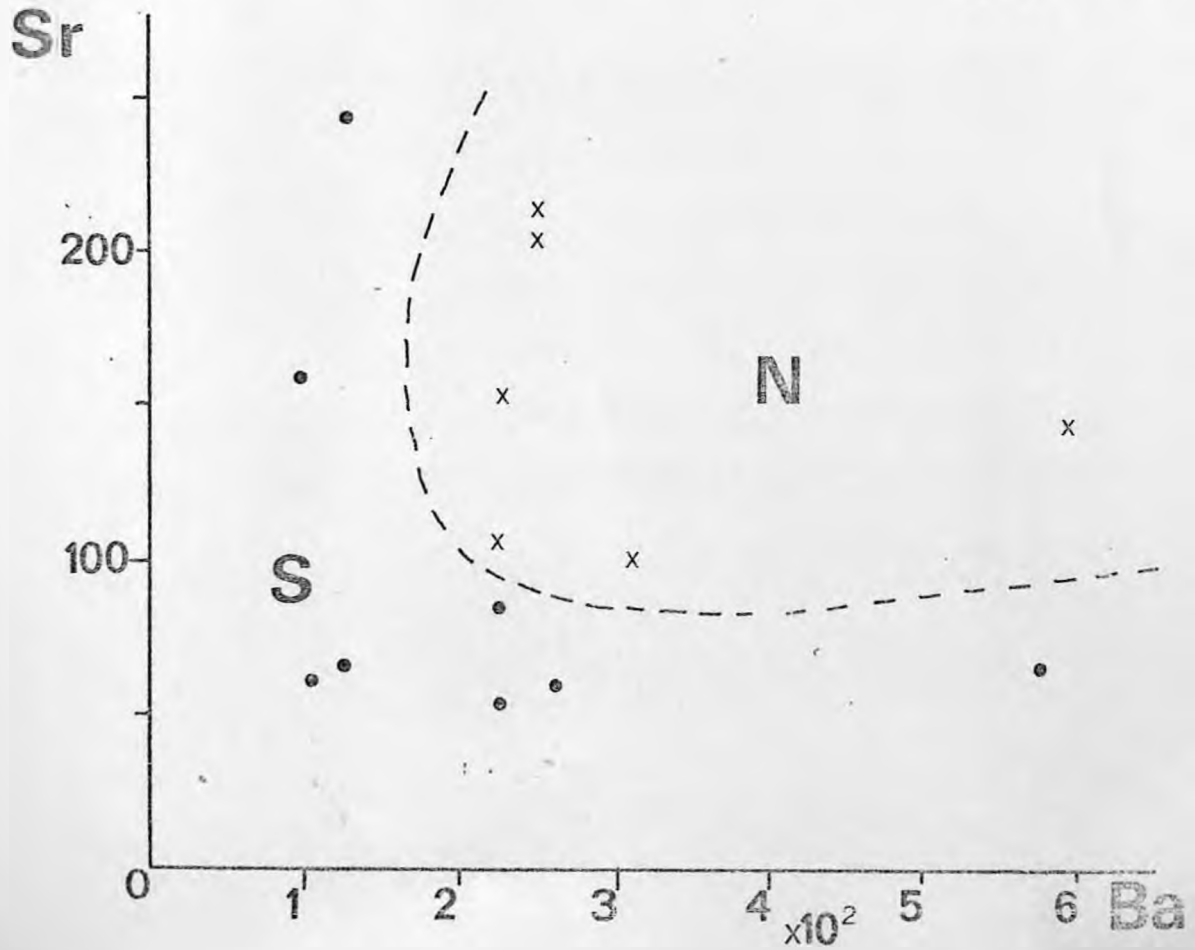

Sr
200
100
0
N
S
1
2
3
4
x10²
5
6
Ba

Fig. 3:1:13. $Al_2O_3/Na_2O + K_2O$ diagram (after Kuno, 1960) shewing the positions of the mean compositions of the N. and S. igneous rocks relative to the 3 classes of basalt erected by Kuno.

Thol. = Tholeiitic basalts

H.A. = High Alumina basalts

Alk. = Alkali basalts.

Solid line marks the class divisions for basalts with SiO_2 50.00% (i.e. S. group). Broken line marks the class divisions for basalts (andesites) with SiO_2 50.00% (i.e. N. group).

Fig. 3:1:15. AFM diagram shewing the relationships of the N. rocks (dots) and the S. rocks (crosses) to the compositional fields described by Fitton and Hughes (1970) for the Eycott Volcanic Group ("N", and solid line) and the main Borrowdale Volcanic Group ("S", and broken line).

A good correlation between my N. igneous rocks and the Eycott Volcanic Group is suggested.

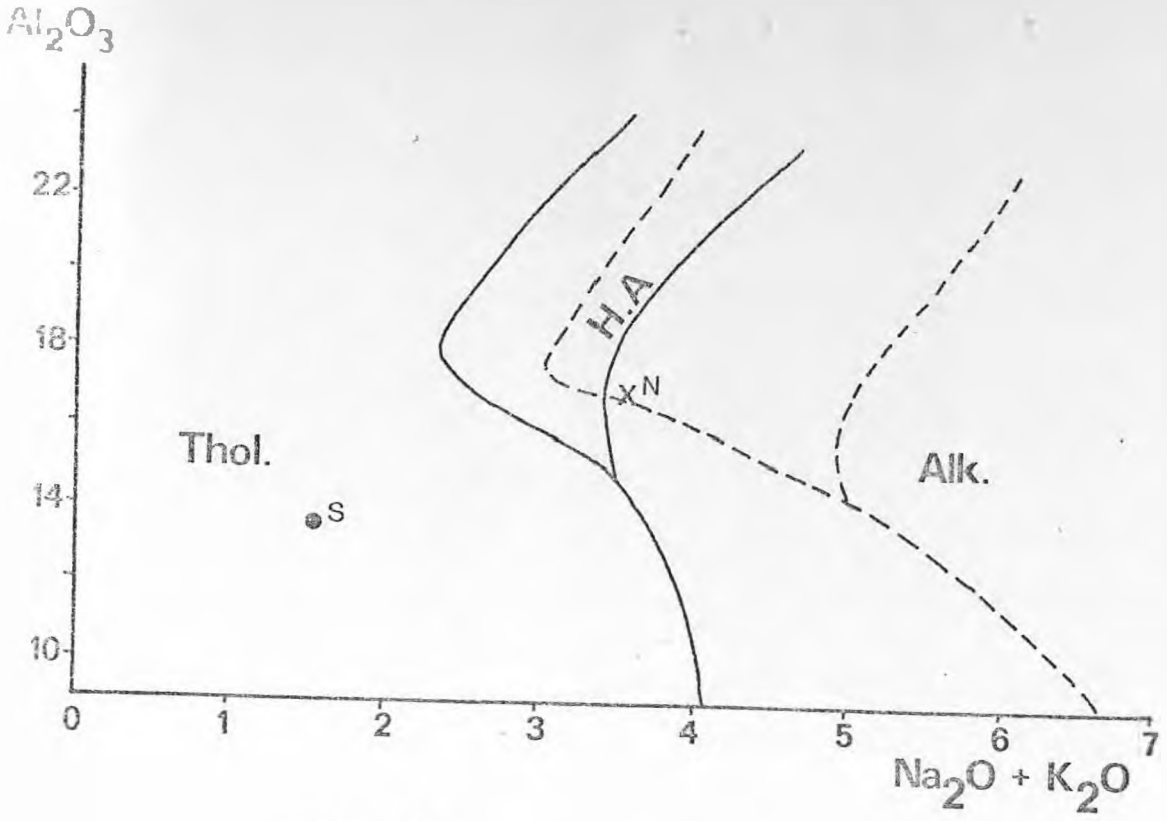


FIG. 3:1:13

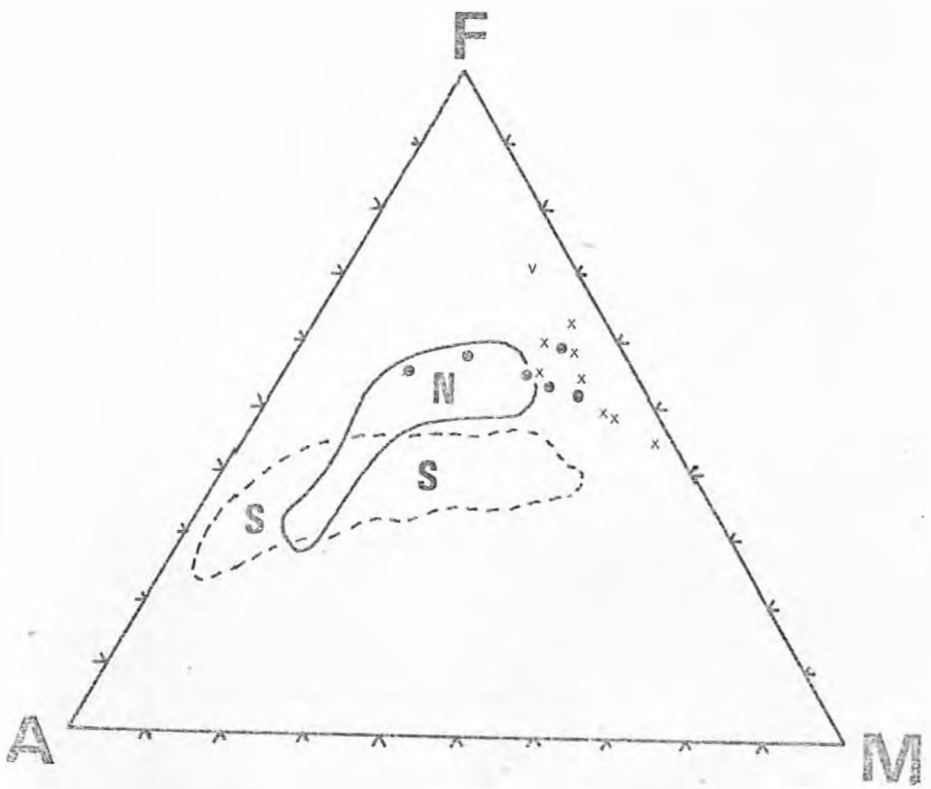
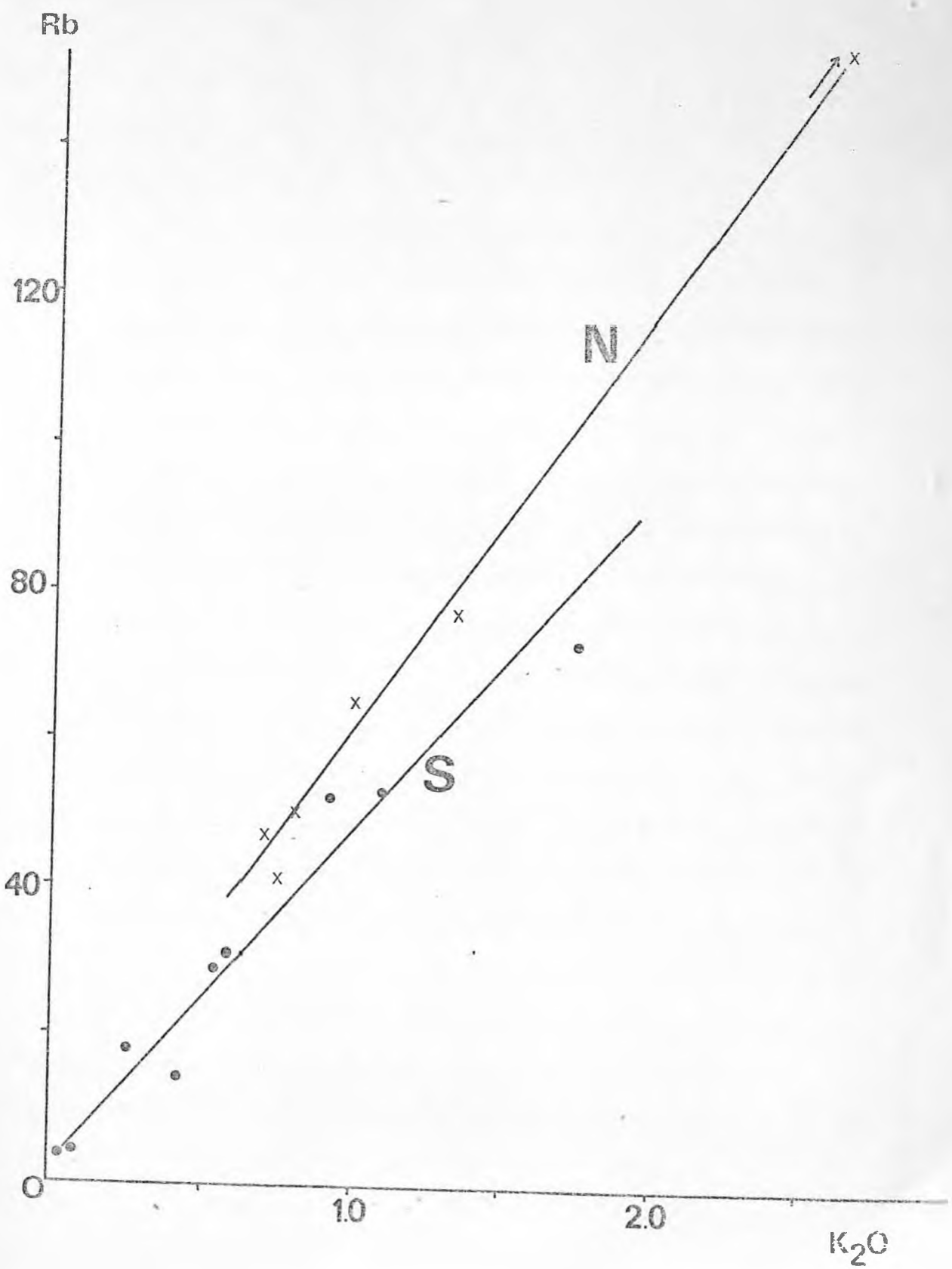


FIG. 3:1:15

Fig. 3:1:14. Graph of K_2O/Rb. The steeper slope of the N. igneous rocks line suggests their more highly differentiated condition.

FIG. 3114



the rocks were so extensively altered that nothing of their original mineralogy could be discerned. The norms must be interpreted with care, however, if misleading conclusions are not to be drawn. The values of OR are too high because it contains all the K_2O recalculated from the modal sericite; likewise, the presence of corundum, C, and spinel, SP, in the norm reflects the recalculations of Al_2O_3 from the modal sericite and chlorite, though this latter is more marked in the northern rocks where CaO concentrations are too low to take up the excess alumina as anorthite. The significance of individual normative analyses is probably low, in the view of the unknown effects of hydrothermal alteration upon the original chemistry, but taken as a group the mean normative analyses probably reflect quite accurately the relative differences between the northern and southern groups of intrusives. Thus, the southern group is much less saturated with silica than the northern group; though that they should be oversaturated at all is remarkable in view of their very low silica contents. Only one specimen (9009) was, in fact, under-saturated, as revealed by the presence of normative olivine. The higher values of normative anorthite and hypersthene in the southern rocks also reflects their more basaltic nature. (The vent rock, 9012, is omitted from these considerations because of the unknown effect its chemical and normative composition of the pelitic inclusions).

Given this data, it is possible for an estimate to be made of the likely original mineralogy of these two groups of rocks. The PER-AN values for the northern group averages 24.7, and for the southern group, 78.9; this is taken to indicate a plagioclase composition of oligoclase-andesine for the northern intrusives, and of labradorite-bytownite for their southern equivalents. The southern rocks were probably slightly undersaturated, with occasional modal olivine; the northern rocks were probably slightly oversaturated with accessory quartz appearing in the

mode, though the anomalously high Q values when compared with the low SiO_2 percentages are difficult to interpret in this respect.

By analogy with Kuno's (1960) work, the southern rocks are intensely tholeiitic in character, whilst the northern rocks, though just within the High Alumina class, also show marked tholeiitic affinities, (Fig. 3:1:13). Following MacDonald and Katsura (1964) the southern rocks would be tholeiitic basalts, and the northern rocks would fall within the hawaiite-mugearite field, despite the fact that the soda : potash ratio is somewhat less than the required 2:1 (in fact is 1.7:1).

That the northern intrusives crystallized from a more highly differentiated magma than those to the south is suggested by several independent factors:

i) The higher differentiation index (D.I.) for the northern group (mean = 52) than for the southern group (mean ; 27.4).

ii) The steeper gradient for the northern line compared with that for the southern rocks on the K_2O/Rb graph, Fig. 3:1:14; this enrichment in Rb in the northern rocks indicates their more highly differentiated condition.

iii) The generally higher concentrations of the lithophilic elements, both major and trace, within the northern rocks, including SiO_2, Al_2O_3, K_2O, Na_2O, Sr, Rb, Ba, Pb, Zn, Ce, La. All of these elements are usually enriched by progressive fractionation, (Taylor, 1964).

iv) Conversely, the southern rocks show higher concentrations of the siderophilic elements, Fe_2O_3, MnO, MgO, CaO, Ni, Cr; i,e. those elements which are usually concentrated in the early crystallization products.

The relationships of these basic intrusions to the Skiddaw Slate

country rock, and to the succeeding Borrowdale Volcanics, are more difficult to define. Both northern and southern groups were emplaced before F2, the first end-Silurian deformation. This is evidenced by the presence of S2 cleavage within the intrusions; the contact strain at their margins; and the truncation of the Gasgale intrusion by an overthrust of F2 age. The intrusions must post-date the deposition of the Skiddaw Slates, and this narrows down the possible time range for their intrusion to lowest Llanvirn to end-Silurian.

Comparison of the chemistry of these basic intrusions with that of the Borrowdale Volcanics is hindered by the lack of published analyses for the latter rocks. Fig. 3:1:15 shows the relationships of the northern and southern groups to the analyses plotted by Fitton and Hughes (1970). It can be seen that the analyses for the northern rocks fall very close to the basic end of the compositional field of the Eycott Volcanic Group (Downie and Soper, 1972); on this basis the suggestion is made that these two groups are co-sanguinous, and probably co-eval. The presence of a vent of similar composition, albeit in the south of the area, is appealing in this respect, and the age of the northern intrusives, on this basis, would be of lower Llanvirn age (D. bifidus zone).

The relations of the southern intrusives are more difficult to ascertain. They show no marked relationship to either the Eycott Volcanic field or the main Borrowdale Volcanic field. In view of the lack of pillow structures and their very low Na_2O contents, I think it unlikely that they are of spilitic origin. More likely it is that they represent small subjacent plutons associated with either the Eycott or Borrowdale vulcanicity, but being composed of a magma that had been depleted in the lithophilic elements by fractionation and volcanic eruption.

Section 3:2 Quartz and Tourmaline Bearing Veins.

Quartz veins are present in great numbers throughout the area but they are small, irregular features and have not been studied in detail. They are found along the bounding surfaces of several of the dolerite intrusions; along the major joint planes (dominantly of a northern trend); and along the horizontal cleavage and shear planes formed by the second deformation. The latter are often found to terminate at the bed boundary and these are probably a "sweat" type of vein, the quartz being derived from the immediately adjacent rocks: the larger quartz veins and the tourmaline-bearing veins described below, are probably of more distinct origin.

Tourmaline-bearing veins are restricted to the central, aureole area (Fig. 3:1:1), being concentrated on Whiteless Pike, Low Bank, and Lad Hows; only scattered fragments were found on the screes below Dove Crags. The veins are matt black in colour, though often having white, quartz, margins or small white veinlets cross-cutting normal to the walls of the vein, (Fig. 3:2:1). The vein patterns are complex, with much anastomozing of small offshoots, but the vein systems can often be traced for considerable distances, up to 400' vertically in the case of the Whiteless Pike example. The veins vary in width from 1cm - 10cm and cut across fold and cleavage structures of F2 age, but they are occasionally seen to be folded into open, steep-plunging, F4 buckles, and hence their age can be narrowed down quite closely.

A specimen of a folded vein was obtained and in thin section most of the tourmaline crystals were seen to be orientated with their crystallographic c axes parallel to the tectonic b axis of the fold (Fig. 3:2:2). This indicates synchronous deformation and crystallization during F4. The tourmalines are light green in colour, up to 300-400u in size, and

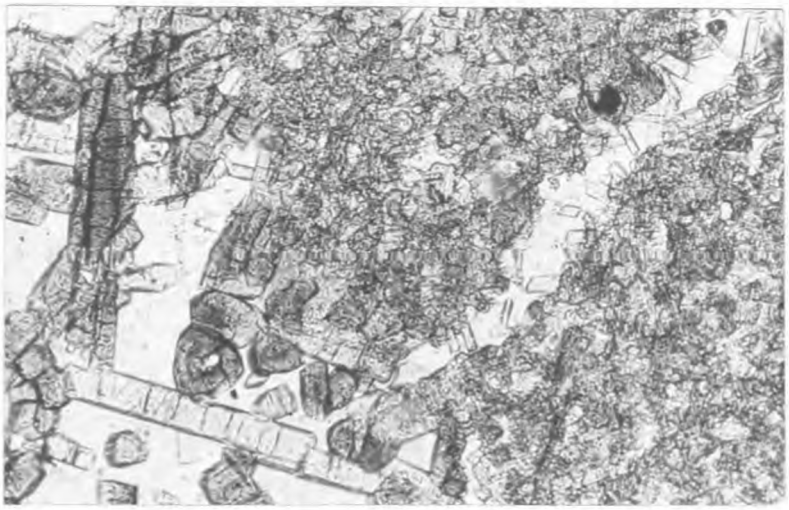
Fig. 3:2:1. Field appearance of tourmaline vein.

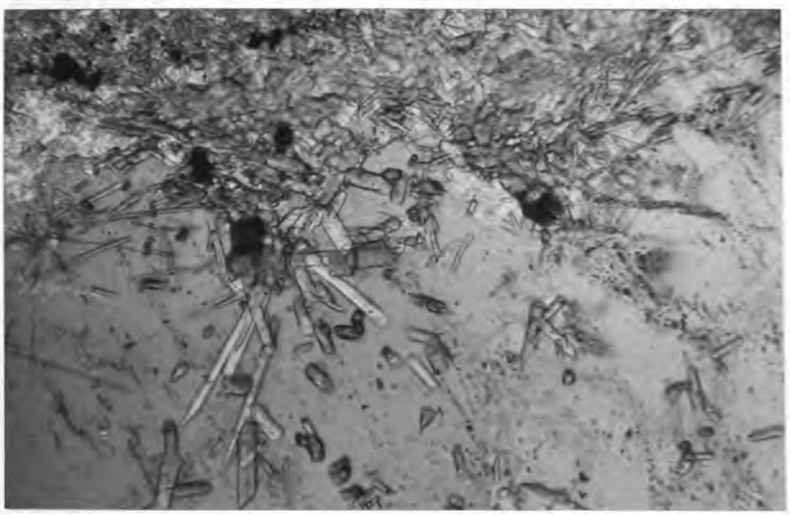
Fig. 3:2:2. Thin section from folded tourmaline vein, showing the orientation of the crystallographic c-axes of the crystals parallel to the tectonic b-axis of the fold.

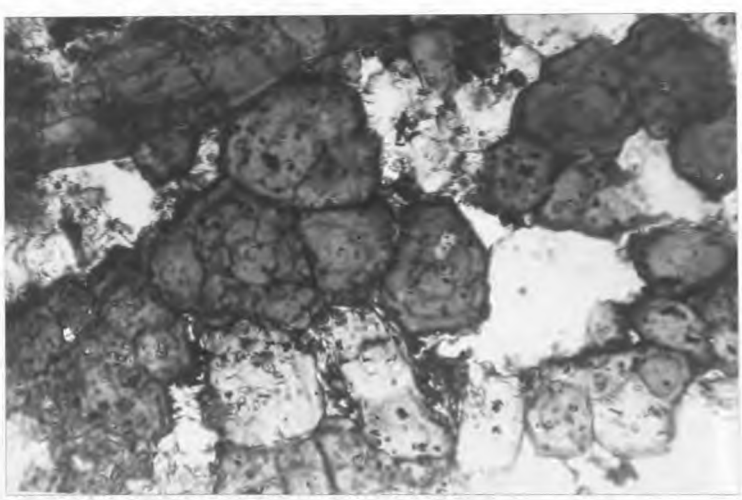
(x 40. PPL)

Fig. 3:2:3. Tourmaline needles growing out from wall-rock into quartz vein (x 40. PPL).

Fig. 3:2:4. Tourmalinized country rock. (x 60. PPL)









disposed in more or less regular bands across the vein, with the coarser crystals towards the centre. Several unfolded veins were sampled and they showed the more common habit of tourmaline to be a very finely crystalline groundmass, with interstitial quartz. The wall rocks were also tourmalinized to a limited extent. Where these tourmaline-rich areas are broken by smaller later, transverse veins of pure quartz, there has occurred at the interface between the quartz vein and the tourmaline vein, the growth of tourmaline needles from the wall rock out into the quartz vein (Fig. 3:2:3). Some of the veins contain included blocks of country rock; these are also heavily tourmalized with virtually all the initially chloritic matrix converted to tourmaline, only the detrital quartz remaining unchanged (Fig. 3:2:4).

The presence of this tourmaline is an important factor in considering the affinities of the proposed granitic body of this area with the other granitic intrusions of the Lake District. The only granites characterized by tourmaline are the Skiddaw and Eskdale Granites; the much closer Ennerdale Granophyre has none.

CHAPTER FOUR

METAMORPHIC GEOLOGY.

The rocks of this area have been affected by two independent metamorphic events. Firstly, there is the low-grade regional metamorphism of the area as a whole; and subsequently, the rocks of the central region were subjected to low-grade thermal metamorphism. In addition, the sediments around the Dodd Dolerite also show evidence of thermal metamorphism.

Section 4:1 Regional Metamorphism.

The correlation of fabric and metamorphic events, as seen in thin section, with the sequence of deformational episodes, is shown in Table 4:1:1 (note that Simpson's F1 is equivalent to my F2). From this it can be seen that there was no fabric-forming element developed between the deposition of the Skiddaw Slates and the regional metamorphic climax, the pre-Borrowdale F1 event having no visible metamorphic or fabric effect on these rocks.

All the sediments of the area have mineral assemblages containing quartz, chlorite and sericite, which are characteristic of the lower greenschist facies (B.1.1. in Winkler's (1967) nomenclature). Wherever the detrital quartz grains are in mutual contact, there solution, migration and recrystallization has occurred, producing mosaics of quartz crystals with interlocking sutured margins. Usually, however, the quartz grains are dispersed in the chloritic matrix and then they show quite corroded outlines (Fig. 4:1:2) indicating that reaction between the quartz and the chlorite of the matrix has taken place.

Sericite is the principal phyllosilicate present. In the pelitic bands the sericite has crystallized as lepidoblastic felts mimetic after bedding (Fig. 4:1:3); in the silty bands sericite is less dominant and

AGE	DEFORMATION EVENTS	IGNEOUS EVENTS	REGIONAL METAMORPHISM AND DIAGENESIS AND FABRICS	CONTACT METAMORPHISM
END - SILURIAN			QUARTZ - CHLORITE VEINS	
				TOURMALINE BEARING VEINS
	F4		NO FABRIC ELEMENT	
	F3		LIMITED RECRYSTALLIZATION OF SERICITE PARALLEL TO S3	
	F2 ENE -WSW folding		DEFORMATION OF CHLORITE PORPHYRO - BLASTS : SEGREGATION OF CHLORITE AND MINOR SERICITE PARALLEL TO S2	DEFORMATION OF DODD SPOTS
		GRANITE INTRUSION 395 m.y.		CONTACT METAMORPHISM OF GRASMOOR AREA: DEVELOPMENT OF BIOTITE AND SPOTS
			REGIONAL METAMORPHIC CLIMAX - GROWTH OF CHLORITE PORPHYRO - BLASTS, FELTS AND LENSES	
andeilo		Borr. Volc. Gp. N. Igneous bodies & breccia pipe		DODD CONTACT METAMORPHISM
	F1 N-S folding		NO FABRIC ELEMENT	
lanvirn	DEPOSITION OF SKIDDAW SLATES & SLUMP FOLDING		CRYSTALLIZATION OF DIAGENETIC PRODUCTS (ILLITE AND CHLORITE) PARALLEL TO BEDDING	

FIG. 4:1

tends to form discrete isolated plates of quite large size which do not show the same high degree of parallel orientation as do the fine sericite felts in the pelitic bands. This is probably a function of the lower ductility of the quartzose bands.

The other phyllosilicate mineral present is chlorite, usually of the penninitic variety, characterized by the light green-yellow pleochroism and the anomalous "Berlin blue" interference colours. It is present in several forms:-

i) As the very finely crystallized groundmass in the silty beds, where it is practically irresolvable, and is probably the dominant mineral, (Fig. 4:1:2).

ii) As discrete segregations made up of small tabular crystals; these bodies are rarely ellipsoidal in shape and elongate parallel to the bedding; more usually they are irregular in shape and transverse to the bedding - they are probably "sweat" segregations, (Fig. 2:2:4).

iii) As tabular crystals mantling the large pyrite crystals described above, having the cleavage orientated perpendicular to the pyrite faces. This suggests initiation of the growth of chlorite at the pyrite/matrix interface due to the availability of iron, and continued growth outwards having the path of easiest diffusion (the cleavage in the chlorite) perpendicular to the pyrite faces (Fig. 2:3:6).

iv) As discrete crystals, often up to 100u in length, oriented roughly parallel to the bedding, and occasionally showing alteration along their cleavage to a brown, pleochroic, highly birefringent mineral (Fig. 4:1:4). The identity of this mineral is discussed in more detail below.

Also found within these rocks, though not universally present, are epidote and the above-mentioned brown pleochroic mineral that forms along the cleavage planes of the large chlorite porphyroblasts. Epidote, where present, occurs as a light green, granular, very fine-grained matrix

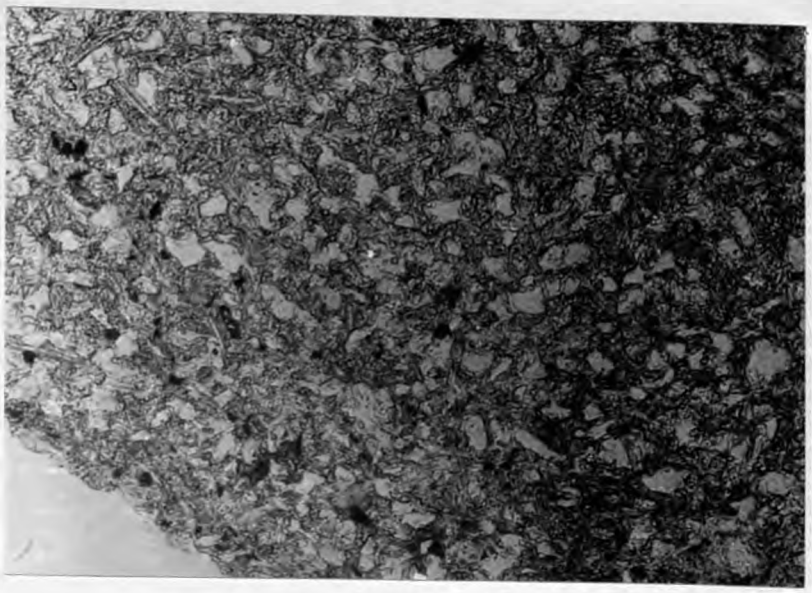
Fig. 4:1:2. Detrital quartz grains showing cusped margins indicative of resorption and reaction with the chlorite matrix. (x 100. PPL) (see also Fig. 2:2:4)

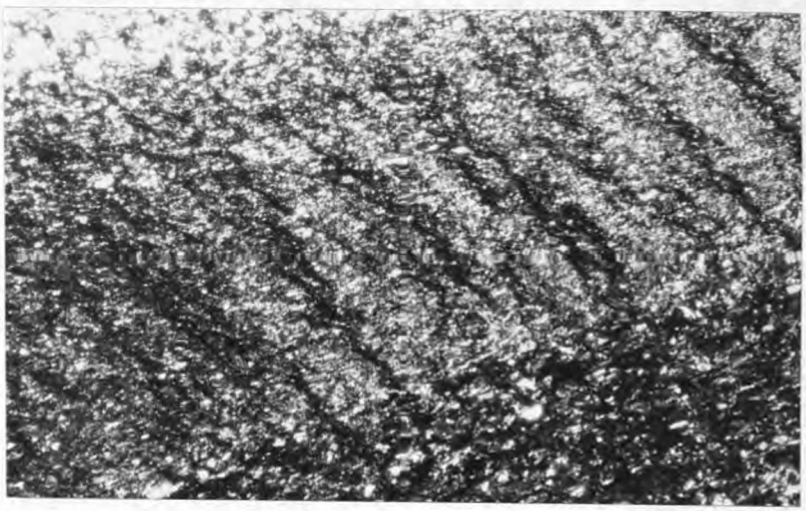
Fig. 4:1:3. Growth of mica and chlorite porphyroblasts parallel to bedding.

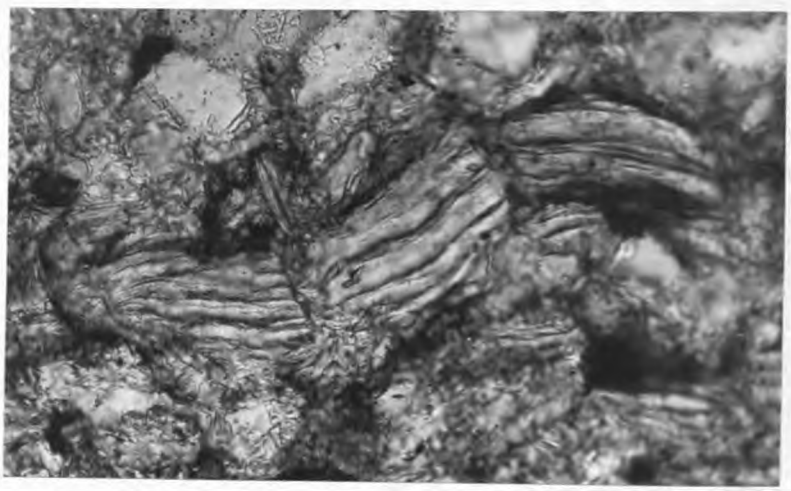
A. Pelitic band; sericite dominant, deformed by subsequent S_2 crenulated cleavage (x 25; x Nicols)

B. More silty band; chlorite megacryst deformed by S_2 fracture cleavage (x 25; PPL).
Note alteration of chlorite to dark oxy-chlorite parallel to the mineral cleavage.

(see also Figs. 2:2:2; 2:3:4)







mineral and is usually confined to the coarser silty bands where it has crystallized from the earlier detrital matrix, possibly due to the metamorphic breakdown of contained felspar fragments.

The brown pleochroic mineral could be either biotite or an oxydized chlorite-vermiculite. Reports by McNamara (1965, 1966) of widespread biotite in the chlorite zone of the Otago Schists were based on the mis-identification of a highly pleochroic chlorite-vermiculite. Chatterjee (1967) found a similar material in the Italian Alps which was shown by X-ray studies to be oxydized, iron-rich chlorite: this same material was also identified, wrongly, as stilpnomelane growing at the expense of chlorite by Turner (1934). Mather (1970) mentions biotite growing interleaved with chlorite in the chlorite zone of the Scottish Dalradian, and this would seem the same material as Chatterjee has shown to be oxydized chlorite. This elusive brown mineral would therefore seem to have a considerable history of mis-identification; it occurs in the Skiddaw Slates as cleavage-parallel growths within large chlorite crystals, often in close proximity to haematite-limonite veins, and hence Chatterjee's (1967) identification of the mineral as oxydized chlorite is followed. The transformation of the penninite into this oxydized state results from the oxidation of $MgFe^{2+}$ to $MgFe^{3+}$ with consequent expulsion of H^{+} to maintain charge balance.

The metamorphic assemblage characteristic of these sediments is therefore quartz-chlorite-sericite$^{+}_{-}$epidote, and the difficulty lies in deciding what part of the mineral growth was metamorphic and which was of earlier, diagenetic, origin.

The changes in mineral composition of pelitic sediments with increasingly deep burial have been summarized by Muller (1970), and are shown diagrammatically in Fig. 4:1:5. It can be seen from his diagram that both illite (sericite) and chlorite are present as initial detrital

Fig. 4:1:5. Diagenetic changes in mineral composition of pelitic sediments with increasingly deep burial. (After Muller, 1970).

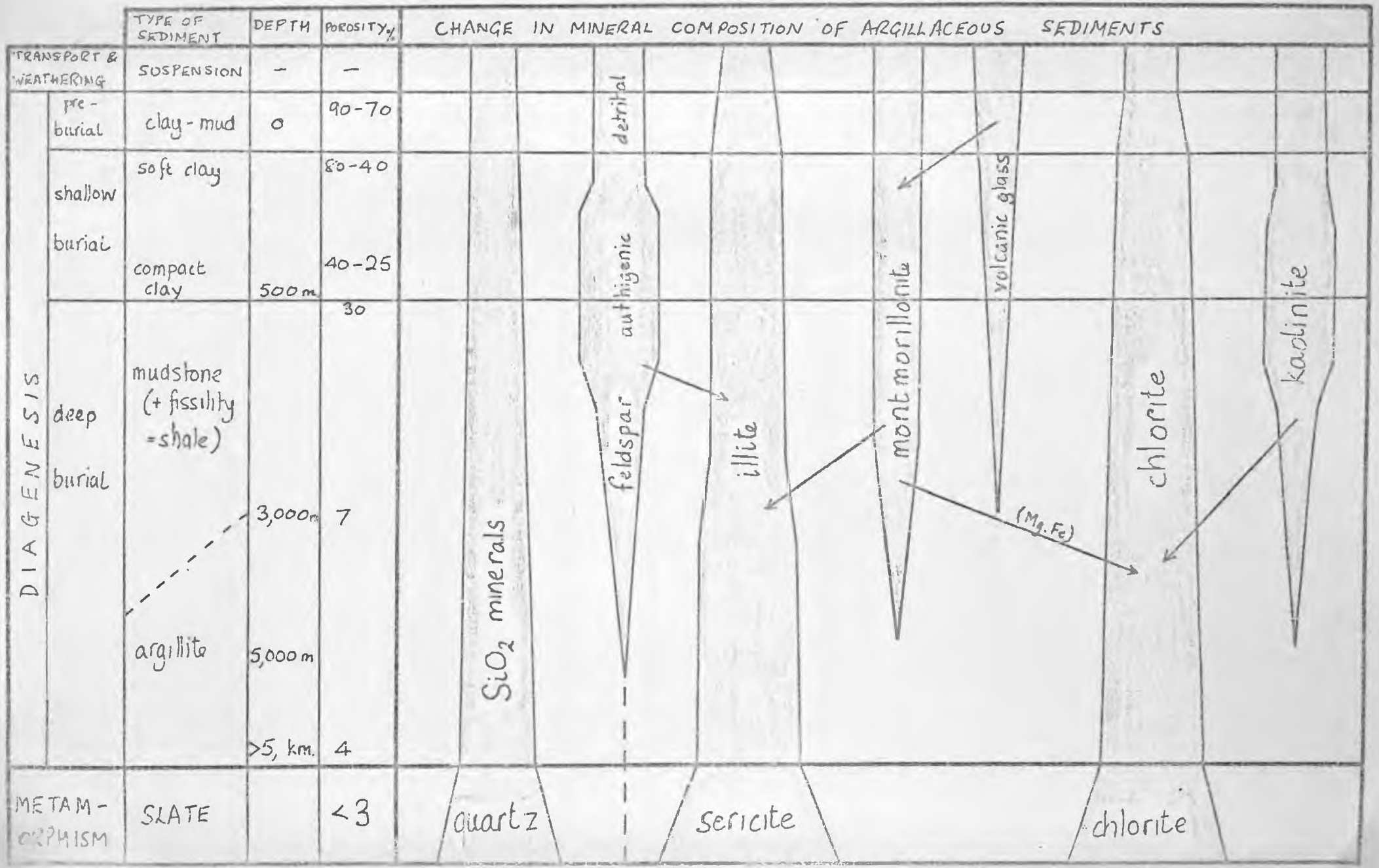

TYPE OF SEDIMENT
DEPTH
POROSITY %
CHANGE IN MINERAL COMPOSITION OF ARGILLACEOUS SEDIMENTS
TRANSPORT & WEATHERING
SUSPENSION
—
—
DIAGENESIS
pre-burial
clay-mud
0
90-70
shallow burial
soft clay
80-40
compact clay
500m
40-25
deep burial
30
mudstone (+ fissility = shale)
3,000m
7
argillite
5,000m
>5, km.
4
METAMORPHISM
SLATE
<3
detrital
authigenic
feldspar
SiO_2 minerals
illite
montmorillonite
volcanic glass
(Mg. Fe)
chlorite
kaolinite
quartz
sericite
chlorite

FIG 4 1 5

components, but that their proportion is small when the other detrital minerals - montmorillonite, kaolinite, feldspar and silica - are considered. With progressively deeper burial Muller describes the proportions of illite, chlorite and silica increasing at the expense of the remaining constituents. Illite forms at the expense of montmorillonite (via intermediate mixed-layer stages) due to simple fixation of K^+ from interstitial water, or possibly as a result of the release of K from the breakdown of detrital feldspar. The alkali earths (Mg, Fe) from the montmorillonite are liberated into the interstitial water and their presence there promotes the kaolinite–chlorite transformation, by simple hydrolysis. These illite - and chlorite-forming reactions liberate large quantities of silica, hence the increase in SiO_2 minerals with depth. At deep burial levels, authigenic and detrital feldspar start to alter illite, though in the Skiddaw Slates this reaction is probably not completed at this stage because the presence of K-feldspar is postulated as necessary to the biotite-forming reactions to be discussed in the following section.

The critical transition temperature between diagenesis and metamorphism is 300°C, and this is almost independent of pressure (Winkler, 1967); above this point the P-T conditions of the lower greenschist facies are realized, with the transformation of illite to sericite and the formation of the characteristic quartz–chlorite–sericite–(albite)* assemblages.

