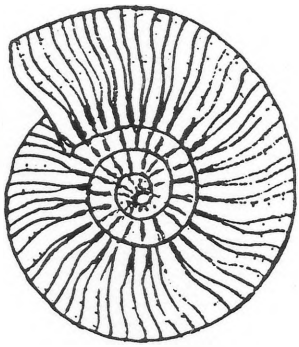
THE CUMBERLAND GEOLOGICAL SOCIETY



PROCEEDINGS

Volume 5 1988-89 Part 2

THE CUMBERLAND GEOLOGICAL SOCIETY

Honorary Life President:

E. H. Shackleton, F.G.S.

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EDITOR'S NOTE

I would like to thank all who have contributed articles, provided lecture and excursion notes and assisted in the preparation of these Proceedings. The art-work and page layouts have again been undertaken by Mr.J.D.Hinde to whom I extend thanks on behalf of the Society.

The Proceedings is a biennial publication that caters for papers on the geology of Cumbria and Northern England. Contributions are welcome and should be submitted to the Editor.

Eric Skipsey

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THE SCAR LIMESTONE AT FAULDS BROW QUARRY, CALDBECK, NORTHERN LAKE DISTRICT

Steve Kershaw & David Gaudoin

SUMMARY

The Brigantian age (Lower Carboniferous) Scar Limestone at Faulds Brow and Old Quarries, north-west of Caldbeck in the northern Lake District, has been studied and logs of the sequences in these quarries are presented. The sequence is dominated by limestones, with interbedded shales and some sandstone, and the rock is mostly fine-grained and contains fully marine fossils. This evidence and the abundance of the trace fossil *Zoophycos* suggests that the rocks were laid down in a protected shallow marine environment while crinoidal and shell debris bands indicate occasional storm deposition.

The sequence in Faulds Brow Quarry shows an upward increase in shale content, eventually replaced by sandstone, and interpreted as a shallowing-up unit where muds and sands from nearby river deltas spread across the area. A similar sequence in the nearby Old Quarry shows a comparable style of deposition but it has not been possible to correlate between shale horizons in the two quarries. Study of the diagenetic history of the sequence supports the depositional model.

1. Introduction

The Scar Limestone of Brigantian age (George et al, 1976) outcrops in the area north-west of Caldbeck, to the north of the Lower Palaeozoic rocks of the Lake District. This study concentrates principally on one locality in the Scar Limestone, Faulds Brow Quarry (FBQ), with additional information from Old Quarry (OQ) 2 km to the west (Figure 1), both former roadstone quarries. FBQ is featured in a

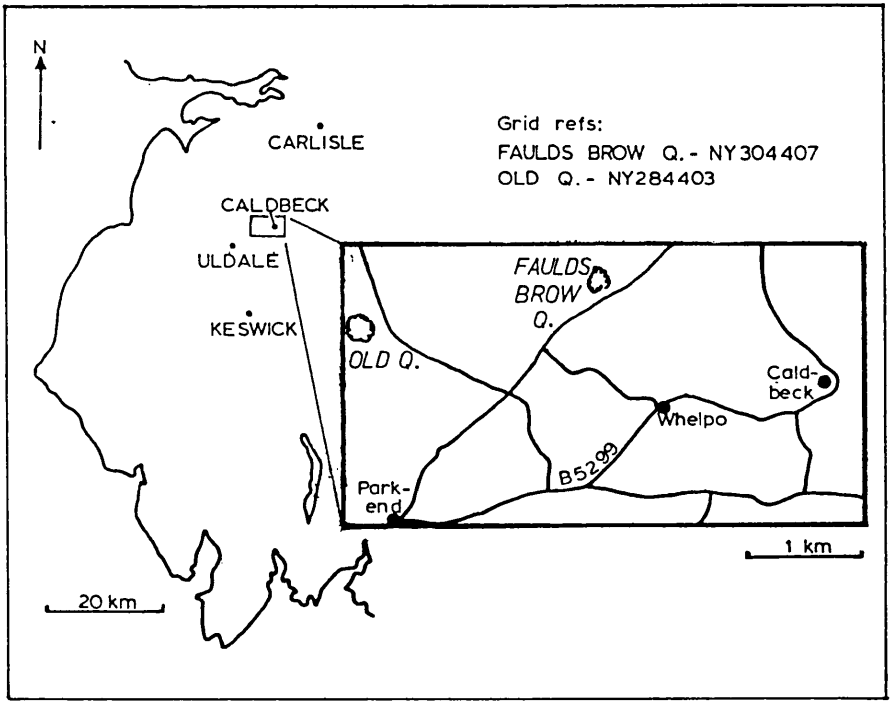


Fig. 1. Locality map showing the position of Faulds Brow and Old Quarries near Caldbeck, northern Lake District.

Lake District excursion guide (Shipp, 1982). The purpose of this paper is to present a detailed study of the sequence in the quarry, which displays interbedded limestones and shales culminating in sandstone, to illustrate the depositional and diagenetic processes preserved. Limestones in West Cumbria are traditionally numbered from top to bottom, First to Seventh Limestones, separated by shales and sands, and forming the Chief Limestone Group (Eastwood et al, 1968).

The Scar Limestone is equivalent to that part of the Fourth Limestone corresponding to the Chert and Junceum Beds further west towards Cockermouth. Interruption of limestone deposition by clastic input from the east split the Fourth Limestone into several smaller units (of which the Scar is one) in the eastern area of its outcrop,

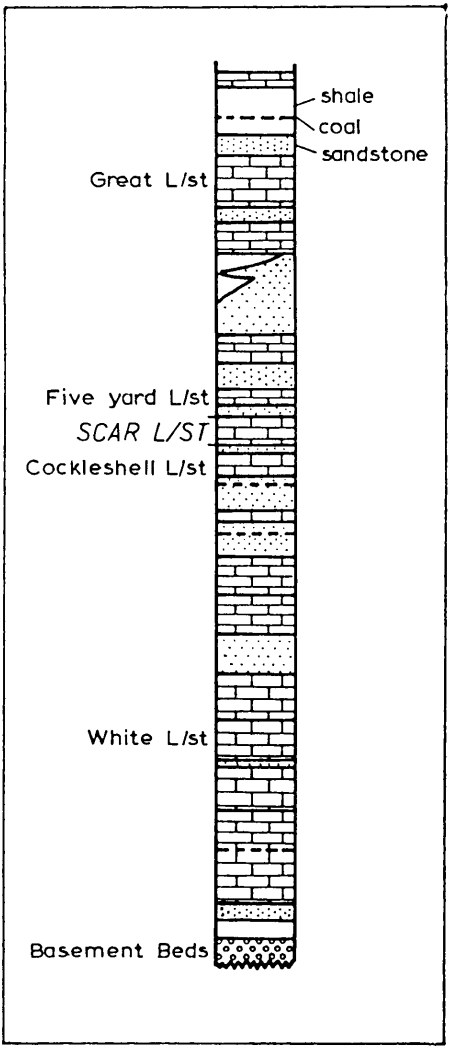


Fig. 2.
Schematic section showing the local sequence of Lower Carboniferous rocks and the stratigraphic position of the Scar Limestone. (not to scale)

imparting a distinct Yoredale cyclicity to the sequence. The Yoredale style of deposition consists of interbedded platform carbonates and clastics, similar to that of the Alston Block (Eastwood et al, 1968; Mitchell et al, 1978). Figure 2 shows a schematic section of the local sequence of Lower Carboniferous rocks in the Caldbeck district, where the Scar Limestone dips gently north-west in the area, part of a heavily faulted, tilted block.

2. Lithologies, Fossils and Measured Sections

Figure 3 is a sketch of FBQ with several features of interest. Note the tabular bedded sediments, prominent shell bands and the abundance of *Zoophycos* trace fossils associated with pyrite at specific horizons. Well developed jointing is visible on exposed bedding planes; lateral stress in the limestone has resulted in sheared joints in places. Figure 4 presents logged sections for the quarries; approximately 17m of section were logged in FBQ while 11m were logged in OQ. Most of the sequences are clearly exposed apart from a

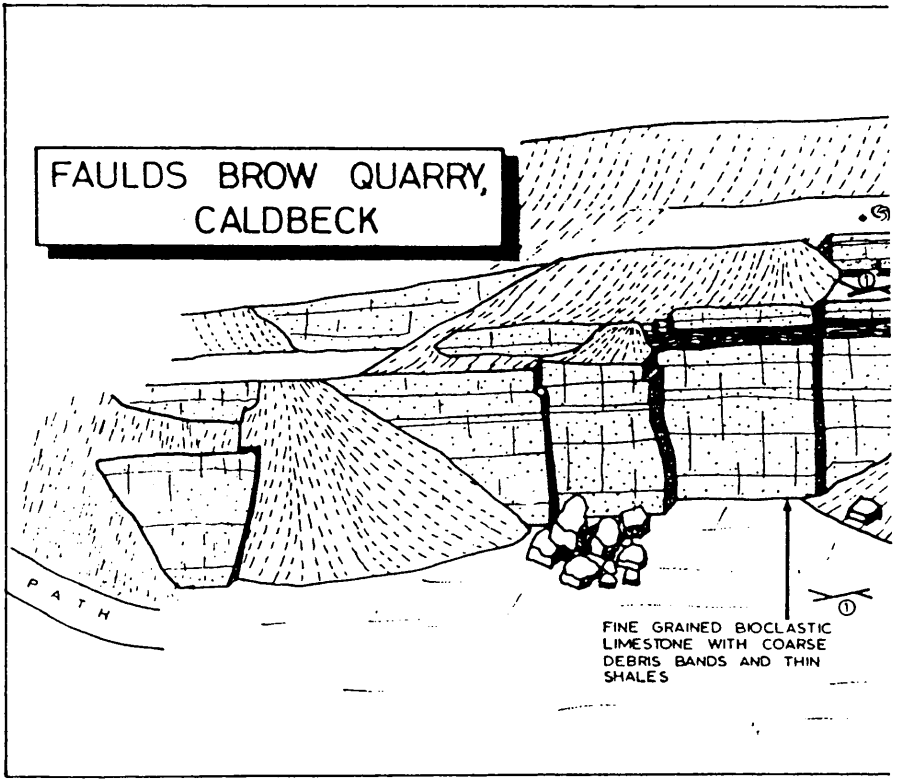
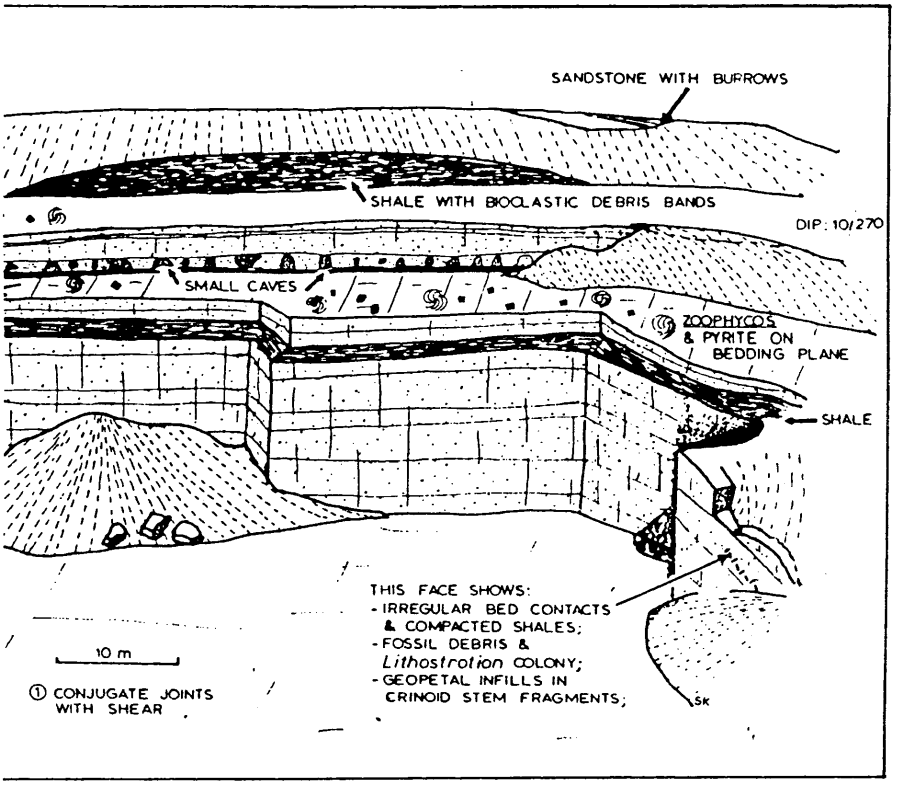


Fig. 3. Field sketch of Faulds Brow Quarry, showing the location of important features discussed in the text. Note that all the sequence is readily accessible, but visitors should not approach the quarry faces; they are loose and dangerous.

small gap at the top of FBQ, where the upper shales were inaccessible due to cover. There appears to be some correlation between the two, in that the two prominent shale bands near the tops of each may be the same beds (with the thick shale and sandstone at the top of FBQ missing from OQ), but thinner bands do not match and so precise correlation could not be assigned with confidence. Weathering masks much of the rock faces and a close sampling interval was required in order to obtain full details. The majority of the rock is fine-grained limestone with varying fossil content; interbedded shales are highly



fossiliferous in places while the medium-grained sandstone at the top of the quarry contains only trace fossils.