The base of the Skiddaw Slates would have been no deeper than 3,000 m. at the time of the early, F1, deformation and hence, from Table 4:1:1, the whole of the Skiddaw Slate succession would still have been

* the absence of albite from these rocks is due to their very low Na_2O content.

within the zone of diagenetic transformation. Since the F1 deformation produced no recognizable fabric element within these rocks (see Chapter 5), there would have been no disruption to the development of the diagenetic fabric. This diagenetic recrystallization was wholly mimetic, i.e. the bedding fabric simply became coarser in grain. In this manner the weak, phyllitic texture of these rocks was attained, and thus the major part of the present fabric of the Skiddaw Slates would have been formed by diagenetic processes prior to the advent of the regional metamorphism.

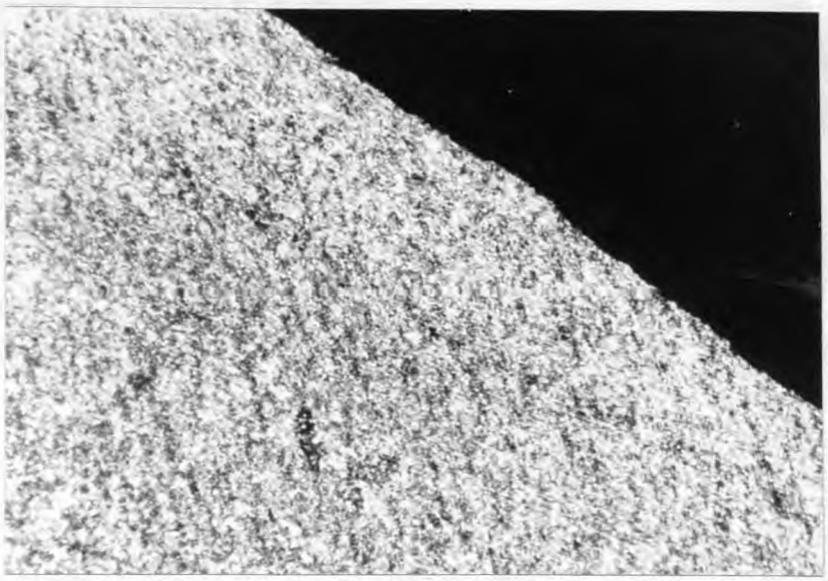
Regional Metamorphism: The only fabric elements within these rocks that are due to the regional metamorphic climax are the chlorite porphyroblasts and the segregation veins and irregular lenticles, also of chlorite. Both these features pre-date the main end-Silurian deformation, F2, because they both are cut by the S2 fracture cleavage (Figs. 4:1:3; 4:1:4).

F2: Further segregation of chlorite occurred during F2 as reflected by the diffuse zones of very fine crystals found parallel to the S2 direction in some of the finer pelites, (Fig. 4:1:6). There was very little recrystallization of sericite at this stage, only occasional grains being found parallel to the axial plane fractures of the S2 crenulation cleavage; further limited recrystallization of sericite also occurred along S3 fracture cleavage planes (Fig. 4:1:8).

Subsequent mobilization of chlorite and quartz is evidenced by the quartz-chlorite veins (Fig. 2:2:4); these veins post-date the contact metamorphism (a vein cutting the spots is shown in Fig. 4:1:9), but evidence necessary to define their age more closely is missing. Helm (1969) records similar veins from the rocks of Black Combe, but these are much earlier, being contemporaneous with the D1 event of pre-Borrowdale age.

Fig. 4:1:6. Thin, diffuse, zones of chlorite recrystallized parallel to S_2 (x 25; x Nicols).

Fig. 4:1:7. S_2 fracture cleavage, with no recrystallization of sericite or chlorite along the cleavage planes.



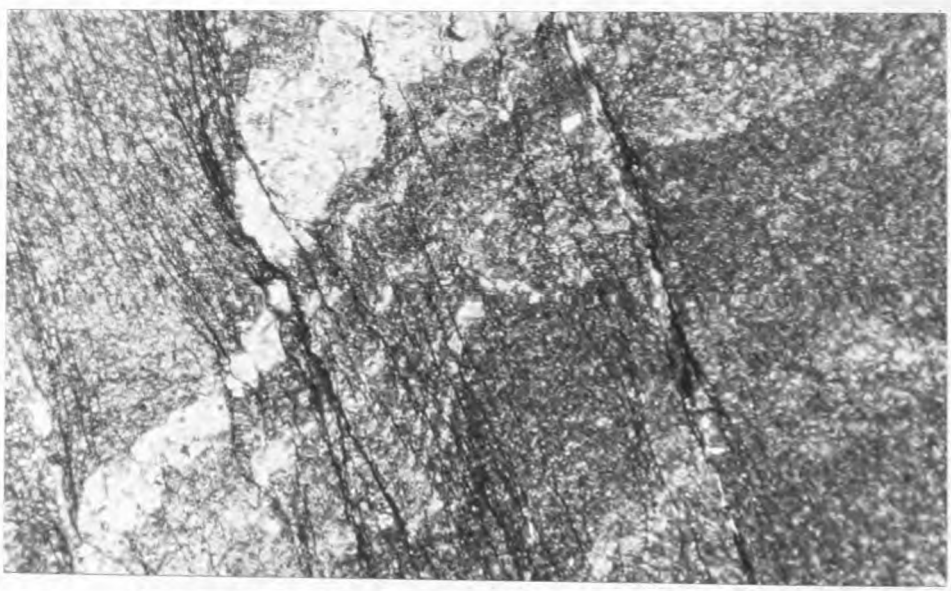
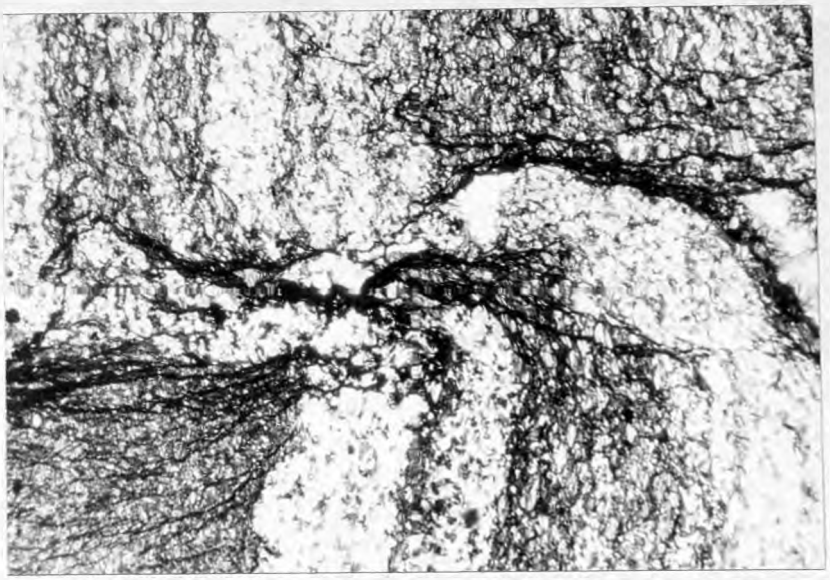
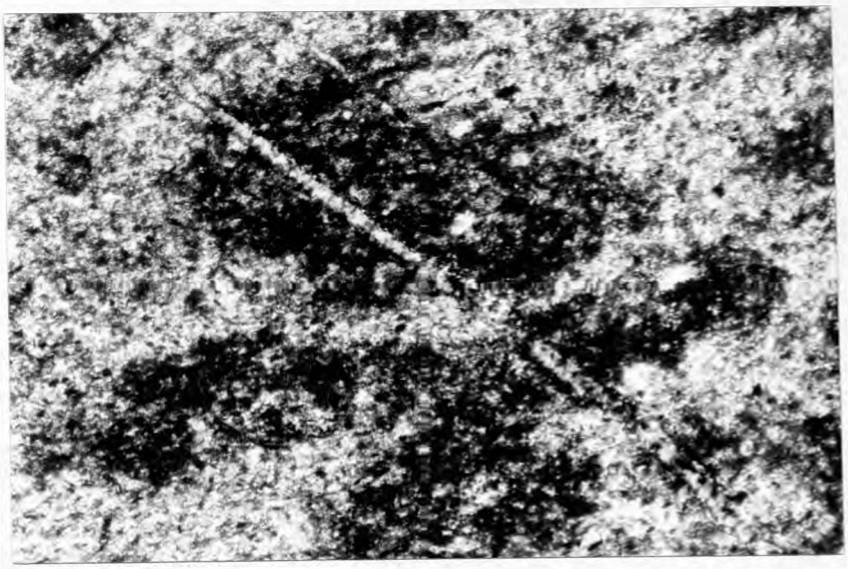


Fig. 4:1:8. S_3 fracture cleavage, with no associated sericite recrystallization. (x 25. PPL)

Fig. 4:1:9. Quartz-chlorite vein cutting anhedral spots. (x 100. x Nicols)





The basic igneous rocks have also suffered from the regional metamorphism, though the effects are difficult to see due to the extensive calcitization. What are thought to have been originally dolerites and quartz-dolerites, comprising plagioclase, hornblende, ?pyroxene, quartz* and K-feldspar*, have been altered, largely by a saussuritization type process into quartz-albite-epidote-chlorite/antigorite-K feldspar-calcite-sericite assemblages. These are the characteristic minerals of basic igneous rocks of the lower greenschist facies.

Section 4:2 Contact Metamorphism.

The evidence suggestive of contact metamorphism was discussed in Sections 2:2 and 3:2, namely the presence of spots and tourmaline veins within the flags and mudstones of the central area. The spots are wholly restricted to the pelitic lithologies (Fig. 2:2:5 and Fig. 4:2:1), average 500u in length, and are always oriented with their long axes parallel to the bedding. The majority are of anhedral, elliptical shape, and appear almost irresolvable under low powers of magnification, due to the presence of myriad tiny inclusions and much chlorite (Fig. 4:2:2). At higher powers, some degree of crystallinity is apparent, sufficient to occasionally indicate parallel extinction, but the spots are generally so highly charged with ore and graphitic material that any mineral identification is almost impossible.

However, a minority of the spots do attain a euhedral, rhombic form (Fig. 4:2:3), and even more rarely, some have grown sufficiently large (700u) to clear themselves of the disseminated inclusions. When in this state, their identification as andalusite is confident. From

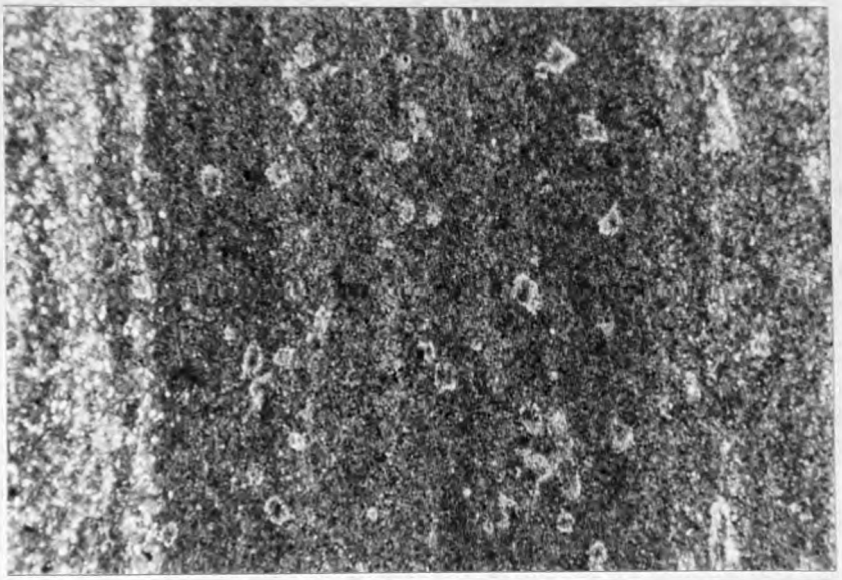
* though these two may possibly be of secondary origin.

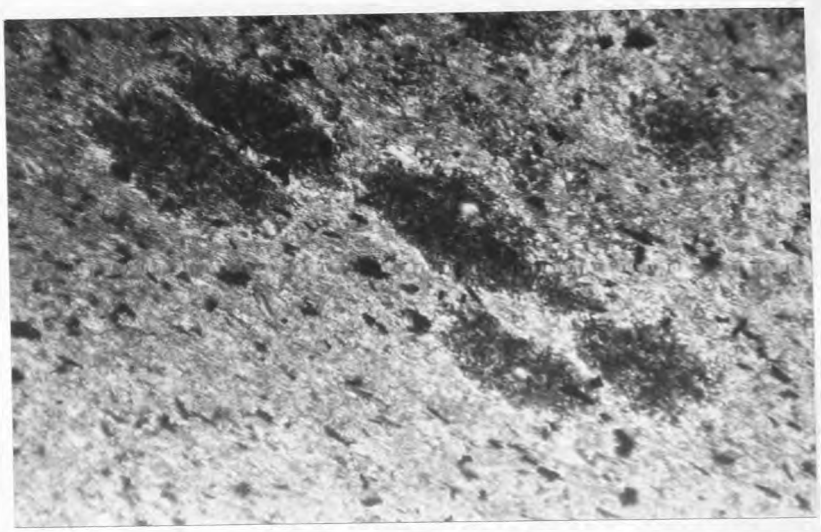
Fig. 4:2:1. Restriction of the spotting to the pelitic horizons (x 25. PPL).
(see also Fig. 2:2:5).

Fig. 4:2:2. Anhedral spots.

A. (x 40. x Nicols)

B. (x 10. x Nicols)





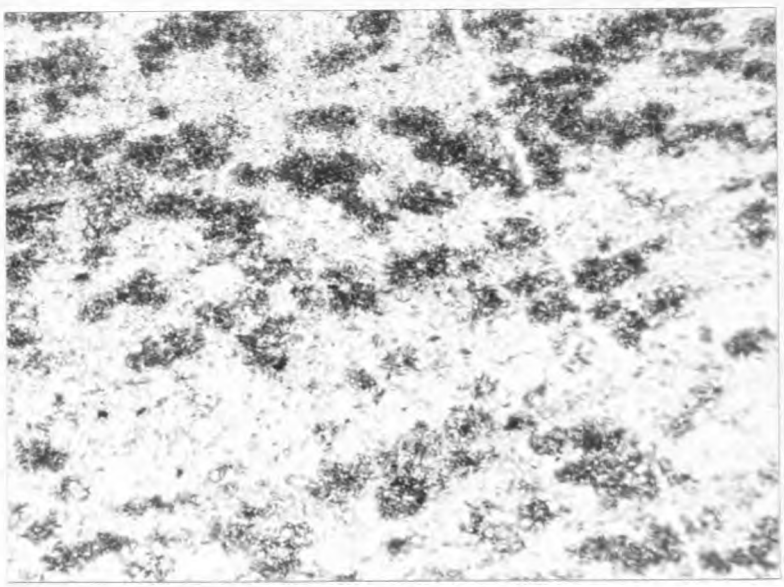
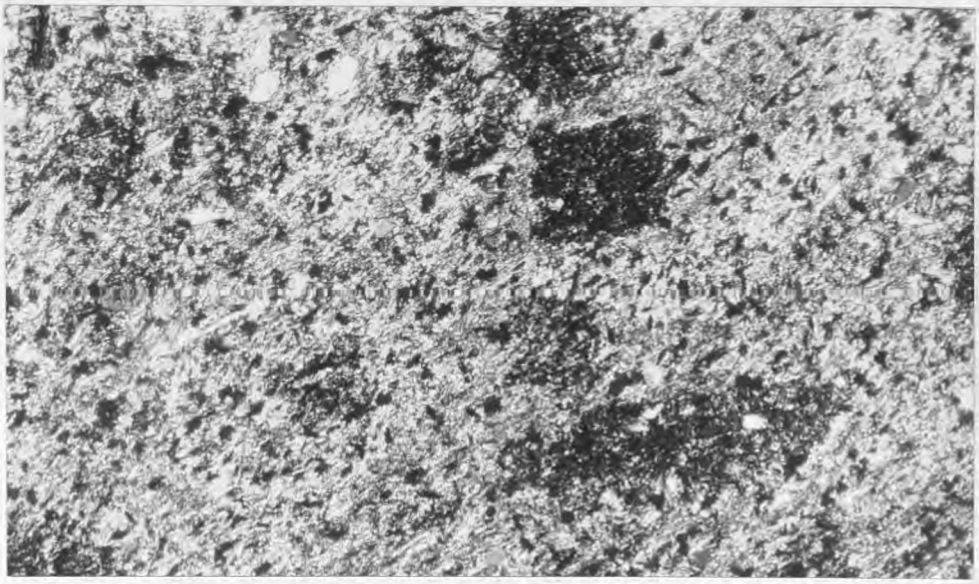


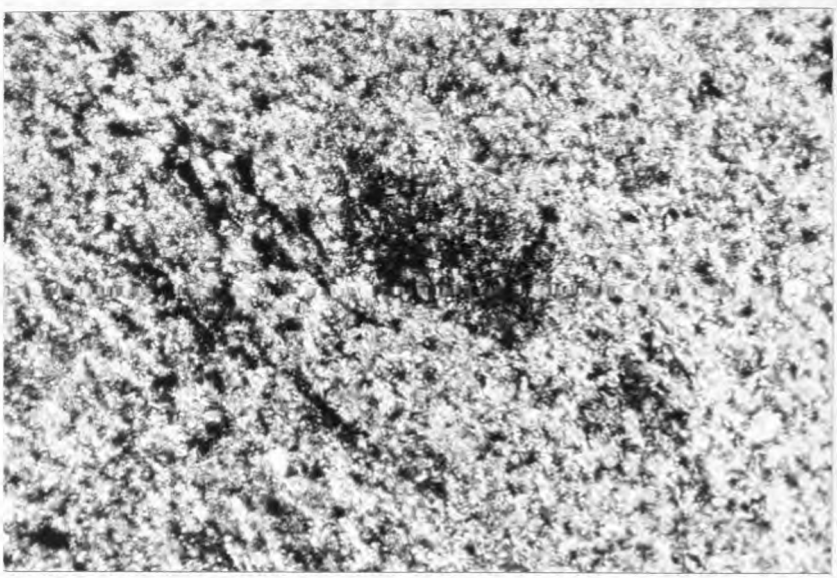
Fig. 4:2:3. Euhedral spots : definate andalusite.

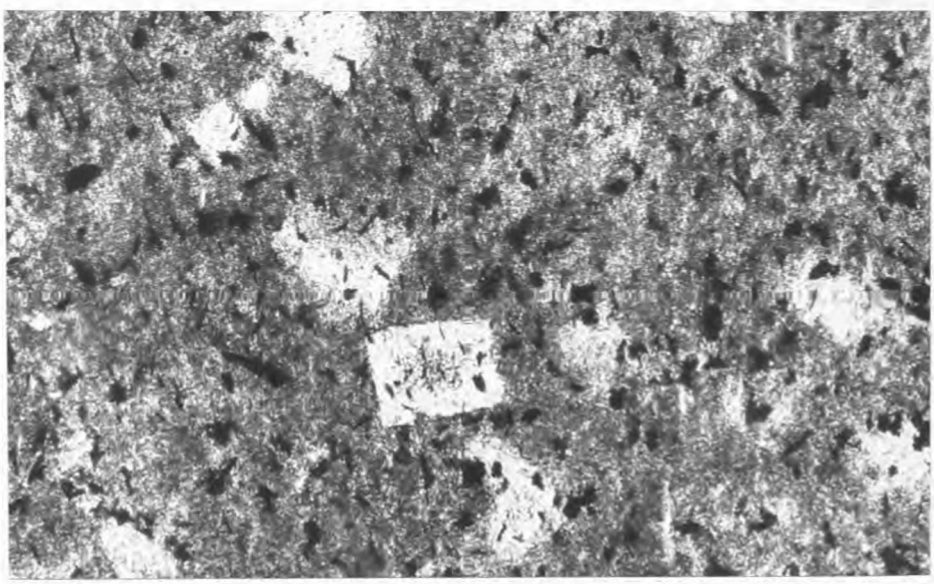
A. x 100 x Nicols

B. x 100 PPL

C. x 100 x Nicols.







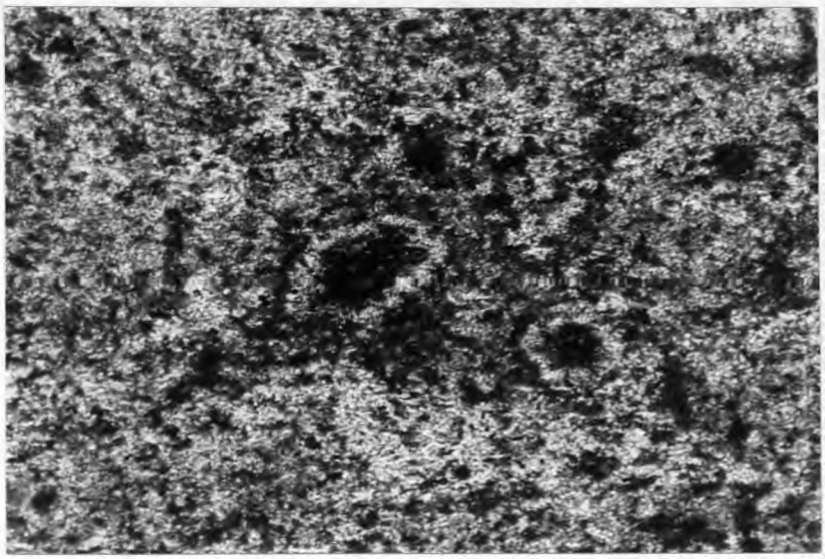
this it would appear possible that the andalusite forms incipiently as anhedral, highly charged spots within the pelitic bands; with continued growth a euhedral, rhombic outline is eventually attained, though still cloudy with inclusions at this stage, and only then does the crystal start to clear itself of inclusions. It is the outer part of the crystal that first becomes clear, and this gives the two-zoned appearance occasionally observed in the larger spots (Fig. 4:2:4).

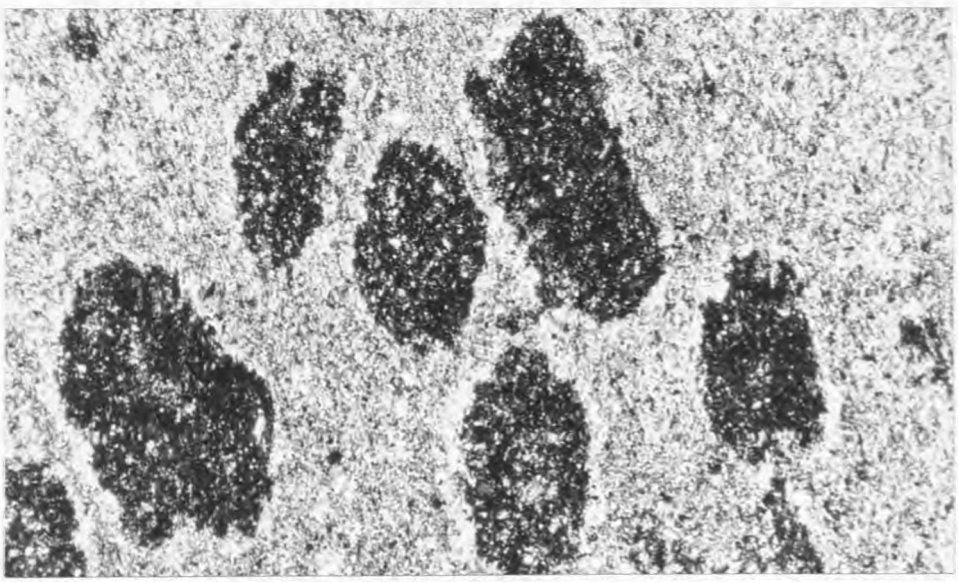
The other possibility is the anhedral, chloritic spots represent pseudomorphs after initial cordierite, and it is interesting to note in this connection the similarity of the chloritic spots from the present area to the chloritized cordierites from the Tan-y-Grisiau aureole (Fig. 4:2:5). If the anhedral spots in these Skiddaw Slates do represent cordierite, then one would expect to find them occurring independently of the euhedral andalusite crystals, a situation described by Rastall (1910) from the Skiddaw Aureole. This, however, is not the case, for in the Grasmoor rocks the chloritic spots and the euhedral andalusites appear to be mutually exclusive, hence suggesting that the initial proposal of sequential andalusite development is substantially correct.

In addition to the andalusite spots, the contact metamorphism was also marked by the development of biotite (Fig. 4:2:6). These small crystals show the characteristic yellow-brown pleochroism and straight extinction of biotite, and are completely different in origin and appearance from the oxydized chlorite-vermiculites mentioned in the previous section. The biotite has grown as tiny (20 - 40u) irregular grains within the chloritic matrix of both pelitic and flaggy horizons, though there is a tendency for preferential development of biotite in the pelites; the grains show a random orientation, with neither bedding nor S2 directions being especially favoured. When the distribution of

Fig. 4:2:4. Zoned anhedral spots. (x 100. x Nicols)

Fig. 4:2:5. Anhedral spots from the Tan-y-Grisiau auerole, N. Wales. Composed of chlorite pseudomorphing after cordierite. (x 25. x Nicols).
Note similarity to anhedral spots from Grasmoor aureole.





those rocks containing euhedral spots, anhedral spots, and those with biotite but no spots, are plotted (Fig. 4:2:7), it can be seen that there is a significant occurrence of biotite outside the limits of the metamorphic aureole as defined by the spotting. This suggests that the biotite was formed before the nucleation of the spots occurred, and the discovery of spots containing tiny biotite grains, provides good corroborating evidence. Deer, et al. (1965) record the same conclusion: .. .. "biotite usually pre-dates the first appearance of andalusite and cordierite" in thermal metamorphic aureoles. The petrogenesis of this biotite is impossible to define: it forms in the matrix, but no evidence of replacement of sericite or chlorite is ever seen. Mather (1970) in discussion of the biotite isograd in the Scottish Dalradians quoted the equation:-

microcline + chlorite + phengitic muscovite — biotite + less phengitic muscovite + quartz + H_2O

and although initially designed to define a regional metamorphic reaction, it may describe the conditions operating in the Skiddaw Slates, although the absence of any observed microcline does militate against it somewhat. The other possibility is that the biotite does form from a sericite-chlorite-iron ore-rutile assemblage, but on too fine a scale to be perceived.

The inference to be drawn from this evidence is that beneath Grasmoor there is to be found a plutonic mass of granitic composition, and bearing tourmaline, which has given rise to the low-grade contact metamorphism of the sediments above. The presence of tourmaline suggests that this granite has a close genetic connection with the granites of Skiddaw and Eskdale, rather than with the Ennerdale Granophyre, which is devoid of tourmaline. The fact that the outcrop of the thermally-altered, Blake Fell Mudstone, facies can be traced from my

Fig. 4:2:6. Microscopic growth of biotite due to the thermal metamorphism.

A. x 25. PPL

B. x 40. PPL



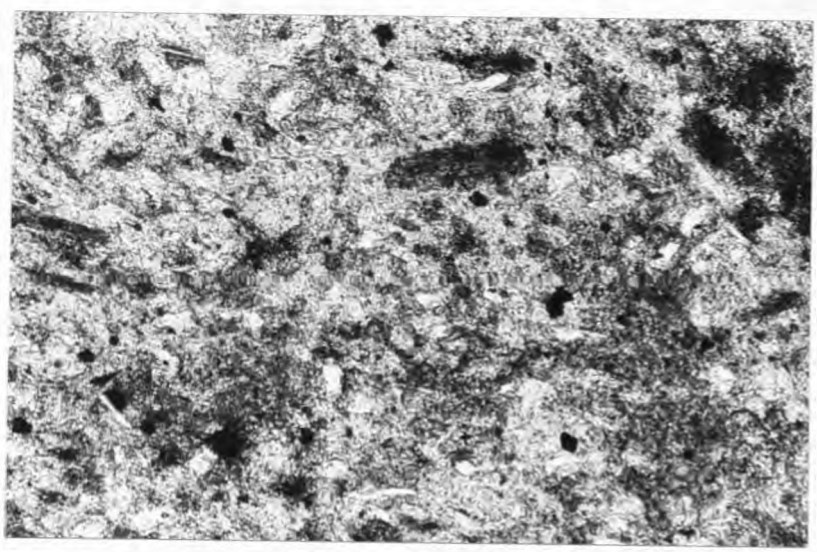


Fig. 4:2:7. Map shewing the distribution of the spots, and the limits of the metamorphic aureole which they define.

Key.

1 = anhedral, cloudy-opaque spots

2 = euhedral, clear, spots - andalusite

3 = biotite crystals found growing in specimens from outside the metamorphic aureole

4 = limits of metamorphic aureole

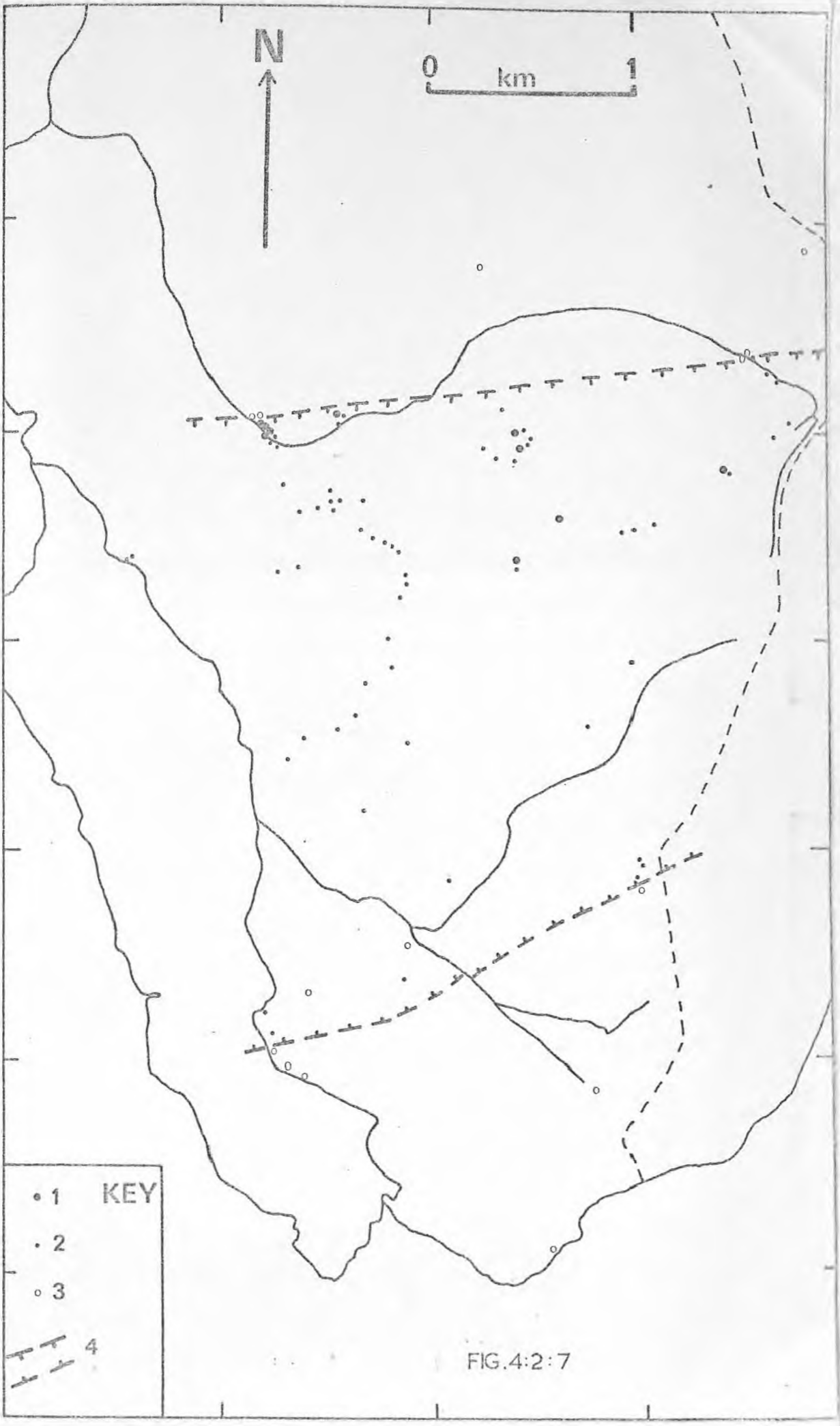


FIG.4:2:7

area across to the western shore of Bassenthwaite (Rose, personal communication) suggests that the Grasmoor body and the Skiddaw Granite may be physically continuous.

The Skiddaw and Eskdale Granites have been dated at 395 $\pm$ 14 m.y. (Miller, 1962); chiastolite crystals in the inner hornfels zone of the Skiddaw Granite are deformed by S2 (Simpson S1); and an aplitic dyke from the Eskdale Granite also suffered the S2 cleavage (Ward, 1876). Hence, both Eskdale and Skiddaw Granites were emplaced prior to the advent of the main end-Silurian deformation, F2, and this would suggest a pre-F2 age for the Grasmoor body as well. The evidence that would confirm this extrapolation directly, namely S2 cutting the metamorphic spots, was not found because the metamorphic aureole was not highly deformed during the F2 deformation. However, this very lack of F2 deformation throughout the aureole is, in itself, suggestive of the prior presence of the granite, acting as a "buffer" for the rocks during F2.

The remarkable feature of this aureole is that there is no discernible difference in hand specimen or thin section between rocks from the top of Grasmoor, and those on the lake-shore, a vertical separation of over 2,500'. This implies that the temperature gradient outwards from the granite was a very shallow one; the conditions necessary for the growth of andalusite and biotite, namely a temperature of 400°C at 2 - 3kb load pressure, were sustained throughout a vertical height of at least 2,500' and probably much further. Copious outward flow of magmatic fluids may well have contributed to the extent of the aureole, and their activity is indicated by the numerous quartz-tourmaline veins.

The presence of the Grasmoor Granite is therefore substantiated by several independent factors, and it is to be hoped that a gravity profile will be run across the area, as soon as the regional background becomes available, in an attempt to deduce the depth and underground

configuration of the pluton.

The pelitic rocks around the Dodd Dolerite also show signs of the development of contact metamorphic spots. These spots, of the anhedral, chloritic type, are highly deformed by S2 (Fig. 4:2:8). Geochemical evidence (Section 3:1) indicates the close genetic connection between the Borrowdale Volcanic Group and the Dodd Dolerite, which suggests a Borrowdale Volcanic age for its intrusion and aureole development, thus providing yet another indication of the end-Silurian age of F2.

Fig. 4:2:8. Chloritized anhedral spots from the aureole of the Dodd dolerite, deformed by S_2. (x 25. PPL)

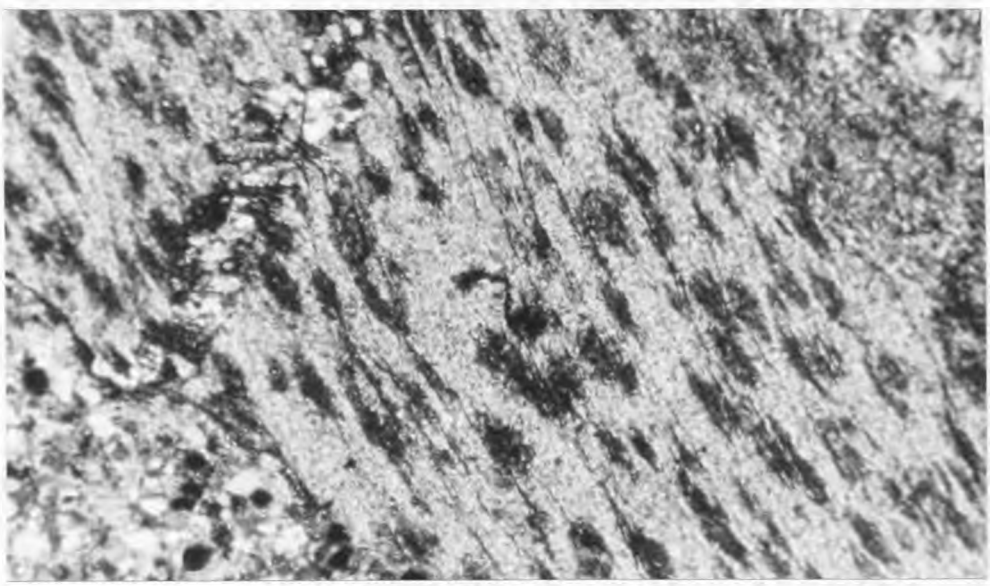


TABLE 5:1

This account	Simpson equivalent	Style and orientation of axes	Facing direction	Cleavage attitude	Age
F4	F3	Variable; generally NW - SE. Steep plunge	-	NW - SE Steep	
F3	F2	Horizontal, recumbent; generally ENE - WSW	-	Horizontal	end-Silurian
F2	F1	Upright, open-tight ENE - WSW	both N and S	ENE - WSW Variable	
F1	-	Overturned N - S	generally west	-	pre-Borrowdale Volcanic
-	-	Generally E - W, recumbent	generally north	-	syn-depositional

CHAPTER FIVE

INTRODUCTION AND PRE-BORROWDALE VOLCANIC FOLDING.

Section 5:1 Introduction.

A sequence of four phases of deformation has been worked out within the present area and, in addition, evidence is also found of a phase of unconsolidated, syn-depositional slumping.

The earliest tectonic event, F1, is thought to be of pre-Borrowdale age, and is held to be responsible for the widely reported unconformity between the Skiddaw Slates and the Borrowdale Volcanics (see Section 1:2, 1:3); the remaining fold phases, F2 - F4, are equivalent to Simpson's (1967) F1 - F3 in style and orientation, but are of end-Silurian age. These details are tabulated below (Table 5:1).

It is obvious from this discussion that the Skiddaw Slates have been subjected to complex polyphase deformation and metamorphism during the Caledonian orogeny. The sequence of these deformations has been worked out from the evidence afforded by the relationships of the minor structures due to each episode, including minor folds, cleavages, and lineations - combined with fabric studies in thin section. Two major criteria were used in the assessment of the chronological sequence of the minor structures:-

i) Structures showing a similarity of style (allowing for lithological variation) and with similar patterns of preferred orientation were grouped together as members of a single generation.

ii) Structures of one generation were considered to be later than those of another generation when they consistently overprint or deform the structures of the earlier generation.

In addition, the relative intensity of the structures developed by successive episodes of folding usually decreases with time; i.e. F2

is a more intense deformation than F3, and so on. The exception to this is the F1 structures which are less intense and less widely developed than those of the succeeding F2 deformation.

Section 5:2 Syn-depositional Slump Folding.

Folds due to unconsolidated slumping could be described as sedimentary structures, but they first strike the eye as tectonic structures and it is with these that they may be confused, so it is thought better to discuss them here, in order that their relationships to the tectonic structures may be more easily appreciated.

The problems of the origin of these structures have attracted an extensive literature; the folds seem to be restricted to areas of rapid sedimentary accumulation such as the continental slope (Moore, 1961), and to be triggered off either by earthquake shocks (Helwig, 1970); by the frictional drag of turbidites passing overhead (Gregory, 1969); or by the shear stress induced by the weight of accumulated sediment, (Terzaghi, 1956). The sands, silts, and muds would still be water-saturated at this stage and when the stress, however developed, was applied, the sediments would have responded in a saturated, hydroplastic manner. Helwig (1970) believes that the slumping mass had a free upper surface, and a bedding-plane decollement as the lower surface. In contrast Gregory (1969) attributes slump folds to sub-surficial movement, the slumped mass being bounded above and below by slip planes. In either case, however, these syn-depositional folds may be recognised by their lack of vertical persistence, being bounded above and below by plane beds. Further characteristics, as discussed by D. Roberts (1972) and Helwig (1970), are their curved axial surfaces; variable wavelength and amplitude; and very consistent axial trend. A mean axial trend direction, however, does tend to lie normal to the sedimentary transport

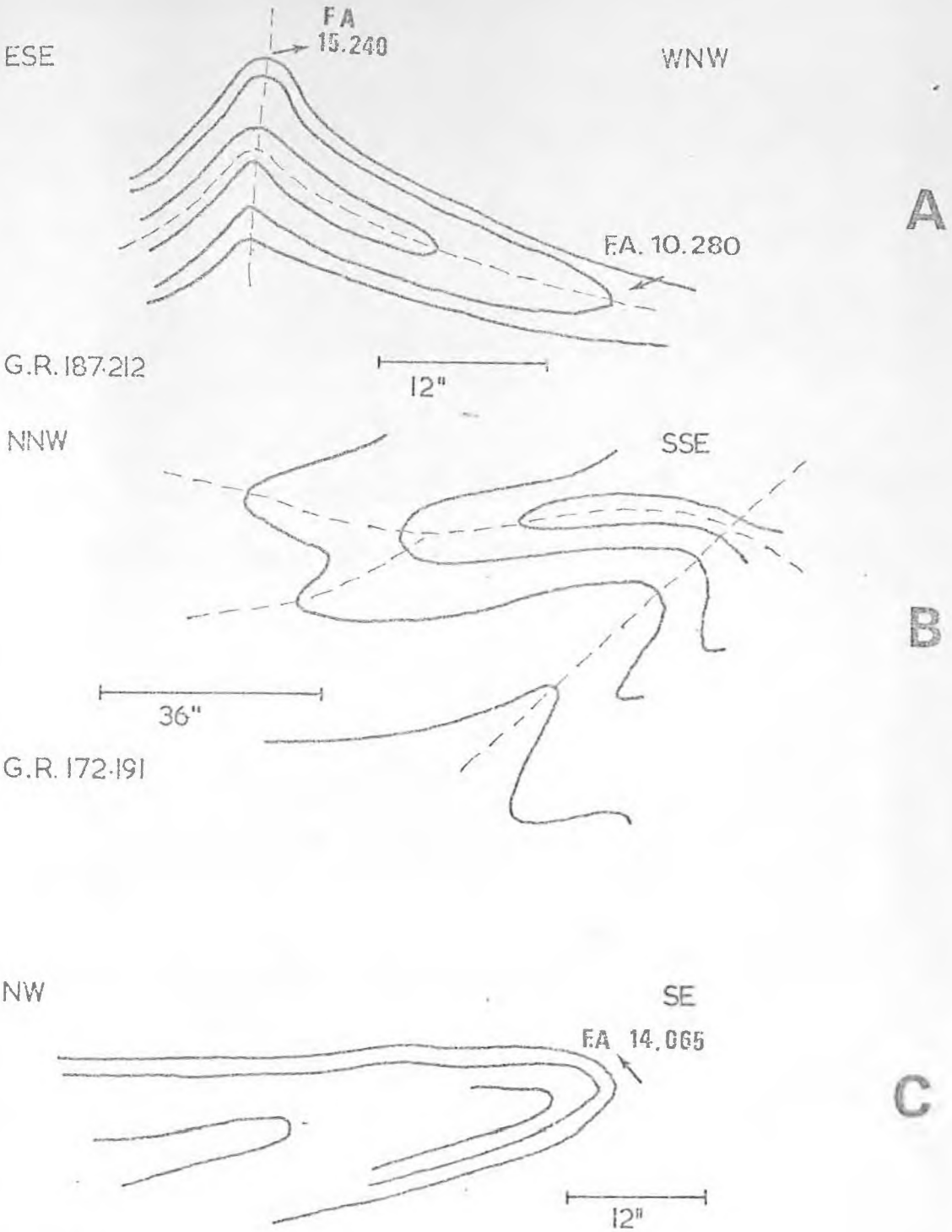
direction as deduced from flute casts, (Helwig, op. cit.).

Many examples of these folds have been found, mostly within the flaggy lithology, especially around Grasmoor and Lad Hows. This is a reflection of the hornfelsing these rocks have suffered, often rendering clearly visible bedding details that are difficult to see in more highly deformed rocks. The possible variation in size and shape of these folds is large; Figs. 2:3:3 and 5:2:1 show several examples that illustrate the wide range of amplitude, attitude, inter-limb angle, and wavelength. Notable also is that two of the latter examples have subsequently been refolded by F2.

In addition to slump folding there are found instances of slump faulting, (Fig. 5:2:2); these two structures (folding and faulting) can be linked as the compressional and extensional forms, respectively, of the process described by Dott (1963) as "sub-aqueous slumping and plastic mass-flow". In 5:2:2, two examples of these structures are compared with Dott's diagram; diagram 1 (drawn from G.R. 182215) shows the simple arcuate extensional faulting characteristic of the up-slope source point of the slump; but diagram 2 (from G.R. 166207) shows similar faults cutting an earlier slump fold. This is taken to indicate that the folds, which once formed the compressive "toe" of an early slump, became the source of a later movement and hence show the extensional characteristics of this part of the slump.

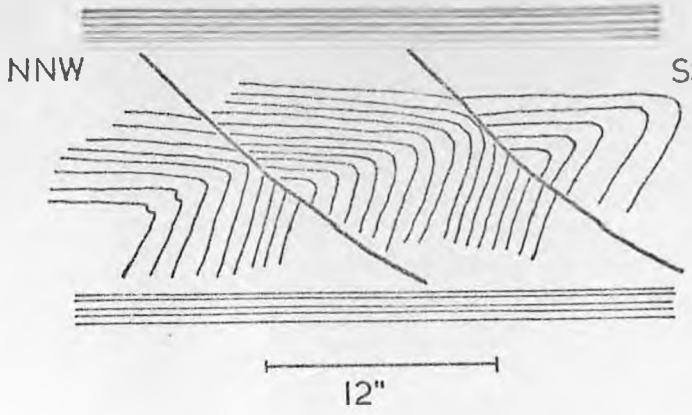
When the orientation of the axes of the slump folds is plotted on a stereographic net (Fig. 5:2:2) it is interesting to note that the fields they occupy are very similar to those within which the majority of the F2 folds are seen to fall, i.e. they have a dominant strike to the E.N.E. quadrant, this in contrast to the dominant northerly strike of the tectonic F1 folds with which they are most likely to be confused.

This E.N.E. orientation of the slump folds suggests that if one



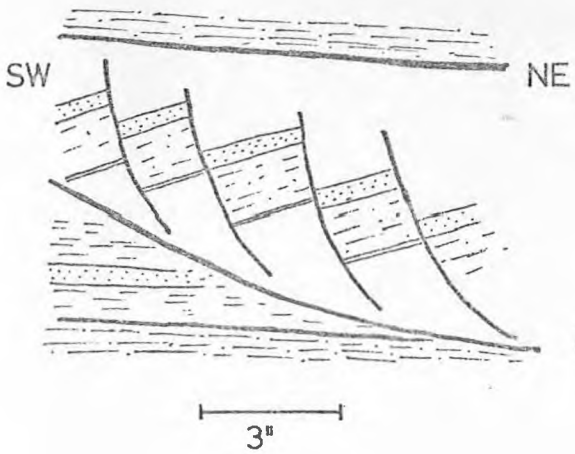
SLUMP FOLDS ; A & B REFOLDED BY F2

FIG 5:2:1



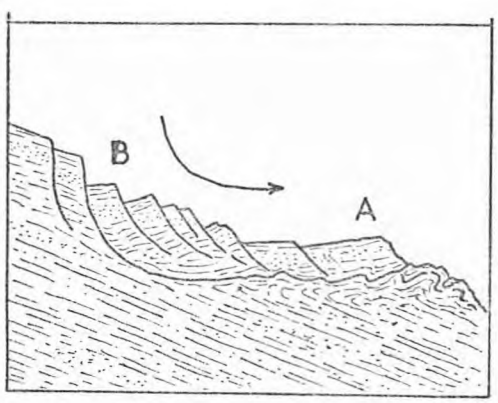
G.R. 166·207

1. Slump folds (A) followed by slump faults (B)

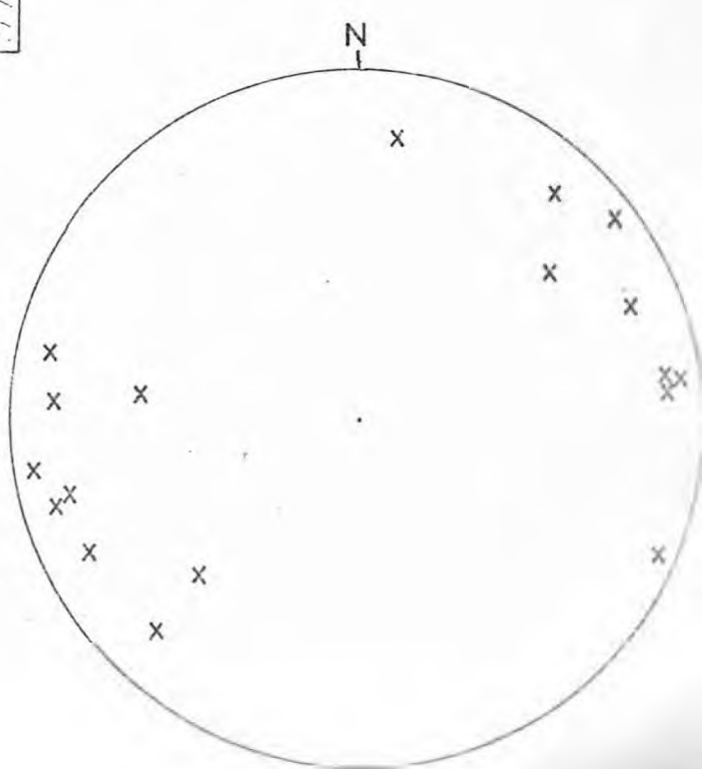


G.R. 183·214

2. Slump faults



3. Features associated with subaqueous slumping, (after Dott, 1963)



Plunges of Slump fold axes

FIG. 5:2:2

assumes the dominant F2 trend to be parallel to the axis of the depositional basin in which the sediments accumulated, then the majority of the slump folds indicate a transport direction or "overthrusting" down the slope of the basin, at right angles to its axis. A similar pattern was described by Norman (1965) for part of the Ludlovian succession of the S. Lake District.

Section 5:3 The First Deformation, F1.

(i) Style of F1 folds.

F1 minor folds are characterized by:

a) N-S striking fold axes and axial planes; fold axes usually plunge steeply down the bedding dip;

b) an overturned-recumbent style,

c) their deformation by succeeding F2 folds and its associated cleavage, S2.

F1 minor folds show a marked change in style and tightness from north to south across the area; in the north and centre they are quite open, usually overturned, though some approach a monoclinal form (Figs. 5:3:1; 5:3:2); in the south of the area they are much tighter and wholly recumbent, often approaching an isoclinal condition (Fig. 5:3:3). This difference in style is due to two inter-related factors:

a) The predominance of massive psammitic units in the north and pelitic units in the south, and

b) because of this lithological variation, the flattening of the F1 folds by the succeeding F2 deformation was much greater in the pelites of the south than in the northern psammites.

(ii) Other F1 Characteristics.

A further feature of the F1 deformation is the lack of any congruent S-surface or b-lineation. No S-surface was formed, penetrative

Fig. 5:3:1. Photos shewing the more open style of F_1 folding in the psammitic lithologies.







FIG. 5:3:2

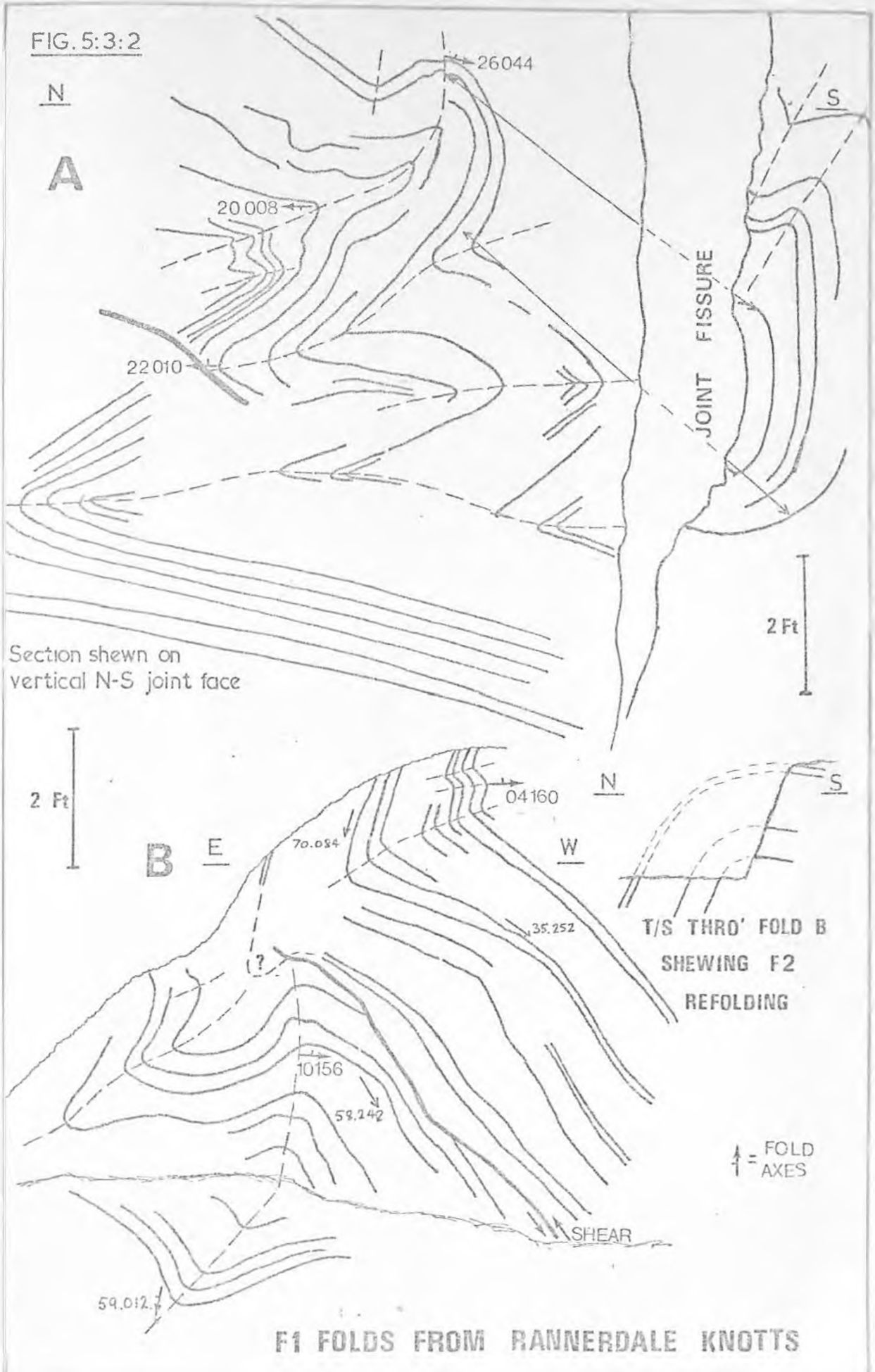

N
A
26044
20 008
22 010
S
JOINT FISSURE
2 Ft
Section shewn on
vertical N-S joint face
2 Ft
B
E
W
04160
70.084
35.252
10156
58.242
SHEAR
59.012
N
S
T/S THRO' FOLD B
SHEWING F2
REFOLDING
= FOLD
AXES
F1 FOLDS FROM RANNERDALE KNOTTS

Fig. 5:3:3. Photos shewing the tighter style of F_1 folds from the pelitic rocks of the south of the area.

A. Shews refolding by F_4.

B. Shews characteristic Z shape of many of the Buttermere folds.

C. A particularly tight example.







or otherwise, and any recrystallization that did occur was mimetic after bedding. Even in the hinges of F1 folds, no disruption of the bedding fabric was observed.

The only other F1 minor structure found was the occasional boudinage of this psammitic bands in the flaggy units. The boudins were of elongate, "sausage" shape, with their long axes striking N - S; their rounded form indicates that the ductility contrast between interbedded psammitic and pelitic bands, during F1, was quite low.

(iii) Present Distribution of the F1 Folds.

Most of the F1 folds were found in the area around Buttermere. This is a result of:

a) The pelitic nature of the lithologies, their higher ductility, compared with the more massive rocks to the north, permitting more intense folding, and

b) the shape of the outcrops themselves, which are glacially-scoured hummocks of sub-horizontal or whaleback form.

Thus the majority of the exposed rock surfaces are flat and more closely approximate to profile sections of the steep-plunging F1 folds, than do the steeper crag exposures to the north.

In the south of the area, including Buttermere, Low Bank, and Whiteless Breast, the F1 folds are nearly all of Z shape and plunge steeply to the SSE down the northern limb of the Buttermere Syncline (Figs. 5:3:4; 5:3:5; Plates 1 and 4). Breaking this regularity of the regional ENE strike, however, there are linear, N - S, belts of steep plunging S shaped F1 folds. This Z to S transition marks the location of the axes of larger F1 folds which from the minor fold style must be asymmetric and west-facing.

Over the remainder of the area, to the north of Whiteless Pike

Fig. 5:3:4. Style of F_1 and F_2 minor folds in the Buttermere area, shewing the way in which changing fold symmetry (S —— Z) demarcates the position of F_1 fold axes.

FIG. 5 3 4

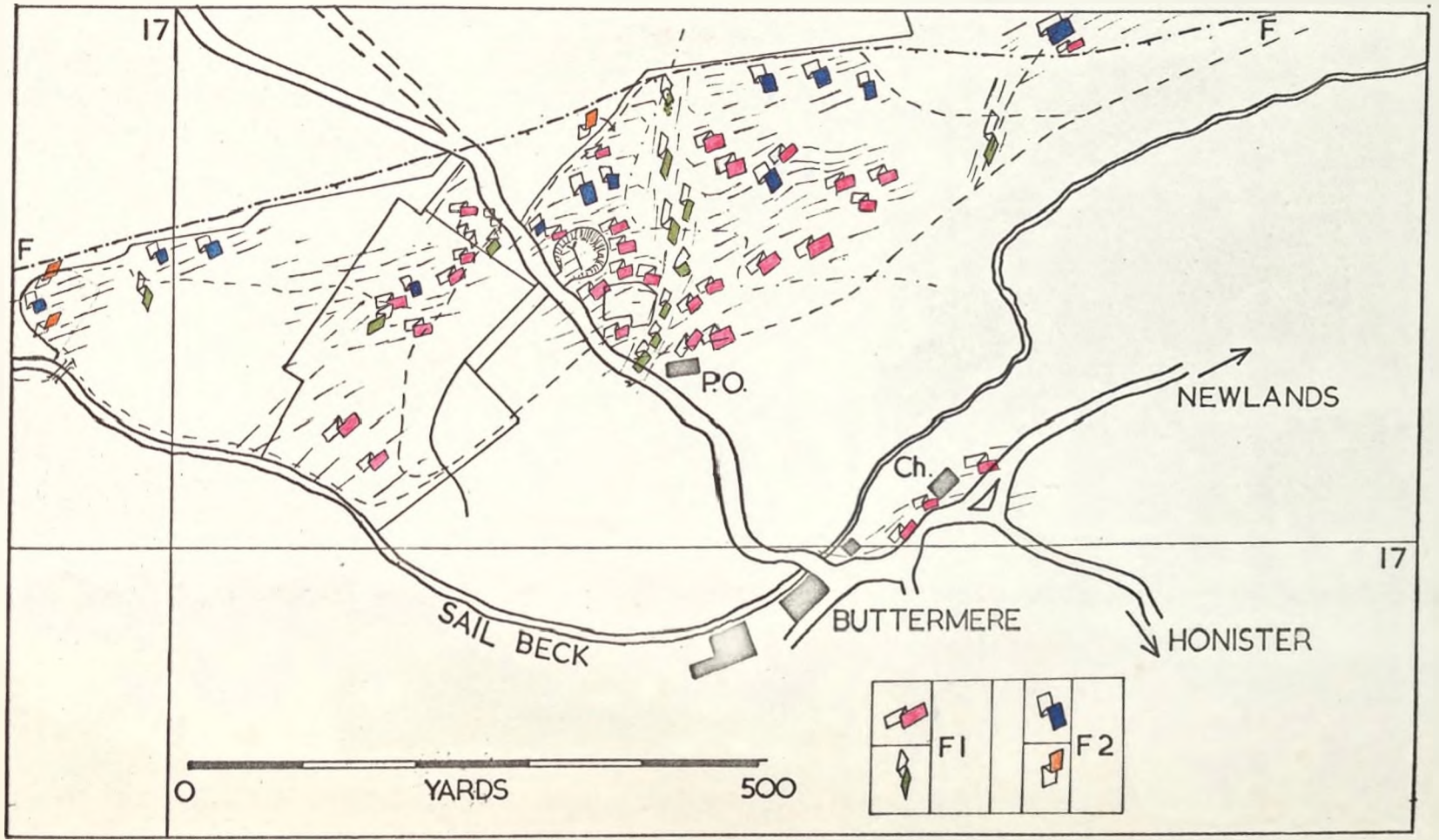
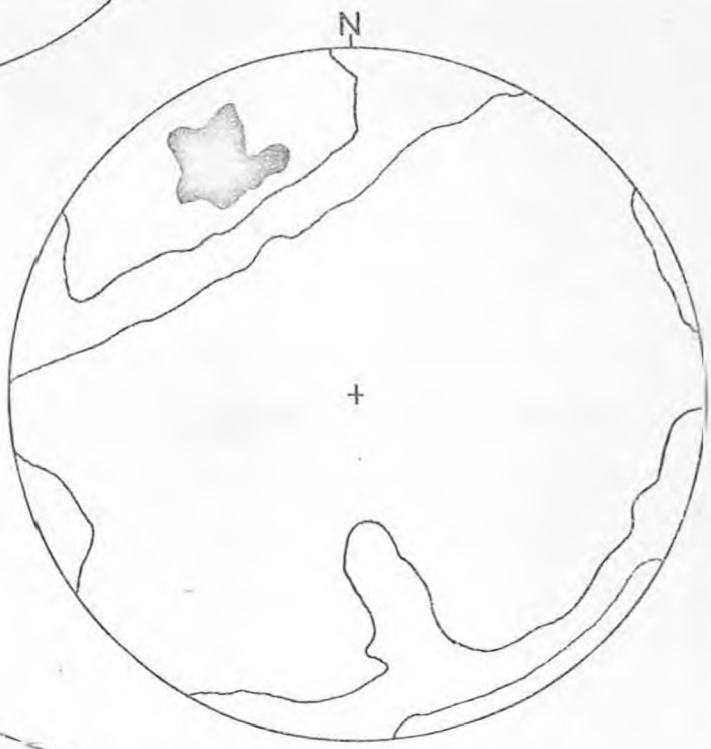

17
F
F
P.O.
NEWLANDS
Ch.
17
SAIL BECK
BUTTERMERE
HONISTER
F1
F2
0
YARDS
500

Fig. 5:3:5. Stereograms to shew the distribution of F_1 fold axes in the Buttermere area (A); and the contrast of the mean ENE - WSW bedding strike (B) with the bedding directions in the steep N - S F_1 zones (C).



A

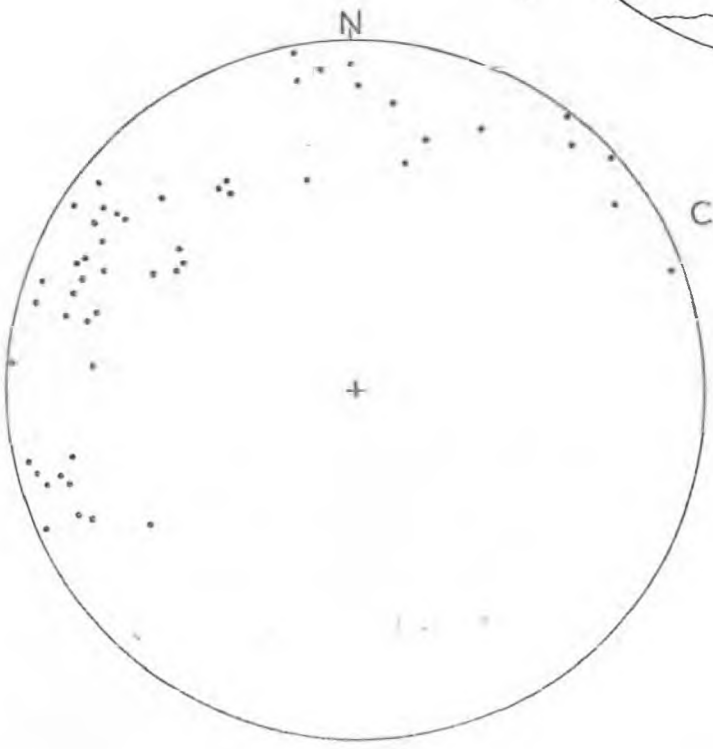
F1 FOLD AXES –
BUTTERMERE (49 points)



B

POLES TO BEDDING -
BUTTERMERE SUB AREA
(see Plate 7)

440 pts. contoured at 0.9,
2.7, 5.5 % per 1% area



C

54 POLES TO BEDDING FROM
N–S STEEP ZONES OF
BUTTERMERE (see Fig. 5:3:4)

note marked contrast between
this and the strong SSE maxima
of the diagram above

FIG. 5:3:5

and Low Bank, the numbers of F1 folds recorded are smaller and their distribution is less clear. Z shaped, south plunging F1 folds occur in the inlier of Loweswater Flags near the head of Rannerdale Beck, and in the broken crags below Lad Hows and Cinderdale Beck. These Z folds are analogous to those of Buttermere (e.g. examples at G.R. 176.170, and 173.172), having their long limbs parallel to the dominant ENE bedding strike. Zones of steep, N - S, striking bedding are also found in these areas, and although the corresponding S folds, which would have completed the Buttermere analogy, were not found, the overall picture of west-facing F1 over-folds appears to be the same.

Just to the east of the summit of Whiteless Pike (G.R. 180.190), there is a small east-facing, S shaped F1 fold which plunges to the south and has its long limbs parallel to the NE bedding strike. This is a reversal of the Buttermere situation, where bedding-parallel, south-plunging F1 folds were of Z shape, facing west, and this suggests the presence of an F1 closure in the vicinity of Whiteless Pike.

In Gasgale Gill and on Whiteside Crags, the F1 minor folds are much less common, but all those found were east-facing and much more open in style, a reversal of the situation in the south of the area. Furthermore, there were no linear N - S zones of steeply inclined bedding, with the exception of the steep middle limb of the large F1 fold which was described from the top of Gasgale Gill (186213) in Section 2:4.

West-facing structures were only recorded at two localities in this part of the area:

a) at 163208, on the south bank of the mouth of Gasgale Gill, and

b) at 162214, in a small gully at the foot of Whiteside End.

As discussed in Section 2:4, these two outcrops may represent part of the same structure, but offset along a WNW - ESE fault line. In the Whiteside gully exposure, the plunges of the west-facing F1 minor folds

diverge about an F2 anticlinal axis, which itself plunges to the east, down the dip of the gentle F1 long limb (Fig. 5:3:6). In the Gasgale exposure, the westerly dip of a series of silty flags steepens from 30° - 70° and this may represent, if movement along the aforementioned fault is reversed, the steep westerly limb of a large F1 fold, with the Whiteside Gully exposures occupying the gentler eastern limb.

Further north than this, only a single F1 fold was found - a small S shaped, west-facing fold plunging gently to the north, which outcrops on the western slopes of Dodd (G.R. 167.234), with the intervening crags of Penn and Whiteside End being apparently devoid of any F1 structures.

(iv) The Form of the Initial F1 Surface and the Effects of F2 Refolding.

The initial form of the F1 surface, prior to the F2 deformation, can be deduced from three lines of evidence. The first, discussed above, is the use of changes in the F1 minor fold geometry to locate the position of the major F1 hinges (Fig. 5:3:4). The second method (Fig. 5:3:7) equates the plunge of F2 minor folds with the initial dip of the F1 surface upon which they were imposed; this does, however, assume that no variation in compressive strain has occurred along the length of the F2 axial surfaces.

Whether this is a valid assumption may be dubious, because when stereographic plots of the F1 and F2 fold plunges are compared (Figs. 5:3:8; 5:3:9), the perfect correlation which the above theory demands, between the pole to the great circle containing the F1 plunges and the maxima of F2 fold axes does not occur. Instead, the F1 fold plunges lie on one (or possibly two) great circle(s) which strike N.10.E., whereas the F2 fold axis maxima lie at either end of a diametric great circle striking N.80.E. - S.80.W. Similarly, the pole to the diametric

Fig. 5:3:6. Structure of Whiteside Gully, shewing re-folding of F_1 minor folds by an open F_2 anticline.

STRUCTURE OF WHITESIDE GULLY

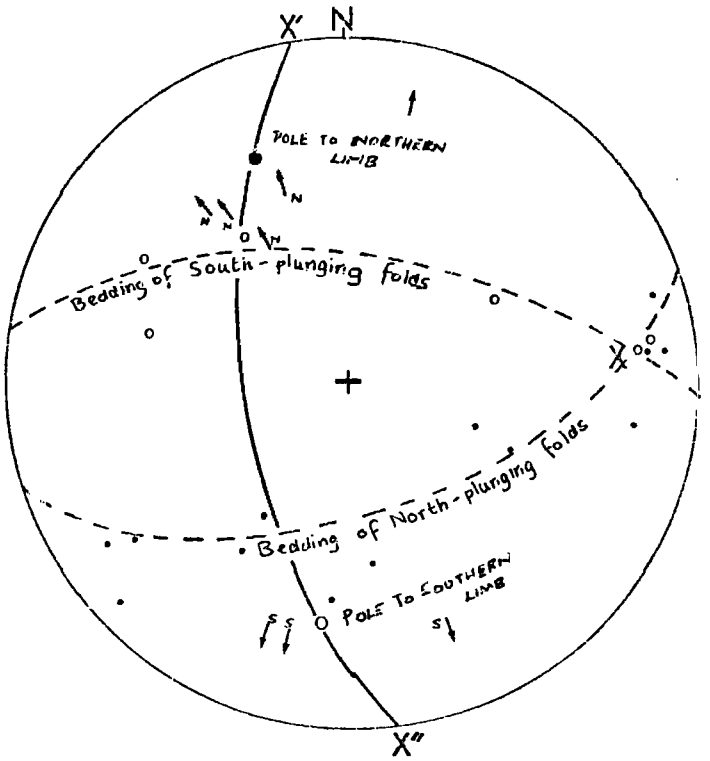
Poles to bedding and fold axes of North-and South-dipping limbs of F 2 anticline

° bedding

\ F 1 fold axes (both N. & S. limbs)

X'-X" F 2 great circle

X Pole to above, representing the F 2 fold plunge



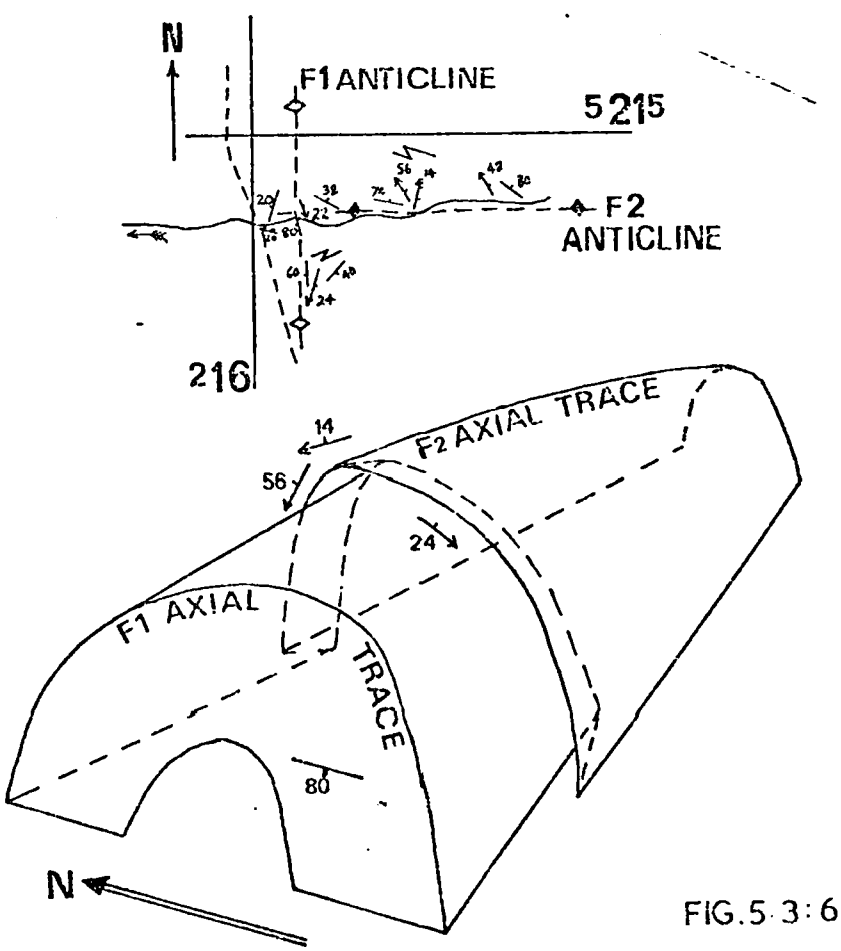
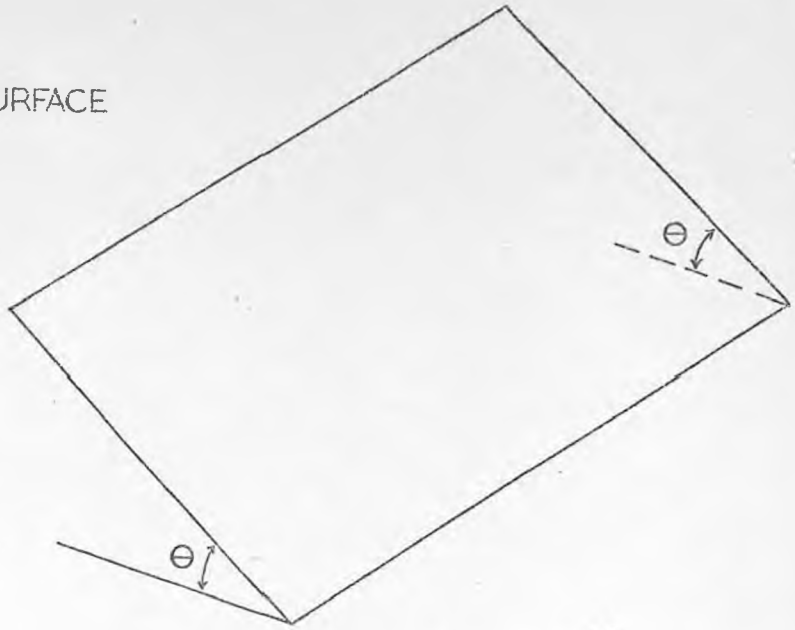
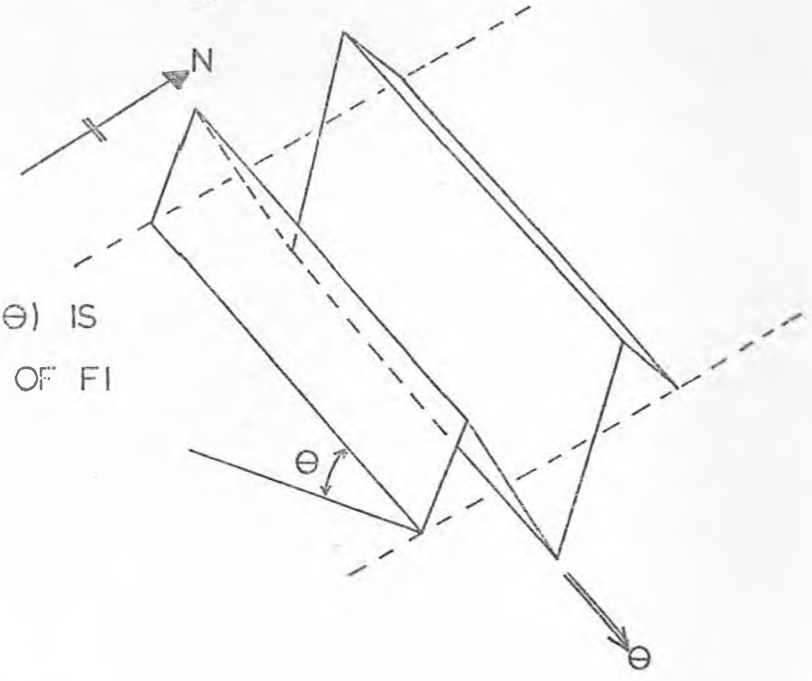


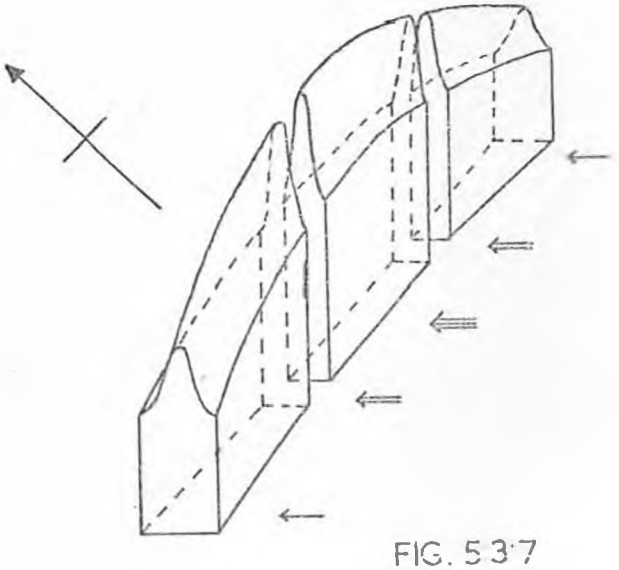
FIG. 5·3:6





F2 FOLD PLUNGE (Θ) IS EQUIVALENT TO DIP OF F1 SURFACE

ASSUMING THERE IS NO



VARIATION IN COMPRESSIVE STRAIN ALONG THE FOLD AXIAL SURFACE (after RAMSAY, 1967)

FIG. 5.3.7

EQUAL AREA PLOT OF
F1 FOLD AXES
(165 points; contours at
1.2, 2.4, 3.6, & 6.0 %
per 1% area)



FIG. 5 3.8

EQUAL AREA PLOT OF
F2 FOLD AXES
(312 points; contours at
0 65, 1.3, 2.6, & 4.0 %
per 1% area)



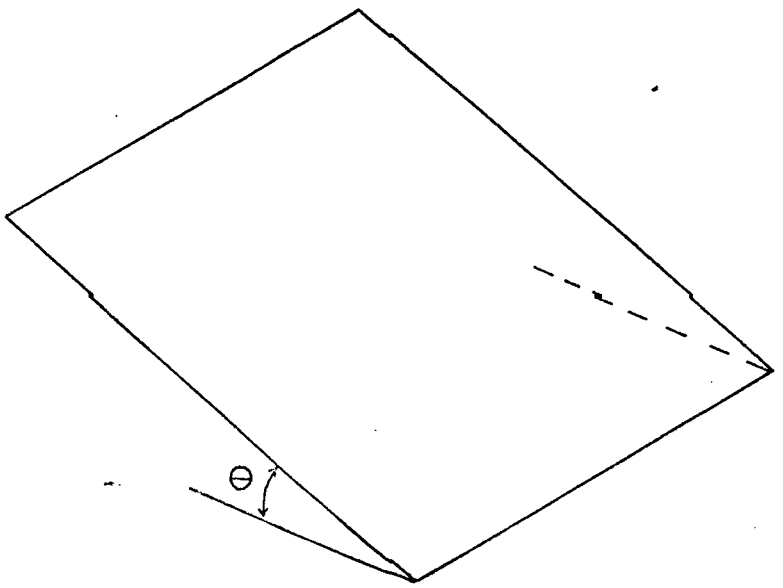
FIG. 5 3 9

great circle containing the F2 fold plunges should correspond to the regional F1 fold axis, but again there is this 20° arc of discrepancy, which again, may be the result of a variation in the F2 compressive strain.

Thirdly, by considering the intersection lineation between the bedding and S2, a further estimation of the dip of the F1 surface may be made, (Fig. 5:3:10). This only applies in the pelites of the south of the area, because only there does the dominant folding mechanism approximate to simple shear, i.e. the angular relationships of the shear planes and the bedding remain largely unchanged during folding (the "pack of cards" analogy). If, therefore, we plot these indices of the F1 surface onto a map of the area, we find that the F1, S to Z transitions mark the major F1 axes, and the F2 minor fold plunges and lineations provide a measure of the limb dips (Fig. 5:3:11), thus giving the shape of the F1 surface, (Fig. 5:3:12).

There is only one major drawback, however, to such a reconstruction, which may be allowed for, but it can not be entirely circumvented. This is the effect of the flattening component in the F2 deformation. If one assumes that the style of F1 folds would originally have varied in the same way as the F2 folds are seen to do, then the F1 folds in the psammites would be asymmetric and quite open, whilst in the pelites they would be much tighter, with overturned middle limbs. The effect of the F2 deformation on these varying fold styles would then depend on the dip of the middle limb of the F1 folds (Fig.5:3:13). With an asymmetric open F1 fold, F2 deformation would tend to squash the fold, making it more open and less asymmetric; with an overturned F1 fold, however, the converse would occur, the extent of the overturning and the tightness of the fold both increasing.

Thus the superimposed section across the F1 surface given in Fig.



ASSUMING A SIMPLE SHEAR MECHANISM
THE PLUNGE OF L2 (Θ) IS EQUIVALENT
TO THE INITIAL DIP OF THE F1 SURFACE

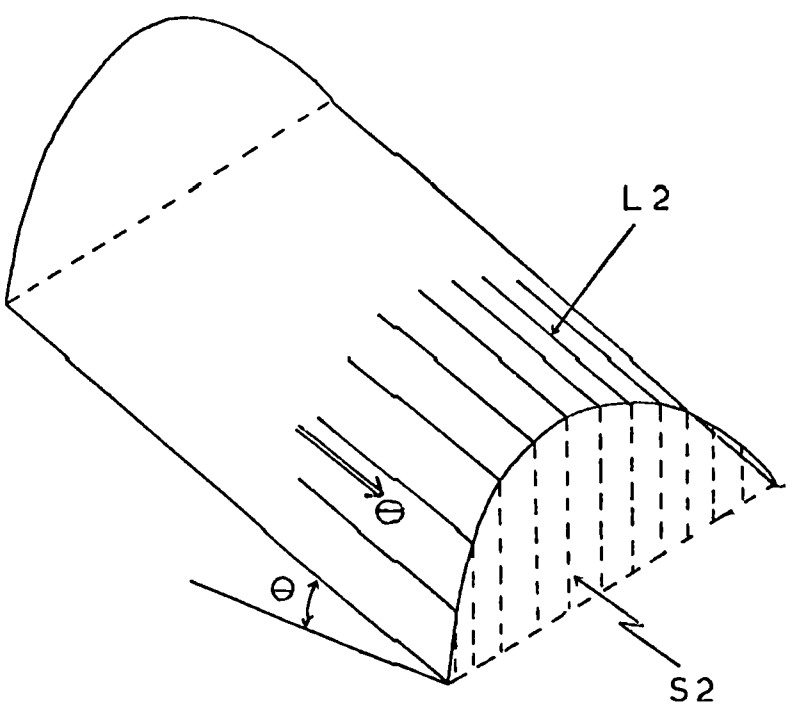


FIG. 5:3:10

Fig. 5:3:11. Variation in style facing direction of F_1 across the area which, when taken with F_2 fold plunges, gives an indication of the F_2 form of the F_1 surface.