2.1. Faulds Brow Quarry

Figure 4 shows that the FBQ sequence is composed largely of fine-grained limestones, interspersed with coarse bands of shell debris. Progressive upward increase in the amount of shale ultimately replaces the limestone, and in turn is replaced by sandstone. The sequence can be considered as several sub-units as follows, from base to top :

a) 0 - 3m : massive fine-grained limestone with coarse bioclastic band, strongly dolomitised in places (Figure 6A)

b) 3 - 5.6m : bedded fine-grained limestone with several prominent fossiliferous thin shale bands. Many thin shale bands (5 - 10mm) are too thin to record on the log.

c) 5.6 - 7.1m : massive fine-grained limestone (Figures 6B,C) with coarse bioclastic band.

d) 7.1 - 7.9m : fossiliferous shale

e) 7.9 - 9.2m : massive fine-grained limestone with coarse bioclastic band (Figures 6D,E). At 9.2m a prominent bedding plane exposes large numbers of *Zoophycos* associated with pyrite.

f) 9.2 - 9.8m : interbedded shale and shaly fine-grained limestone (Figure 6F), with small cave system.

g) 9.8 - 11.2m : massive fine-grained limestone with some bioclastic material (Figure 6G).

h) 11.2 - 11.6m : thin interbedded shale and limestone. At 11.6m further *Zoophycos* and pyrite are present.

i) 11.6 - ?15.9m : thick shale with several coarse bioclastic bands (this sub-unit includes a small gap where the shales were inaccessible).

j) ?15.9 - 17.3m : medium-grained quartzite (Figure 6H), parallel laminated and with horizontal burrows.

Because of weathering, detailed examination of the rock was required to distinguish these units and the fossils were better recognised in thin sections, illustrated in Figures 6A-G.

2.2. Old Quarry

Figure 4 shows that Old Quarry has a similar style of limestone deposition as FBQ with thick shale bands towards the top. There is no sandstone here and the sequence displays several large *Lithostrotion* colonies (Figure 5), more impressive than the smaller ones at FBQ. There are also several coarse shell debris bands, abundant interbedded shales and a horizon of associated *Zoophycos* and pyrite.

In both quarries the fossils formed a benthic fauna and the environment can only be interpreted as fully marine. Most fossils were found to be debris in thin sections and the number of complete

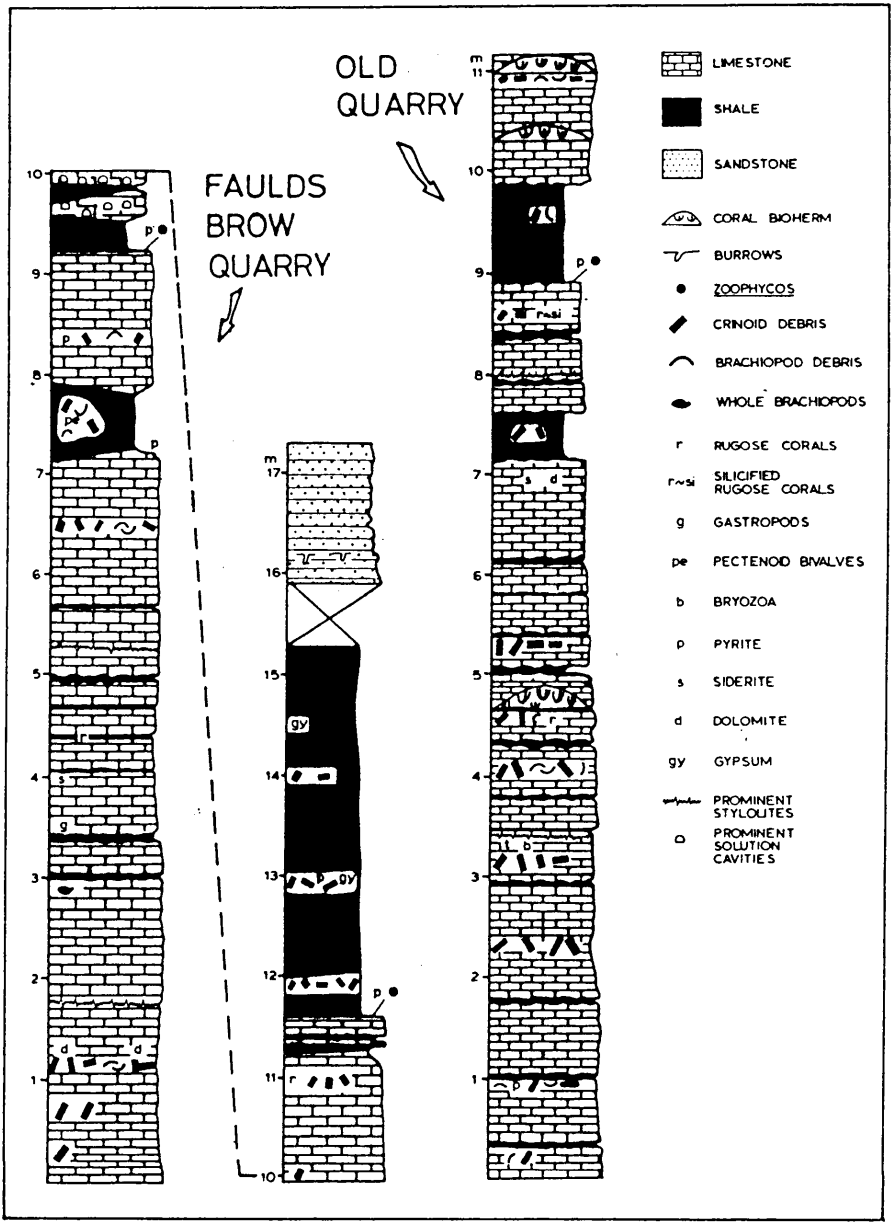


Fig. 4. Logged sections of Faulds Brow and Old Quarries, illustrating the dominant limestones with interbedded shale and sandstone.

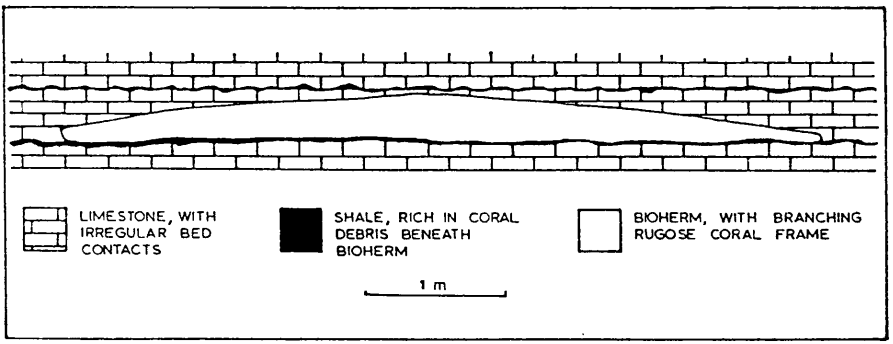


Fig. 5. Outline of a small patch reef at Old Quarry (5.7 m. above base), dominated by *Lithostrotion junceum* as a baffle. Note the thinning of limestone bed over the reef, where sediment accumulated slowly and covered the reef.

specimens was relatively few. In addition to those in the key of Figure 4, numerous foraminifera were present in thin sections.

3. Depositional Model

The information provided by the logs leads to an interpretation of the sequence in terms of a depositional model in which the Scar Limestone at FBQ and OQ displays the following characteristics :

- well developed tabular bedding, where beds are traceable across the quarries; wavy bed contacts, nodular in places.
- fine-grained wackestones and packstones (limestones where sediments are matrix-supported or grain-supported respectively) containing much bioclastic material.
- coarser-grained shell debris bands.
- thin shales.
- stenohaline marine fauna (fauna only able to tolerate slight changes in salinity and indicating a fully marine environment).
- small, low profile *Lithostrotion* reefs.
- trace fossils of low energy and low sedimentation rate environments (*Zoophycos* in abundance).

Lacking from the sequences is the variety of rock types commonly associated with a shallow water environment (eg cross-stratified oolites, thick beds of coarse, mud-free bioclastic limestone) and other common indicators of very shallow water, such as stromatolites, evaporites and omission surfaces. These observations appear to be consistent with the interpretation of a medium shelf setting, well above storm wave base but away from the influence of shallow water conditions. Environmental energy and sedimentation rates were relatively low, with occasional storm input of bioclastic material (Wilson & Jordan, 1983). However, work on the trace fossil *Zoophycos* (Frey & Pemberton, 1986) has shown that it is frequently abundant in shallow quiet environments behind a barrier to open circulation where the water's oxygen level was reduced, and as a result pyrite commonly forms in association. Although no such barrier can be identified at FBQ or OQ, the upward change from limestone to shale to sandstone, provides evidence that the whole sequence was laid down in shallow rather than deeper water.

The interruption of the Fourth Limestone by clastics in the eastern area reflects major sedimentary processes at work during the Brigantian. The main part of the Lower Carboniferous limestone deposition was complete by that time, and this area of Cumbria was subject to clastic deposition from general southward progradation of fluviodeltaic systems spilling westwards across the Lake District, i.e. deltas spreading across the area depositing mud and sand in the sea (Leeder, 1987). This large clastic input was due to uplift whose causes are unknown but may be related to lithospheric stretching during subduction of ocean crust south of Britain (Leeder, 1987).

The transport of the clastics during this time was assisted by increasing climatic humidity due to higher rainfall. Therefore the interbedded shales can be regarded as distal deltaic deposits (ie the outer parts of a delta); their greater frequency and increasing thickness towards the top of FBQ reflects periodic and encroaching delta progradation (advance) until the top of FBQ is inundated with deltaic sand. The sandstone has not yielded any marine body fossils and no evidence was discovered to confirm or reject the possibility of emergence of the delta top above sea-level.

Shackleton (1977) noted that the limestones of the area became more estuarine to the north and east, indicating shallowing in those areas, although the fauna of the limestone in the quarries are marine. Eastwood et al (1968) show that the limestone members of the Upper Chief Limestone Group (which includes the Scar) maintain an approximately constant thickness from east to west; thickness variations of the whole sequence are evidently due to differential geographical distribution of clastic input. Thinning in the Cockermouth district appears to be due to lack of clastics there.

A clastic-influenced carbonate shelf therefore is proposed, as opposed to a carbonate ramp as in South Wales (Wright, 1986). A ramp is formed on a sloping continental shelf with water deepening downslope towards the open sea until the limestone is replaced laterally by muds, while in contrast a shelf develops on an approximately horizontal sea bed, which consequently lies in shallow water. Often a ramp builds up into a shelf as limestone accumulates in an area starved of clastics (Wilson, 1975).

Thus the FBQ and OQ sequences, although lacking in many of the characteristics often regarded as shallow water indicators mentioned above, are quite-consistent with a shallow marine setting protected from high energy waves and currents, where delta deposition simply swamped the limestones.

4. Solution features

Just below the 10m level at FBQ (Figure 4), small solution cavities occur (see Figure 3 and Shipp, 1982 - Figure 25). These are vertically elongated and merge directly with vertical joints in the limestone. Shale underlies this line of solution holes. Such solution is a karst feature but no palaeokarst surface (where the limestone was dissolved by being exposed on a land surface during an uplift phase) could be recognised at this level of the quarry. This is interpreted as solution taking place beneath the surface, due to water flowing down joints and, being unable to continue downwards because of the shale blocking its path, flowed laterally dissolving the limestone at this level. The lack of an ancient karst surface means

that it is difficult to determine the time that the solution occurred, as discussed by Wright (1982) in his account of Lower Carboniferous palaeokarsts in South Wales. The solution could have occurred soon after uplift of the area, with consequent downflow of fresh water, and could therefore be consistent with the depositional processes described above. However, solution could have happened at a later date, perhaps even in relatively recent times. It is curious that solution only occurred at one level; it might have been expected at all levels where limestone overlaid shale. Although this part of the sequence provided important pathways for groundwater movement, unfortunately the solution features cannot be satisfactorily dated.