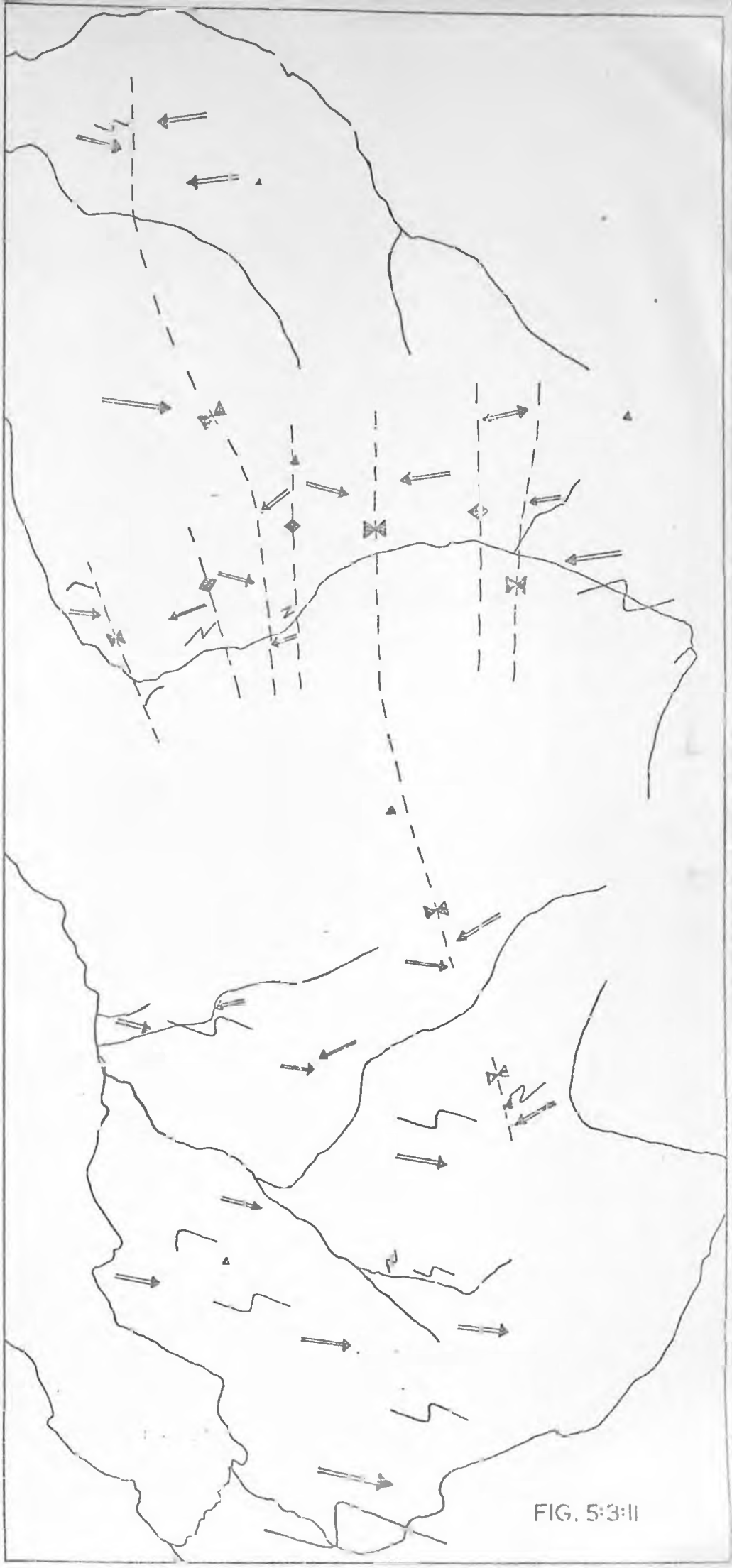


FIG. 5:3:11

Fig. 5:3:12. A superimposed section W – E across the area, shewing the reconstruction of the F_1 surface prior to F_2 deformation.

Fig. 5:3:13. Diagram shewing the way in which the attitude of the middle limb of the F_1 folds controls the effect upon them of F_2 d flattening, with some shewing increasing, and others decreasing, tightness and limb-dip.

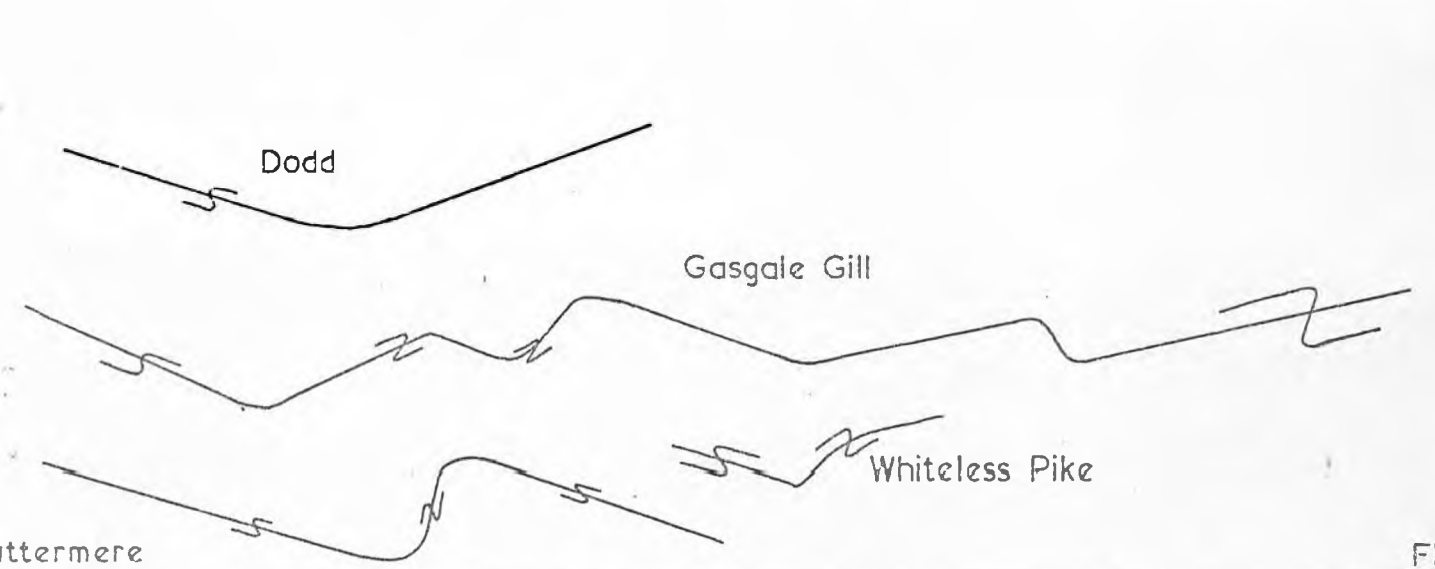


FIG. 5:3:12

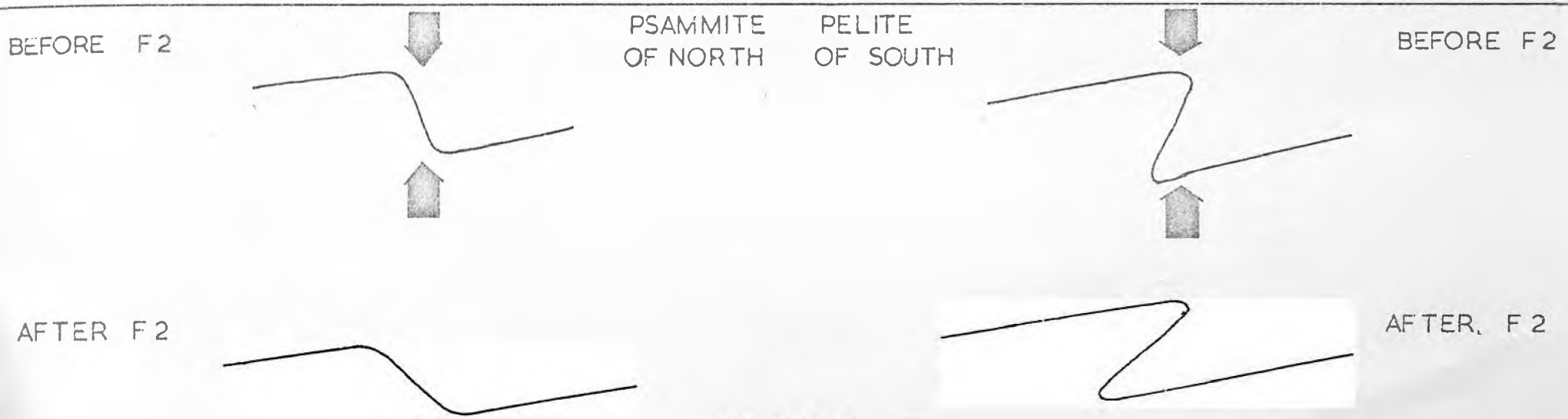


FIG. 5:3:13

5:3:12 must be interpreted in this light; namely that F1 fold tightness in the north was probably reduced, and in the south was increased, by the F2 compression.

(v) Distinction between F1 Folds and F4 Folds.

Although it is quite an easy distinction to make in the field, these two fold phases have been confused in the past, most notably by Simpson (1967), who interpreted my F1 structures either as F4 folds (his F3), or as F2 (his F1) "cross-folds". Although the fold axes plunge in similar directions (NW-SE), F1 and F4 folds may be distinguished by their relationships to S2, i.e. F1 folds are cut by S2, whereas S2 is itself deformed into open warps around the F4 axes. Once the relations to S2 were known, two further differences between F1 and F4 folds were found to be of universal application:

a) F1 folds are much tighter and overturned, whereas F4 folds are open, steep-plunging warps.

b) F4 folds have a NW - SE striking fracture cleavage and possibly a steep plunging lineation associated with them; F1 folds have none.

CHAPTER SIX

THE FIRST END-SILURIAN DEFORMATION, F2.

The end-Silurian age of this deformation has only been established outside the present area. Direct evidence comes from the overprinting of contact metamorphic minerals by F2 in the aureole of Skiddaw Granite (Soper and Roberts, 1971). More directly, an end-Silurian age is indicated by the fact that F2 structures are penetrative, on a large scale, through not only the Skiddaw Slates but also an additional 8,000m. of strata above, from the base of the Borrowdale Volcanic Group to the top of the Silurian succession.

Within the area mapped, the only indications of the age of F2 are:

i) the deformation of the spots in the contact aureole of the Dodd intrusion - of Borrowdale Volcanic affinities (see Section 4:2)

ii) the deformation of the sediments around the vent-breccia pipe on Whiteless Pike (G.R. 179187) which is again of Borrowdale Volcanic affinities

iii) the truncation by F2 overthrusting of the small doleritic stock (of Borrowdale affinities) in a tributary gully of Gasgale Gill (G.R. 182217).

This evidence only gives a maximum possible age to the F2 deformation of post-Borrowdale Volcanic; however, the constancy of style and orientation of F2 structures across the Skiddaw Slate outcrop makes extrapolation of the end-Silurian age from the Skiddaw aureole back to the Crummock area quite firmly based.

It was unfortunate that none of the spots from the Grasmoor aureole were found to shew F2 deformation, for that would have confirmed an end-Silurian age for the folding, assuming the Grasmoor pluton to be co-eval with that of Skiddaw.

Section 6:1 Major F2 Structures.

The major F2 folds of the area have been previously discussed in the Geological Description, Section 2:4 (see Plates 1 and 2). In brief, we have to the north of the Grasmoor Dome, the Dodd Monocline, the Braithwaite and Sand Hill fold-pairs, and the Whiteside Syncline, all of which face to the south. To the south of Grasmoor there is the Hause Point - Whiteless Pike fold-pair, and the Buttermere Syncline, all of which are north-facing structures. Thus the hypothesis is proposed that the granitic pluton underlying Grasmoor was emplaced (like the Skiddaw Granite) prior to the F2 deformation, and that so re-inforced, the rocks of Grasmoor were only gently buckled by F2. Conversely, the rocks to the north and south of Grasmoor were deformed against this resistant mass, and hence thrown into large-amplitude folds, and over-turned so as to face each other across the intervening undeformed area.

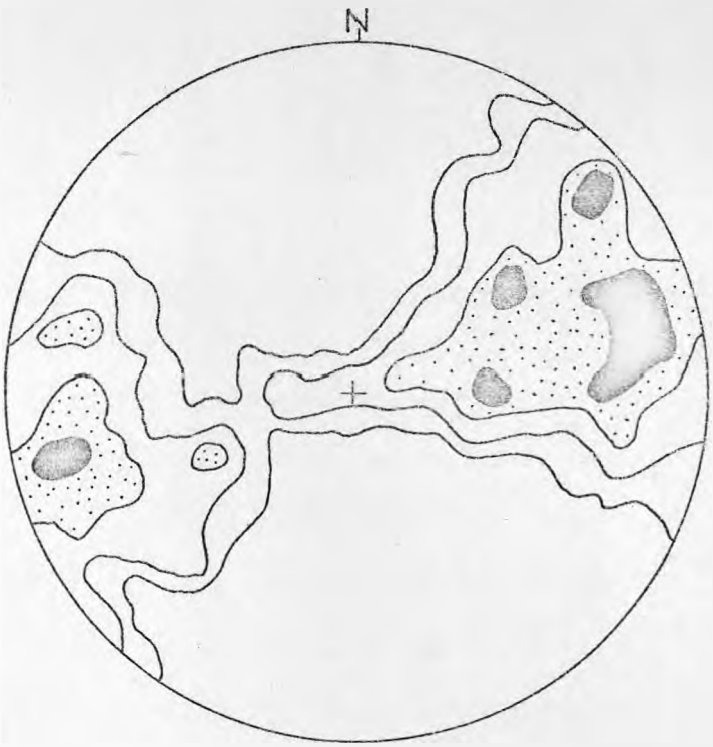
The regional strike of the F2 fold hinges and their attendant axial plane cleavage, S2, are quite constant across the entire area, any deviations from the mean ENE-WSW direction being largely the result of subsequent F4 warping.

Section 6:2 F2 Minor Folds.

The typical F2 minor fold is an upright, moderately tight fold having a variable, though usually gentle, plunge to the ENE or WSW quarters (Fig. 6:2:1). However, variation in lithology and local strain conditions have resulted in a great variety of F2 styles and attendant minor structures (cleavage, boudinage, etc.) as shewn in Figs. 6:2:2 to 6:2:5.

i) F2 structures in the psammitic rocks.

The massive psammitic beds of the middle part of the Loweswater Flag succession show minor folds of buckle type only, with little



F2 FOLD AXES : 312 pts. contoured at 0·65, 1·3, 2·6, 4·0%

FIG. 6·21

Fig. 6:2:2. Typical field appearance of F_2 minor folds in the pelitic lithologies.

A. Mosser-Kirkstile Slates of Gasgale Gill.

(G.R. 170.214).

B. Mosser-Kirkstile Slates of Low Bank. Steep plunging F_2 fold, with marked shearing and disharmony in the axial region. (part shaded by the author).

G.R. 172.1891.

C. S-shaped F_2 folds above Buttermere Village, indicating proposed style of F_2 major structures. G.R. 177.174.



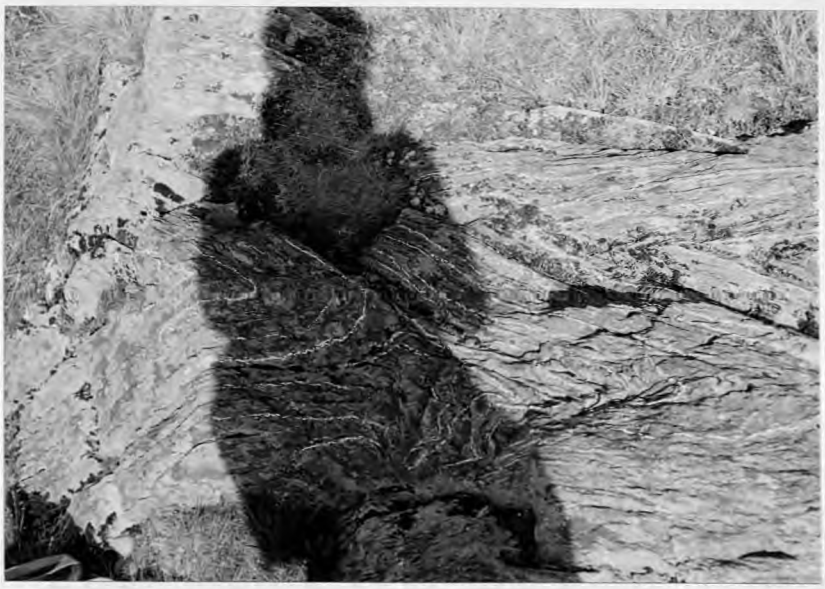




Fig. 6:2:3. F_2 fold styles in the thinly-interbedded, flaggy lithology.

Note the boudinage of the psammitic bands and the marked development of an axial planar shear cleavage in the fold hinges.







Fig. 6:2:4. F_2 fold styles in the flaggy lithology (continued).

S_2 cleavage development well shewn in A. and C.

B. shews the steep middle limb and anticlinal hinge of an S-shaped F_2 fold from Whiteless Breast.



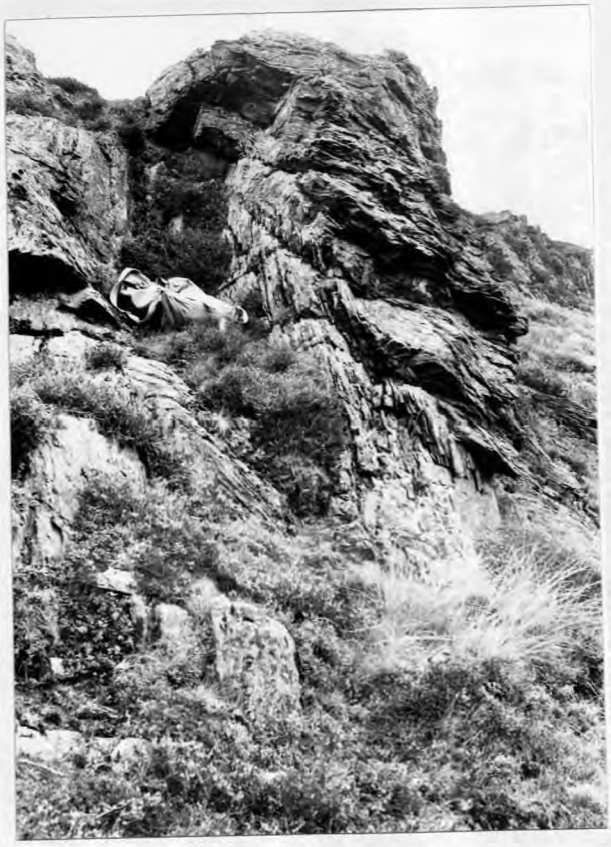




Fig. 6:2:5. Typical F_2 folds from the massive psammitic lithology.

A. from the Whiteless Pike Flags, is shewn again in Fig. 6:2:6, in connection with the T' and t' plots made from it.

B. note extreme flattening of the pelitic band in the inner arc of the fold.

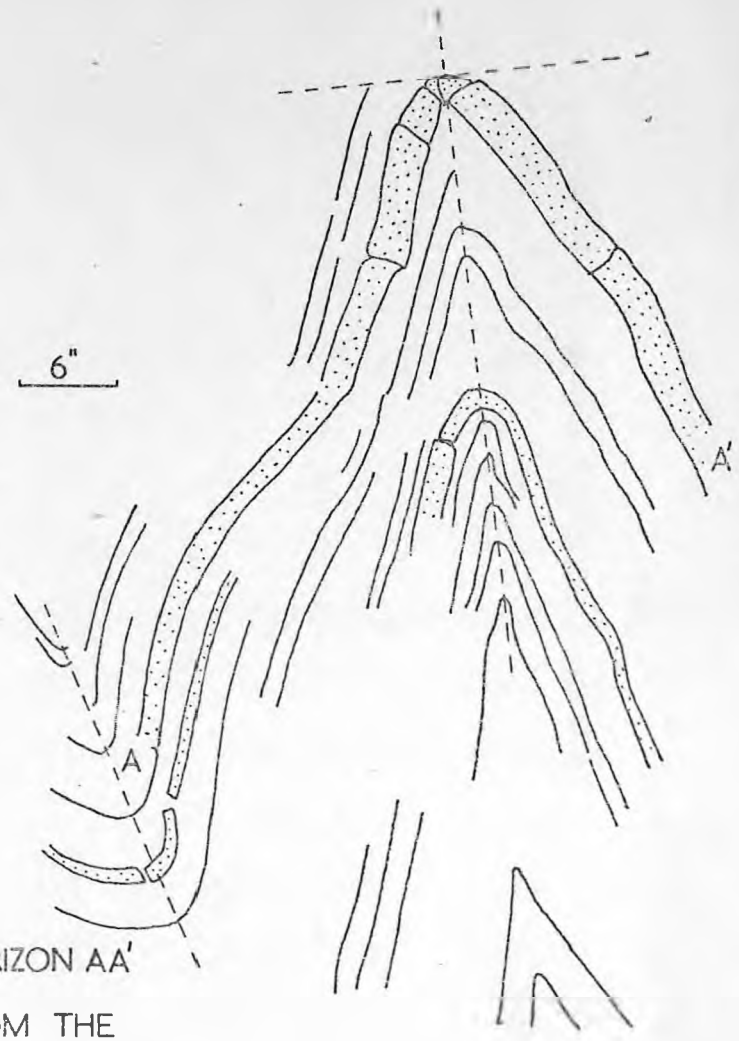
C. excellent axial-zone thrusting; a means of accommodating further deformation once the buckling limit has been reached.

flattening of the limbs. Many of these folds show excellent structures due to continued strain past the buckling limit - such as the axial zone thrusts of De Sitter (1964), (Fig. 6:2:5). Inter-limb angles are rarely less than 90°, and dip isogons are very sharply convergent due to the formation of very tight inner arcs of the folds (Fig. 6:2:5), which has resulted in a thickening of the limbs and a thinning at the hinge. When plotted on T' and t' graphs of Ramsay (1967), these folds are seen to lie very largely with Class 1A, though certain zones within the folds may be of 1B or 1C type - possibly due to the initial effects of flattening (Fig. 6:2:6).

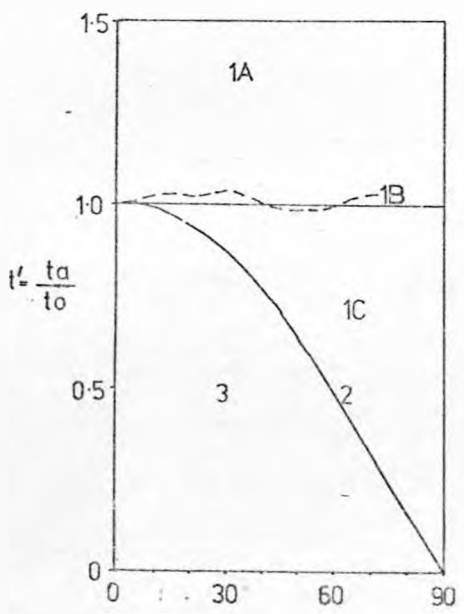
Between these strongly buckled psammite bands, the pelitic bands, where they occur, are characterized by an intense flattening-type deformation; the pelitic material flows to accommodate the changing form of the psammitic layer during buckling, and the intense strain in the inner arc leads to the development of a shear cleavage in the pelite parallel to the axial plane of the fold, (see Fig. 6:2:3, and Section 6:3).

With an increasing proportion of pelitic horizons, the psammites grade into the flaggy rocks and the Loweswater and Whiteless Pike flags show the most varied suite of F2 structures in the whole area. The ductility contrast between the psammitic and pelitic layers has a strong influence on fold shape, as does the thickness of individual layers, e.g. isolated, relatively thick psammitic bands buckle to give arcuate, or even semi-circular folds, whereas thin multi-layered units tend to form more angular folds, having straight limbs and tight hinges. Folding is dominantly by buckling, but in the multi-layered units flattening and shearing parallel to the axial plane do occur, as in the tightly appressed Whiteless Pike Syncline.

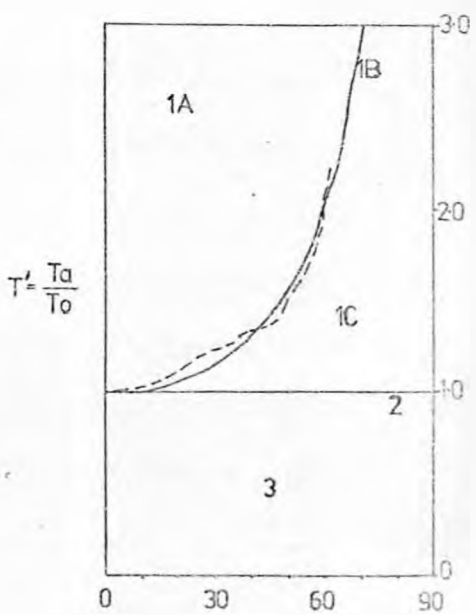
FIG. 6 2 6



T' AND t' PLOTS OF HORIZON AA'
IN A TYPICAL FOLD FROM THE
WHITELESS PIKE FLAGS (G.R. 177187)



a



Two features common in the flaggy lithology are disharmonic folding, and boudinage of thin psammite bands in the limbs of the F2 minor folds. Disharmonic parasitic folding is well seen in the flags of Dodd, and is thought to be the result of ductility contrast and the high component of thrusting generated in the northern limb of the Monocline. These folds are usually developed in the more ductile pelites between two massive bedding units, and the folds pass from their greatest amplitude in the centre of the bed into bedding-parallel thrusts at their extremities, (Fig. 6:2:7).

In contrast to the boudins of F1 age, which were rounded and "sausage-shaped", the boudins found in the limbs of the F2 minor folds are usually little deformed, having square-cut, flat ends and no signs of necking (Fig. 6:2:3). This is taken to indicate an increase in ductility contrast between the psammitic and pelitic layers with time, the psammites compacting and expelling their contained water more rapidly than the pelites and hence becoming less ductile. By way of confirmation, evidence is described in the next section which indicates the presence of some quantity of interstitial water still present in the pelites at the time of F2, despite 100 m.y. of burial and an intervening deformation.

In these flaggy rocks, the zonal nature of some of the folding is quite remarkable; gently folded, uncleaved rocks suddenly give way to tight, highly cleaved and sheared zones, which then return to gentle folding after a distance of 10 - 50 yds. Fig. 6:2:8 shews an example of this from Whiteside, which was initially interpreted as a zone up-thrust along shear planes, but close examination shows the individual lithological units to be essentially continuous across the shear planes, no large translation having taken place.

Fig. 6:2:7. Disharmonic folding from the Loweswater Flags of Dodd.



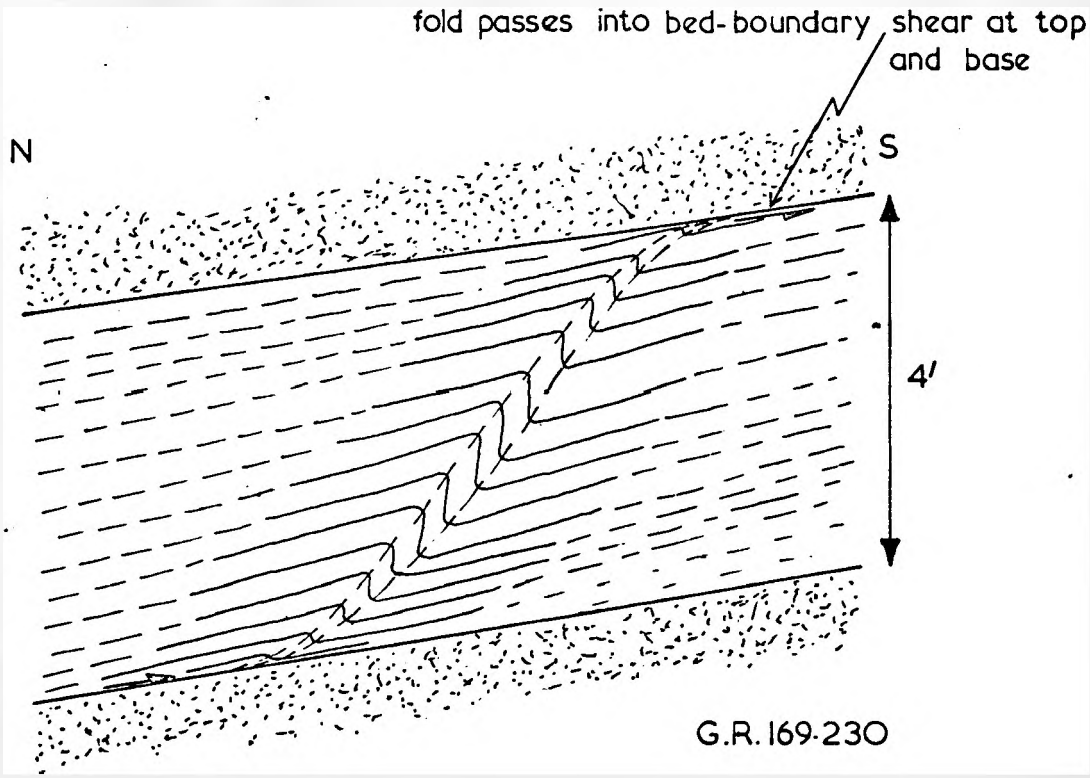

fold passes into bed-boundary shear at top and base
N
S
4'
G.R. 169·230

FIG. 6·2·7

ii) F2 structures in the pelitic rocks.

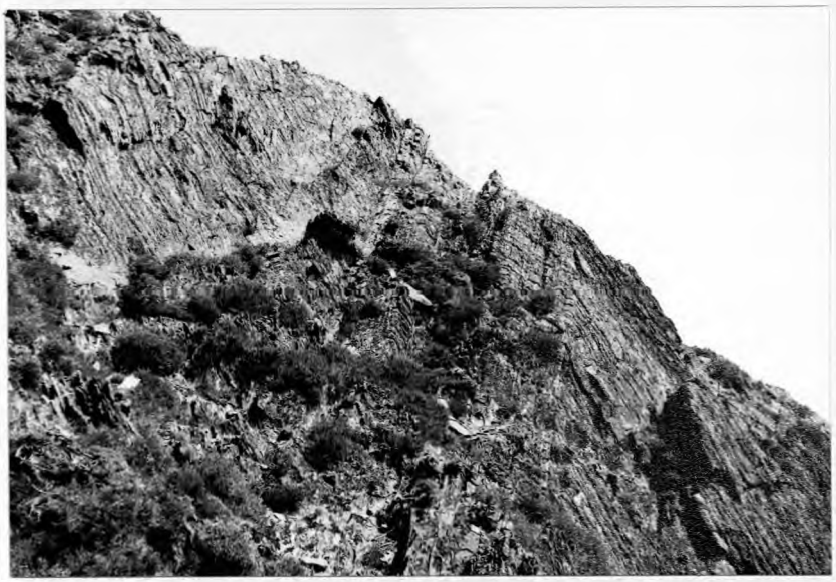
Unlike the psammitic and flaggy rocks, which shew a reasonable uniformity of fold styles across the area (with the exception of the relatively undeformed zone of the Grasmoor Dome), the nature of the F2 minor folding in the pelites varies quite markedly across the area.

The Hope Beck Slates are little deformed, being flexed into open buckle folds with horizontal ENE-WSW striking axes, and are cut by a shallow north-dipping fracture cleavage. The Slates are succeeded by the Loweswater Flags before the more intense folding of the shallow northern limb of the Dodd Monocline is seen. Hope Beck Slates are exposed in the steep, southern limb of the Monocline, however, and here they stand vertical and are cut by an intense horizontal fracture cleavage, which is so strongly developed as to be easily mistaken for bedding at first glance.

The Mosser-Kirkstile Slates of the Grasmoor Dome appear little deformed, though occasional F2 minor folds are seen, generally quite open in style and little appressed. To the north of Grasmoor, these same rocks are much more highly deformed; on Gasgale Crags, tight angular folds of flattened buckle (Type 1C) or similar (Type 2) style are found, though in the core of Braithwaite Syncline more open, rounded folds are the norm. Due to the absence of ductility contrasts, the marked axial-planar shear cleavage which characterizes the hinge regions of minor folds in the flags, is not found in these more homogenous pelites. For the first time, minor folds start to take on a definite S or Z geometry, thus enabling major fold axes to be positioned by these means.

South of Grasmoor, the Mosser-Kirkstile and Buttermere Slates are very highly deformed. The intense shearing of the axis and northern limb of the Whiteless Pike Syncline is a reflection of the large displace-

Fig. 6:2:8. Zonal development of intense S_2 cleavage in the Mosser-Kirkstile Slates along the axis of the Braithwaite Syncline.



ment that is thought to have occurred along this line. South of the axis of the Whiteless Pike-Hause Point Anticline, the rocks dip SSE towards the Buttermere Syncline, and the F2 minor fold geometry reflects their location on this south-dipping limb. The folds are dominantly of S shape (looking East), tight, and of variable plunge - this latter depending upon the dip of the pre-existing F1 surface (see Section 6:5). On Whiteless Breast, quite large S-folds can be seen, often with their steep-overturned middle limbs faulted out, though in the illustrated example (Fig. 6:2:2) this limb is preserved intact.

In Buttermere, the F2 minor folds were more difficult to distinguish, because being basically upright folds, they are best revealed on vertical surfaces, and in these Buttermere is conspicuously lacking. However, at G.R. 177174 is exposed the fold illustrated in Fig. 6:2:2; the tight, angular overturned style is typical of the F2 folds of this area, as is the lack of a marked axial plane cleavage. These folds are probably quite highly flattened if one is to judge from the shape-tightness of the F1 folds which they deform, (see Fig. 6:5:1). As mentioned above, the large majority of F2 minor folds in this area are of S-shape; only occasionally are small Z-folds discovered, these representing congrous minor folds in the steep-overturned middle limbs of larger F2 folds, such as is shewn in Fig. 6:2:2.

Section 6:3 The F2 cleavage (S2) and the "bedding cleavage".

Two distinct cleavages have, in the past, been assigned to the F2 deformation, namely the axial-planar, shear- or fracture-cleavage, S2, and what has been termed the "bedding cleavage" (Simpson, 1967; Helm, 1970).

i) "Bedding Cleavage".

This term has been used to describe the property of these rocks,

especially the pelites, to retain the bedding as the principal plane of fissility, rather than the tectonic, S2, direction. In my opinion, however, "bedding cleavage" is a strictly erroneous term, for the planar fabric which it describes is not a cleavage, although it has its uses as a convenient descriptive term. Bedding cleavage merely represents the original detrital fabric - from the first an S-tectonite due to the high phyllosilicate content of the sediments - which was reinforced by mimetic recrystallization of the sericite and chlorite during diagenesis, and subsequently during the regional metamorphism.

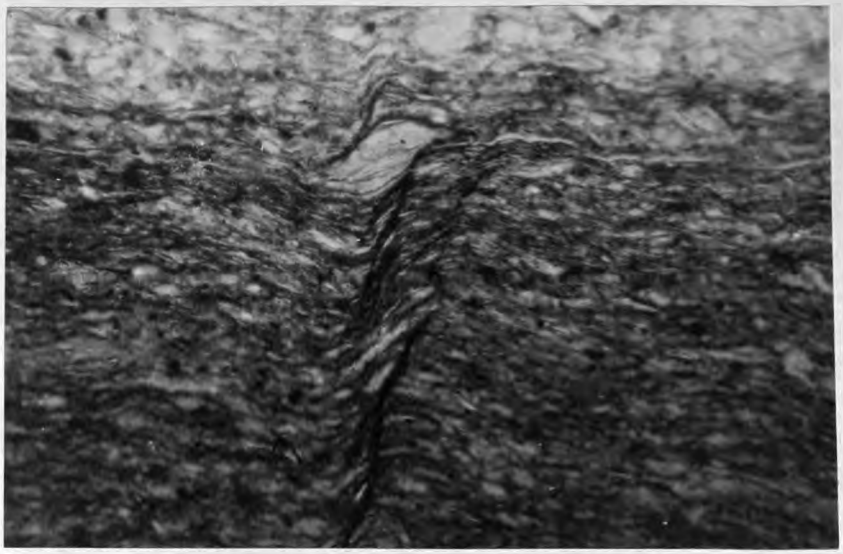
In thin sections of pelitic bands, the bedding cleavage is represented by lepidoblastic felts of fine-grained sericite shewing a very strong orientation parallel to the bedding. Discrete laths and lensoid aggregates of chlorite are also oriented in this direction. In the more competent, psammitic layers the matrix has recrystallized to form larger, more isolated, laths of sericite and chlorite, and these do not possess the same high degree of parallel orientation as is seen in the pelites, though this is only a function of the much lower ductility of the psammites. That virtually no component of F1 or F2 has contributed to the formation of this fabric can be seen in thin section, where the bedding passes around F1 and F2 fold hinges (Fig. 6:3:1) with no change, apart from slight flexing or more intense fracturing of the fabric by S2.

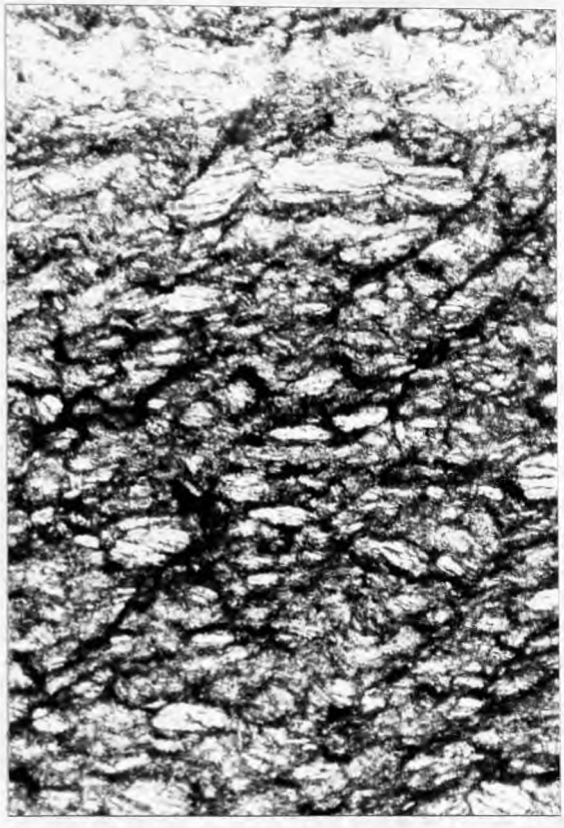
The mechanism of F2 flexural slip folding has been advocated by Simpson (1967) and Helm (1970) as a universal means of producing the "bedding cleavage", but this, I think, could only have operated given a high ductility contrast between adjacent rock layers. As has been shown above, however, the pelitic interbeds in the flaggy units are so highly deformed in the inner arcs of the resultant folds that much of

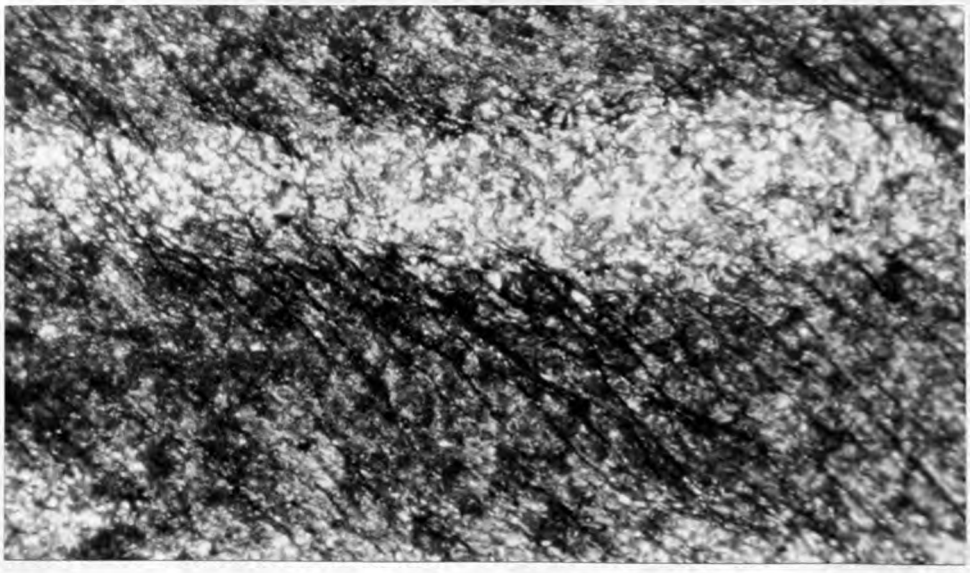
Fig. 6:3:1. "Bedding cleavage" in the hinge of an open F_2 fold.

Only slightly deformed by an S_2 crenulation, the presence of the "bedding cleavage" in the F_2 hinge indicates that the flexural-slip mechanism proposed for its formation can not be upheld.

Fig. 6:3:3. A. and B. S_2 fracture cleavage. In A, regional metamorphic chlorite porphyroblasts are deformed; in B, the cleavage is refracted through a silty band.







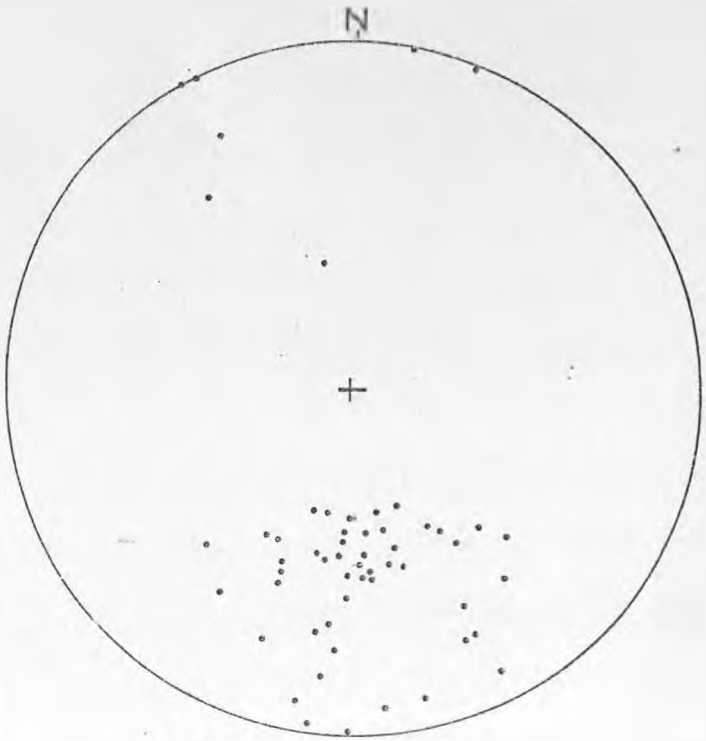
the bedding would be totally transposed - thus destroying rather than enhancing the bedding fabric.

It is admittedly difficult to conceive of a situation in which such a well-developed schistosity as that shewn steeply dipping in Fig. 7:1:2 can form entirely from passive recrystallization. The relative weakness of the S2 cleavage in these rocks has been taken to indicate that accommodation during the folding process took place by shearing along the bedding, once the latter was relatively close in orientation to the axial planes of the F2 folds. However, despite the fact that no evidence is visible in thin section to support such a contention, and although no operative mechanism can be proven, the use of the term "bedding cleavage" will be continued, but only in a descriptive capacity, there being no genetic or mechanical implications involved.

ii) S2: On a regional scale, S2 is a uniform and penetrative structure, and its occurrence through the entire thickness of Skiddaw, Borrowdale, and Silurian rocks is one of the criteria by which the end-Silurian age of the F2 deformation is judged. On the smaller scale of individual outcrops and sub-areas, however, S2 is variable both in intensity and orientation, but never approaches the condition of a slaty cleavage as has been claimed by some authors.

South of the Braithwaite Syncline, S2 is a steeply inclined fracture- or shear-cleavage, statistically parallel to the F2 axial planes and striking uniformly to the ENE - WSW quadrants. This is shewn in Fig. 6:3:2; the wide spread of poles about the ENE - WSW axis being due to the subsequent F4 refolding. The intensity of S2 varies widely throughout this sub-area, being best developed in the tightly appressed inner arcs of folds in the Loweswater Flags. S2 is usually

57 poles to S2
Gasgale Crags and
Hopegill Head



306 poles to S2
Grasmoor - Buttermere
Contoured at 1·3, 2·6,
4·0, 5·3 % per 1% area



FIG. 6·3·2

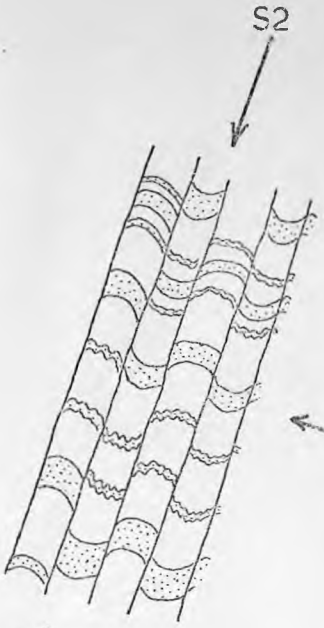
axial-planar in the hinges of these folds but, where present, fans divergently into the limbs, thus following the predicted orientation of the principal strain axes for this situation, which suggests that mechanical deformation was the dominant factor governing the formation of S2, and that solution and recrystallization of the platy minerals played a relatively minor part.

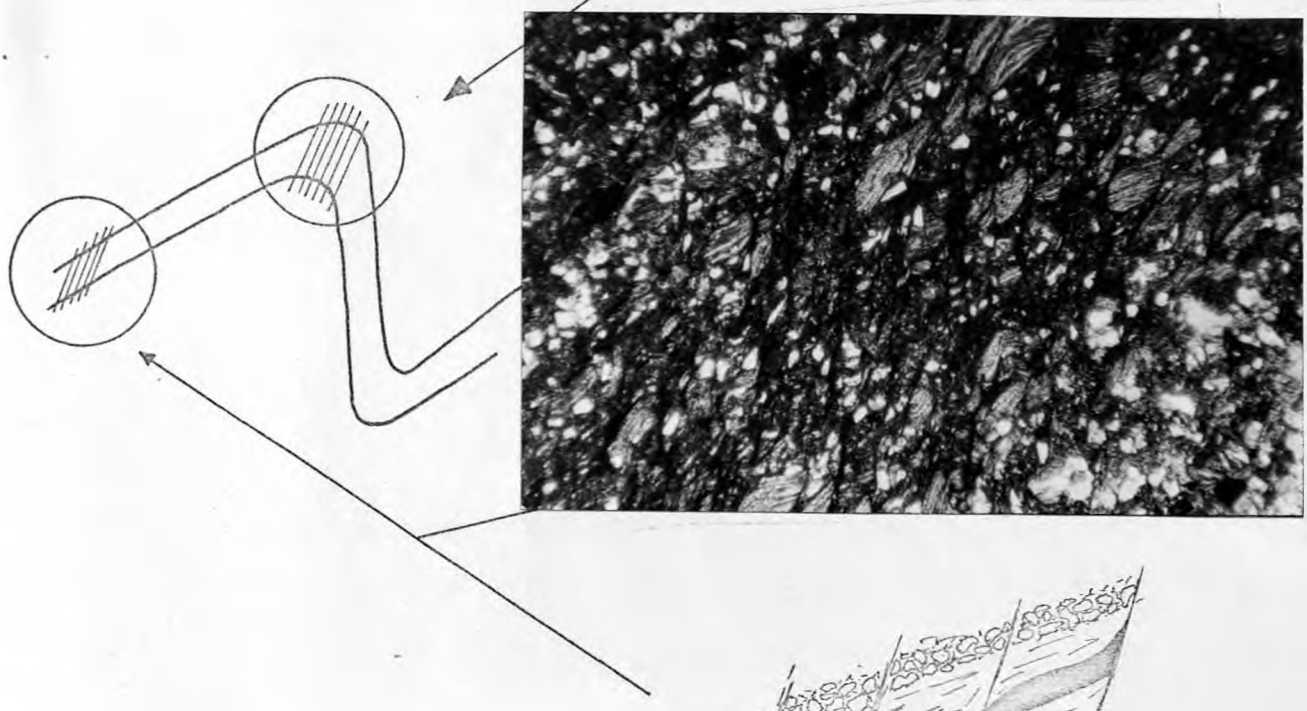
In thin section the axial planar, S2, shear cleavage is quite similar to that described by Beavis (1970), in that it is not composed of discrete fracture or shear surfaces, but consists of domains of finer crystalline material passing through which are a series of fine anastomozing shear planes, (Fig. 6:3:3). The cleavage is not controlled by mineral orientation though occasional oriented crystals do occur within the cleavage domains. No transposition of bedding into S2 (as described by Helm, 1970) has been discovered.

In addition to this shear cleavage developed in the F2 hinges, two other forms of bedding / S2 intersection have been observed. In the very thin multi-layered units occasionally found on Whiteside, one sees the development of "gleitbretter" or micro-lithons. These were formed by the extension of the limbs of small buckle folds beyond the ductile limit, at which point the limbs fracture and further hinge separation is accomplished by shearing along these fractures. The rounded hinges of the original folds are still preserved within the micro-lithons (Fig. 6:3:4). This micro-lithon formation can only occur when the bedding and S2 are roughly perpendicular; when the angle between them is small, as on the limbs of a fold where S2 fans divergently from the hinge, a flaser-like interference pattern is found. The chlorite and sericite plates in the more pelitic bands are flexed into gentle sigmoidal curves, bedding parallel in their central sections, but curving

Fig. 6:3:4. Change in style of S_2/bedding intersections with position in the folded layer.

GLEITBRETTER FORMED BY THE INTERSECTION OF BEDDING AND S2 AT A HIGH ANGLE





GENTLE SIGMOIDAL FLEXURE OF CHLORITE LATHS DUE TO LOW ANGLE INTERSECTION OF BEDDING AND S2. LITTLE RECRYSTALLIZATION OF MICA ALONG S2.

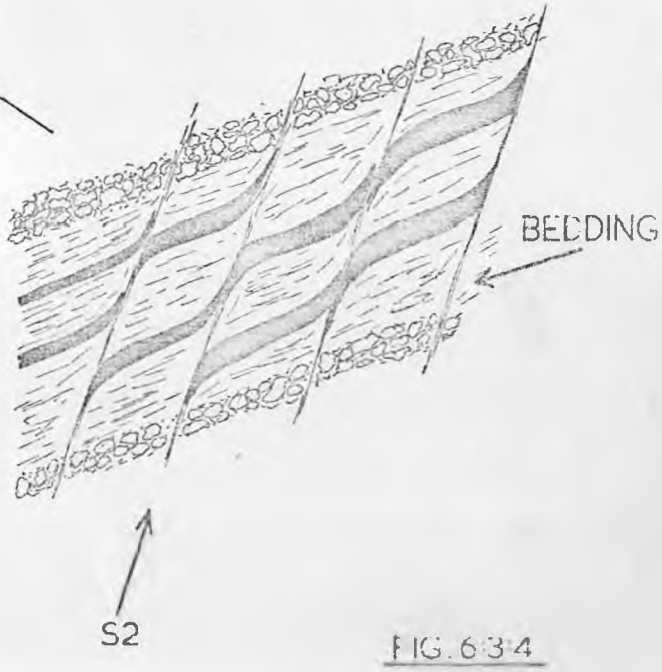


FIG. 6:3:4

into the gently-inclined S2 shear planes at their extremities, (Fig. 6:3:4).

As discussed above, the strongest development of S2 south of the Braithwaite Syncline occurs in the hinges of tight F2 folds in the flags of Whiteside. In the basically pelitic succession in the south of the area, around Buttermere, S2 is still present, though less obvious, as a fine, semi-penetrative, shear cleavage, and is best seen cross-cutting the steep, north-south striking F1 fold limbs (see Fig. 6:5:1). Over most of this area, however, S2 is largely overshadowed by the intensity of the "bedding cleavage", which closely approximates to the F2 axial plane in strike and orientation.

The weakest development of S2 occurs over the Grasmoor Dome, where the only evidence for its existence is a widely-spaced, ENE-striking, set of joints. These can occasionally be confused with the bedding because of the massive indurated appearance of the rocks of this area.

Throughout the area to the south of the Braithwaite Syncline, therefore, only a single cleavage of F2 age is present, lying roughly parallel to the axial planes of the major F2 folds. To the north of the Braithwaite Syncline, however, S2 is completely different in style and orientation and reflects a much more complex structural history, which will now be described.

On the fell-sides of Dodd and in the vicinity of Hope Beck, the slates and flags of the gentle north-dipping limb of the Dodd Monocline are cut by a shallow, north-dipping cleavage. This cleavage cuts, and therefore post-dates, the open ENE - WSW trending warps found in Hope Beck, and is approximately axial-planar to the overturned F2 minor folds described from the East face of Dodd (see Section 2:4). However, the equivalent overturned F2 folds on the west face of Dodd are seen to

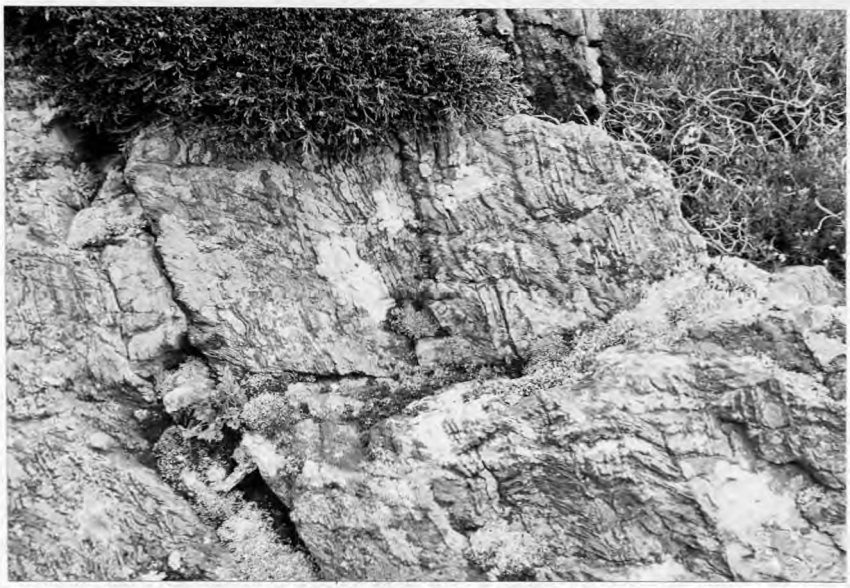
refold this low angle cleavage which becomes rotated to a moderately steep, south-dipping attitude in the middle limbs of these folds, (Figs. 6:3:5, 6:3:8).

To the south of Dodd Pass the faulted-out axial trace of the Dodd Monocline is crossed and the outcrops of Cold Gill, Penn, and Whiteside End occupy the steep southern limb of the fold. Here, too, this anomalous refolded cleavage is present in the pelitic interbeds of the Loweswater Flags; it dips southwards at 30° - 60°, roughly normal to the axial plane of the Dodd Monocline. This refolded cleavage is the dominant structural element in these rocks up to within 300 yards of the axis of the Braithwaite Syncline, at which point it is replaced by the standard, steep, north-dipping S2 cleavage of this fold.

In this north-south traverse, therefore, we have three distinct orientations of the S2 cleavage; initially gently north-dipping; it is refolded into a moderately steep, south-dipping attitude, which is then superceded by the steep, north-dipping cleavage of the Braithwaite Syncline; this sequence is shewn stereographically in Fig. 6:3:6.

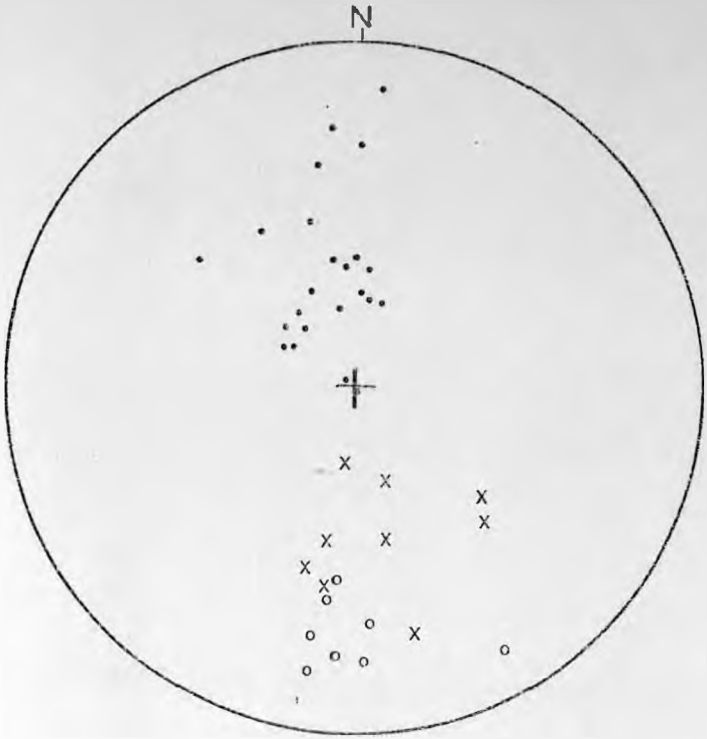
Following the principles outlined by Ramsay (1967), it is possible to plot the modification of the dihedral angle between the bedding and the inclined cleavage plane around the hinge of these monoclinal folds. The particular example chosen for this study was the minor fold on the west side of Dodd, referred to above; Fig. 6:3:8 shews the outline of the fold as traced from a photo-mosaic, and the orientation of the refolded cleavage around the hinge. The inset diagram is the graph of θ against ϕ (θ angle between bedding at any point and the axial trace; ϕ angle between bedding and cleavage at the same point). On the basis of comparison between this graph and those in Ramsay (1967) the best fit is with the theoretical curve for the

Fig. 6:3:5. Photos shewing the anomalous refolded S_2 cleavage in the Flags of the steep limb of the Dodd monocline.





VARIATION IN S2 ORIENTATION NORTH OF THE BRAITHWAITE SYNCLINE



SUB AREA A

× S2 shallow to N

· S2 refolded to S

SUB AREA B

∘ S2 steep to N

FIG. 6:3:6

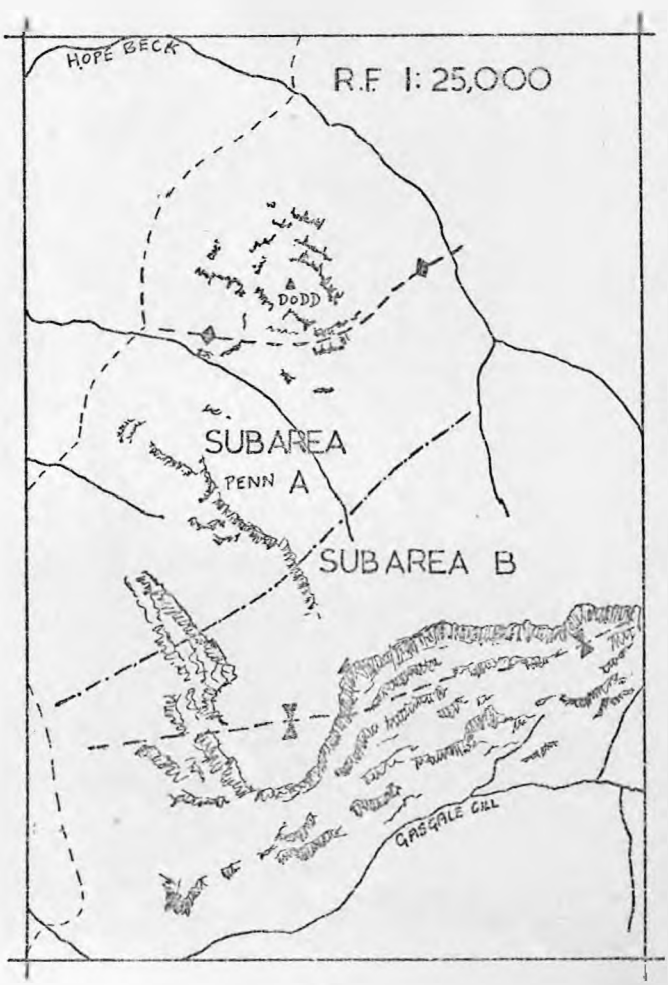
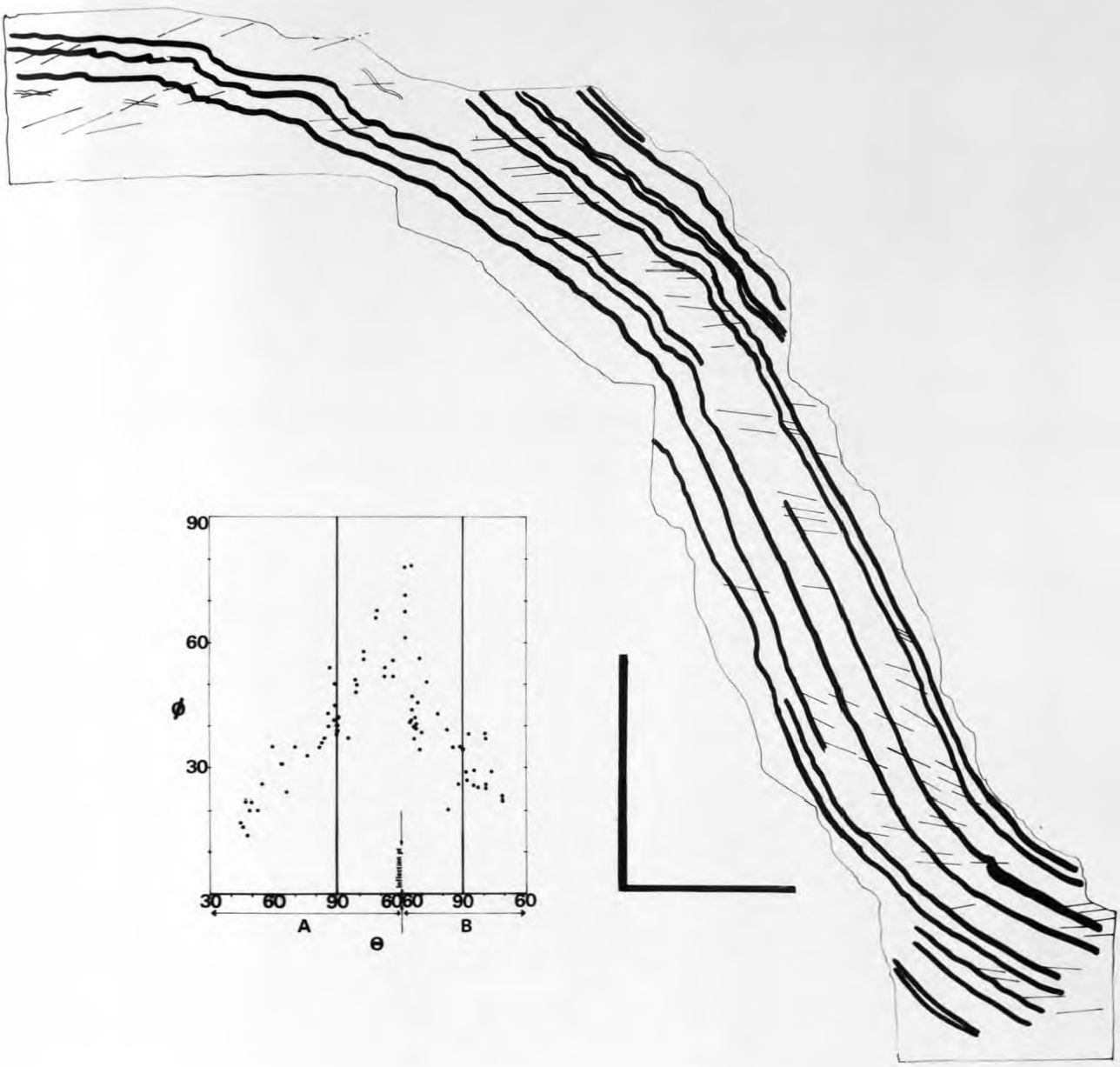


Fig. 6:3:7. The single outcrop that suggests evidence of tectonic dewatering and hydraulic injection as a mechanism for the formation of S_2; on the N. face of Dodd (G.R. 168.234). The psammitic bands are broken by pelitic injections, with the resultant cleavage in the pelites fanning away from the inner-arc hinge.



Fig. 6:3:8. Bedding and cleavage orientations as traced from a profile photo-mosaic of the overturned fold on the W. side of Dodd (described in the text).
The graph inset is a plot of ϕ (angle between cleavage bedding) and θ (angle between bedding and the axial surface) and indicates the formation of the fold by flexural slip and slight subsequent flattening.


90
60
30
ϕ
30
60
90
60
60
90
60
A
B
Θ

deformation of two inclined surfaces by flexural slip and a homogenous, flattening, strain. However, the inclined surfaces which Ramsay used were homogenous structures (such as current bedding) whereas the present example utilizes discrete fracture planes, so the two situations are not directly comparable because of the possibility in the latter of shear movement along the cleavage planes during refolding.

It follows from the above discussion that within this northern area there must have been several phases of folding within the time-span that, to the south, is represented by the single, uniform, F2 event. During this time, however, only one cleavage was formed in each area : the gentle north-dipping one in the north, and the steep north-dipping one in the south. The sequence of structures in the north can therefore be tabulated using the shallow cleavage as a structural datum:-

YOUNGEST 5) Refolding of shallow cleavage by a second episode of minor overfolding (W. side of Dodd)

4) Refolding of shallow cleavage by major Dodd Monocline

3) FORMATION OF SHALLOW, N-DIPPING CLEAVAGE

2) Larger minor overfolds on the east side of Dodd

OLDEST 1) Open warps of Hope Beck.

The conventional steep, north-dipping cleavage of the Braithwaite Syncline and the south is thought to have formed just after the shallow cleavage was refolded by the Dodd Monocline, for although no evidence was seen in outcrop of the anomalous reversed cleavage being cut by the steep S2 of the Braithwaite Syncline, this is the only explanation that is compatible with the geometry of the structures.

The major problem in this northern area is how to account for the formation of the low-angle cleavage prior to the formation of the Dodd Monocline which deforms it. Evidence from the disharmonic folding (Fig. 6:3:7) would seem to indicate that a high component of shearing

was operative during the generation of the cleavage, at least within the pelitic units of the flaggy lithology. Further evidence (Fig. 6:3:8) suggests that some tectonic dewatering may also have occurred at this stage, but since none of the injection structures described by Maxwell (1962), Powell (1972) or Roberts (1973), were found in these rocks, this process of cleavage generation is not considered to have been of major significance.

What is thought to have been the likely sequence of events will now be discussed, bearing in mind the sequence of structures described above:-

1) asymmetric compression from the north which induced open warping in the Hope Beck Slates,

2) meanwhile, the more heterogeneous flags were reacting to this imposed stress in two ways - in the psammite bands, layer shortening was the dominant process, whilst to accommodate this contraction in the psammites, the pelitic interbeds began to shear along planes inclined at roughly 30° to the bedding, and also to fold disharmonically with some, though little, tectonic dewatering,

3) this shear cleavage was axial-planar to occasional larger F2 minor folds that were developing also at this time. These folds were generated in what must still have been an essentially flat surface (a long F1 fold limb), because the low-angle cleavage extended considerably beyond the present position of the axial trace of the Dodd Monocline,

4) only at this stage could the major Dodd monoclinal structure form, when the low-angle cleavage was established for a considerable distance along what was to become the steep southern limb of the Monocline. The cleavage was refolded around this structure and the resultant pattern is that which is observed today, i.e. shallow, north-

dipping cleavage to the north of the axis and a steeper, south-dipping cleavage to the south. Minor folding also occurred, as is seen in the outcrops of the west side of Dodd (Fig. 6:3:8), and continued horizontal compression induced the formation of small conjugate folds in the gentle northern limb of the Monocline.

In summary, therefore, we have a situation similar to that envisaged by Price in his Tectonic Studies Group discussion of serial folding, in that small warps were followed by larger overturned folds, which in turn were followed by the major monoclinal flexure, with the axial trace of each structure occurring in southwards succession. The lack of the anomalous reversed cleavage in the region of the Braithwaite Syncline suggests that the compressive stress which induced the low-angle shear cleavage initially, was not transmitted for an infinite distance through the rocks but was effectively absorbed by the northern rocks, or was only transmitted at depth, thus resulting in the absence of the refolded cleavage in the vicinity of the Braithwaite Syncline.

Section 6:4 The F2 lineations.

The dominant F2 lineation in this area is that defined by the intersection of the bedding and S2, but due to the frequency with which "bedding cleavage" occurs, this lineation, L2, is only rarely developed. L2 is most commonly found in the highly deformed pelites of the southern area, and here it can vary in intensity from vague, linear traces of bedding on a cleavage face, to the intense local development of pencil slates (as found near the head Squatt Beck (G.R. 187.179)).

L2 was found to be a good b-lineation, being parallel to the F2 axes at all times, and hence the plunge variations of L2 outline the form of the initial F1 surface in the same way as do the F2 axes. The

mechanism by which this occurs was described in Section 5:3 (see Fig. 5:3:10), though this, admittedly, only strictly applies to folds formed by simple shear (i.e. similar or parallel folding). This assumption is reasonably valid in the south of the area, for no discrepancies occurred when L2 was assumed to be a perfect b-lineation, and hence L2 is combined with readings of F2 fold axes for the purposes of stereographic projection (see Fig. 5:3:9).

One further linear structure may be of F2 age; this is the NW-SE striking lineation which lies almost parallel to the axes of open F4 warps on Dodd (G.R. 170.230). As will be discussed later in section 7:2, this linear structure may represent a "slickside lineation" developed perpendicular to the axes of F2 buckle folds as a result of the latter's formation by flexural slip.

Section 6:5 The form of the F1/F2 refolded surface.

Earlier in this account, Section 5:3 outlined the ways in which the F2 axes and lineations could provide a picture of the form of the F1 folded surface. This present section discusses the converse process, i.e. the use of F1 data to define the shape of the F2 surface, and, also, the way in which S2 can aid the geometric interpretation of the refolded surface in the field. The principles to be described and the field evidence from which they were evolved are shewn diagrammatically in Fig. 6:5:1; this shows the features associated with the refolding of an overturned west-facing F1 fold-pair by an upright F2 syncline. Completion of the F2 fold-pair would have rendered the diagram too confusing, but the same principles would apply to the anticline as to the syncline illustrated, but in the reversed sense.

The field evidence from the Buttermere area indicated that the majority of the F1 folds were of Z shape, plunging to the south-east or

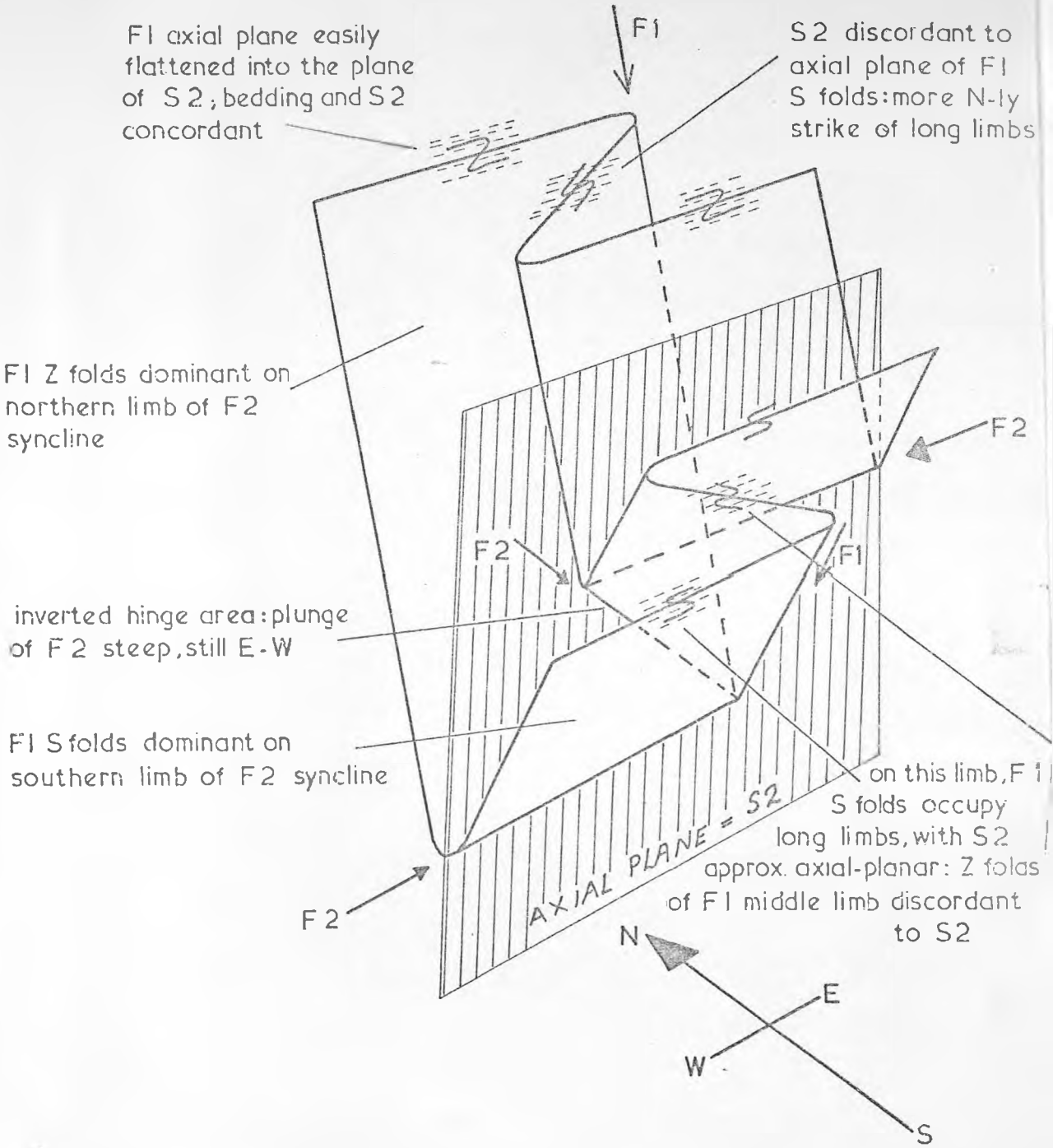
south, with their long-limbs striking roughly ENE, and their axial planes roughly coincident with the S2 direction, where seen. The large areas of Z folds were broken by linear north-south belts in which the F1 folds were of S shape, plunging south, and their long limbs, oriented north-south, were oblique to the S2 cleavage. The F2 folds over this area were all of S shape, facing north, though small Z folds occurred in the steep middle limbs of the larger of these F2 S folds.

The southerly plunge of both S and Z style F1 minor folds, coupled with the dominant bedding dip to the SSE, known to be right-way up (from Buttermere quarry), and the S shape of the north facing F2 folds locates the whole of the Buttermere area on the northern limb of the Buttermere Syncline, and this situation is represented on Fig. 6:5:1 by the northern limb of the F2 syncline.

Refolding of the F1 fold-pair about the F2 axis results in the structural pattern shown for the southern, north-dipping limb of the F2 syncline. In this case, the S2 cleavage would be roughly parallel to the axial planes of S shaped F1 minor folds, and discordant to the Z folds, with these Z folds being located in the linear north-south zones. The fact that this situation was not recorded in this area signifies the very limited outcrop area over which the steep, north-dipping, middle limb of the S shaped F2 minor folds extended. South of the axis of the Buttermere Syncline, however, this pattern would be the dominant one, and from my limited excursions over the ground to the south, this would seem to be the case.

The diagram also illustrates the way in which any change in the F1 plunge direction indicates the presence of an F2 axis, and it shows the assistance which S2 can offer in the field, by indicating one's place in such a refolded structure, in combination with such features as bedding dip or F1 fold style.

FIG. 6:5:1



∴

ANY CHANGE IN F1 PLUNGE DIRECTION INDICATES PRESENCE OF F2 AXIS
" " " F1 STYLE " " " F1 "
" " " F2 PLUNGE DIRECTION " " " F1 "

RELATIONSHIP BETWEEN F1 AND F2 IN THE BUTTERMERE AREA

CHAPTER SEVEN

F3 AND F4 DEFORMATIONS; FAULTS AND JOINTS.

Section 7:1 The F3 Deformation.

The F3 deformation is characterized by open, recumbent, sideways-closing folds having sub-horizontal axial planes (Fig. 7:1:1). There is an associated axial-planar fracture cleavage, S3, and the intersection of S3 with the bedding or S2 yields a perfect b-lineation (L3) parallel to the F3 axes (Fig. 7:1:2).

By virtue of their sub-horizontal axial surfaces, the trend of F3 fold axes and L3, and the dip of the axial planes and S3, varies according to the orientation of the bedding surface on which they are imposed. Hence, where bedding strikes are largely a function of F2 deformation, the F3 fold axes and L3 plunge gently (0-25°) to the ENE or WSW quarters, with the S3 surfaces dipping gently to the north or south, but where F3 is imposed upon bedding which inherits a N. - S. strike from F1 folding, then F3 and L3 plunge gently north or south, and S3 dips to the east or west (Fig. 7:1:3). This bi-modal distribution, though heavily biased in favour of the F2 directions, can still be seen in the stereographic plot of F3 axes, L3 and S3 (Figs. 7:1:4 and 7:1:5).

In certain instances, as on Whiteside, it can be difficult distinguishing between F2 and F3 folding, where minor folds of the F2 "clinoria" approach the vertical at the inflexion point of the median surface. At most localities though, F2 and F3 are clearly distinguishable, and indeed, it is the deformation of F2 by the open recumbent warps of F3 (Fig. 7:1:6), that fixes the latter's place in the structural sequence.

No major F3 folds have been found; Simpson's (1967) Crummock

Fig. 7:1:1. Open, recumbent F_3 folds.

Note horizontal axial planes.

A. N. end of Rannerdale Knotts (G.R. 164.184)

B. On Rannerdale Knotts, 200' above the road. (G.R. 163.183).





Recumbent Syncline (see Fig. 1:3:3) does not exist, though its form may possibly have been misunderstood from the geniculate pattern of major joints in the eastern face of Mellbreak on the opposite side of the valley.

The intensity and style of F3 folding and S3 cleavage vary widely with lithology. In the massive, metamorphosed, mudstones of Rannerdale Knotts and Low Bank, they occur as large open warps with very little associated cleavage (Fig. 7:1:1) though occasionally they do form angular flexures having a weak fracture cleavage concentrated in the hinges (Fig. 7:1:3). In the finely bedded, more ductile, pelites the D3 folds are tighter, more often angular in profile, and the associated S3 becomes quite an intense crenulation or strain-slip cleavage (Fig. 7:1:2).

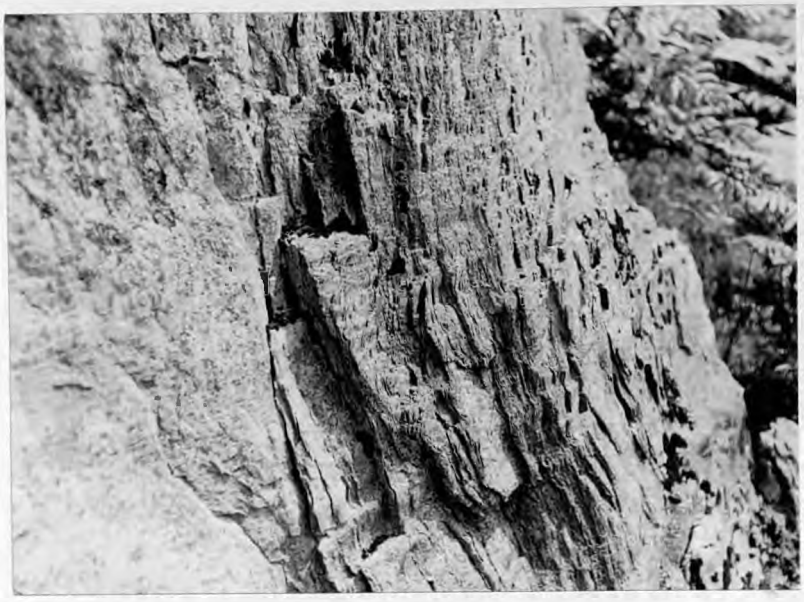
The S3 cleavage is everywhere flat-lying (Fig. 7:1:5), and is parallel to the axial planes of the F3 folds. There is great variation in the intensity of S3 development within the area, and this is seemingly independent of the presence of F3 folding. In Buttermere, where relatively few F3 folds occur, a penetrative S3 is very strongly developed and in some cases almost obliterates the bedding. Such S3 cleavage surfaces as these have an almost phyllitic sheen, suggestive of quite a high degree of sericite recrystallization along the plane of schistosity, though this is rarely seen in thin section.

Although, in hand specimen, S3 frequently appears as a perfect crenulation cleavage, in thin section the bedding or S2 fabric between the individual S3 fractures is little deformed (Fig. 7:1:7), and this lack of crenulated cross-foliation between the cleavage domain means that S3 does not conform to the Rickards' (1962) definition of a true crenulation cleavage.

To the north of the Buttermere-Low Bank-Rannerdale Knotts area, F3 folding is little in evidence, inhibited by the increasingly massive

Fig. 7:1:2. Steep dipping Buttermere Slates on High House Crag, shewing strong S_3 crenulation cleavage, and the bedding/S_3 intersection lineation, L3.

Fig. 7:1:3. F_3 folding is imposed on N - S striking bedding on the N. end of Rannerdale Knotts. Both F_3 fold axis and the S_3/bedding intersection lineation plunge to the N (towards the reader).





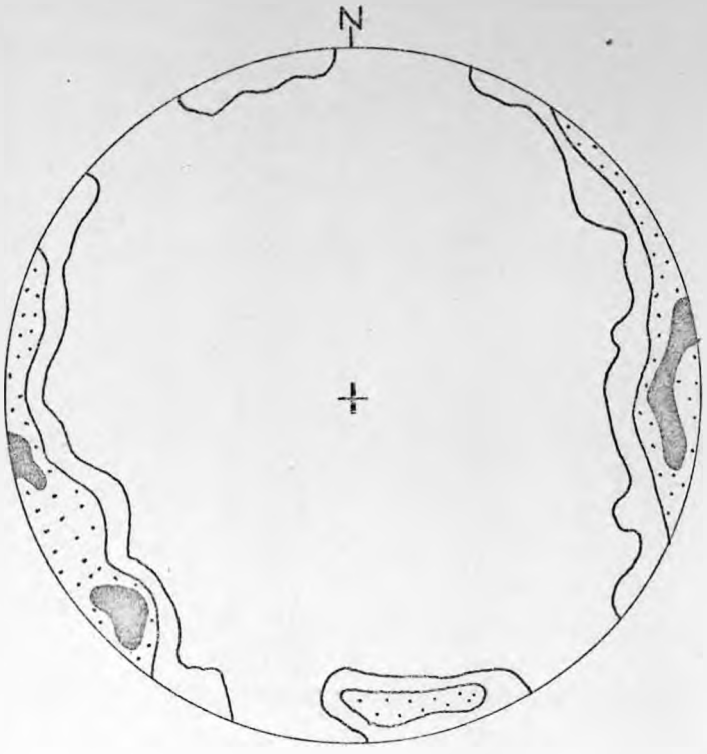


FIG. 7.1:4

F3 FOLD AXES & L3

120 points: contours at

1.5, 6.0, 10.0 % per

1% area

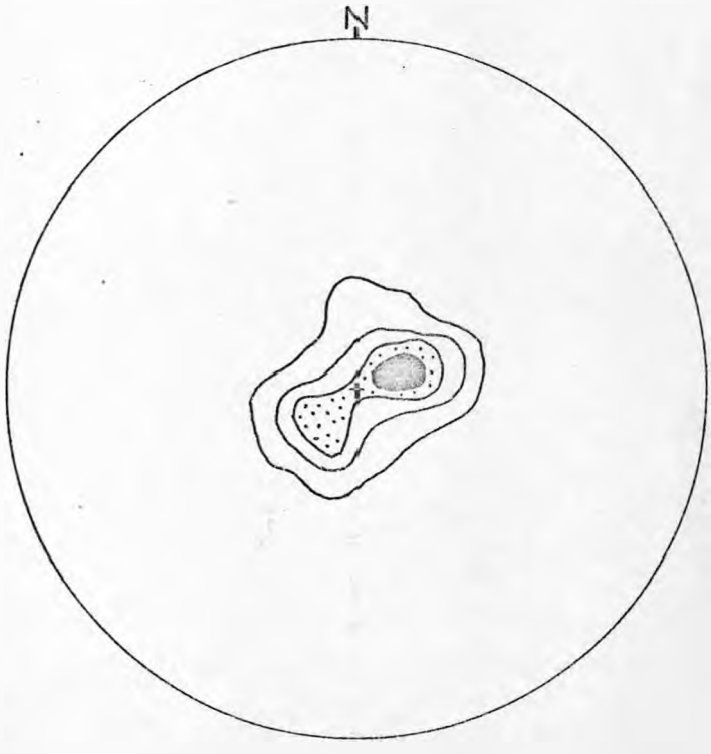


FIG. 7.1:5

POLES TO F3 AXIAL

PLANES & S3

67 points: contours at

3; 6; 9; 12 % per 1% area

Fig. 7:1:6. Tight F_2 syncline deformed by recumbent warps of F_3 and its associated axial plane parallel fracture cleavage, S_3.