5. Diagenesis

The depositional model presented fits well with observations and the literature but the diagenesis of the sediments should also reflect the processes described. Thin section studies indicate a complex diagenetic history in these rocks but the broad pattern can be related to the vertical facies changes at FBQ. The diagenetic features are typical of shallow marine deposits and full descriptions of the processes can be found in standard text books, such as Leeder (1982) and Scoffin (1986).

5.1. Early diagenesis

Following deposition of lime muds and bioclastic material, the earliest formed calcium carbonate cements are likely to have been formed in marine phreatic (below the water table) conditions within 1m. of the sea bed surface. Such cements are often present as rims of calcite crystals around grains. However in this limestone such rims have not been observed and much of the sediment now consists of fine-grained calcite crystals, called micro- and pseudo-sparite, which appear as a fine mosaic of crystals in thin section (Figure 6B). They are usually regarded as recrystallisation products of a micrite (very fine-grained calcium carbonate) sediment. Here they are preserved as non-ferroan calcite, which can be identified by a common limestone-staining technique. In oxygenated environments, any iron in the water is oxidised to insoluble detrital material and

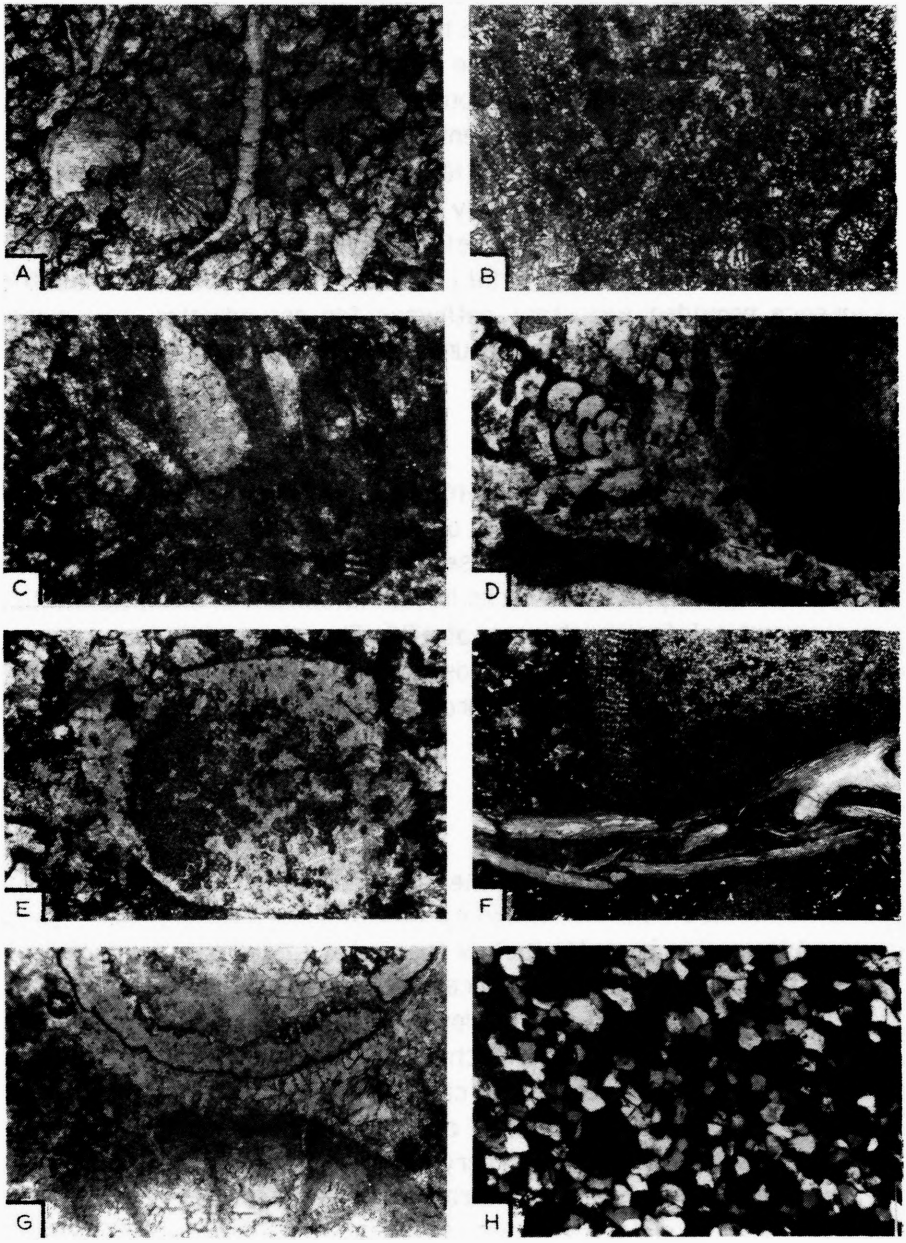


Figure 6 - see opposite.

is not incorporated into the calcite crystal lattice; in reducing conditions, iron stays in the reduced (soluble) form and is included in the calcite. Non-ferroan calcite can therefore indicate either deposition in an oxygenated environment or in an iron-depleted reducing environment. Distinguishing between these two is problematic and requires isotope analysis, impracticable in this study. Thus it is not known with certainty whether the non-ferroan, fine-grained calcite is an early (oxygenated environment) or a late (iron-depleted environment) product of diagenesis.

Fig. 6. Photomicrographs of samples from Faulds Brow Quarry - height above floor is indicated. All are in plane polarised light except H, which is in cross polarised light - photographs at X 30 magnification.

A. Dolomitised bioclastic material, showing effects of pressure solution : 1.80m.

B. Bioclastic wackestone showing pronounced pseudo-sparite texture : 5.80m.

C. Recrystallised bioclasts and foraminifera in wackestone: 6.00m.

D. Crinoidal and foraminiferal packstone : 7.90m.

E. Crinoid ossicle with micro-dolomite inclusions : 8.40m.

F. Crushed brachiopod and altered crinoid : 7.95m.

G. Adjacent *Lithostrotion* corallites, upper- silicified, lower - unaffected : 11.00m.

H. Orthoquartzite : from sandstone at16.50m.

More certain early calcite cements are commonly observed on crinoid stem fragments. Individual crinoid ossicles are composed of single calcite crystals and attract cement readily to grow on their surfaces in optical continuity (called syntaxial cement overgrowths on the crinoids). These are usually taken to indicate a period of uplift (regression), whereby lithifying processes occurred in meteoric phreatic (below water table in fresh water) environments because such overgrowths are inhibited in marine diagenesis by sea water poisoning. Syntaxial overgrowths were commonly observed in thin sections from FBQ and OQ. Hence such an interpretation is consistent with the depositional scenario outlined, whereby water shallowed upwards in the section and marine water was replaced by fresh to allow these cements to grow.

Dolomitized beds are rarely met in either quarry, occurring at 1.4m. above the base in FBQ (Figure 4). This dolomite is coarsely crystalline (sucrosic) in texture and is more likely to be of secondary formation (primary dolomites are often cryptocrystalline and associated with evaporites, none of which were seen in the sequence). The most likely model for formation of these dolomites is the ground water mixing model (Leeder,1982), which indicated that dolomites can be formed where sea water and fresh water may be saturated with respect to dolomite and calcite, but mixtures with between 5 and 50% sea water are under-saturated with dolomite. Thus dolomitization is expected to have taken place where marine and fresh waters mix and may well have taken place during the regressive phases. Ramsbottom (1973) has indicated that dolomitization is related to uplift in many Dinantian cycles. Figure 6A illustrates how dolomite has replaced parts of the limestone.

5.2. Later diagenesis

A variety of other diagenetic effects were observed, involving both pore-filling cements and the types of replacements of original fabric, which are usually taken to indicate later diagenesis during and after the sediments have been buried under piles of accumulating material.

Formation of inter- & intra-granular (between and within) sparite was of both non-ferroan and ferroan calcite in these rocks. Since ferroan calcite forms in reducing conditions, it is often regarded as a late cement. However the ferroan cements are restricted to limestones directly below shale beds (eg. Figure 6F). The shale beds provided a barrier to percolating ground waters, and stagnant pore water may have existed during the early diagenetic history; hence ferroan calcite could be an early formed cement in these places. Siderite found below a shale in OQ (Figure 4) can be interpreted as developed in anaerobic fresh water where carbonates combine with iron in preference to sulphides, which has low concentration in fresh water. Examples were found of whole brachiopods which had been partly crushed by compaction of overlying sediments (Figure 6F), indicating that they were not cemented until later in diagenesis. These brachiopods were filled with ferroan calcite and prove that at least part of this cement was a late event after burial.

The calcium carbonate minerals (chiefly aragonite and calcite) behave differently in diagenesis, with aragonite being prone to alteration and dissolution while calcite is more stable. However calcite stability depends on the magnesium content of the $CaCO_3$ lattice and since fossils contain various amounts of Mg, they are affected to different extents by diagenesis. Brachiopods are composed of low magnesium calcite (LoMg calcite) while crinoids are of high magnesium calcite (HiMg calcite); crinoids are consequently less stable than brachiopods. Crinoid alteration leads to expulsion of Mg from the calcite lattice and the local formation of tiny dolomite crystals called micro-dolomite within the crinoid grains. Such inclusions (Figure 6E) are common in the FBQ sequence but only in echinoderm grains. This suggests that the supply of magnesium required for dolomite formation is from within the grains themselves. Thus excess Mg ions are produced as HiMg structures are recrystallised to more stable LoMg calcite. Since there is no evidence of the migration of magnesium ions (micro-dolomite is only seen within the grain), the method of formation is likely to have been solid state exsolution. This results in two stable phases - LoMg calcite and dolomite - formed directly by solid-state diffusion of Mg within the metastable HiMg calcites. Thus an intermediate

aqueous phase is not required as the Mg ions cannot migrate. Figure 6G shows patches of sparite between corals where fine-grained calcite grades into coarser sparite. In this case the finer material has recrystallised to larger crystals by the well known process of aggrading neomorphism, where newer crystals show an increase in size, usually taken to be a later diagenetic event. Silicified bioclastic fragments are quite common throughout the sequence but are particularly noticeable in the coral bands. Corals do not contain quartz as a primary part of their skeletal structure, so the source of the quartz is possibly the overlying orthoquartzite or siliceous components of the fauna, such as sponges, not observed as preserved fossils in this sequence. The concentration of silica within the coral skeleton and not in the sediment suggests an acid micro-environment in the coral walls, precipitating the silica from solution. Such conditions can be developed in places where organic matter remains in the coral; the breakdown of such matter producing local acid conditions.

Sporadic occurrence of gypsum in shales (Figure 4) might suggest high evaporation rates leading to the formation of evaporite deposits, but such gypsum is better interpreted as being the oxidation product of pyrite in recent times, whereby sulphide is converted to sulphate and combined with calcium ions in the groundwater.

5.3. *Zoophycos*

In reducing environments just below muddy sea floors, modern (and therefore, presumably, ancient) sediment is laden with sulphate-reducing bacteria, which reduce sulphates to sulphides. The sulphides generated combine with reduced iron in pore water to deposit pyrite. The association of abundant large pyrite nodules and the trace fossil *Zoophycos* is notable in this context in places where limestone passes upwards into shale in both quarries (Figures 3 and 4). *Zoophycos* is an elaborate ichnofossil (Shipp 1982 - Figure 24). It has the shape of a low angled helicoidal sheet constructed of spirally radiating burrows, each lined with re-textured sediment derived from adjacent beds. Samples that are found in the Scar Limestone are

approximately 200mm in diameter and flattened by compaction. The soft-bodied organism that made the trace is unknown but its activity clearly involved burrowing and processing sediment. Such a mode of life is referred to as 'deposit feeding' where the animal ate sediment as it burrowed, so that the burrows are both feeding and dwelling structures.

The dwelling environment of *Zoophycos* is generally considered to be one of loose sediment and muddy water condition, nearly always associated with the interface between black shales and other sediments (Frey & Pemberton, 1986), as seen in the Scar Limestone. Pyrite is likely to be formed early in diagenesis by sulphate-reducing bacteria and signifies reducing conditions in the sediments. Thus abundant pyrite at specific levels in the sequence indicates periods when the sea bed became stagnant, signifying pronounced lack of water circulation with a low sedimentation rate. The fact that shales overlie the *Zoophycos*-bearing beds suggests that uplift was occurring and that just before shales started to accumulate, the locality became shallower and more dis-connected from the open sea, leading to a lagoon-like environment.