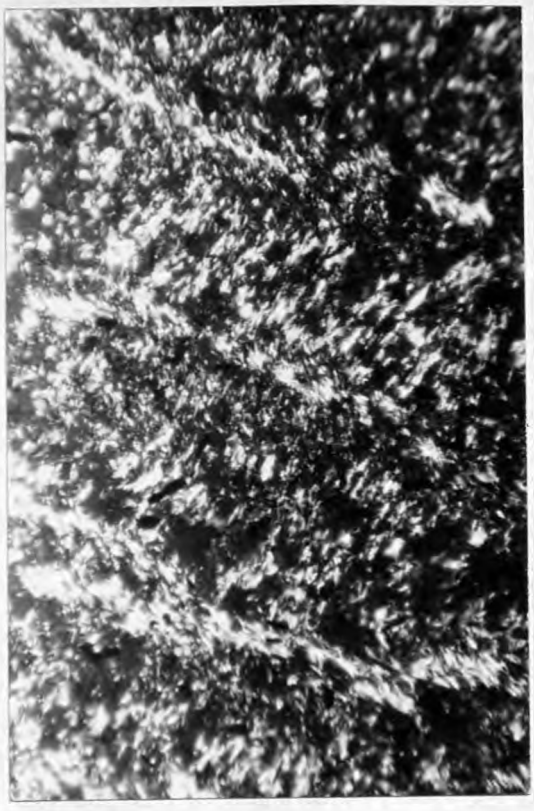
Fig. 7:1:7. S_3 fracture cleavage cutting bedding in thin-section.

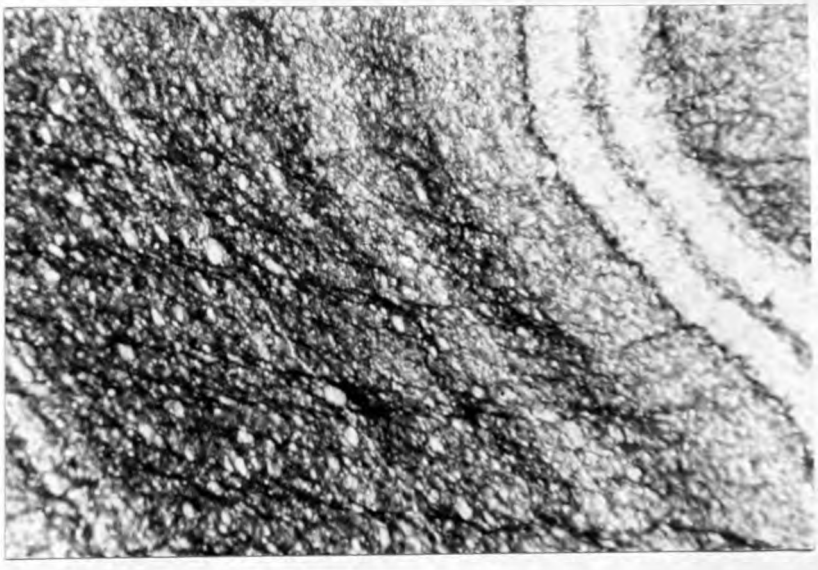
(see also Fig. 4:1:8).

A. x 40, x Nicols

B. x 25. PPL







nature of the rocks outcropping in that direction.

As mentioned above, the age of the F3 deformation relative to F2 is easily established both from evidence of refolding and superimposition in the field, and the overprinting of cleavage fabrics in thin section.

The effect of F3 and F2 (and F1) as revealed by stereograms is small: F1 and F2 fold axes are not deviated because F3 is co-axial with them, but the open F3 warping does produce the bi-modal distribution of bedding and S2 poles about the primitive circle.

Section 7:2 The F4 Deformation.

Folds of this episode are of open rounded form, with vertical axial planes, and having fold axes plunging down the dip of the bedding surfaces which they deform. The strong dominance of Caledonoid (F2) strikes means that both F4 axes and axial surfaces most commonly have ENE - WSW strikes (Fig. 7:2:1).

Only rarely is an S4 cleavage associated with the F4 folds, the folding generally being too weak, but there is, strangely, a well-developed b-lineation. This lineation has a pitch of 90° on the bedding surfaces and plunges parallel to the F4 axes, down the bedding dip. Whether or not this lineation is correctly called L4 is uncertain, it being strange for a strong lineation to be produced without the presence of a marked associated cleavage; this anomaly could be explained if the structures in question were slickenside or slip-lineations developed during the formation of F2 folds by flexural slip. This problem is well illustrated by quite a large scale F4 fold-pair from the crags above Dodd Pass (G.R. 170230) shown in Fig. 7:2:2. The F4 axes plunge at 20° to 329°, and the lineation in question plunges at 14° to 324°, which is normal to the strike of the F2 folds in the area. In view of the lack

Fig. 7:2:1. Equal-area stereographic plot of F_4 fold axes and poles to S_4.

Fig. 7:2:2. A. Poles to bedding (small dots) recorded from F_4 folds at G.R. 170.230, lie on a great circle the pole of which co-incides with the F_4 plunge. The position of the $?L_4$ lineation is also shewn.

B. Block diagram illustrating the relations between F_4 and F_2 at G.R. 170.230, and shewing the possible formation of the lineation as a "slickenside" due to F_2 flexural slip.

• F4 fold axes
• poles to S4

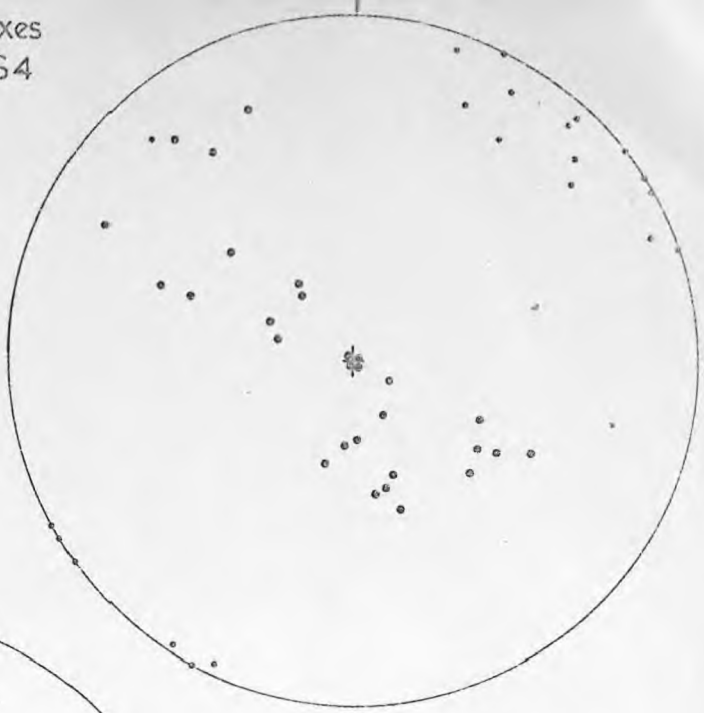


FIG. 7:2:1

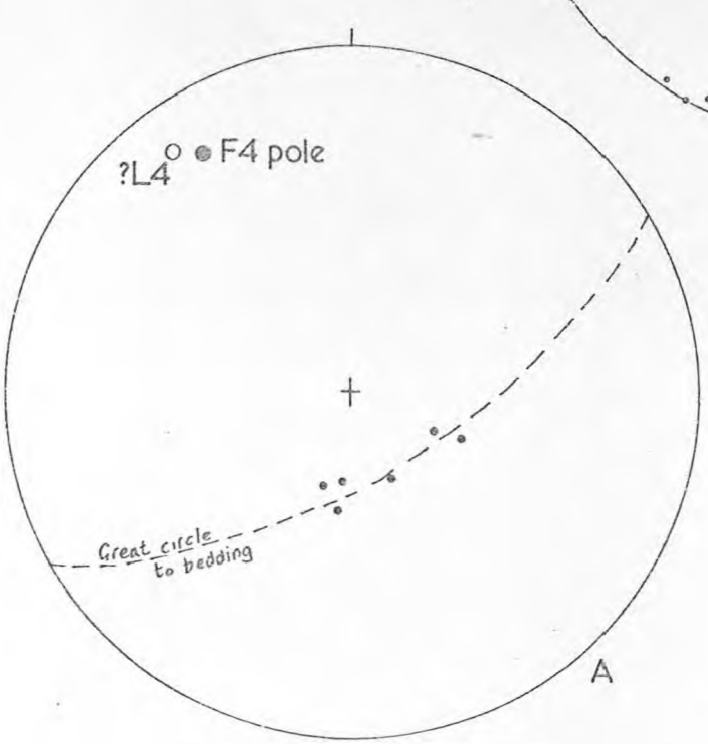


FIG. 7:2:2

F4 & ?L4 FROM DODD PASS

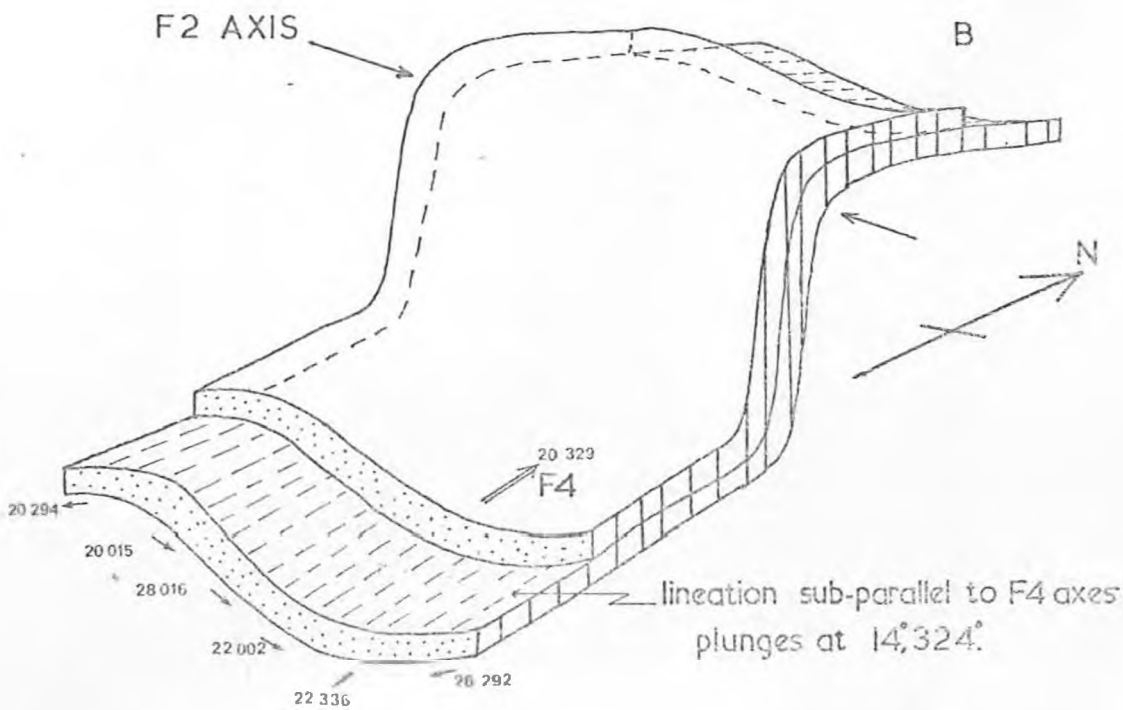
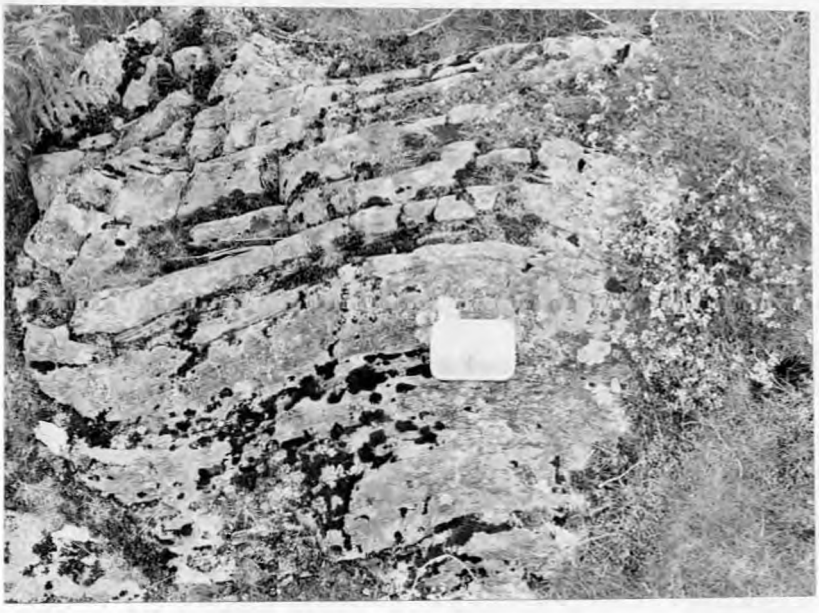


Fig. 7:2:3. Photo shewing the open, steep-plunging style of F_4 folding. (G.R. 178.176).

(see also Fig. 5:3:3).



of a distinct S4 cleavage, I think the most likely explanation of this, and possibly other occurrences where a lineation sub-parallel to the F4 axes is found, is that the lineation represents a slickenside lineation consequent upon the formation of the F2 folds by flexural slip.

F4 folds are not commonly found within this area, but most often occur in the flaggy lithologies, where they are usually small scale features (wavelength of 18"; amplitude of 6"). Larger scale examples are found in the more pelitic units, but rarely as discrete folds; instead they appear as wave-like variations in bedding strike, with a wavelength of 20 feet or more.

No regional-scale structures of F4 age are thought to occur. Whether the periclinal form of the Mosser-Kirkstile Slate outcrops in the hinge of the Braithwaite Syncline are the result of F4 warping, rather than the result of variation in F2 compressive strain, is not known, but either of these mechanisms would account equally well for the observed outcrop pattern.

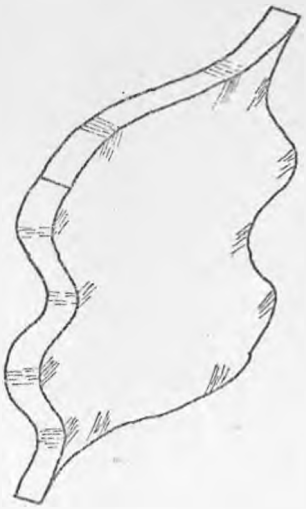
The effect of F4 refolding on the distribution of bedding, F2 axes and S2, is well seen in the relevant stereograms (Figs. 6:2:1; 6:3:2; Plate 3), with F4 responsible for the wide spread of these structures around the primitive circle within the ENE - WSW quadrants.

The distinction of F4 from F1 folds has been discussed earlier in Section 5:2 and will not be further elaborated.

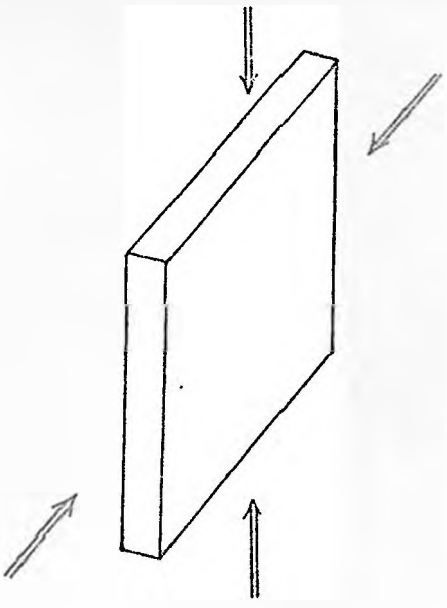
Section 7:3 Relationship of F3 and F4 to the Main Deformation, F2.

F2 tectonism was characterized by strong NNW - SSE compression which resulted in quite intense folding and flattening of the Skiddaw Slate pile, and especially of the ductile pelitic units in the south of the area. Such a flattening-type F2 deformation (Flinn, 1962) would have resulted in the production of a large scale S-tectonite, with

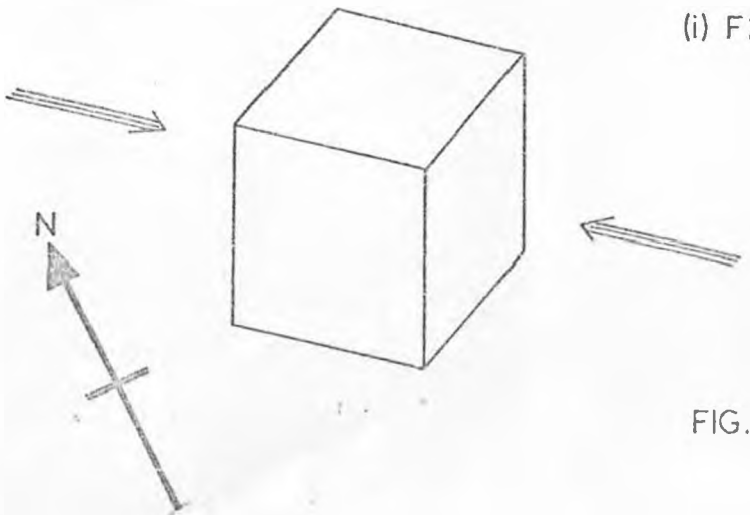
Fig. 7:3:1. Formation of F_3 and F_4 by vertical + axial compression consequent upon the extensive forces exerted in these directions by the F_2 flattening.



(iii) FORMATION OF F3 & F4 BY VERTICAL & AXIAL COMPRESSION CONSEQUENT UPON (ii)



(ii) CONSEQUENT VERTICAL & AXIAL EXTENSION



(i) F2 COMPRESSION

FIG. 7:31

compensatory extension of the rock both vertically and ENE-WSW along the axis of the Skiddaw trough, i.e. within the XY plane of the finite-strain ellipsoid (Fig. 7:3:1). The disposition of large masses of rock in a steep to vertical orientation is gravitationally unstable, and the recumbent, open F3 buckle folds indicate a regime of vertical shortening consequent upon the gravitation collapse of these F2 structures.

Superimposed upon F2 and F3 are the steeply-plunging, open buckles of F4 - indicative of a late stage bulk axial shortening, counteracting the extension in that direction induced by the strong F2 flattening.

Section 7:4 Faults and Joints.

It is very difficult to map out the faults in the Skiddaw Slates, because of the absence of distinct lithological horizons; hence unless the fault reveals itself by topographic expression, or is actually exposed, it may be easily missed.

All of the faults mapped in the present age were, with one exception, located by a combination of topographic expression and anomalies in the outcrop distribution, i.e. on Gasgale Crags the steep northern limb of the Whiteside Syncline is absent, and a marked linear feature corresponds with the line along which it should outcrop, thus the presence of the fault may reasonably be postulated.

The only fault actually exposed in the area is the overthrust at G.R. 182217, where a corrugated succession of flags and slates is thrust southwards over a small stock of dolerite.

The majority of faults recorded were strike faults, but occasional NW - SE (Squatt Beck and Rowantree Beck Faults) and N - S (Dove Crags Faults) faults were found.

Joints are the latest tectonic phenomena recorded from the rocks of this area. Without exception, the master joint system trends N - S

Fig. 7:4:1. Low mist over Gasgale Crags revealing the topographic expression of the N - S master joints as linear, parallel gullies.



and the topographic expression of these joint-controlled zones of weakness is well seen in the zig-zag form of Gasgale Crags - best revealed by mist (Fig. 7:4:1). Occasional joint faces on Grasmoor have yielded specimens of wavellite - the hydrated phosphate of alumina.

Very little evidence of the complementary direction to the north-south set was found, and this prompts the question as to whether re-activation of N - S lineaments in the (presumed) Pre-Cambrian basement might be responsible for the N - S directions of both the F1 folding and the master joints. There is a considerable history of basement re-activation along N - S lines in the Welsh Borders (Malvern axis), Pennines (Cross Fell Fault), and the Southern Uplands (Leadhills line), therefore extension of such movements into the sub-Lake District basement is not impossible. It assumes that Pre-Cambrian basement does occur beneath NW England and the evidence for this is presented in the palaeogeographic reconstruction (Section 2:5).

SUMMARY

"..... two sets of gritty and flaggy beds in the Skiddaw Slate series, the lower being by far the most important, and mainly forming the mountains along the NE - SW axis of the Skiddaw Slate area, while the more shaly beds, with sometimes the upper series of sandy strata, roll away to the SE and NW on either side. Thus it happens that no estimate can be made of the true thickness of the series, since the underlying formation, whatever that may be, is never exposed. Moreover ..... (with) few definite horizons being determinable it becomes hard to make even a fair approximation of the thickness of the series exposed."

Ward (1876, p.42).

TWO POINTS FOR FUTURE RESEARCH

1) Test my stratigraphic succession by sampling the three pelitic units (Hope Beck Slates, Mosser-Kirkstile and Buttermere Slates) and, by using both micro-palaeontology and X.R.F. analysis, attempt to show that the three units are faunally and geochemically distinct.

2) Further investigation of the unusual and complex deformation sequence in the limbs of the Dodd Monocline. Does the serial folding hypothesis apply?

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APPENDIX A

Sample Preparation

The sediments and intrusives sampled were all fine grained and had a high degree of internal homogeneity, hence relatively small specimens (2-3lbs.) would provide statistically representative samples. The sample was first crushed down to fragments of 5-10mm size in a fly-press having hardened steel crushing faces to reduce contamination. During this crushing process as much of the weathered crust as possible was removed from the sample, although in the case of many of the pelites there was extensive weathering and iron-staining along joints and cleavage planes which it was impossible to eliminate.

The crushed, non-weathered material was then transferred to a "Tema" swing mill, having non-contaminating, tungsten carbide grinding surfaces, and milled for 30-40 seconds. This produced a coarse powder from which a split of 50-60 grams was taken by quartering, to be further reduced by grinding for 3 intervals of 40 seconds, with a cooling interval of 60 seconds between each run, to prevent fusion of the powder to the "Tema" walls. The resultant powder, now of 50-60u mean grain size, was stored in self-sealing, airtight plastic bags, and the mill was wiped or washed clean before starting the next sample.

Seven grams of the fine powder was then taken and mixed into a dry paste with 12-14 drops of "Mowiol" binding agent, (Mowiol = polyvinyl alcohol, dispersed in a 1:5 mixture of ethanol and water). The paste was then placed between polished steel plattens in an evacuated die and pressed under a hydraulic ram at a jaw load of 12-13 tons for 30-60 seconds. The resultant pellet, roughly 3cm in diameter and 3-4mm thick, was then labelled, dried, and stored in a self-sealing plastic bag.

The determination of 25 elements was carried out on the departmental Phillips PW 1212 automatic X-ray spectrometer, following essentially the methods of Leake et al. (1969). Two measurements were made on each pellet and the resultant counts were printed out and also punched out on 5-hole computer tape, this latter then being processed using a computer programme written by A.C. Skinner. This programme subtracts the background count from the peak count of each element, and divides the result by the number of counts on the ratio standard. The result is then multiplied by the appropriate calibration coefficients and corrections for interfering elements are performed. The mean number of ratiod counts is then printed out, together with the concentration of the element in weight percent oxide (major elements) or ppm element (trace elements).

The detection limits for the analyzed elements are given below:

Elements	ppm	Elements	ppm	Elements	ppm
Si	20	P	4	Zr	1
Ti	10	S	30	Ba	6
Al	70	Cr	2	La	6
Fe	1	Nc	1	Ce	6
Mn	1	Ca	1	Pb	8
Mg	50	Zn	1	Th	5
Ca	2	Rb	1	Ga	2
Na	200	Sr	1		
K	2	Y	2		

The accuracy and precision of the X.R.F. method of analysis has been discussed in detail by Leake et al (1969) and comparison of the Birmingham University X.R.F. results with the accepted values and ranges for the international standards G-1, G-2, AGV-1, BCR-L (Flanagan, 1969; Fleischer, 1969) shews reasonable to good agreement for most of the elements, though Mg and Rb tend to be slightly low, and Si slightly high. Al is the most variable element.

APPENDIX B

Geochemical Results - IGNEOUS ROCKS.

BASIC IGNEOUS INTRUSIONS (NORTHERN AREA and ? VENT)

SAMPLE NO.	J	Lx	L13	L35	L41	L24	VENT
ANALYSIS NO.	9007	9015	9016	9017	9018	9019	9012
SiO_2	54.32	53.38	48.79	49.67	54.84	54.92	57.11
TiO_2	1.32	1.15	1.67	1.29	1.15	1.17	1.10
Al_2O_3	15.46	22.72	16.60	18.53	15.54	14.74	13.81
Fe_2O_3(total Fe)	13.90	11.34	14.26	14.77	13.96	13.85	15.89
MnO	0.26	0.19	0.75	0.17	0.16	0.29	0.23
MgO	7.20	2.84	5.05	7.63	7.27	6.95	5.15
CaO	0.96	0.79	2.36	0.80	0.87	0.92	0.90
Na_2O	0.48	2.05	3.66	1.76	1.75	2.19	0.38
K_2O	0.99	3.09	0.74	1.29	0.80	0.67	0.53
P_2O_5	0.12	0.09	0.08	0.08	0.11	0.10	0.04
Trace Elements in p.p.m.							
S	78	246	254	239	168	202	1069
Cr	340	133	25	140	115	116	932
Ni	86	33	52	56	52	51	191
Cu	216	115	225	193	201	184	205
Zn	140	85	156	153	168	103	99
Ga	14	18	12	13	12	12	11
Rb	65	180	41	77	50	47	29
Sr	102	142	203	213	150	110	62
Y	18	30	25	25	20	21	19
Ba	311	598	252	250	231	226	128
La	28	44	19	19	20	20	12
Ce	30	95	26	16	40	16	-
Pb	51	28	36	10	23	15	21
Th	7	7	3	-	6	6	4

BASIC IGNEOUS INTRUSIONS (SOUTHERN AREA)

SAMPLE NO.	S.80	S.103	605	G	965	596	918	948
ANALYSIS NO.	9002	9006	9008	9009	9010	9011	9013	9020
SiO_2	53.79	47.74	45.94	41.61	46.53	44.14	42.88	47.22
TiO_2	1.18	1.06	1.16	1.03	1.18	1.33	1.42	1.05
Al_2O_3	13.17	13.68	12.57	12.77	13.78	14.67	16.83	12.45
Fe_2O_3	13.05	17.22	14.33	14.36	14.61	17.01	15.52	12.96
MnO	0.21	0.53	0.37	0.25	0.52	0.29	0.25	0.27
MgO	7.24	8.30	15.31	12.26	6.68	10.95	10.43	10.46
CaO	4.51	3.86	3.78	8.46	7.39	3.69	4.79	7.22
Na_2O	0.31	0.35	0.46	1.50	0.68	1.64	0.29	1.16
K_2O	0.87	0.58	0.07	0.04	1.09	0.37	1.74	0.25
P_2O_5	0.09	0.10	0.13	0.10	0.08	0.12	0.11	0.09
Trace Elements in p.p.m.								
S	95	117	148	473	469	224	241	581
Cr	247	559	849	317	371	248	276	333
Ni	122	123	209	92	99	62	62	49
Cu	185	186	178	191	158	253	183	102
Zn	121	136	110	97	106	139	123	109
Ga	8	13	12	11	10	14	14	12
Rb	52	31	4	4	53	14	73	18
Sr	57	53	62	155	87	243	68	158
Y	29	20	21	19	22	26	23	23
Ba	259	223	101	70	252	228	575	93
La	40	24	17	-	8	27	21	23
Ce	6	34 *	30 *	11	16	44	22	41
Pb	9	30	10	10	15	20	16	15
Th	4	3.7	11	-	-	4	8	10

* Ce values to be ignored if Cr 500p.p.m.

APPENDIX C

CHEMICAL ANALYSES OF SKIDDAW SLATES FROM RANNERDALE AND GRASMOOR AREAS

MAJOR ELEMENTS (wt. %)

PEL.NO.	SAMPLE NO.	SIO2	TIO2	AL203	Total Iron as FE203	MNO	MGO	CAO	NA20	K20	P205	H20+	H20-	CO2	TOTAL
70020	M1 PJFJ	49.978	1.089	25.153	11.356	0.575	2.169	1.360	1.383	3.370	0.121				965.540
70030	M2 PJFJ	51.664	1.288	25.135	11.365	0.387	1.821	0.862	1.499	2.986	0.072				970.790
70040	M3 PJFJ	57.452	1.366	21.052	9.688	0.167	2.136	0.821	1.211	2.755	0.105				967.530
70050	M4 PJFJ	59.580	0.995	15.230	13.929	0.232	2.633	0.783	0.527	1.588	0.082				955.490
70060	M5 PJFJ	50.840	1.317	21.075	14.007	0.334	2.687	1.009	0.893	2.822	0.160				951.440
70070	M6 PJFJ	55.630	1.399	24.285	8.227	0.208	1.168	0.729	0.543	4.806	0.104				975.490
70080	M8 PJFJ	56.366	1.376	25.059	6.925	0.045	1.639	0.919	1.231	3.615	0.128	Not determined			972.730
70090	M11 PJFJ	53.278	1.334	25.660	8.644	0.201	1.559	1.043	1.311	3.674	0.139				968.430
84410	M9 PJFJ	52.023	1.372	25.196	9.667	0.092	1.785	0.833	1.273	3.650	0.086				959.770
84420	M13 PJFJ	53.139	1.341	25.791	9.192	0.201	1.174	0.925	1.379	3.363	0.125				966.300
84430	M14 PJFJ	52.534	1.311	26.664	8.875	0.178	1.564	0.946	1.138	3.796	0.109				971.170
84440	M15 PJFJ	53.909	1.377	26.144	8.533	0.288	1.435	1.057	1.446	3.261	0.145				975.950
84450	M16 PJFJ	52.311	1.344	24.016	10.123	0.328	1.585	0.840	1.223	3.415	0.102				952.870
84460	M17 PJFJ	53.281	1.448	23.387	9.656	0.097	1.890	0.799	1.498	3.137	0.064				952.570
84470	M18 PJFJ	53.825	1.407	22.077	9.468	0.138	1.674	0.776	1.590	2.753	0.096				938.040
84480	MU. PJFJ	49.461	1.334	23.636	13.556	0.139	2.406	0.898	0.960	3.161	0.120				956.630
84490	M21 PJFJ	50.901	1.170	19.286	6.954	0.059	1.878	0.891	1.009	2.724	0.080				850.840
84500	M22 PJFJ	54.517	1.297	23.099	9.527	0.064	2.107	0.963	0.923	3.327	0.108				959.320
84510	M23 PJFJ	55.334	1.320	23.452	9.138	0.054	1.890	0.950	1.048	3.262	0.054				965.020
84520	M26 PJFJ	51.342	1.236	25.589	10.471	0.120	2.026	0.797	1.171	4.295	0.094				971.410
84530	M28 PJFJ	50.904	1.329	26.680	9.282	0.193	2.020	0.794	1.257	4.406	0.107				969.720
84540	M29 PJFJ	53.450	1.319	44.495	9.948	0.072	1.899	0.846	1.190	3.625	0.087				969.310
84550	M31 PJFJ	51.892	1.388	27.425	7.977	0.051	1.819	0.734	1.183	5.322	0.072				978.630
84560	M32 PJFJ	50.717	1.278	25.803	10.310	0.367	1.863	0.938	0.867	4.469	0.158				967.700
84570	M33 PJFJ	53.024	1.278	24.094	10.385	0.192	1.898	0.776	0.667	4.132	0.132				965.680
84580	M34 PJFJ	54.671	1.366	25.112	8.768	0.076	1.808	0.860	1.222	3.567	0.104				975.740
84590	M35 PJFJ	54.046	1.387	24.133	9.958	0.099	1.867	0.866	1.139	3.100	0.087	Not determined			966.820
84600	M36 PJFJ	59.883	1.276	25.392	5.079	0.035	0.894	0.678	1.218	4.308	0.045				988.080
84610	M37 PJFJ	57.448	1.279	22.681	8.422	0.058	1.564	0.856	1.137	3.655	0.142				972.420
84620	M39 PJFJ	52.491	1.335	26.992	7.720	0.070	1.947	0.934	1.512	4.580	0.108				972.210
84630	M41 PJFJ	51.114	1.215	25.240	11.533	0.281	1.902	0.975	1.226	3.326	0.140				969.520
84640	M42 PJFJ	50.925	1.228	26.651	10.115	0.243	1.832	0.877	1.137	4.232	0.113				926.080
84650	M.00 PJFJ	54.840	1.231	20.790	11.272	0.488	2.140	1.025	0.912	3.144	0.238				960.800
MEAN OF	33 ROCKS	53.417	1.304	24.136	9.699	0.186	1.852	0.890	1.149	3.564	0.110				961.516

TRACE ELEMENTS (ppm)

PEL NO.	S	CR	NI	CU	ZN	RB	SR	Y	ZR	BA	LA	CE	PB	TH	GA
70020	59.	129.	72.	106.	51.	186.	141.	28.	150.	864.	81.	114.	2.	18.	19.
70030	32.	130.	71.	101.	57.	171.	147.	19.	139.	744.	15.	26.	13.	17.	23.
70040	87.	199.	52.	81.	104.	150.	149.	20.	273.	636.	14.	35.	41.	17.	16.
70050	349.	96.	52.	65.	117.	84.	58.	10.	161.	475.	13.	70.	4.	11.	13.
70060	166.	128.	82.	138.	143.	135.	118.	19.	174.	782.	32.	93.	30.	10.	17.
70070	242.	128.	26.	112.	139.	266.	63.	23.	257.	1012.	33.	68.	203	15.	20.
70080	264.	124.	39.	109.	101.	206.	168.	35.	328.	990.	62.	123.	44.	22.	23.
70090	198.	129.	42.	101.	99.	197.	191.	39.	194.	1027.	60.	137.	97.	16.	25.
84410	379.	133.	51.	108.	343.	206.	161.	23.	195.	1053.	55.	112.	208.	18.	19.
84420	49	132.	60.	118.	111.	203.	161.	31.	191.	922.	66.	136.	39.	17.	24.
84430	58.	127.	44.	106.	74.	221.	134.	27.	180.	1011.	64.	134.	23.	21.	24.
84440	60.	131.	49.	93.	106.	211.	171.	30.	196.	914.	71.	130.	276.	21.	26.
84450	60.	131.	67.	107.	86.	211.	158.	29.	168.	948.	53.	117.	24.	19.	23.
84460	33.	143.	40.	113.	90.	193.	172.	31.	232.	1016.	71.	154.	343.	16.	22.
84470	57.	133.	47.	100.	68.	178.	163.	32.	255.	773.	62.	125.	12.	14.	200.
84480	65.	135.	38.	89.	99.	159.	117.	23.	168.	899.	53.	110.	25.	16.	20.
84490	1315.	117.	43.	109.	81.	198.	172.	31.	339.	879.	55.	115.	34.	19.	20.
84500	1722.	120.	62.	117.	141.	180.	121.	26.	243.	900.	49.	98.	22.	17.	19.
84510	871.	125.	36.	132.	89.	175.	146.	26.	271.	901.	44.	84.	324.	16.	22.
84520	59.	132.	60.	79.	28.	232.	116.	14.	159.	1012.	16.	65.	2.	20.	24.
84530	76.	136.	49.	109.	28.	277.	153.	13.	175.	964.	21.	97.	18.	21.	21.
84540	65.	126.	36.	85.	33.	215.	158.	18.	232.	837.	23.	46.	21.	13.	21.
84550	286.	134.	28.	119.	29.	319.	104.	24.	236.	953.	38.	75.	5.	19.	27.
84560	275.	127.	49.	82.	72.	268.	115.	25.	172.	882.	63.	124.	8.	19.	24.
84570	143.	128.	57.	108.	68.	266.	90.	20.	209.	919.	52.	108.	29.	15.	20.
84580	18.	138.	45.	108.	108.	246.	145.	29.	253.	943.	54.	116.	70.	17.	22.
84590	160.	134.	33.	140.	81.	188.	144.	30.	243.	877.	50.	113.	75.	10.	23.
84600	868.	120.	17.	96.	25.	321.	142.	25.	299.	772.	41.	92.	0.	24.	26.
84610	573.	116.	21.	96.	23.	237.	100.	25.	287.	737.	18.	28.	5.	20.	18.
84620	35.	132.	64.	94.	74.	286.	150.	25.	234.	1060.	63.	137.	13.	20.	23.
84630	114.	127.	48.	84.	73.	184.	148.	28.	197.	811.	45.	91.	3.	5.	21.
84640	293.	126.	47.	111.	81.	238.	117.	21.	205.	961.	26.	55.	2.	10.	20.
84650	108.	120.	68.	114.	72.	165.	97.	25.	220.	904.	32.	107.	5.	14.	16.
MEAN	277.	130.	48.	104.	88.	211.	136.	25.	219.	890.	45.	98.	61.	17.	27.

PLATE 1

LANTHWAITE

RANNERDALE

GRASMOOR

▲ 2791'

BUTTERMERE

CRUMMOCK WATER

2 anomalous cleavages but both S2

bedding elongate E-W plunge N.: F1.

"Comb" (61)

melt-water channel

terrace

intense N-S jnts.

marked Fe staining

W. fall.

"bomb" (705)

Fe + tourmaline mineralization along faults

N-S jnts.

Complex of F_2 climbing to S. & F_1 plunge inc to E

F_1 plunges ine into F_2 steep zone

N-S fault gully

Qtz veins ringing breccia pipe

marked N-S jnts

Tourmaline veins

F_1 steep zone

F_1 flat F_2 recumb. facing N

"Bomb" (28)

KEY

BEDDING

" CONTORTED

" HORIZONTAL

" VERTICAL

FOLD AXES

F1, F2 — TICK INDICATES DIP OF AXIAL PLANE

F3

F4

LINEATIONS

L1, L2, L3, L4

CLEAVAGES

S2, S3, S4

S2 vertical

AXIAL TRACES

F2

F1

STYLE OF FOLD

YOUNGING DIRECTION

F FAULT

Th THRUST

T TOURMALINE

16 17 18

23

22

21

520

N

MILLER PLACE

HOLLINS

LANTHWAITE

2791'
GRASMOOR

erratic blocks of pink granophyre - Ennerdale?

low bedding dips to N. & S: open F_2 warps pre-dates low angle cleavage.

gorge section starts

boulder clay on a terrace bank

DRIFT-FILLED LAKE BASIN

laminated sands, silts & clay

NO EXPOSURE

anomalous cleavage in middle limb of monocline refolded?

open N.W. F_4 folds

? anomalous low angle cleavage

sink

sinks

white stones

massive pyramids

possibly S_2

vertical pyramids

steep F_3 axes

v. intense cleavage + highly appressed folds at junction

to N of thrust folds climb to S.

to S of thrust folds descend to north.