It is likely that organic detritus was accumulating in the stagnant sediments into which the *Zoophycos* animal was burrowing. This animal was clearly an active metazoan and can be presumed to have required an oxygen supply to maintain its metabolism; so the water above must have been oxygenated to some degree. Therefore *Zoophycos* seems to have been produced by an animal that thrived in poorly oxygenated water but populations of the animal were wiped out as the shales smothered them and the soft bodies of the *Zoophycos* animals added to the organic detritus. Frey & Pemberton (1986) provide a full discussion of *Zoophycos* and stress its importance in shallow environments of this type.

Information from the diagenetic fabrics and the *Zoophycos*-pyrite beds are consistent with the depositional processes at work in these sediments. In the next section the wider aspects of Lower Carboniferous deposition are considered.

6. Controls on depositional processes

Controls on cyclicity evident in the Lower Carboniferous of northern Cumbria are, as elsewhere, subject to more than one interpretation and a study of this cyclicity provides clues to understand FBQ. Leeder and Strudwick (1987) provide a succinct treatment of the processes causing carbonate-clastic cyclicity. They propose four mechanisms of control on such shallowing-up cycles :

- **Sedimentary mechanisms** involve the dynamics of fluviodeltaic processes, whereby deltas switch direction from one place to another and lead to areas suddenly becoming inundated with clastics, resulting in an upward progression of limestone-shale-sandstone. Repeated delta switching, taking deposition elsewhere, allows a return to carbonate deposition as the area once again becomes starved of clastic supply.

- **Eustatic mechanisms** involve simple sea level fall and rise, bringing the shoreline closer to the area of carbonate deposition and resulting in clastic progradation across the lime mud. Subsequent sea level rise would push back the shoreline, allowing a return to carbonate deposition.

- **Tectonic mechanisms** where contemporaneous fault movement affects the relative position of a prograding delta with respect to sea level.

- **Tecto-sedimentary mechanisms** where fault-controlled uplift or subsidence affects the direction in which a delta advances.

Leeder and Strudwick (1987) indicate that research has not progressed far enough to differentiate between their suggested four controlling mechanisms, and the following comments relating to FBQ and the Brigantian sequences in the northern Lake District bear this out. The sequence at FBQ can be justifiably assumed to be a shallowing-up feature, with prograding clastics. Also the presence of non-ferroan syntaxial cements on crinoids in many places in the sequence strongly suggests influence of meteoric water in the early

diagenetic history. The limestones in the Caldbeck area are often overlaid by shales and then sands, but a sharp return to limestone deposition above the sand (without intervening shales) is a persistent feature of the local stratigraphy (Eastwood et al, 1968; Figure 2), and is presumably due to an increase in sea level at those points. Delta progradation itself could be responsible for shallowing-up, if clastic deposition exceeded subsidence, as has been suggested for Yoredale cyclothems (Wilson, 1975; p.203-6). The following sea level rise indicates abrupt cut-off of the clastic supply, since the sands are directly overlaid by limestone, and could be due either to a tectonic or eustatic change, or to rapid delta abandonment.

Figure 4 shows different numbers and thicknesses of thin shale bands in the lower parts of each of the two quarries. The fact that these could not be correlated over only 2 km. could reflect local variations of distribution of fine clastics deposited from distal delta lobes. Thus some of the bands may be part of local patches of mud developed on the sea bed; others may be more extensive and the difference could relate to varying current activity or different amounts of mud input. Incomplete shale beds would also generate varying pathways for pore waters through the rocks, affecting the diagenetic textures seen.

7. Conclusions

a) The vertical facies changes at Faulds Brow Quarry reflect increased clastic input and provide evidence of uplift and delta progradation across a shallow water carbonate platform in the late Lower Carboniferous.

b) Diagenesis of the sequence reveals a period of post-depositional uplift followed by burial, consistent with the depositional model.

c) Groundwater movement during diagenesis, blocked by shales in places, produced varying diagenetic environments at different stratigraphical levels in FBQ and OQ, particularly development of ferroan cements directly beneath shale bands.

d) Association between *Zoophycos* and large amounts of pyrite indicate changing sea bed conditions in the vertical development of the sequence.

e) FBQ makes an excellent venue for undergraduate parties studying carbonate-clastic facies.

7. Acknowledgements

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RECENT DEVELOPMENTS CONCERNING THE BORROWDALE VOLCANIC GROUP OF THE ENGLISH LAKE DISTRICT

M.G.Petterson

SUMMARY

The Borrowdale Volcanic Group (BVG) is part of a Caledonian magmatic province which is located south of the Iapetus Suture. The application of modern volcanological principles to the BVG has led to new interpretations which attempt to explain the BVG as products of volcanic processes.

The BVG can be divided into two phases of volcanicity; a first phase which comprises the products of effusive lava-dominated volcanism and a second phase largely comprised of ignimbrites which are the products of explosive volcanism. Re-working of the volcanic material produced fluvial & lacustrine epiclastic sediments.

One of the most exciting recent discoveries has been the realisation that many faults within the BVG are volcano-tectonic in origin and are part of the larger process of volcanic-induced subsidence and caldera collapse.

1. Introduction

During the past seven years the Borrowdale Volcanic Group (BVG) has been the subject of geological research by the British Geological Survey, in collaboration with several university geologists. This detailed survey has applied the principles of physical volcanology and has greatly increased understanding about the Group. The aim of the paper is to give a brief review of current ideas on the

palaeo-environment and structure of the volcanics and to discuss the significance of the more common lithologies found within the Borrowdale volcanic pile.

2. Geological setting

The Borrowdale (Llandeilo-Caradoc age) volcanics are part of an elongate chain of Caledonian magmatism, lying south of the Iapetus

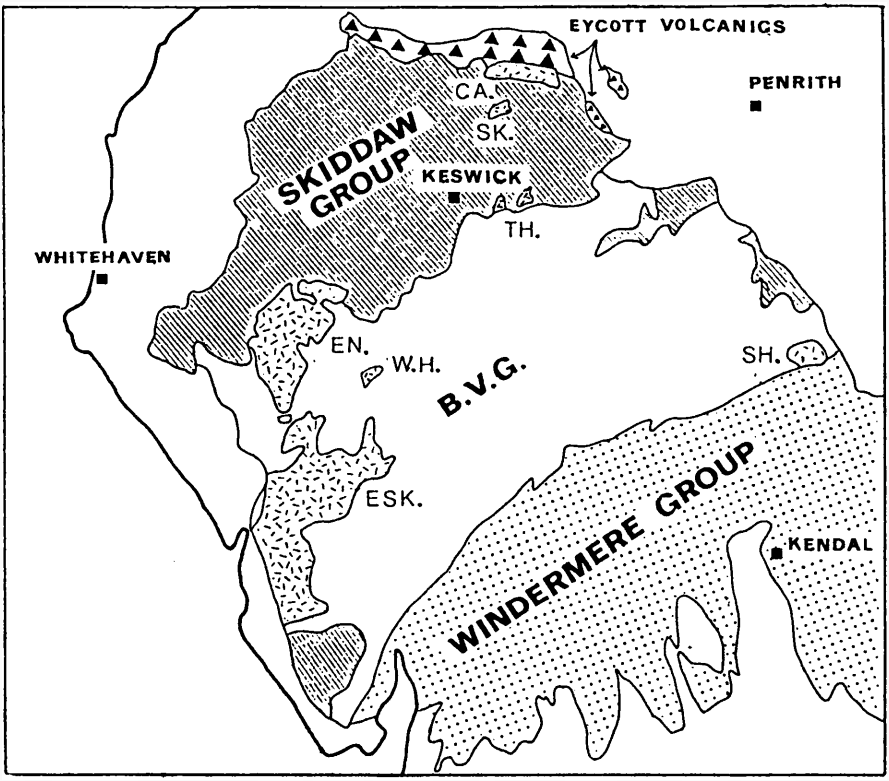


FIG. 1 Geological map of the Lower Palaeozoic inlier of the English Lake District, comprising the Skiddaw, Eycott Volcanic, Borrowdale Volcanic and Windermere Groups. These groups are cut by the Carrock (Ca), Skiddaw (Sk), Threlkeld (Th), Ennerdale (En), Wasdale Head (WH), Eskdale (Esk) and Shap (Sh) intrusions.

Suture, and stretching from Dingle, S.W. Ireland, through Leinster, S.E. Ireland, to the Lake District and North Pennines. Within the Lower Palaeozoic inlier of the Lake District (Figure 1), rocks of the Borrowdale Volcanic Group crop over an area of 800 km^2 and attain a maximum thickness of 6 km. They were erupted on to a basement of siltstone/sandstone turbidites (Skiddaw Group) of Tremadoc-Llanvirn age. Volcanism had ended by the Silurian and major subsidence in the southern part of the Lake District led to the deposition of a thick (6 km.) sequence of sediments which comprise the Windermere Group. Thus the Borrowdale Volcanic Group unconformably overlies the Skiddaw Group and is itself unconformably overlain by the Windermere Group, shown in Figure 1.

3. Effusive and explosive volcanism

In order to explain the significance of volcanic deposits, volcanologists must attempt to explain the genesis of the deposits in term of volcanic processes. Much of this information has been gained by observations of active volcanoes in Hawaii, Etna (Sicily) and Mount St. Helens (U.S.A.). By observing volcanoes in every possible geological environment, deductions can be made about the style and variation of volcanic processes.

However the formation of some types of deposits which occur in the geological record have never been observed, for example large scale ignimbrite eruptions. In order to explain these rock types a combination of theory backed by experimental physico-chemical and field data is required. The aim therefore is to explain volcanic deposits in terms of the physical and chemical processes which produce them, and how these processes vary in time and space.

3.1. Effusive volcanism

Effusive and relatively quiet volcanism with the eruption of lava is perhaps the process with which we are most familiar. The viscosity of lava is affected by factors such as chemical composition, dissolved gas content and temperature. Generally the most fluid lavas are low silica basalts extruded at high temperatures, and the most

viscous lavas are high silica rhyolites extruded at relatively low temperatures.

Basalts can flow over relatively large areas and produce lava flows with a low thickness/length ratio, whereas rhyolites tend to form thick dome structures of limited lateral extent. There is a full spectrum of lava types between these two extremes. The overall morphology of lava flows is largely controlled by effusion rate, physical properties and slope.

3.2 Explosive volcanism

Deposits resulting from explosive eruptions which fragment and

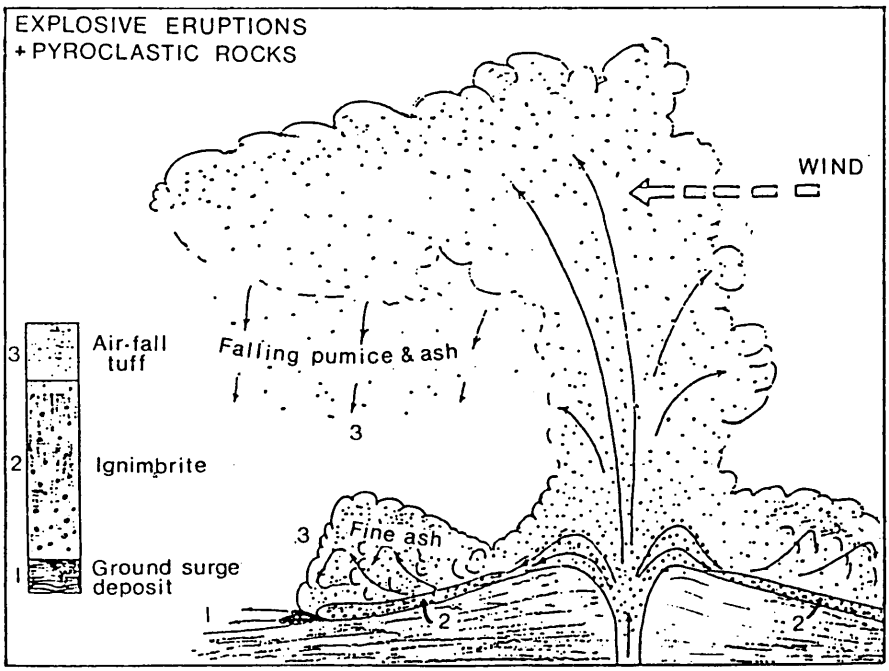


Fig. 2 Deposits associated with an explosive (Plinian) volcanic eruption. Material thrown upwards to form a volcanic plume falls as a pyroclastic fall deposit. Collapse of the plume leads to pyroclastic surge & pyroclastic flow (ignimbrite) deposits.

disperse solidified magma are termed "pyroclastic"; they can be sub-divided into "pyroclastic fall", "pyroclastic flow" (ignimbrite) and "pyroclastic surge" deposits according to their mode of origin (see below and Figure 2). Explosive eruptions themselves are classified as "Hawaiian", "Strombolian" and "Plinian" according to the degree of fragmentation and dispersal of their erupted material, illustrated in Figure 3. Other descriptive terms are used for more specific types of eruption, for example "Surtseyan" and "Phreatoplinian" eruptions involve the inter-action of water and magma which increases the violence of the eruption (Figure 3).

Explosive volcanic activity is especially common in high viscosity, high silica magma systems, that is andesite, dacite and rhyolite volcanoes. The greater viscosity of these magmas often leads to a violently explosive de-gassing of magmas at low pressure; explosive activity is favoured by a high magma gas content and the inter-action of magma with water. Very violent Plinian volcanic eruptions fragment solidified magma to a high degree and disperse these fragments over a very wide area (Figure 3). (The term "Plinian" is derived from Pliny the Younger who wrote a relatively objective account of the eruption of Vesuvius in A.D.79, the eruption which destroyed Pompeii and killed Pliny the Younger's uncle, Pliny the Elder).

Pyroclastic eruptions produce volcanic plumes which vary in size according to the violence of the eruption (Figures 2 and 3). Plinian eruption columns can reach heights of 50 km. or so and disperse over a radius of 500-5000 km. Great volumes of air are ingested by the plume which reduces its density to less than the surrounding air, enabling it to convect. Ultimately material within the plume reaches a point where it ceases to rise through the atmosphere and starts to fall, to be deposited as air fall tuff, (pyroclastic fall deposits - Figure 2)

In some circumstances a volcanic plume does not form, or cannot be maintained, and the volcano appears to "boil over". This occurs when, for example, a large volume of material is erupted over a short period of time, resulting in the density of the erupted material being

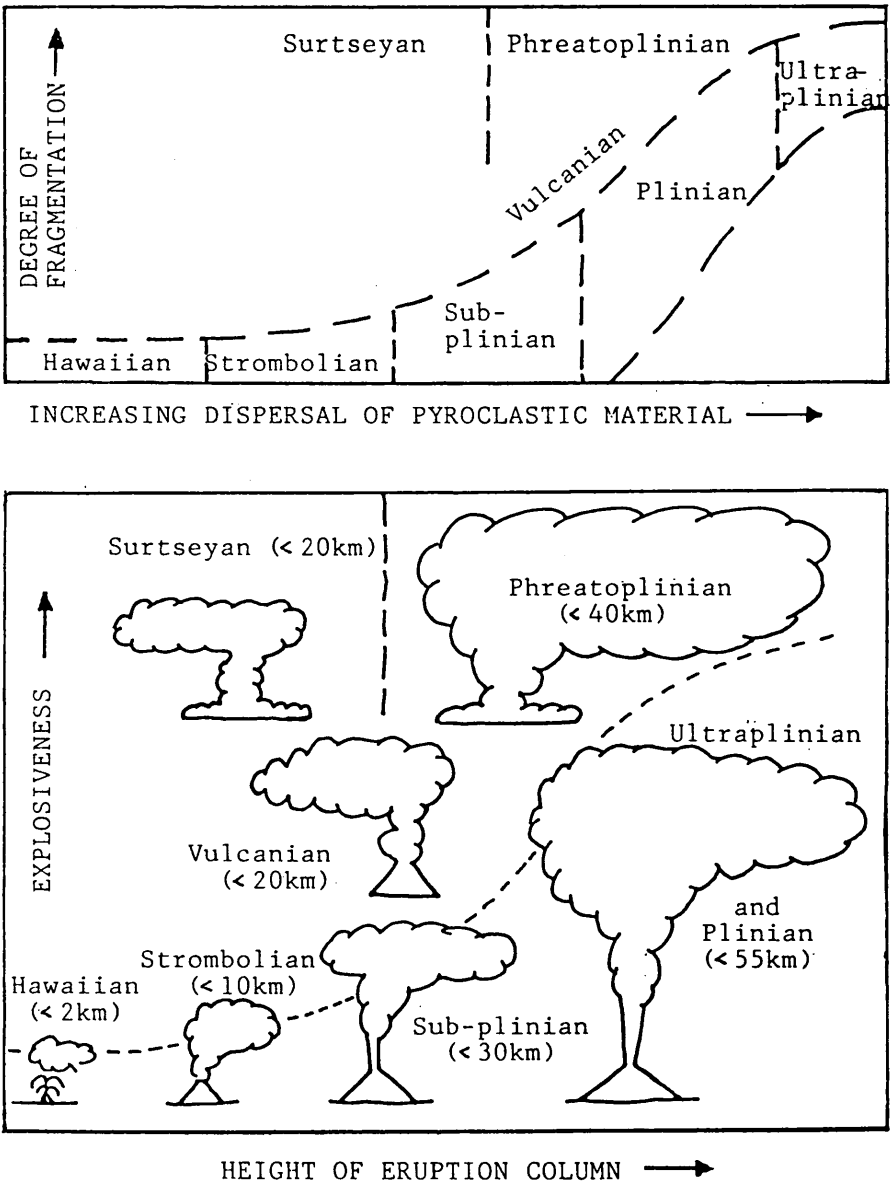


Fig. 3 Classification of explosive volcanic eruptions, using degree of fragmentation, explosiveness, height of eruption column and material dispersed as parameters (after Cas & Wright, 1987).

greater than the surrounding atmosphere. This style of eruption flows very quickly over the surrounding terrain as a low viscosity pyroclastic flow - also termed ignimbrite (Figure 2). Ignimbrites are a fluidised mixture of solid fragments, such as pumice, ash and crystals, and gas with a high ratio of solid to gas which are generally erupted at high temperatures (500-600 ^{0}C), temperatures high enough to allow the fragments within the deposit to weld together. When the ignimbrite finally comes to rest the sheer weight of the deposit may flatten many of the pumice fragments within the ignimbrite. These flattened fragments are termed "fiamme" and a welded ignimbrite with a moderate to high proportion of fiamme exhibits a "eutaxitic" fabric, which gives the rock a streaky appearance. Ignimbrites can cover large areas in a short period of time; for example, it is estimated that the 20,000 year old Taupo (New Zealand) ignimbrite eruption discharged 30 km^{3} of magma in 6-7 minutes and the flow took 12 minutes to travel some 80 km. from the vent (Wilson, 1986). Much smaller pyroclastic flows have led to high casualties, for example the 1902 eruptions of Mount Pelee (Martinique) and Soufriere (St.Vincent) killed over 30,000 people. Many ignimbrite eruptions result in large volumes of magma being evacuated from sub-volcanic magma chambers over a short period of time, a process which commonly leads to major subsidence and the resultant formation of volcano-tectonic faults and caldera structures. A very large scale example of this process is the Valles Caldera in the San Juan volcanic field, New Mexico. This caldera has a volume of some 330km^{3} which is equivalent to the volume of the pyroclastic flow ejecta forming the Lower Bandelier Tuff. More typically, eruptions involving the release of 25-50 km^{3} of magma lead to caldera collapse structures (Lipman,1984).

Pyroclastic surges are dilute, low particle density eruptions which are highly fluidised and turbulent and have a low solid to gas ratio. They form in a number of different ways but are known to be associated with pyroclastic flows, pyroclastic falls and water-magma inter-actions. Surges can underlie (or overlie) ignimbrites (Figure 2) and exhibit features such as low-angle cross bedding and accretionary lapilli tuff (these have a similar concentric

structure to ooliths and are formed by the repeated accretion of ash layers around a central nucleus).

3.3 Epiclastic deposits

Volcanoes are normally regions of positive topographic relief with an abundant supply of unconsolidated sediment. This makes them prone to sub-aerial (or sub-aqueous) erosive forces. Torrential downpours may accompany volcanic eruptions and lead to mud flows (lahars); over-steepening of slopes results in landslides, slumps and mass wastage. Rills and canyons are cut into the volcanic structures and erode, transport and re-deposit volcanic material within a developing river network. Ephemeral lakes may form within depressions surrounding the central cone. River systems draining into these lakes will produce alluvial fans while fine clays suspended in lake waters will slowly be deposited in quiet conditions. Depending upon the latitude and position relative to sea level, marine or glacial sedimentological processes may affect the volcano. In arid climates, wind erosion, transport and depositional processes will dominate.

During the time a volcano is active it will be affected by erosive and depositional processes punctuated by volcanic eruptions. This re-working of primary volcanic material leads to the formation of epiclastic deposits. Primary volcanic material survives to become part of the geological record only if it is produced in large enough volumes and/or it is capped by younger deposits which protect it from erosion. Volcanic deposits are well preserved in the marine environment, but sub-aerial volcanic deposits like the BVG have an extremely low preservation potential unless they are preserved within a subsiding area.

4. The Borrowdale Volcanics

The Borrowdale Volcanic Group is an example of a sequence of volcanic deposits which have survived the ravages of time. In broad terms they can be divided into lower and upper sub-groups which reflect different types of volcanic processes; the lower BVG is

dominated by the products of effusive volcanism and the upper BVG by the products of explosive volcanism. The lower and upper sub-groups, illustrated in Figure 4, will be referred to as "phase 1" and "phase 2" respectively.

4.1 A sub-aerial environment for the BVG

There is abundant evidence to suggest that the BVG were erupted in a sub-aerial environment. Negative evidence exists in the absence of marine sediments, pillow lavas and marine fossils, one exception being the Holehouse Gill Formation of Dunnerdale (Johnson, pers. comm.). Positive evidence is provided by the abundance of unconformities within the phase 2 succession, mantling and draping structures within air fall tuff sequences, common occurrences of fluviatile sediments, and the existence of pyroclastic surge and thin welded pyroclastic flow horizons (Branney, 1988a). Some tuff formations were however deposited sub-aqueously, probably in ephemeral lakes which formed in basins and topographic depressions within the volcanic field, for example parts of the Honister and Seathwaite Fell tuffs (Suthren, 1977).

4.2 Phase 1 BVG

The lower 2.5 km. or so of the stratigraphy is dominated by andesite lavas and sills with intervening bedded volcaniclastic sedimentary and primary tuff horizons. Individual lava flows are by their nature laterally impersistent, and their present-day morphology is dependent upon the volume and rate of lava erupted, their physical properties, the relief of the terrain over which they flowed (i.e. some lavas were channelised for part or all their length) and the effects of erosion. Thus the stratigraphy of the phase 1 BVG is highly variable, and any stratigraphy that is consistent over the small scale (e.g. the Wasdale area) may not be applicable over the whole of the lower BVG crop. This characteristic lack of widespread marker horizons is one of the major problems of working in ancient volcanic terrains like the BVG. As a result it is often difficult to establish even local correlations and in consequence many structural features may be missed. This has been one of the major stumbling