Th Th

marked Fe staining

w. fall

2 anomalous but both S_2 cleavages

boudins elongate E-W plunge N. ~ F_4

"bomb" (61)

"bomb" (703)

melt water channel

ice contact terrace

intense N-S joints

Fe + tourmaline mineralization along faults

N-S jnts.

quartz veins ringing breccia pipe

marked N-S jnts

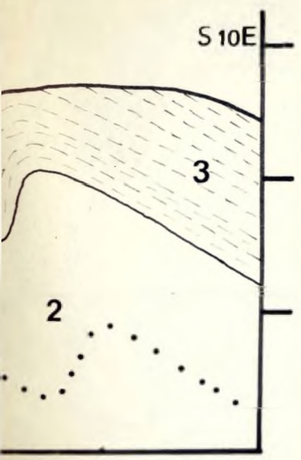
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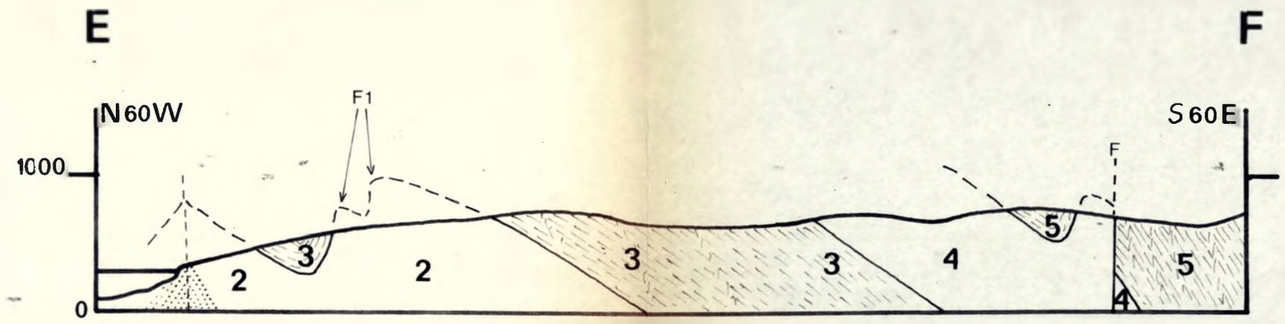
T

KEY


BUTTERMERE SLATES
WHITELESS PIKE FLAGS
MOSSER-KIRKSTILE SLATES
LOWESWATER FLAGS
HOPE BECK SLATES
BASIC IGNEOUS INTRUSIONS







1 HOPE BECK SLATES

2 LOWESWATER FLAGS

3 MOSSER KIRKSTILE SLATES

4 WHITELESS PIKE FLAGS

5 BUTTERMERE SLATES

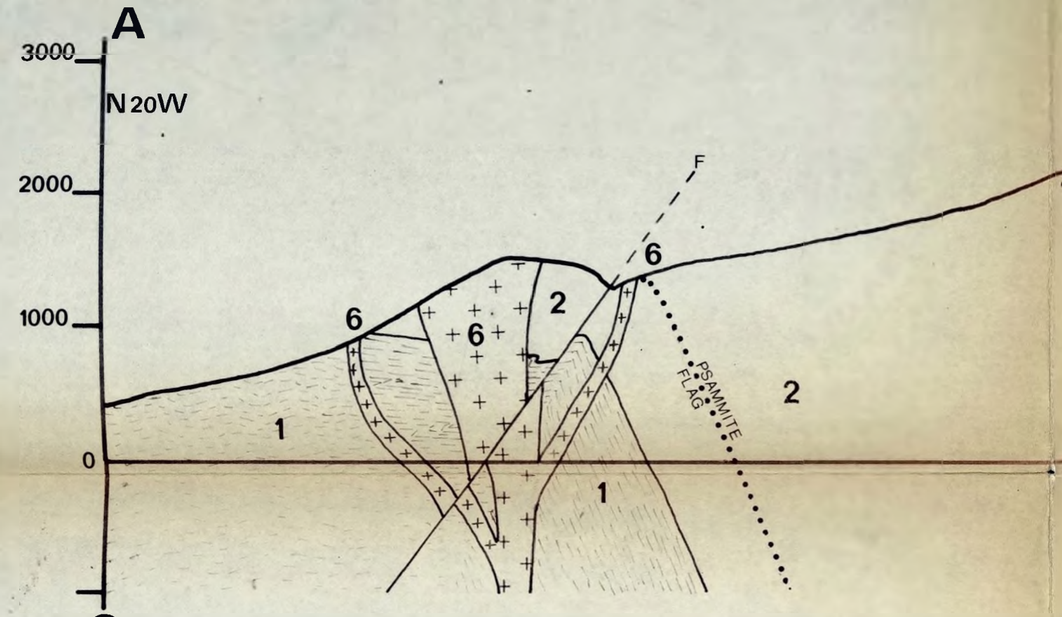
6 BASIC INTRUSIONS

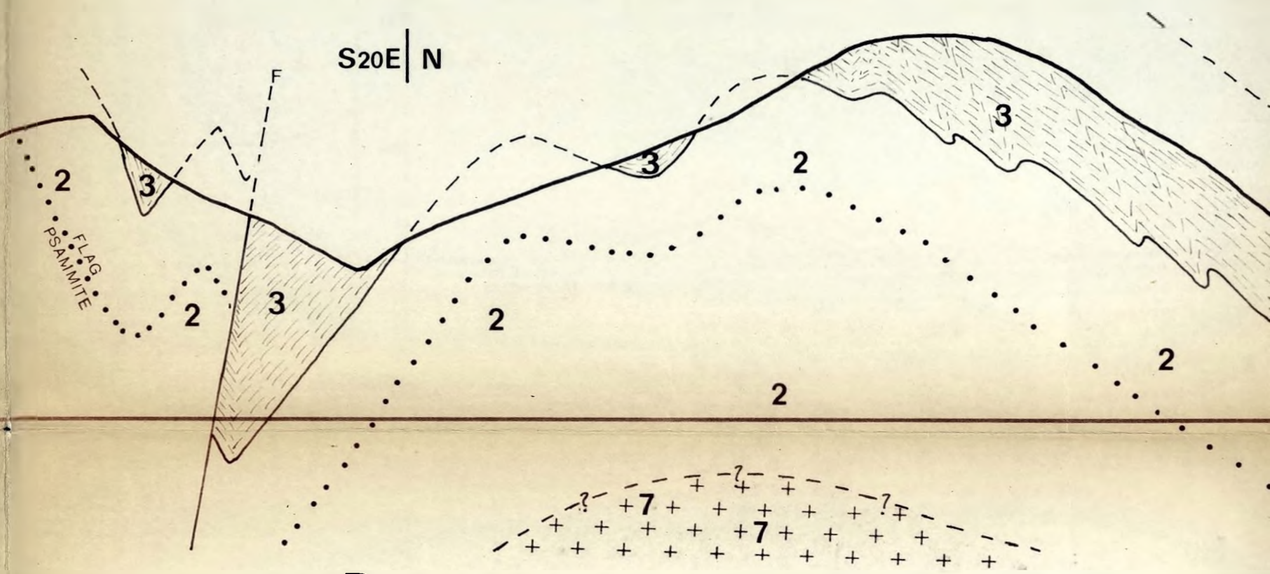
7 ?GRASMOOR GRANITE ?

VERTICAL EXAG = 0·88

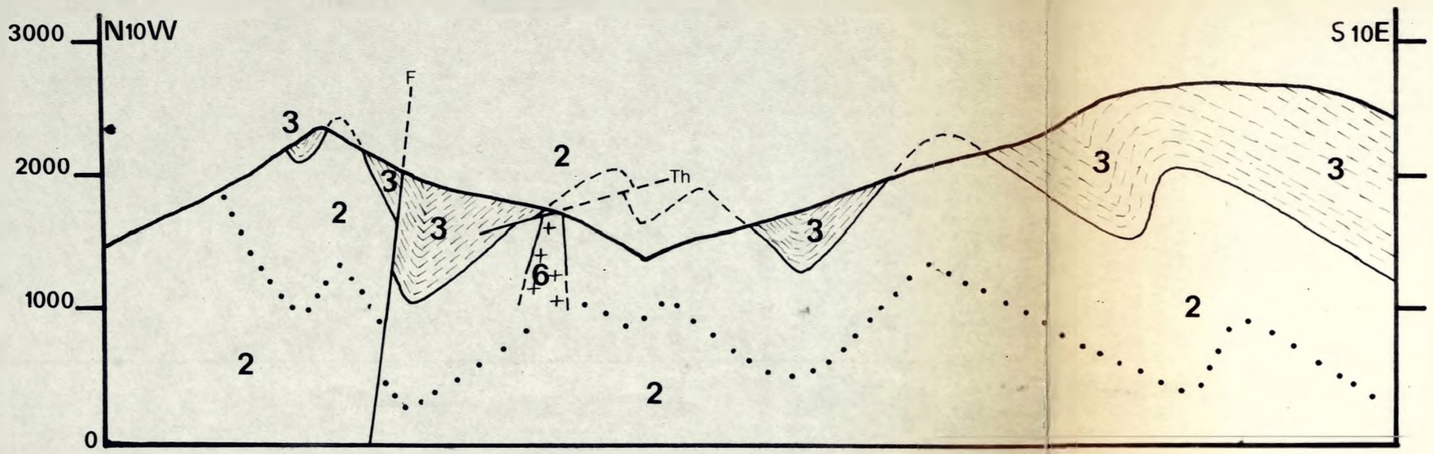
ALL FOLDS F2 UNLESS INDICATED OTHERWISE

PLATE 2



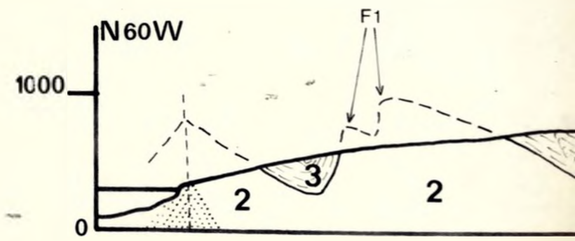


C



D

E



VERTICAL EXAG = 0·88

ALL FOLDS F2 UNL


PLATE 3
HOPE BECK
GRASMOOR
F1 & F2 AXES
F1
F2
24
18
23
22
21
20
19 16
17
19

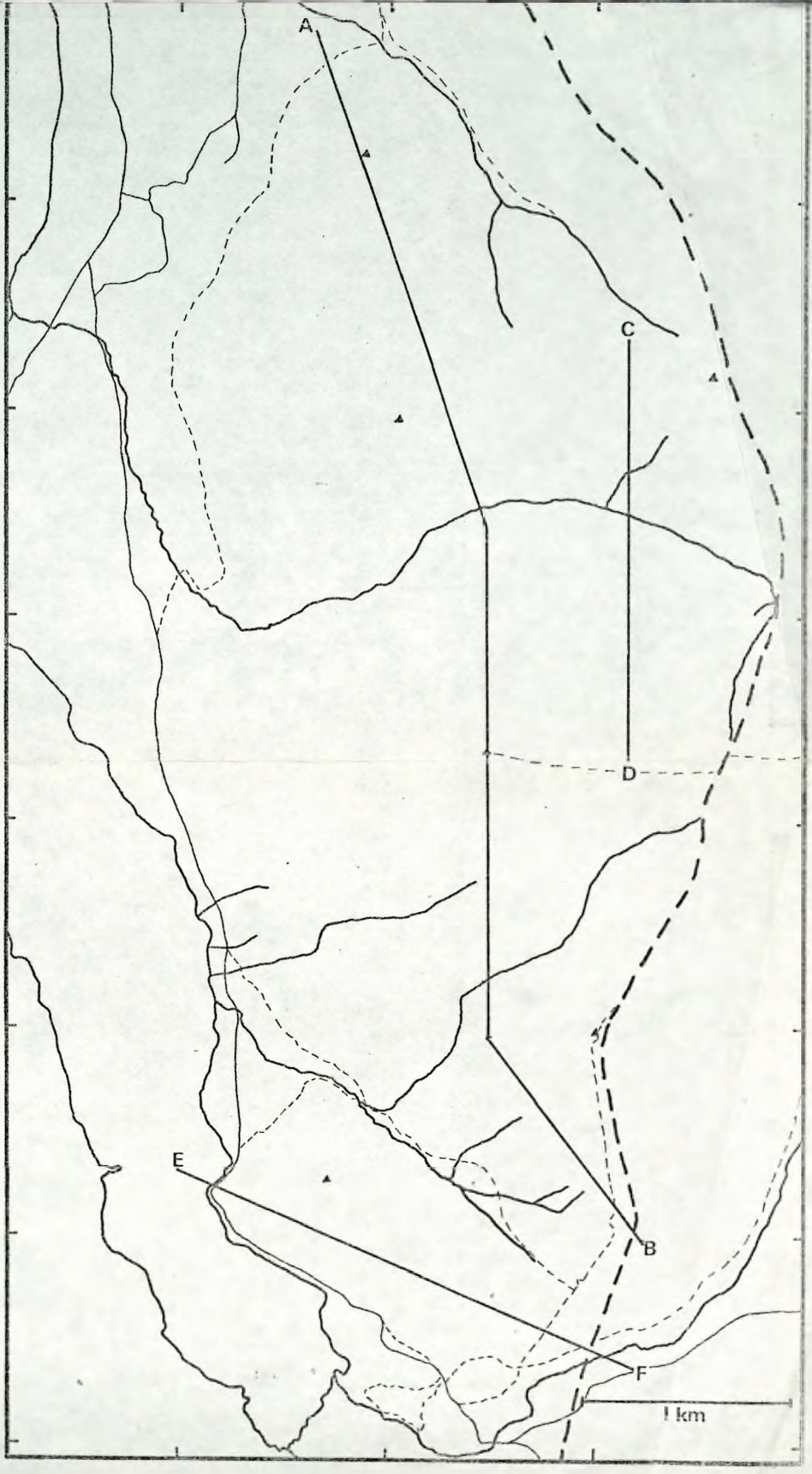
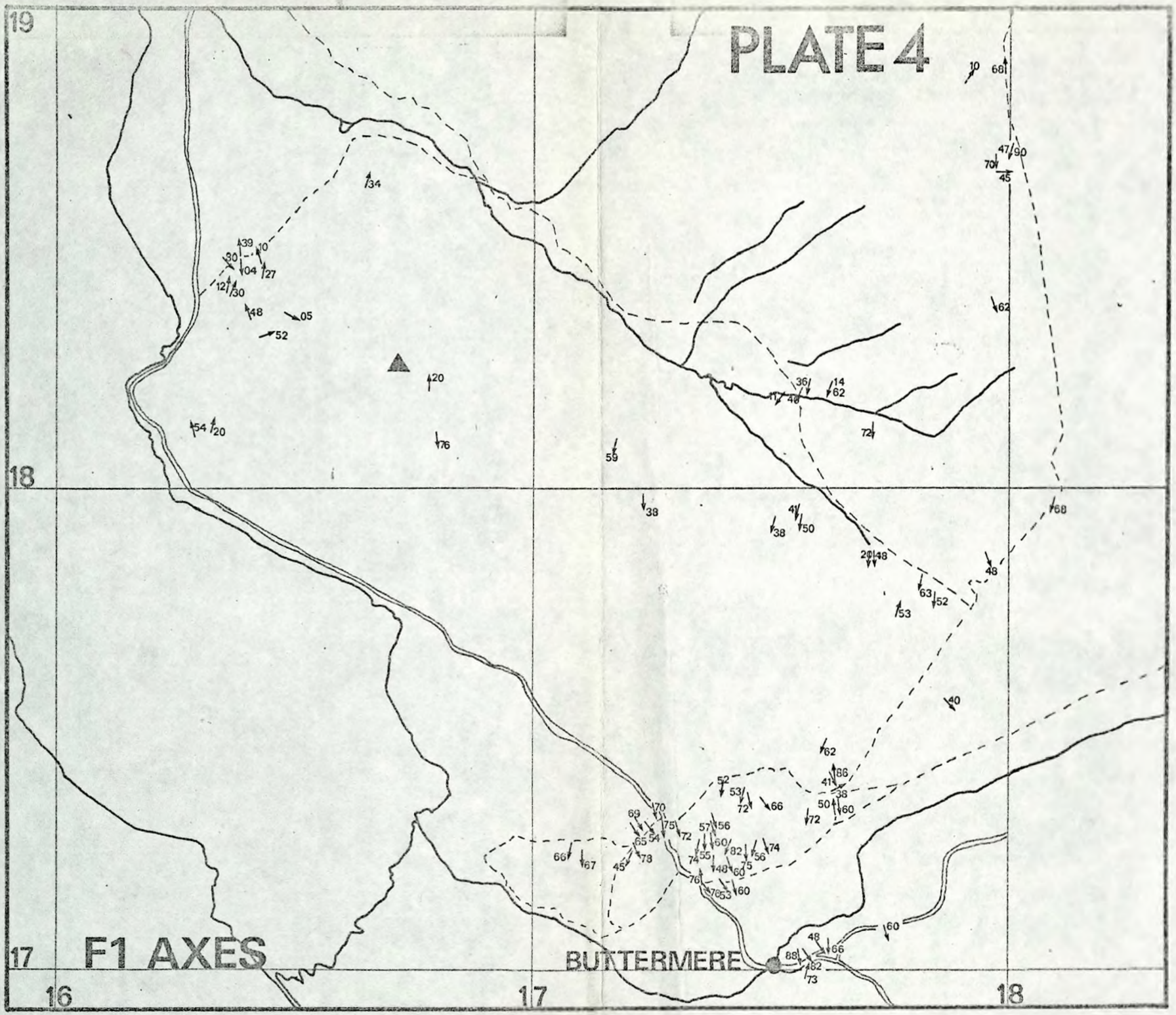
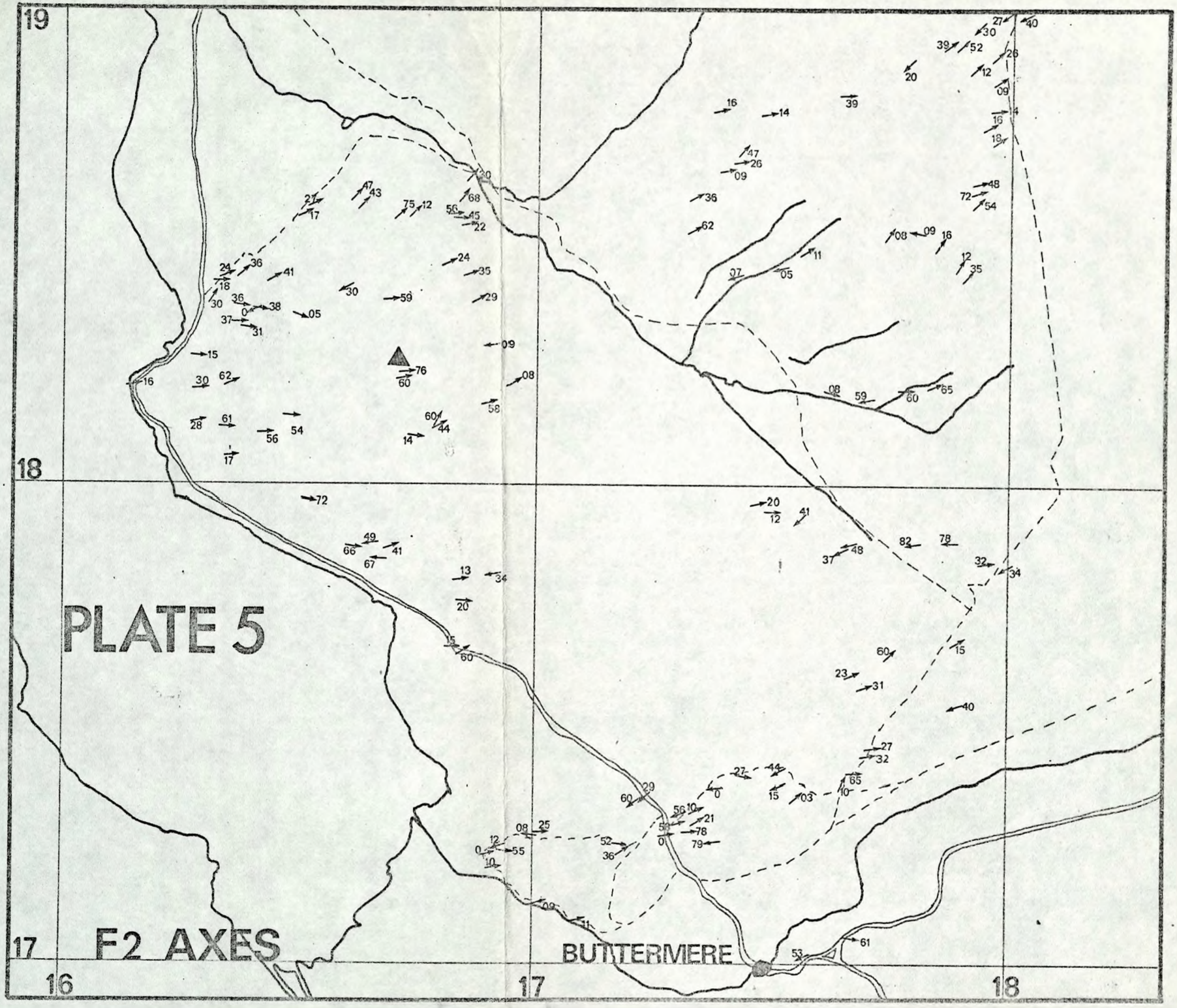


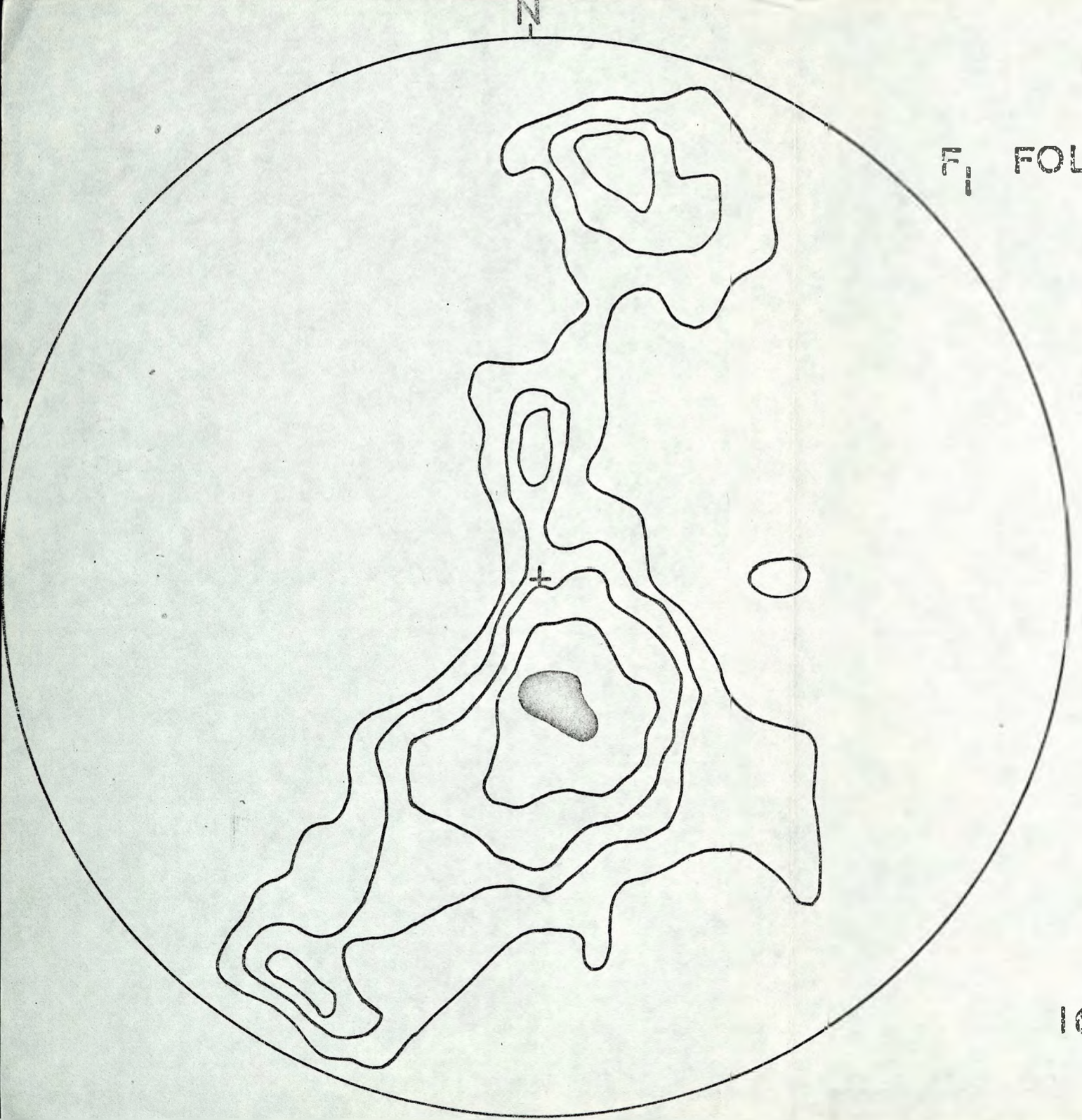
PLATE 3

SHEWING LOCATION

OF SECTION LINES


PLATE 4
F1 AXES
BUTTERMERE
19
18
17
16
17
18


19
18
17
16
17
18
PLATE 5
F2 AXES
BUTTERMERE

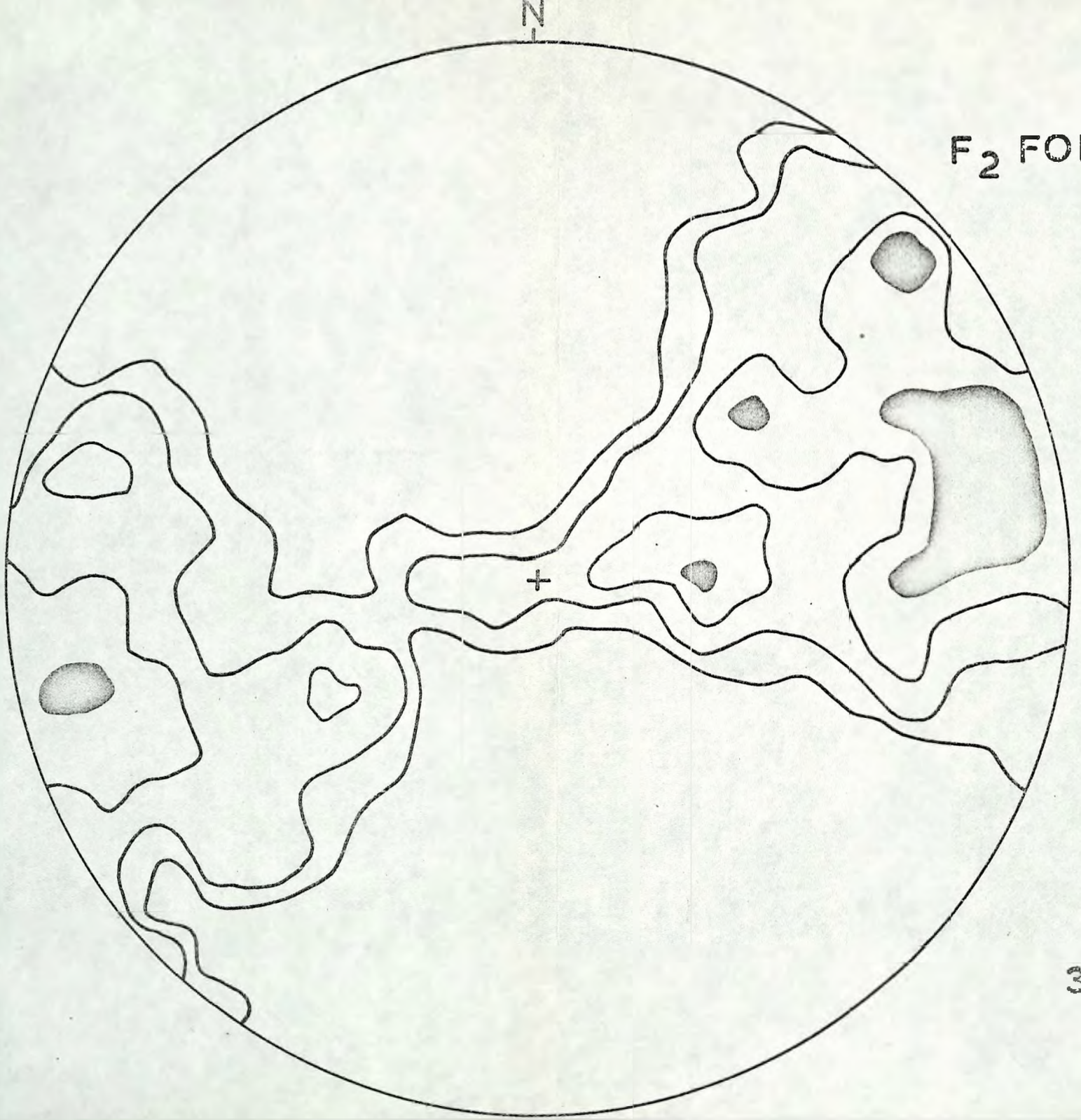


F_1 FOLD AXIS

Contours at

1·2% }
2·4% } per
3·6% } 1% area
6·0% }
10·3% }

165 readings



F_2 FOLD AXES

Contours at
- .64%
- 1.28%
- 2.56%
- 3.84%

per 1% area

312 readings

(*Reprinted from Nature Physical Science, Vol.* 234, *No.* 46, *p.* 59, *November* 15, 1971)

Relationship between the Skiddaw Slates and the Borrowdale Volcanics

THE communication by Helm and Roberts[1] conveys a misleading picture of the geological situation at Scarbrow Wood and Hollows Farm and, as this is vitally important to interpretation of the junction between the Skiddaw Slates and Borrowdale Volcanics, I should like to place their observations in perspective. Two years ago I mapped this small area in detail and have revisited it several times since then; it is clear to me that there is no evidence for an unconformity of the scale they suggest anywhere in the area.

As they state, the existence of local unconformities is not now doubted (Crookley Beck[2], Newlands Valley[3]). Furthermore, the evidence from Brampton and the West Lake District cited by Wadge and Burgess (both in ref. 1) suggests the presence of a regional break between the Skiddaw Slates and Borrowdale Volcanics, but this break will manifest itself in several ways depending on the local geographical conditions and the structural response to deformation of the rocks in the area concerned.

In each of the three examples from Scarbrow–Hollows figured by Helm and Roberts, their major unconformity is made to pass through areas of no exposure. Their backing-up evidence, though correct, is insufficient and misleadingly represented, and their conclusions would be impossible to sustain in the field. The critical evidence necessary to decide the nature of these junctions is lacking.

For example, exposure (1) in Scarbrow Wood is situated some thirty feet below, and sixty yards distant from the junction, which in Skiddaw Slates is ample room for disharmony of considerable magnitude to develop, sufficient to account for a single, poorly-exposed, inverted fold-pair. There is no structural evidence for a major unconformity, for both the Skiddaw Slates and the Borrowdale Volcanics show S_1 and S_2 cleavages (end-Silurian), and at exposure 3 (wrongly marked on their map) they admit that the S_1 cleavage passes from slates to volcanics without deviation. They explain this by representing the slates at this locality not as Skiddaw Slates but as the lowest beds of the Borrowdale Volcanics, hence leaving room below these slates for their unconformity. They present no evidence for this assertion.

The way for a major unconformity to be proved in this area is to find some break in the slates at or below the junction, either by structural or possibly geochemical means, but until this is

done the possibility of the unconformity should not be represented as fact, as Helm and Roberts claim it to be.

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Received October 11, 1971.

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Printed in Great Britain by Flarepath Printers Limited, Watling Street, Colney Street, St. Albans, Herts.

(Reprinted from Nature Physical Science, Vol. 245, No. 147, pp. 120–122, October 22, 1973)

Plate Tectonic Reconstruction of the Southern Caledonides of Great Britain

GUNN[1] has suggested an elegant solution to the problem of accommodating two subduction zones, one dipping north under the Highlands[2], the other dipping south under the Lake District[3] between the continental crust thought to underlie the Southern Uplands[4] and that of the Highlands. He considers the Midland Valley of Scotland "to be the oceanic remnant of the proto-Atlantic Ocean", and suggests that the Highland Boundary Fault and the Southern Uplands Fault "mark the positions of diverging Benioff zones". The ophiolite complex at Ballantrae and the ophiolitic rocks of the Highland Border Series, both of which may be traced aeromagnetically across into North Ireland, are considered to be fragments of ocean floor emplaced by thrusting and welding along the zones where oceanic crust was overridden by the converging continental plates.

Here I extend this hypothesis to account for the geological features of the Southern Caledonides of Britain (Wales, Lake District and the Southern Uplands). As mentioned above there is strong evidence[4] to indicate that the Southern Uplands are underlain by continental crust, and so it seems reasonable to postulate that a single continuous continental plate underlay the whole of the Southern Caledonides; certainly there is no evidence for any crustal suture line between the postulated plate margin at Ballantrae and the Welsh Borders. Two major depositional basins would therefore be located on this plate, the Lake District-Southern Uplands basin and the Welsh basin, which were separated by the often-proposed Irish Sea geanticline. Evidence for the northwards continuation of the Manx-Skiddaw sequence beneath the post-Llanvirn rocks of the Southern Uplands is possibly found in the Coldingham Beds of Berwickshire, especially as they show very similar refolding patterns to those of the Skiddaw Slates[5–7]. It is necessary to postulate the presence of the Irish Sea geanticline as an actively eroding upland area to provide the detritus for the Skiddaw-Manx and Welsh sediments, and palaeoslope and sediment transport directions from the Skiddaw Slates[7,8] do indicate the presence of a shoreline to the south and supply of sediment from that direction.

Recent radiometric dating of the Ingletonian rocks has yielded Cambro-Ordovician ages[9], which indicates their likely deposition in the Manx-Skiddaw trough, as the possible time-equivalents of the lowest Manx Slates.

I suggest that during Arenig times (Fig. 1), the south dipping oceanic plate was gently angled, giving the andesite volcanism of Cader Idris, Rhobell Fawr and Pembroke, whilst beneath the Manx-Skiddaw trough the descending plate was insufficiently deep for partial melting and volcanism to occur. Following Rast[10], the Bala Fault is shown as the feeder to both North and South Wales volcanics. Eruptions in these areas continued into the Llanvirn and were characterised by either spilites and andesites, or rhyolites and keratophyres, the latter representing the products of crustal fusion and differentiation consequent upon the emplacement within the crust of large bodies of hot, mantle-derived magma.

By Llandeilo times (Fig. 2) the inclination of the Benioff zone must have increased to permit the formation of andesite melts beneath the Manx-Skiddaw trough; on rising to the surface these andesites built up an island arc, now represented by the Eycott and Borrowdale Volcanic Groups. Although dominantly composed of andesite, the more extreme differentiation products occur towards the top of the Borrowdale sequence. The Borrowdale Volcanics were wholly subaerial in eruption[11]; this suggests that the compressive

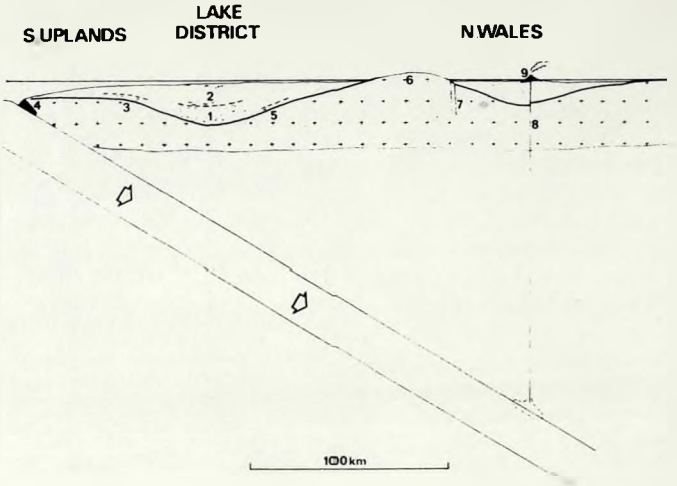


Fig. 1 Situation during Arenig times. 1, Manx Slates; 2, Skiddaw Slates; 3, Coldingham Beds, Berwick; 4, Ballantrae Ophiolite Complex; 5, Ingletonian; 6, Irish Sea Geanticline; 7, Menai Straits Fault Complex; 8, Bala Fault—acting as feeder for: 9, Cader Idris, Rhobell Fawr and Pembroke Volcanics.

forces responsible for the closing of the proto-Ocean were sufficient to raise the top of the Skiddaw Slates above sea level, and this is reflected in the widely reported sub-Volcanic unconformity[12], and pre-Volcanic, albeit north-south orientated, deformation phases[6,13,11].

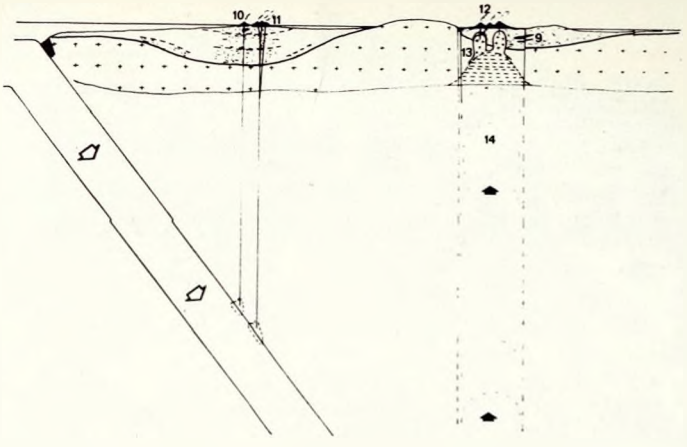


Fig. 2 Volcanism during Llandeilo-Caradoc times. 10, Eycott Volcanic Group; 11, Borrowdale Volcanic Group; 12, Snowdon and Conway acid volcanics; 13, subjacent granitic plutons, such as Tan-y-Grisiau; 14, proposed 'heat plume', fault controlled, rising from Benioff zone at depth, causing melting at base of crust.

The inclinations of the downgoing slabs required for this model are compatible with present day observations. The diagrams show the situation in which no crustal shortening is assumed; taking 150 km as a reasonable depth for andesite genesis, this gives lithospheric dips of 30° and 52° for North Wales and the Lake District, respectively. If a 50% crustal shortening is admitted, the respective dips would be decreased to 15° and 32°. These values compare favourably with present day observations presented by Oliver[19], who records

lithospheric dips ranging from 10° to 20° beneath Peru and North Chile, to 60° for the majority of the Western Pacific Benioff zones, such as Kermadoc, New Hebrides and Mindanao.

Cessation of the Volcanic episode by the end of the Llandeilo was accompanied by deformation and subsidence of the island arc.[9] This subsidence and marine incursion occurred earlier in the east than in the west, for the Volcanics at Cross Fell are overlain unconformably by the corona Beds which have no equivalents in the main outcrop, the Volcanics there being overlain by later members of the Coniston Limestone Group. Progressive deepening of this shelf sea would then lead to the initiation of the miogeosynclinal trough in which the thick succession of Silurian greywackes and slates was deposited.

Meanwhile, during Caradocian times, volcanicity in North Wales was renewed on a huge scale, starting in Conway and the Capel Curig area east of Snowdon. Snowdon became a major centre of volcanism, with the eruption of vast thicknesses of rhyolite and ignimbrite, and the emplacement of subjacent granitic plutons such as Tan-y-Grisiau. There is evidence that the whole of Snowdonia is underlain by a large sub-Volcanic granitic body[10], and geochemical analyses from the ignimbrites (Hendry, personal communication) suggest their origin by partial melting of sediments at no great depth below the surface (<10 km). The mechanism by which this crustal melting occurred is not known, but it has been suggested[10,15] that the crust below Snowdonia was invaded by a large, mantle-derived basic pluton, and that the heat from this basic body melted the continental crust and sediments above, thus providing the acid magma for eruption and epizonal injection. Whether the basic material was derived directly from the Benioff zone at depth, or whether only heat was transferred upwards to melt mantle material nearer the base of the crust, is not known, but the latter hypothesis is illustrated in Fig. 2, the rising 'heat plume' being controlled by the deep faults of Menai and Bala.

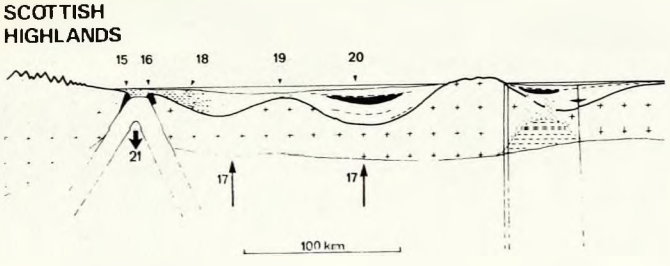


Fig. 3 Final stages of ocean closure. Silurian times. 15, Highland Border Series and Boundary Fault; 16, Southern Uplands Fault and Ballantrae Ophiolites; 17, End-Silurian intrusion of granites in the Southern Uplands and Lakes due to rising 'heat plume'; 18, prograding delta fan depositing the Girvan greywacke sequence; 19, thin Moffat sequence; axial zone progressively swamped by northerly derived greywacke deposits; 20, Silurian rocks of Lake District and Cross Fell; 21, downpinching ocean crust.

By the end of Ordovician and beginning of Silurian times the situation as envisaged in Fig. 3 was reached, with the closing rafts of continental crust beginning to pinch down the remnant ocean crust of the Midland Valley. By lowest Llandovery times (acuminatus zone[20]) coarse grained terrigenous sediment from the rising Highlands to the north was beginning to appear in the prograding greywacke fans of the Girvan succession, whilst the central Moffat region at this time was the site of very slow pelagic deposition. This axial zone was progressively encroached upon throughout the Silurian, as the whole Southern Uplands basin was rapidly filled by greywacke turbidites of northerly derivation[16].

In the Lake District the Coniston Limestone Group was succeeded conformably by Silurian rocks, again of greywacke lithology, and derived from the south, where the Irish Sea Geanticline remained a positive feature. The recent dating of part of the Kirkcudbright succession of greywackes and shales as Upper Llandovery by Walton[17] means that there is only 30 miles of unexposed ground between the Lower Silurian of the Cross Fell Inlier and the most southerly outcrop of the Southern Uplands succession, so the postulated continuity of the Skiddaw-Southern Uplands depositional basin at this time is not unreasonable.

The final episode in the evolution of the Southern Caledonides came at the end of the Silurian with the emplacement of high level granitic plutons into the Skiddaw and Southern Upland successions. As in North Wales, where the Arenig andesite vulcanism was followed after roughly 50 m.y. by granitic intrusion and ignimbrite eruption (Caradoc age), so in the Lakes and the Southern Uplands. The andesite vulcanism (Llandeilo) is succeeded, again after 50 m.y., by such granites as Criffel, Dalbeattie, Skiddaw, and Eskdale, and so on, of end-Silurian age[18]. Despite the absence of the contemporaneous ignimbritic eruptions in the Lakes or the Southern Uplands, the parallels with North Wales are still so similar as to suggest the existence of a magmatic cycle of roughly 50 m.y. duration, and the operation beneath the Lakes and the Southern Uplands of a similar mechanism to that invoked for North Wales, namely a rising 'heat plume' from the downgoing slab which induced partial melting at the base of, or within, the continental crust.

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Received July 20; revised September 4, 1973.