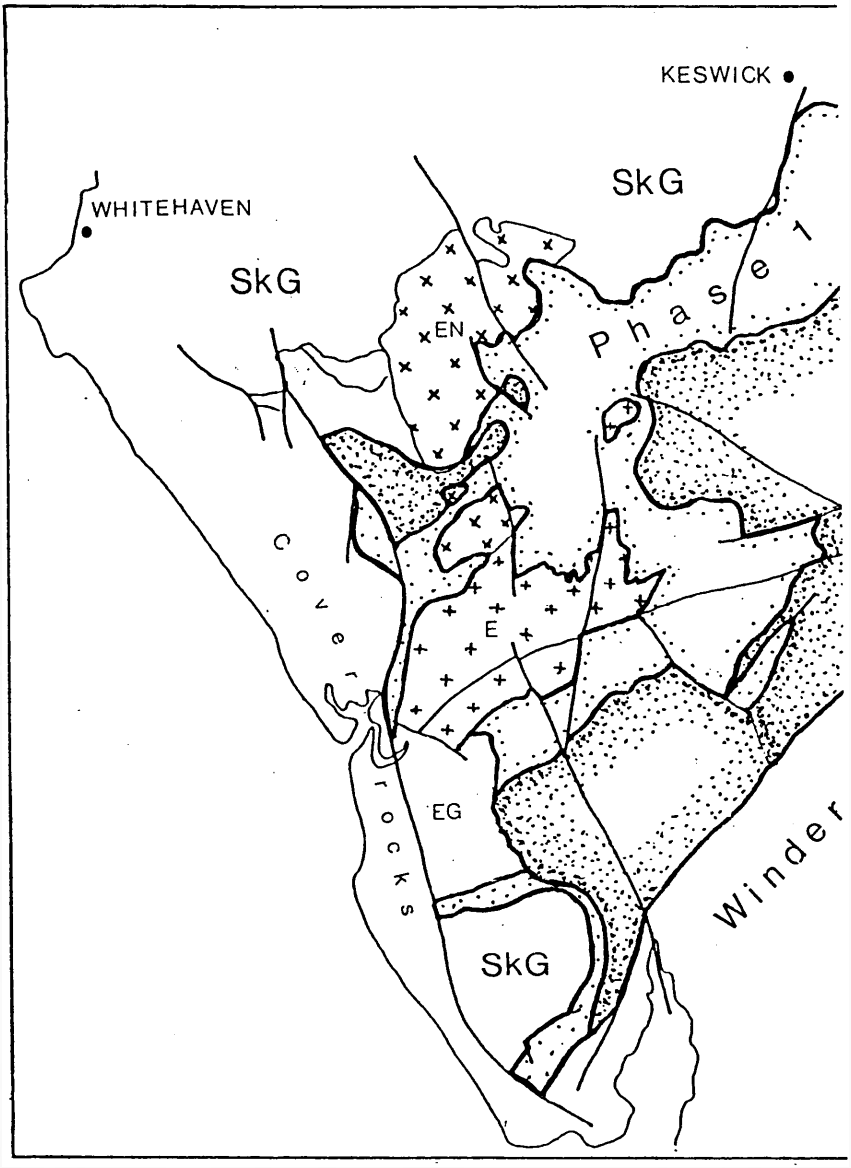
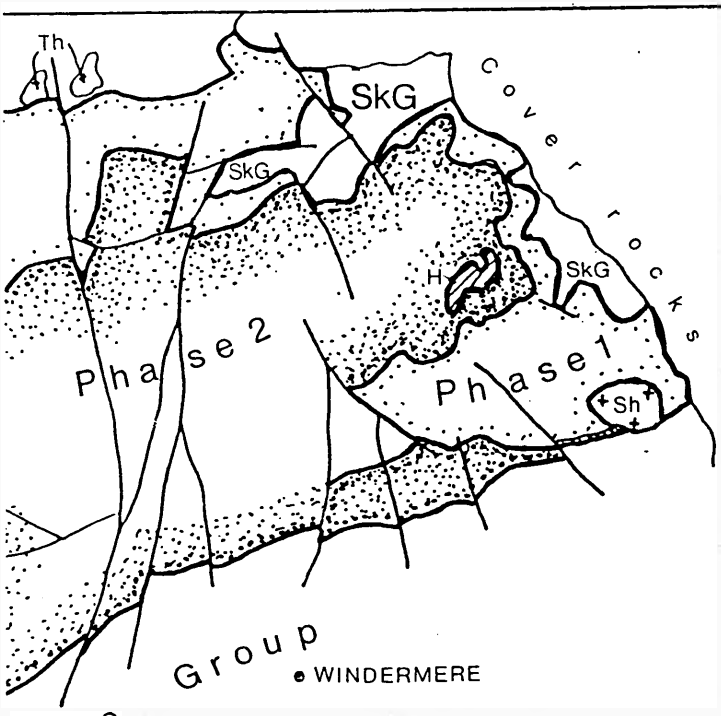


Fig. 4 Geological sketch map of the BVG showing the distribution of the lava-dominated Lower BVG (phase 1) and the pyroclastics-dominated Upper BVG (phase 2) - compiled by D. Millward.



Phase 2 } Borrowdale Volcanic Group
Phase 1 }

SkG Skiddaw Group

INTRUSIONS:

EN	Ennerdale Granophyre	E	Eskdale Granite
EG	Granodiorite	H	Haweswater Complex
Th	Threlkeld Microgranite	Sh	Shap Granite

blocks to our understanding of the BVG and the only way to resolve this is by a detailed and comprehensive survey.

Perhaps the most typical rock type of the phase 1 BVG is flow-brecciated andesite composed of clasts of all shapes and sizes, though usually angular, with volcaniclastic material occupying the interstices between the angular clasts. Classic examples of these rocks are located about $^{1}/_{2}$ km. north of Middle Fell summit, Wasdale (NY152,075). Figure 5 illustrates how these rocks were formed; as andesite lava was erupted the exposed surfaces cooled and froze but the interior of the flow remained liquid. This liquid interior continually over-rode the solidified edges, producing flow-brecciated lava margins. The centres of the flows are often homogeneous and exhibit structures such as flow joints and flow folds - formed as the lava cooled and solidified, with their orientations controlled by the final movements of the lava. Laminar flow produced joints parallel to the top and bottom of the flow while flow folds formed as a result of internal shearing within a viscous lava.

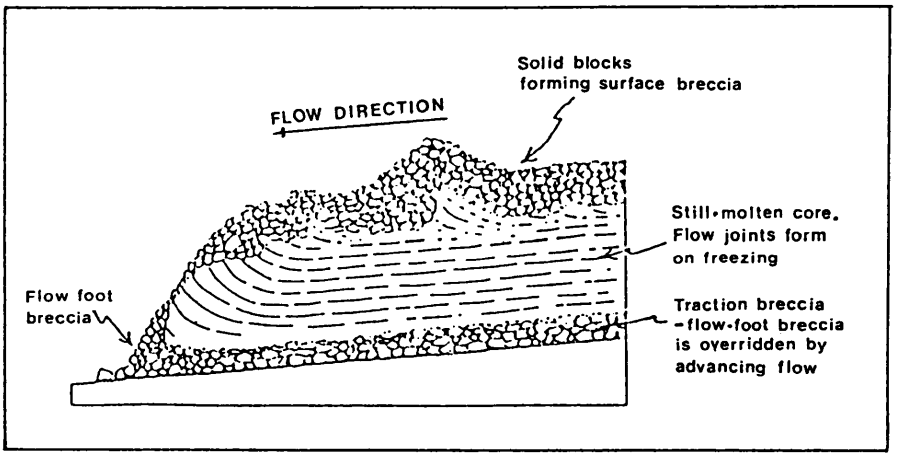


Fig. 5 Formation of flow-brecciated andesites. A still molten lava interior continuously disrupts and bulldozes through the solidified lava front and top surfaces. Flow joints form parting planes as the lava finally cools and solidifies.

Volcaniclastic sedimentary and primary airfall tuff sequences are locally important within the phase 1 BVG, for example the lacustrine sediments at Honister and a possible surviving sector of a tuff cone at Devoke Water (Millward, pers. comm.). Tuff and epiclastic sequences of varying thicknesses are interbedded with the lavas although it has recently been realised that some "lavas" are in fact

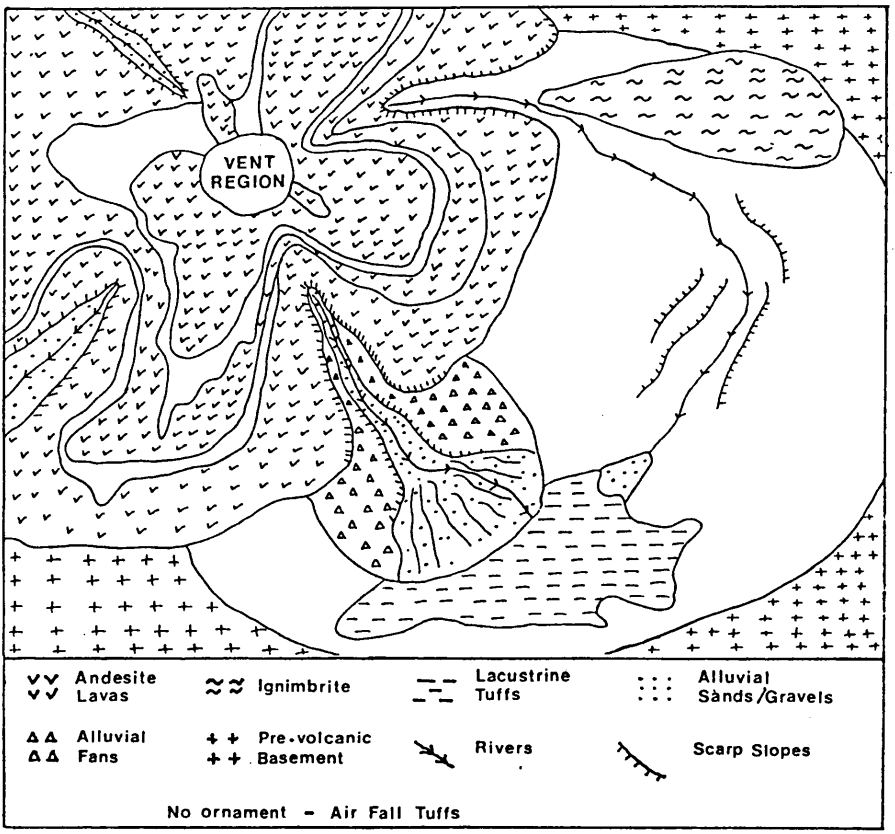


Fig. 6 Palaeo-geological reconstruction of phase 1 BVG time. A lava cone is built upon a pre-volcanic basement and comprises andesite lava with occasional air fall tuffs and ignimbrites. Streams deeply incise the volcanic cone and deposit alluvial sands, gravels, cobbles and boulders. Lacustrine tuffs are deposited within ephemeral lakes occupying depressions surrounding the volcano.

shallow sills, providing evidence that some magma did not reach the surface but was intruded into the recently formed volcanic field (Branney, 1988b). Intrusion as well as extrusion are common features of modern volcanic terrains and are observable, for example, within the caldera walls of Hawaiian volcanoes.

The proportions of the BVG formed by sills are the subject of present research. Packages of basalt and basaltic andesite occur within the phase 1 BVG and are characteristically highly amygdoidal with cindery/clinkery tops. Around the Throstle Garth, Upper Eskdale, compound aphyric basalt flows are exposed (Millward, pers. comm.); these are produced by medium to large volumes of magma erupted at a low effusion rate. In this situation numerous, small low-volume flows coalesce over a period of time to form a compound volcanic field, with discontinuous channels of lava in a mass of clinkery rubble.

Phase 1 BVG records the build-up of a volcanic field dominated by effusive volcanism, with an occasional burst of explosive volcanism. These deposits were probably derived from a number of centres within a multi-centred volcanic field. Branney (1988c) suggests that by the end of phase 1 BVG time the volcanic field was relatively low-lying. He has calculated that the Whorneyside Tuffs which overlie the lower BVG swamped and flattened a terrain with a palaeo-relief of only 80-100m. between the lowest and highest points. Figure 6 shows a possible palaeo-geographical /environmental map for phase 1 BVG time. A "lava-cone" was formed on a pre-volcanic basement and an early extensive air fall tuff was gradually covered by lavas, bedded tuffs and occasional thin ignimbrites. River systems drained the lava cone, forming steep scarp slopes and depositing alluvial sands and gravels. Fine material suspended in lakes which occupied the depressions around the volcano settled to form lacustrine tuffs.

4.3 Phase 2 BVG

Phase 2 BVG marked an abrupt change from effusive to explosive volcanism, the dominant products of which were thick ignimbrite

sheets and their associated pyroclastic surge deposits (Millward, 1979). Enormous volumes of magma were extruded over a relatively short period of time to form ignimbrite sequences such as the Airy's Bridge Formation (Oliver,1961), well exposedin the Sca Fell area, and stratigraphically higher formations such as the Lickle and Lag Bank ignimbrites which are exposed in the Coniston and south-western fells (Johnson and Millward, pers. comm.). The thick ignimbrites are not the result of one eruption but are composed of individual flow units which were erupted at different times, though probably with only a short time gap between eruptions (Davis, pers. comm.). The compositions of the pyroclastic flow deposits are dominantly rhyo-dacitic, though some are more basic in composition.

Associated with the ignimbrites are volcano-tectonic faults which formed contemporaneously with the eruption (Branney, 1988c). These faults caused the slumping of unconsolidated tuffs down palaeo-fault scarps, the steepening of eutaxitic fabrics close to fault planes and caused newly consolidated but still plastic ignimbrite sheets to move down fault scarps under gravity; this phenomena is known as "rheomorphism". Some faults cut lower parts of an ignimbrite but not upper parts, implying that they were active when the lower ignimbrites were erupted but not when the upper ones were erupted.

These volcano-tectonic faults are extensional "normal" faults and suggest that subsidence of the volcanic pile was closely linked to the eruption of large volumes of ignimbrites. Caldera collapse and subsidence is commonly associated with the eruption of large pyroclastic flows, recent examples including Crater Lake, Oregon, and Santorini in the Mediterranean (see also section 3.2). Current research is attempting to define the shape of the BVG caldera and evidence to date suggests that it is not a simple circular caldera but a complex network of graben or half-graben-type faults and down-sags.

Although the ignimbrites occupy much of the thickness of the upper BVG stratigraphy, they would have only occupied a small amount of the overall time span. Most of this time would have been occupied

by epiclastic processes, which resulted in the deposition of material within basins and topographic depressions containing ephemeral lakes. Primary pyroclastic air fall material would also have been deposited from time to time. One of the most important epiclastic sedimentary sequences within phase 2 BVG is the laminated slate belt of Langdale and Coniston. These sediments are of silt to sand grade and exhibit a variety of sedimentary structures, such as cross bedding, rip-up clasts and slump folds. The bulk of the sediments formed within a fluviatile to lacustrine environment. At least one sedimentary sequence, at Holehouse Gill, formed in a fully marine environment. These mudrocks contain an acritarch flora (Johnson, pers. comm.)

A series of basic andesite sills intruded the volcaniclastics of the upper phase 2 BVG in the Lickle valley area and andesite lavas and sills are common within the phase 2 of the Ulpha area. Thus basalt/basaltic andesite effusive volcanism was pene-contemporaneous, at least for part of the time, with the explosive rhyo-dacite ignimbrite eruptions (Millward and Johnson, pers. comm.).

5. Future research

This review of current research concerning the BVG has attempted to outline briefly some of the many new developments in the understanding of the Lakeland volcanics. A series of geological maps at varying scales and their related reports will be published in the short- to medium-term future. Volcanological research undertaken in conjunction with the geological surveys will hopefully bring new insights into the relative timing of the volcanic deposits, the "plumbing system" of the Ordovician volcano(es), the nature of the caldera(s) and the geochemical evolution of the BVG magmas. One of the advantages that areas such as the Lake District have over modern volcanoes is that the rocks are exposed to much deeper structural levels so that the "roots" of the volcano are visible. This is one of the reasons why the Borrowdale Volcanics will be of interest for many years to come.

Acknowledgements

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THE DEPOSITIONAL ENVIRONMENT AND THE DUNE MORPHOLOGY OF THE PENRITH SANDSTONE

F.J.Cockersole

1. Introduction

The Penrith Sandstone deposits were laid down about 260 million years ago in Lower Permian times, in a subsiding basin which was roughly coincidental with the present Vale of Eden (see Figure 2). This article is concerned with the central dune facies of the basin which represents the remains of a Sahara-like sand desert. The aim of the study is to attempt a reconstruction of the environment of deposition and an assessment of the dune morphology.

The sandstones have long been regarded as aeolian (wind-blown) in origin and were deposited in an arid desert environment, when the region lay between 10 and 30 degrees latitude north of the Equator. At that time the whole of what is now North-West Europe lay at the western edge of the huge super-continent of Pangaea. The deposits consist of thick, uniform, cross-bedded sandstones, representing the lower levels of eroded sand dune structures superimposed upon each other. Areas of fluviatile sandstones and polygenetic conglomerates, known locally as 'brockrams', are peripheral to the dune-bedded deposits and they are regarded as the remains of alluvial fans and piedmont breccias, formed mainly as a result of the rapid erosion by episodic flash floods of the mountainous basin rim. Although part of the Penrith Sandstone Formation, they will not be considered in this study.

The study is based upon field work and research carried out in 1986, which formed part of an Open University Fourth Level Project Course entitled 'Directed Studies in Earth Science'.

Table 1 - THE GENERALISED SUCCESSION IN THE VALE OF EDEN.

HORIZON		MAX. THICKNESS	ENVIRONMENT
TRIAS	St. Bees Sandstone	600 m.	shallow non-marine waters, seasonal braided rivers.
Upper PERMIAN (Zechstein)	Eden Shales	150 - 200 m.	playa & evaporating playa - red shales, evaporite beds, plant remains
Lower PERMIAN (Rotliegend)	Penrith Sandstone	400 m.	Dune sands with peripheral fluvial sands & conglomerates
CARBONIFEROUS	*unconformity*		

2. A general description and tectonic history of the area.

The generalised succession in the Vale of Eden is summarised in Table 1. The Penrith Sandstone today occupies a broad syncline and crops out on both sides of the Vale (Figures 1 and 2). The principal outcrop is on the western limb in a strip of territory approximately 40 km. from north to south and 9 km. from east to west.

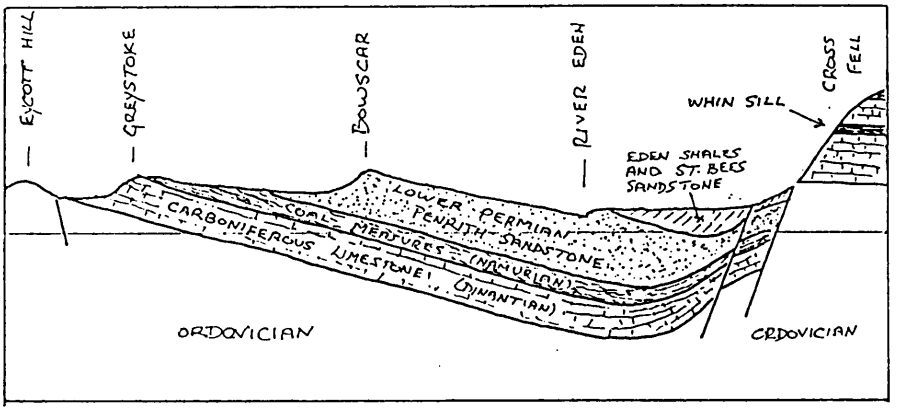


Fig. 1 Sketch cross-section, E - W across the Vale of Eden, approximate line of section shown on Fig. 2. (vertical scale is considerably exaggerated)

Much is drift covered but good exposures of the dune facies occur in the more elevated ground north and south of Lazonby Fell (515,395). (All Map References refer to the 'NY' 100 km. square of the British National Grid.)

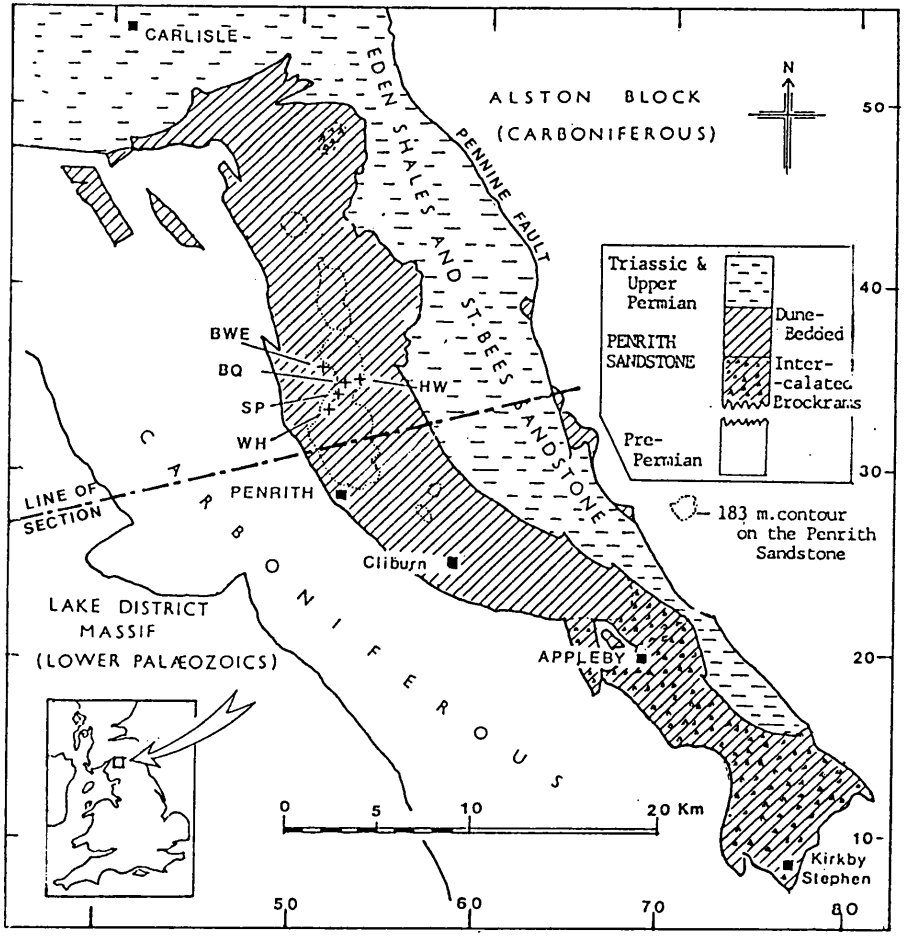


Fig. 2 Geological map of Permian and Triassic rocks in the Vale of Eden
Key to localities mentioned in the text :
HW: Halfwaywell Quarry. BQ: Bowscar Quarries.
BWE: Bowscar Wood Escarpment. SP: Sandstone Products Quarry. WH: Whelpdale Hill Quarry.

In these localities the sandstone is well cemented with quartz and the consequent cohesion of the stone gives it a degree of hardness and durability that has led to it being used extensively as a building material. There are many small exposures in quarries and trial pits and, although only one quarry in the area is working, these multiple exposures facilitate examination of the fossilised dune remnants and allow an attempt to be made to reconstruct the characteristics of the Permian dunefield.

Detailed mapping by British Geological Survey, supported by borehole evidence, shows that the maximum thickness of the deposit is about 400 metres. The probable original area of the basin of deposition is defined on the east by the Pennine Fault system, downthrowing to the west and active in Permian times. To the south the Howgill Fells and the high land of the Askrigg Block were, in part at least, the sources of the brockram clasts in the fan conglomerates. To the north the sandstones thin appreciably and were probably separated from the Carlisle depositional basin. On the west the extent of the basin is unknown but is likely to have been limited by the Lower Palaeozoic mountains of the central Lake District, which were almost certainly an area of positive topography in Permian times (Kimber and Johnson 1986). A clue to the extension of the Permian strata up to the Lake District massif may be provided by the reddening of the underlying Carboniferous beds. The original basin was therefore probably about 50 km. by 30 km.

The formation of the basin can be related to later Variscan extensional movements with renewed down-faulting along the Pennine Fault Line (Figure 1). Similar extensional fault movements in the Mesozoic, for example in the Cheshire Basin, have been ascribed to the opening of the Atlantic Ocean.

The Penrith Sandstone is virtually unfossiliferous and its Permian age can only be deduced from examination of the strata above and below it. The dune deposits lie unconformably upon deeply eroded Dinantian, Namurian and Westphalian strata, so that a considerable period of erosion must have followed the late Carboniferous. The Eden Shales, with the Hilton Plant Bed near its base, lie immediately above

and may be correlated with the lower part of the Upper Permian. The allocation to the Lower Permian is also made on the basis of lithological and depositional analogies with other Lower Permian sandstones which form the Rotleigend Formation of N.W. Europe (Glennie1984). The evidence suggests that the Penrith Sandstone accumulated over a period of about 40 million years between 280 and 240 million years ago. Although unfossiliferous in the ordinary sense, trace fossils in the form of vertebrate footprints have been reported on at least three occasions. Sargeant (1982), in these Proceedings, has suggested that they belong to a certain genus of caseasaur or giant lizard, whose footprints have been widely identified in later early Permian strata in Europe and N. America.

The lack of plant remains and the evidence of the largely aeolian nature of the sandstones indicate an intensely arid climate. Present-day deserts of this nature occur between the latitudes 10 and 30 degrees north and south of the Equator where aridity is a function of the atmospheric circulation. Coriolis forces, due to the circulation of the earth, operate on the circulating air systems to give rise to the Trade Winds north and south of the Equator. The N.W. Trade Winds north of the Equator become increasingly dessicating as they traverse continental land masses, especially towards the western continental margins (Glennie 1970). These circumstances would apply to the area that is now N.W. Europe when it formed part of the super-continent of Pangaea, which had drifted north across the Equator and into the Trade Wind regimes by Permian times. Reconstructions, based upon palaeomagnetic data and deduced polar wandering curves, indicate that Britain by Lower Permian times had reached a latitude of about 10 degrees north.

3. Criteria for aeolian deposition in the Penrith Sandstone

To reconstruct the palaeo-environment it is essential to establish that the Penrith Sandstone is indeed aeolian in origin. This is nowadays generally accepted but it is worth while to review briefly the evidence in support of this hypothesis.

1. Large scale dune bedding is seen uniformly in outcrops and excavations all over the area. The thick, well-sorted foresets and, particularly, the upward increase in inclination of foreset laminae to 20 or 25 degrees before truncation (Glennie 1983), are typical of present day aeolian structures that have been studied (McKee 1966, Brookfield 1979, and Wilson 1972).

2. Petrographic evidence from thin-section microscopic examinations shows large, well-rounded quartz grains (known as "millet seed" grains or "wind" grains) with frosted surfaces from abrasion. Quartz grains constitute 95% of the rock and grain-size analyses reveal a median grain size of about 0.4 mm. with a very high degree of sorting and a size distribution pattern typical of known aeolian sands. In addition, Waugh (1965) has shown that characteristic Scanning Electron Microscope appearances are present - abrasion blocks and conchoidal fractures rounded by abrasion intersect to give a pitted surface, (these features can be reproduced experimentally and are also seen in modern aeolian sand particles).