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Printed in Great Britain by Henry Ling Ltd., at the Dorset Press, Dorchester, Dorset

The junction between the Skiddaw Slates and Borrowdale Volcanics in Newlands Beck, Cumberland

PETER J. F. JEANS

(Plate 1)

Summary. The junction in Newlands Beck between the steeply dipping Skiddaw Slates striking N–S, and Borrowdale Volcanic rocks dipping gently to the SE is interpreted as an angular unconformity, and indicates the existence of pre-Borrowdale Volcanic folding. There is no cleavage associated with this folding, however, the ENE–WSW Caledonoid cleavage being the only one present, and this is common to both Skiddaw Slates and Borrowdale Volcanics.

1. Introduction

There has been recent resurgence of interest in the relationship of the Skiddaw Slates to the Borrowdale Volcanics, kindled by Simpson's (1967) paper in which he described three phases of deformation in the Skiddaw Slates, two of which (F1 and F2 with associated S1 and S2) were claimed to pre-date the formation of the Borrowdale Volcanic rocks. This interpretation has been challenged by Soper (1970), Roberts (1971) and Soper & Roberts (1971) but supported, with modification, by Helm (1970) and Helm & Siddans (1971).

The two first mentioned claim that Simpson's S1 passes upwards into the Borrowdale Volcanics and is therefore end-Silurian, whilst the latter, referring to Black Combe, believed there were important phases of northerly (north-westerly) folding and cleavage prior to the volcanic episode, although they regard the north-easterly trends (parallel to Simpson's F1) as post-volcanic. Roberts suggested that these northerly folds are pre-Caradoc and to be equated with the pre-Caradoc folding of the Borrowdale Volcanics, and thus it will be seen that there is as yet no agreement on the nature of the junction. This paper describes one locality where that junction appears to be displayed, and is presented as evidence to assist in the ultimate interpretation.

The Newlands Valley lies some 5 miles SW of Keswick between Borrowdale and Buttermere (Fig. 1) and the junction is exposed (at G.R. 3230. 5157) about 400 yards above the prominent waterfall in the upper reaches of the beck (Fig. 1 and plate 1).

2. Bedding

The Skiddaw Slates exposed in the beck are dark, blue-grey, finely-banded' slates and mudstones of the type called Mosser Slates by Eastwood *et al.* (1968), though in Simpson's (1967) nomenclature they are designated Kirkstile Slates. They strike consistently N–S over the whole length of the section and for at least a further 800 yards downstream. Bedding dips vary from 50° to 90°, predominantly to the east and can be clearly seen on the waterworn surfaces in the beck. At one outcrop (locality A in Fig. 1) there is an upright minor fold

Geol. Mag. 109 (1), 1972, pp. 25–28. Printed in Great Britain.

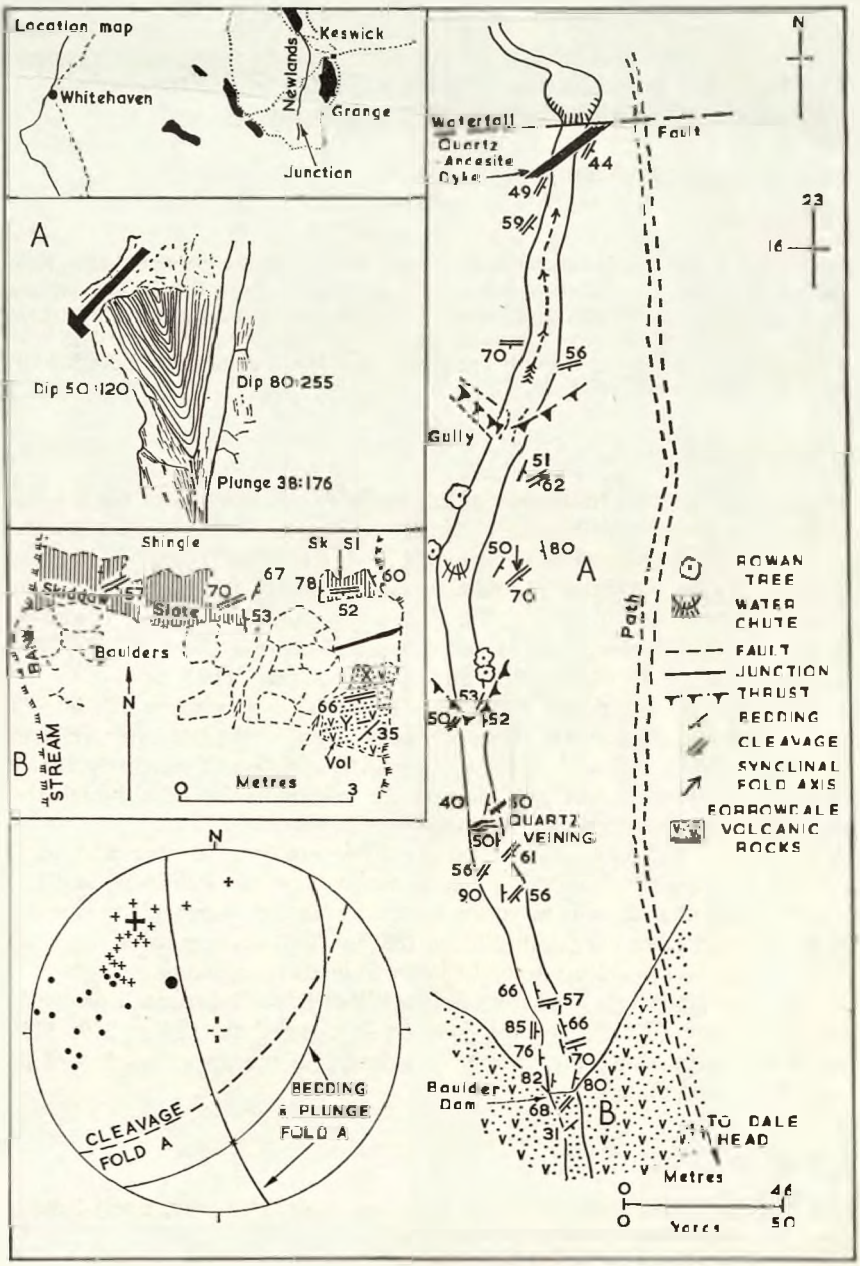


Figure 1. Structural map of the upper part of Newlands Beck. Insets (i) Location map. (ii) Section of the fold at A. (iii) Plan of the junction at B. X—Red silty slate, intensely sheared. No bedding visible. Y—Massive tuff. (iv) Stereogram (equal area) of poles to bedding (Skiddaw Slates small dots, Borrowdale Volcanics large dot), and poles to cleavage (Skiddaw Slates small crosses, Borrowdale Volcanics large cross).

plunging fairly steeply to the south and other folds of this type may be inferred from the opposition of beddings dips measured in the beck, but no other hinges have been seen.

The Borrowdale Volcanic rocks above the junction dip gently away to the SE quarter, and these dips are consistent over the surrounding area of Eel Crags and High Spy (Plate 1).

Between the Borrowdale Volcanic rocks and the Skiddaw Slates, at the junction, there is unfortunately 1·3 m of non-exposure, and the actual contact cannot be seen. However, the absence of any indications of faulting or thrusting in the immediate vicinity considered with the wide extent of the N–S bedding structures below, and the ENE–WSE bedding structures above the junction, suggests that one may reasonably infer the presence of an angular unconformity beneath the stream debris.

3. Cleavage

There is only one cleavage present throughout the whole section and this is the ENE–WSW Caledonoid cleavage (F1 of Simpson; D4 of Helm). It is present in both the slates and volcanics and there is no change in cleavage orientation in passing from one group to the other.

4. Conclusions

It is believed that an important angular unconformity separates the Skiddaw Slates from the Borrowdale Volcanics in the Newlands Valley. There is tight N–S folding in the slates which is not to be seen in the volcanics, and it has been shown that this northerly folding has no related cleavage; and that the only cleavage affecting the slates is the same ENE cleavage which affects the volcanics. It is concluded that the Skiddaw Slates of Newlands have been subjected to two periods of deformation, the first of pre-Borrowdale Volcanic age and the second, a much more severe episode, of post-Borrowdale Volcanic and presumed end-Silurian age. It is possible that the first episode is to be correlated with D1 of Helm (1970) and the early northerly folds of Roberts (1971), although the latter has made out a good case for the folds he describes being post-volcanic but pre-Caradoc, to be correlated with the pre-Caradoc movements which affect the volcanics. The second episode is clearly that which Simpson describes as S1 (F1) (equals D4 of Helm), and the observations recorded in this paper are thus clearly in support of Soper's contention (1970) that Simpson's F1 and F2 are not pre-Borrowdale Volcanic but are end-Silurian.

Acknowledgement. Dr F. Moseley helped at all stages in the preparation of this paper, and Mr. D. E. Roberts took the photographs of Plate 1.

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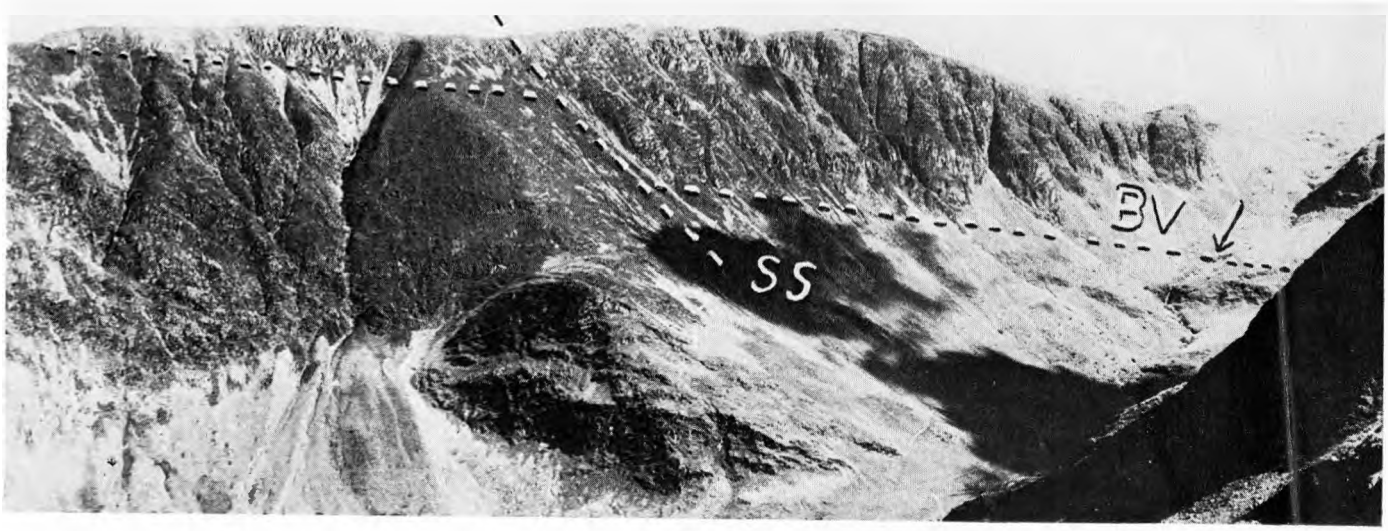
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EXPLANATION OF PLATE 1

View to the south-east of the upper Newlands Valley, Eel Crags and High Spy. The arrow indicates the junction described. The uniform gentle dip of the Borrowdale Volcanic rocks (BV) can be seen. The low angle planes visible in the Skiddaw Slates of Castlenook are joints and shear planes.

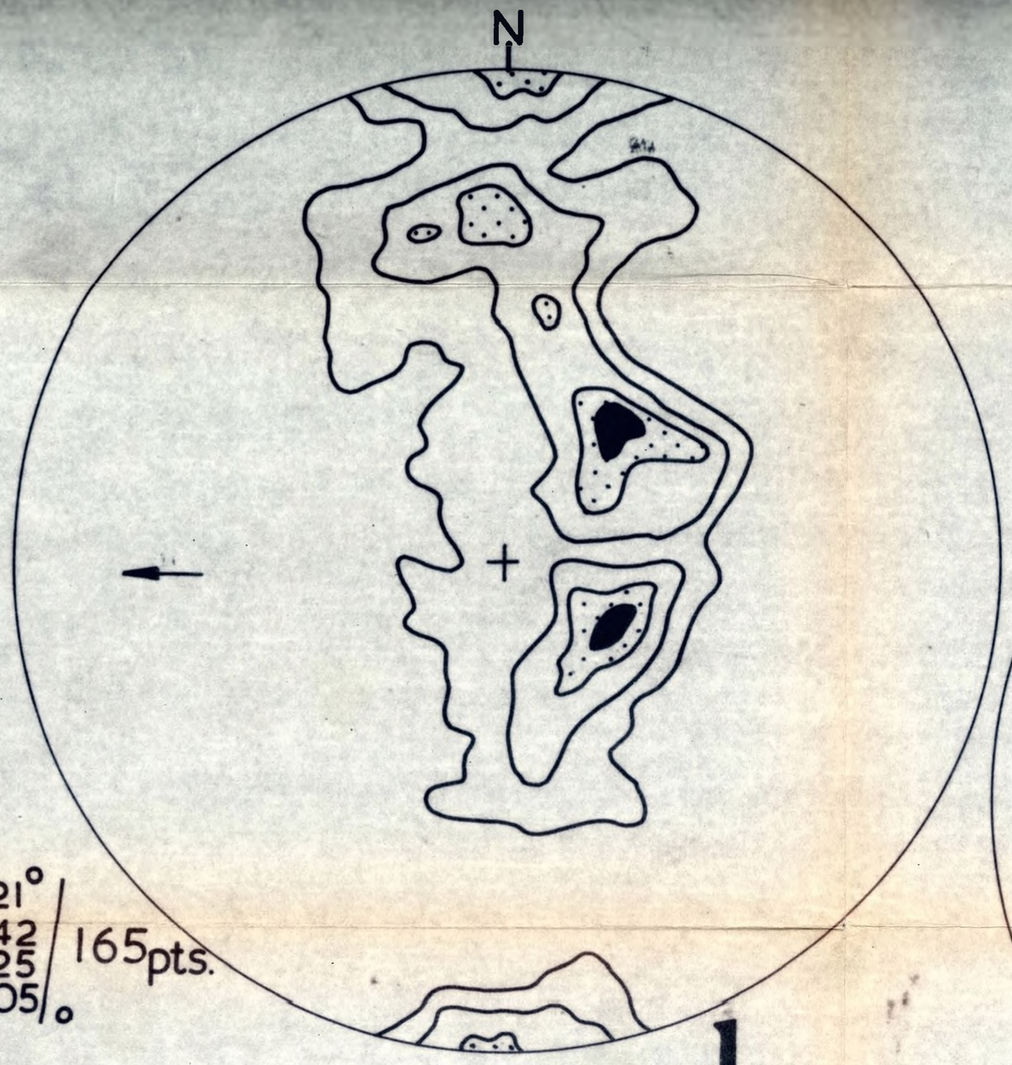
Geog. Mag. Vol. 109, 1972, Jeans, Plate I, facing p. 28.

High Spy

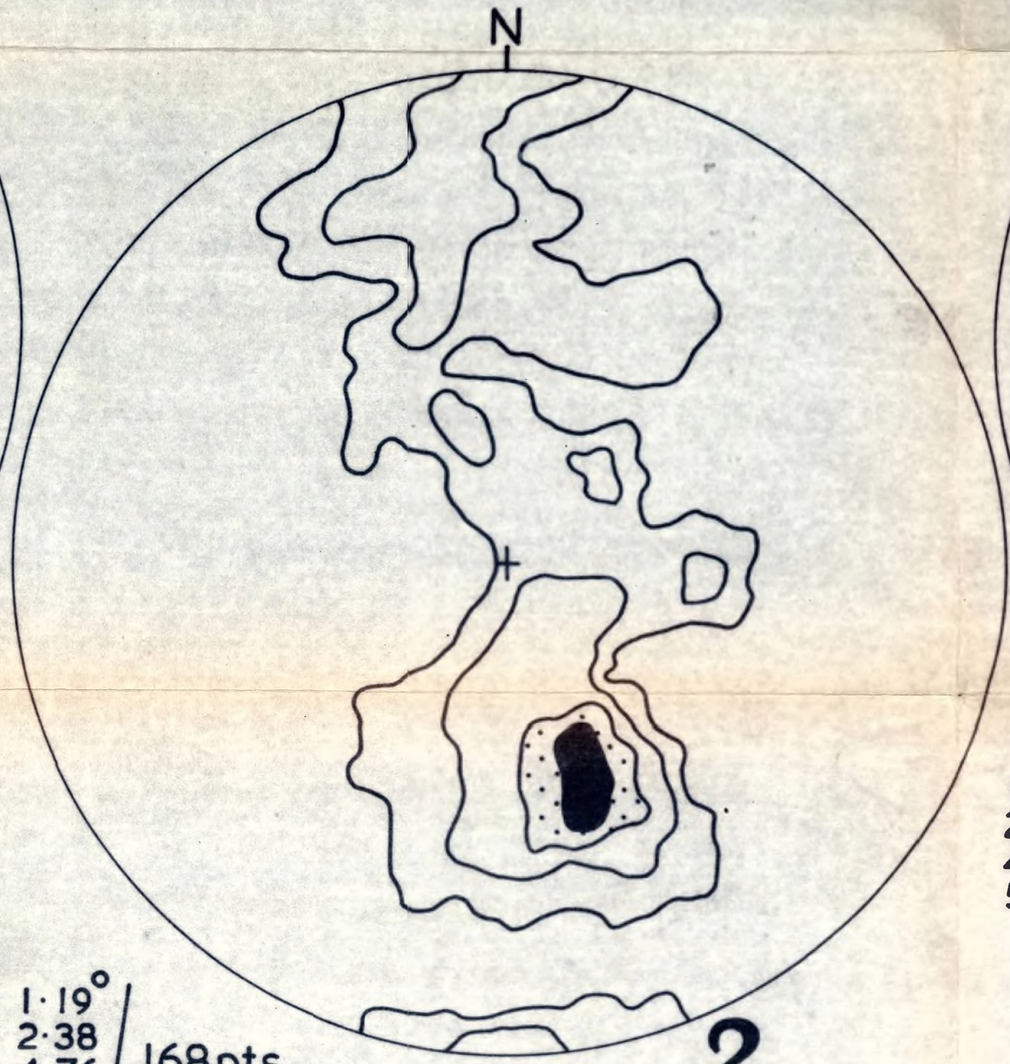


Castle Nook

Upper Newlands Valley, Cumberland.

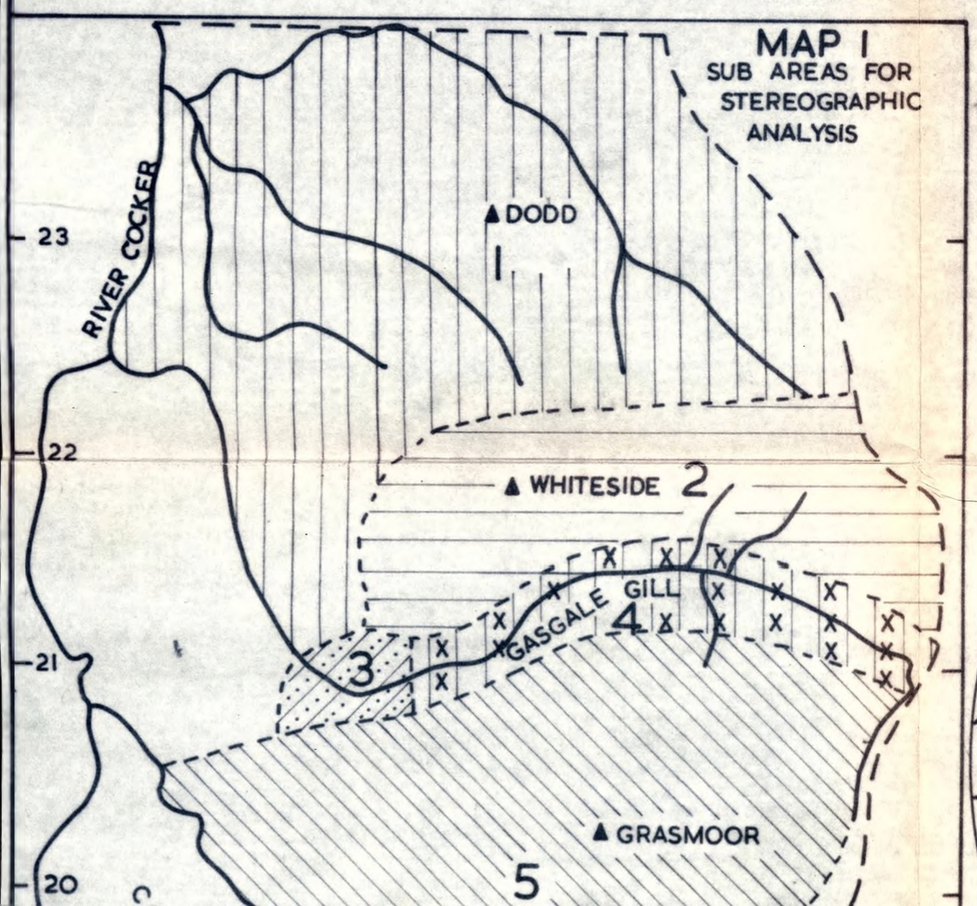

N
1·21°
2·42
4·25
6·05
165pts.

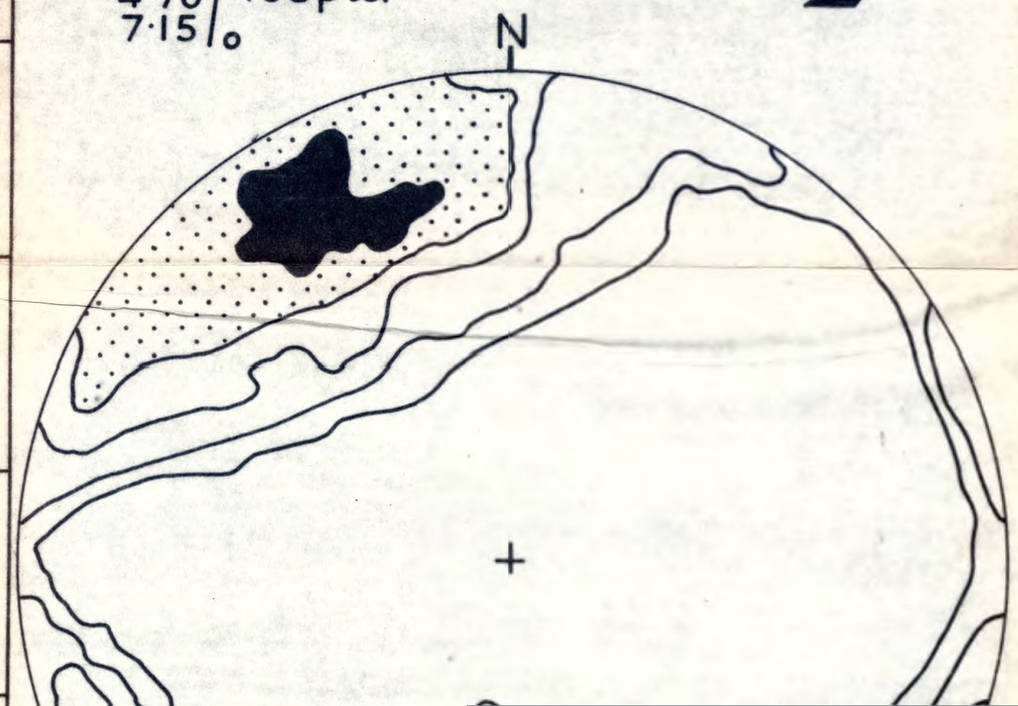
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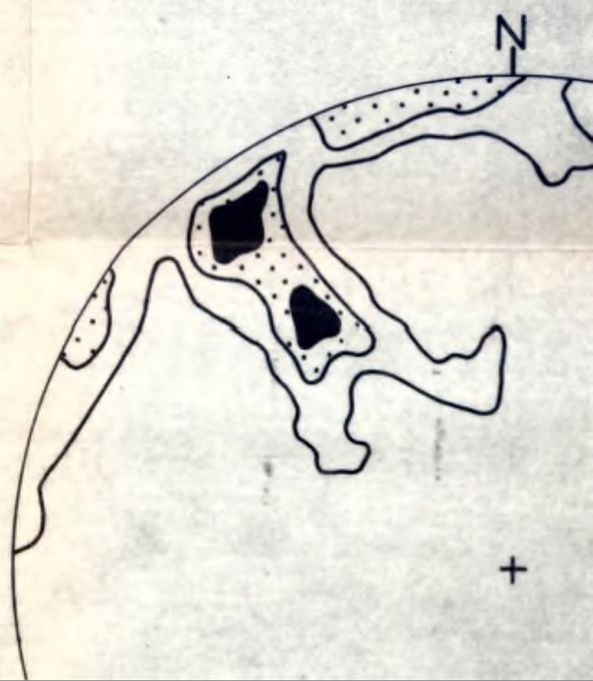

N
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2·38
4·76
7·15
168pts.

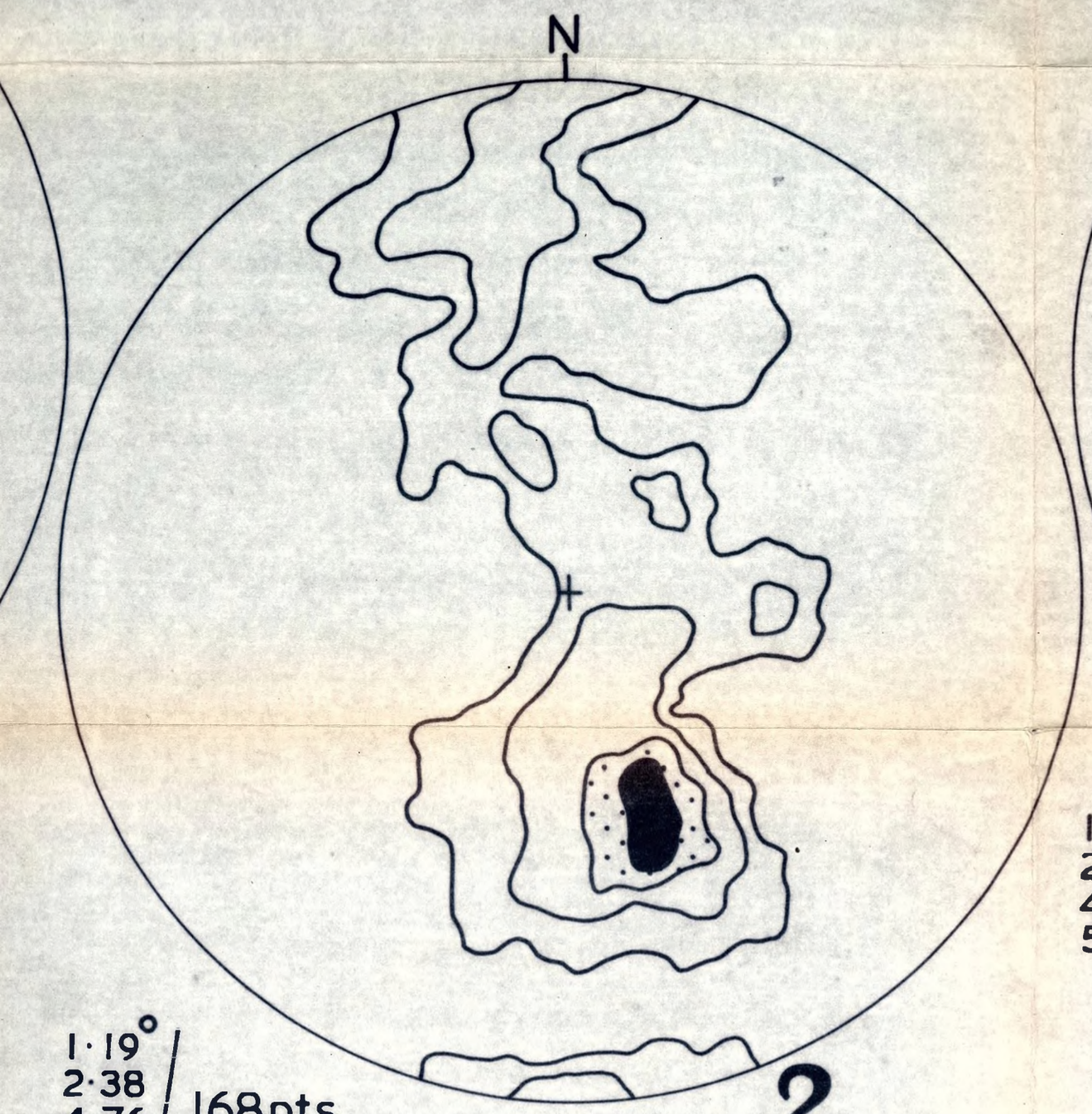
2


N
·86°
1·72
2·87
4·3
5·75
351pts.


MAP 1
SUB AREAS FOR STEREOGRAPHIC ANALYSIS
RIVER COCKER
DODD
1
WHITESIDE
2
GASGALE GILL
3
4
GRASMOOR
5
23
22
21
20

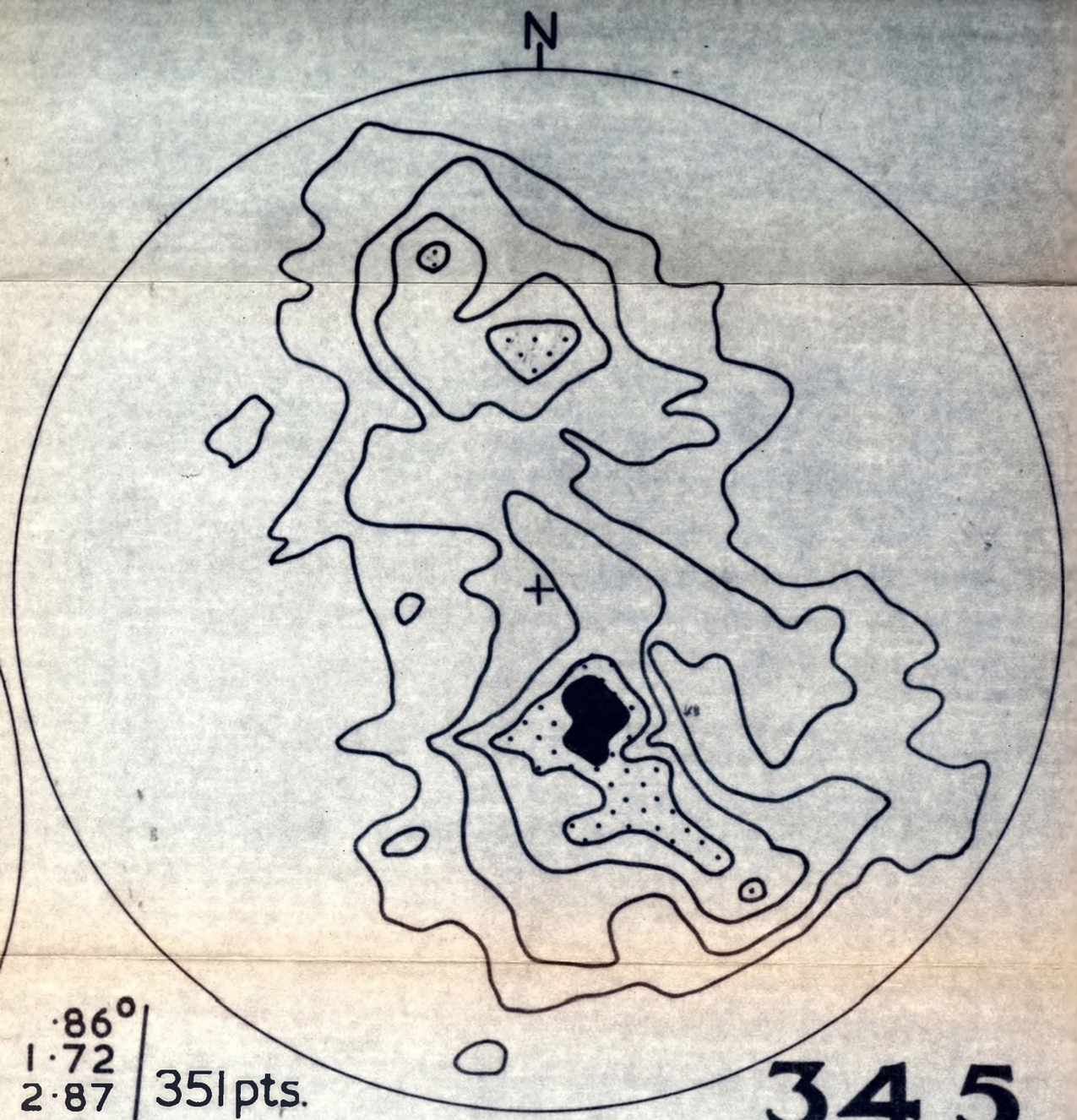

N


N



1·19°
2·38
4·76 / 168pts.
7·15 / o

2

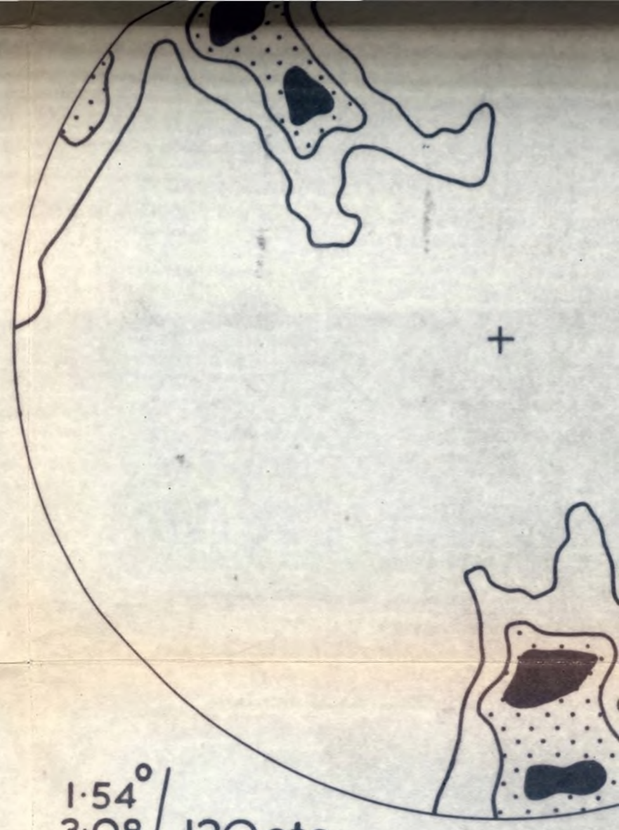


·86°
1·72
2·87 / 351pts.
4·3
5·75 / o

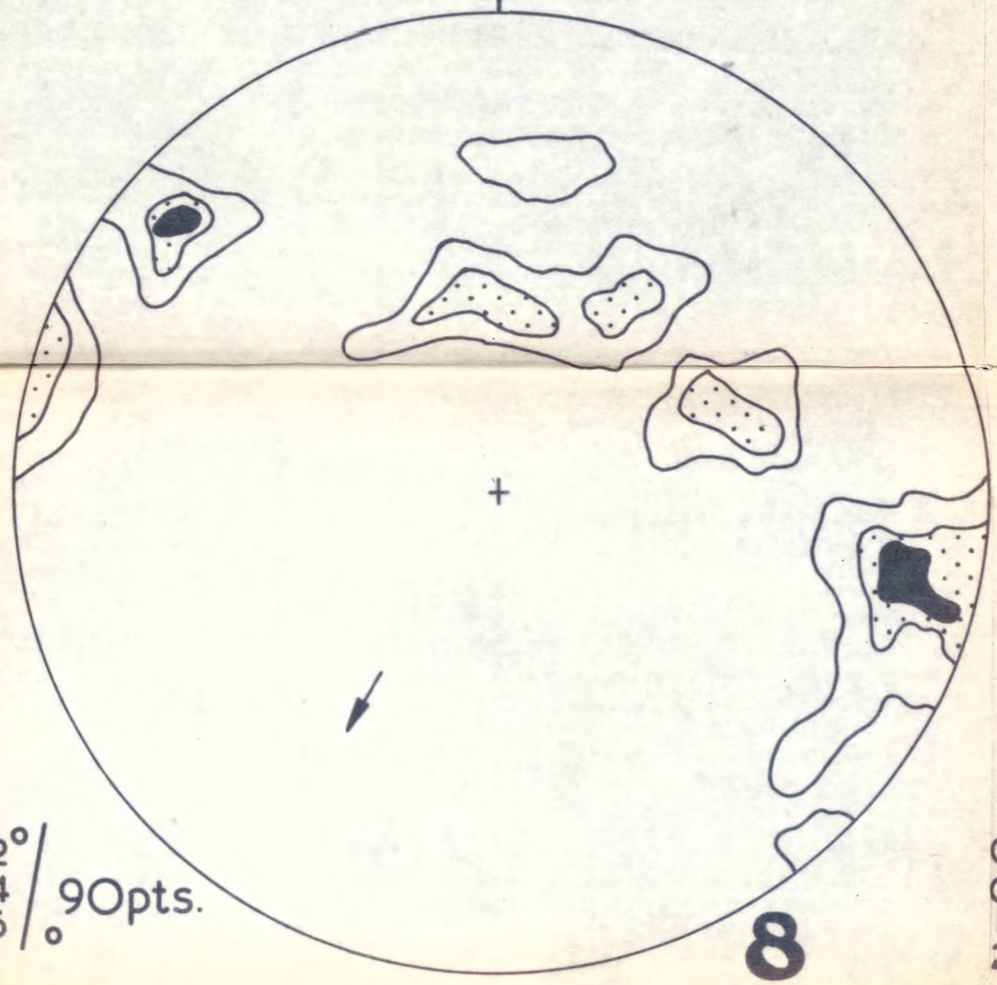
345


WHITESIDE
2
GASGALE GILL
4
3
GRASMOOR
5
CRUMMOCK WATER
RANNERDALE BECK
WHITELESS PIKE
6
7
RANNERDALE KNOTS
8
SAIL BECK
9
BUTTERMERE
2½" — 1 mile
22
21
20
19
18
17
16
17
18
19


0·45°
0·91
1·82
2·72
5·45
440 pts.
9


1·54°
3·08
5·4
130 pts.


·57°
1·14
1·70
2·27
2·84°
1762pts.
Total


N
2·22°
4·44
6·66°
90pts.
8


N
0·48°
0·95
1·9
2·6°
420pts.





0·45°
0·91
1·82
2·72
5·45
/ 440 pts.
o

9

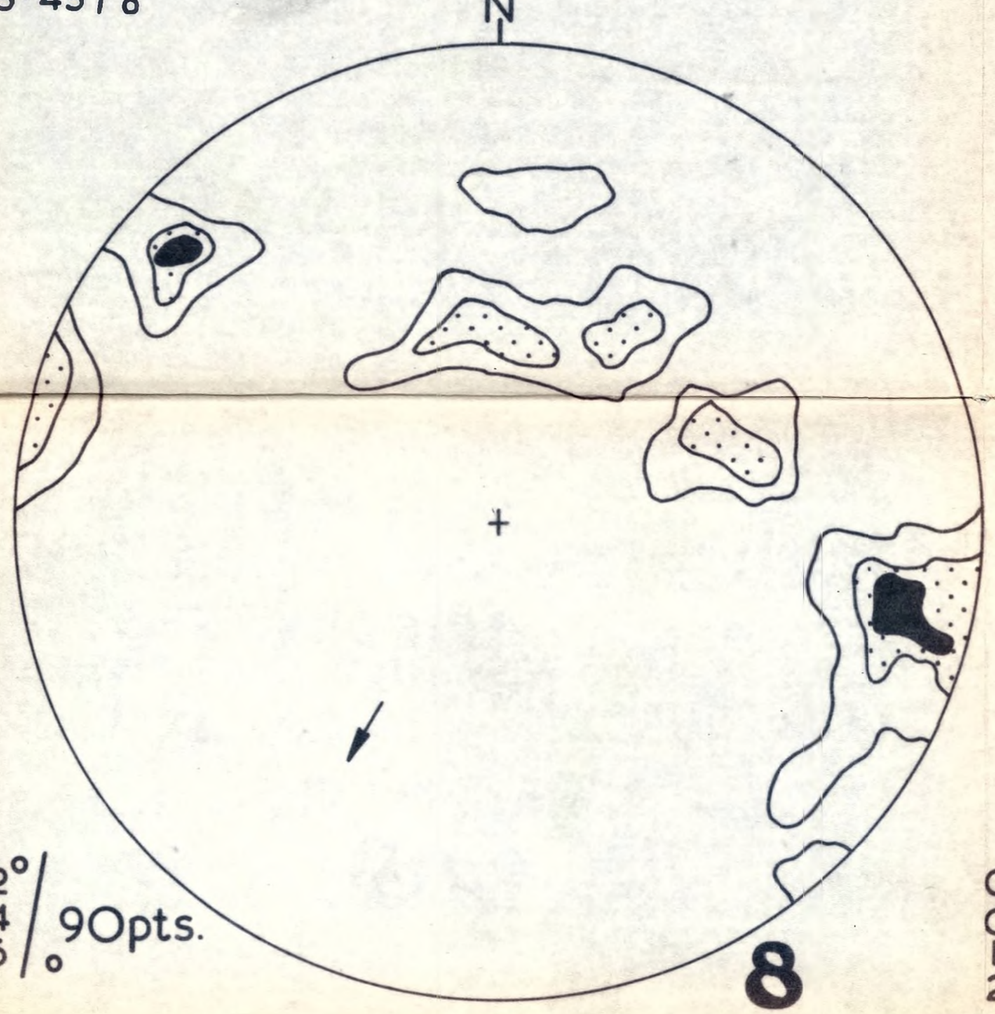


1·54°
3·08
5·4
/ 130 pts.
o

6



Total



2·22°
4·44
6·66
/ 90 pts.
o

8



0·48°
0·95
1·9
2·6
/ 420 pts.
o

7