3. The facies associations also provide evidence in that the Penrith Sandstone is succeeded conformably by the Eden Shales, which include evaporite horizons and evidence of marine incursions in a tropical context, such as dolomitisation, red shale sequences, salt pseudomorphs, dessication cracks and mud-flakes.

4. The complete absence of fossils, apart from the vertebrate footprints already mentioned, denotes the arid terrestrial nature of the environment.

4. The general stratigraphic features.

Wherever the Penrith Sandstone is exposed in the central basinal area, it is characterised by large scale dune bedding. Planar and wedge-planar sets of cross-bedded foreset laminae are seen lying upon, and sharply cut-off above, by low-angle planar bounding surfaces, described by McKee (1953) and Brookfield (1979) among others, Figure 3. These foreset laminae are the preserved roots of dune structures, sandwiched between the eroded remnants of a dune below and the base of the next dune above. The features can be seen in almost every exposure, whatever its orientation with the deduced

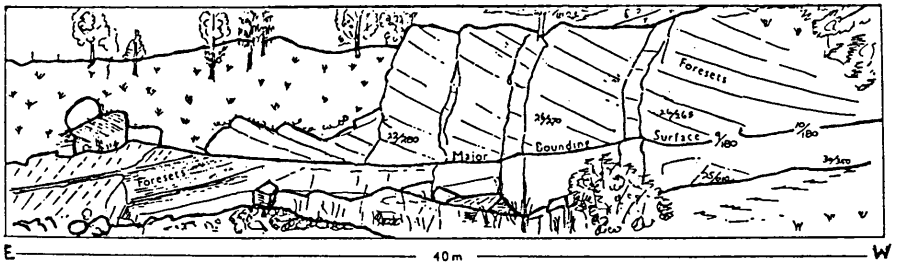


Fig. 3 Explanatory sketch of a face at BOWSCAR QUARRIES (520,343). The basal foresets of a large dune taper away asymptotically to the west & lie above a major bounding surface. The dipping planes of the truncated foresets of another dune are seen below the bounding surface at the bottom left of the section. Differing dips & dip directions indicated by convention, e.g. 26/270.

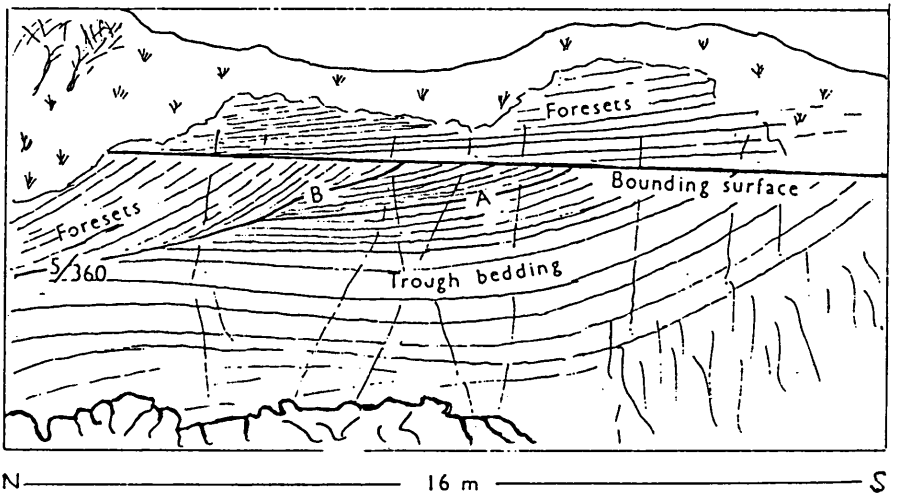


Fig. 4 Explanatory sketch of WHELPDALE HILL QUARRY, (517,338). Quarry face aligned approx. N - S and oblique to palaeo-wind direction. Thin-bedded, asymptotic foreset apron deposits are seen above a major bounding surface which truncates more steeply dipping foresets below. These show large scale, shallow trough cross-bedding. 'A' & 'B' are possible examples of subordinate bounding surfaces (reactivation surfaces).

palaeo-wind direction. Exposures transverse to the palaeo-wind direction, in addition, show examples of large scale, shallow trough cross-bedding (Figure 4). Localities where these typical features can be seen include Halfwaywell Quarry (533,352), Bowscar Quarry (520,343) and Whelpdale Hill Quarry (517,338), all located on Figure 2. The reader is referred to an excursion note for the Penrith Sandstone in the Proceedings (Vol.5, Part 1, p.86), where various exposures are described.

Another characteristic feature is the occurrence of extensive, regular, flat-bedded sandstone laminations between 20 and 100 mm. in thickness. Such strata lie immediately above one of the near horizontal bounding surfaces - good examples can be seen at Bowscar Quarry (520,343) and at the 'Sandstone Products' Quarry (526,348). Where exposures are adequate such flat-bedded deposits, and the subjacent bounding surface, are seen to lie immediately above steeply dipping, truncated foresets. These flat-bedded deposits have been extensively exploited as they supply convenient rectangular blocks and slabs for building purposes. In a few localities such deposits can be traced laterally for sufficient distances to establish that they thicken and assume progressively steeper angles of dip until they become typical dune foresets. It may be concluded therefore that these flat beds represent apron deposits at the foot of the lee-side dune slip-faces and their extent must be considered in assessments of dune size (Figures 4 and 6).

5. Interpretation of the stratigraphy

In the last fifteen to twenty years knowledge of depositional processes in aeolian deserts has become more scientifically based as a result of work on modern and ancient dune systems by McKee, (1966), Hunter, (1977), Wilson, (1972), Brookfield, (1977) and others. Bagnold, (1941) laid the foundations with his analysis of the physics of aeolian particle movement.

In continental environments aeolian sand accumulates in topographic basins, such as the Danakil Depression in Egypt today, or in large intra-cratonic areas, of which the Sahara is an example, to

form 'sand seas' or 'ergs'. These accumulations can vary greatly in size, from a few hundred to several thousand square kilometres in extent. The extent of the Permian 'sand sea' in the Vale of Eden was about 1500 km^2 representing a relatively small area of sand desert.

In the Vale of Eden two basic bedforms can be recognised - sand ripples and dunes. Sand ripples in modern deposits are commonly seen on the lee faces of dunes and in inter-dune areas; ripples do occur in the Penrith Sandstone but are not particularly well preserved. The characteristic wavelength is about 100 to 150 mm., which probably relates to the average distance that medium to large sand particles bounce or jump along the surface in the direction of the prevailing wind. This is the process of saltation, one of the mechanisms by which mass migration of sand occurs downwind.

The most prominent bedform to be seen are the dune remnants. In modern deserts dunes migrate down-wind at rates dependent upon the wind strength and frequency and on the grain size of the aeolian material. The migration process results in the movement of sand particles from the windward or stoss slope of the dune, over the crest and down the steeper lee slope. This occurs by saltation, as described above, but an additional important process is that of avalanching down the lee slope. The movement of particles up the stoss slope results in over-steepening of the lee slope which becomes unstable and avalanches down until the natural angle of rest of dry sand in air is restored. This process of episodic avalanching is probably the basis of the formation of the well-defined foreset laminae seen in the exposures, while the planar bounding surfaces are formed by the mass down-wind movement of the dune over the substrate.

McKee (1966) and others have shown that modern dunes, when sectioned, have a structure of cross-bedded foresets sandwiched between bounding surfaces of different orders. It has been postulated that these well-defined planar features represent erosion surfaces over which the dunes migrate. Thus in aggrading conditions, dunes will migrate over the roots of pre-existing dunes which have been eroded to form the substrate.

Certainly in nearly all the significant exposures of the Penrith Sandstone it is possible to identify deposits made up of dune remnants piled up on top of each other. The dune foresets flatten out asymptotically* at what would be the foot of the steep lee slope of the dune while steepening dips of 25 - 30 degrees, prior to truncation by erosion at the bounding surface above, are approaching the angle of repose of dry quartz sand in air (about 34 degrees). It should be remembered that a correction for a regional post-depositional tilt of 2 - 3 degrees to the N.E. would add a degree or two to most of these dip values and the effects of post-depositional compaction will also have slightly reduced the dip inclinations. These considerations point to lee-slope avalanching and a grain-flow mechanism in the formation of the foresets, supported by the absence of graded beds in the sets.

Different authors have attempted to classify bounding surfaces into various orders and groups, but in the Penrith Sandstone only two orders are clearly recognisable :

a) the low angle planar surfaces separating dune remnants (Figures 3 and 4) and supposedly related to dune migration;
b) a second and lower order of bounding surfaces which separate sets within groups of foresets and which are considered to be reactivation surfaces arising from minor wind shifts between periods of sand deposition (see Figure 4).

A feature already discussed is the ubiquitous occurrence of the asymptotic lower foresets forming almost horizontal laminations of 50 - 100 mm. in thickness. These must have formed extensive aprons at the foot of dune lee-faces, but they seem to have received rather scant attention in the lit"- - flattening of high angle foresets near their bases to form tangents with the flat, silty sub-dune floor", but the laminations he illustrates are very small scale structures compared with those in the Penrith Sandstone. Avalanching material is likely to have spread out to form an apron on reaching the sub-dune floor, although the concave upward profile has been attributed by

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* Asymptote : A curved line converging with but never meeting a straight line.

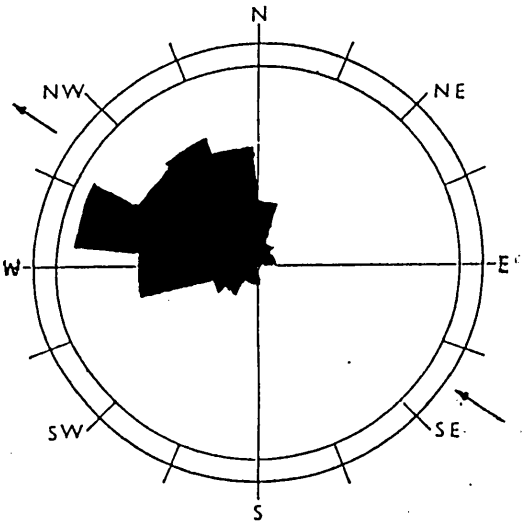


Fig.5
Palaeo-wind direction from foreset dip alignments. A rose diagram constructed from 171 foreset dip direction measurements from a variety of exposures, corrected for magnetic variation. The predominant sand- transporting wind direction is on a vector of 304 degrees (present day azimuth); i.e. E.S.E. to S.E. wind as indicated by the arrows.

Hunter, (1981) to the scouring effects of lee-face eddy currents which have been observed in coastal dune environments in S.Africa. It is also possible that such eddy currents could be responsible for creating an excoriated planar surface in advance of the migrating dune, thus forming the well-defined bounding surface which sharply cuts off the foresets of the dune below.

6. A reconstruction of the desert environment

6.1. The palaeo-wind

Since dunes form and migrate under the influence of the wind, a study of their attitudes, dimensions and morphology should provide evidence of the direction and strength of the palaeo-wind. The azimuths of the dip slopes of the foresets have been used in this respect. However, averages of a large number of observations must be

used because not all dunes will have been formed in winds blowing consistently in one direction while some types of dunes will have curved crest lines, resulting in the avalanching foresets dipping over a range of azimuth directions. The total number of readings of the dip directions from foresets in the Penrith Sandstone, recorded at exposures in the area, has been consolidated and plotted in the form of a wind rose diagram and an example is shown in Figure 5. This indicates a palaeo-wind blowing from a direction between E.S.E and S.E. in terms of the present day compass.

Another indication of the palaeo-wind direction, and perhaps the most accurate, has been proposed to the author by D.B.Smith, who suggested that the trough-bedded structures, which are also seen in present-day deserts, are formed by the scouring action of the wind; hence the trough axes will represent the formative wind direction. In the available exposures of the Penrith Sandstone however, only one cross-section of such structures in an exposure is available, whereas two or more would be needed for precise measurements of the azimuth of the trough axis. However, it can be said that such structures as can be examined are in approximately the expected orientation.

The Penrith Sandstone is one of a number of Lower Permian aeolian sandstone deposits in Britain with other basins occurring in Southern Scotland, Durham, Cheshire, the West Midlands and South Devon as well as the North Sea basins. Palaeo-wind directions have been determined for all these localities and the predominant direction is from the east on present-day azimuths. Palaeo-magnetic reconstructions (Scotese 1979; Van de Voo et al. 1974 and Briden et al. 1973) indicate that in addition to drifting north since Permian times, Britain and N.W. Europe have undergone a certain amount of anti-clockwise rotation, variously assessed at between 20 and 45 degrees. This would convert the easterly palaeo-wind direction into a N.E. or N.N.E. direction, which is what would be predicted in an area between 10 and 30 degrees north of the Equator in the Trade Wind belt.

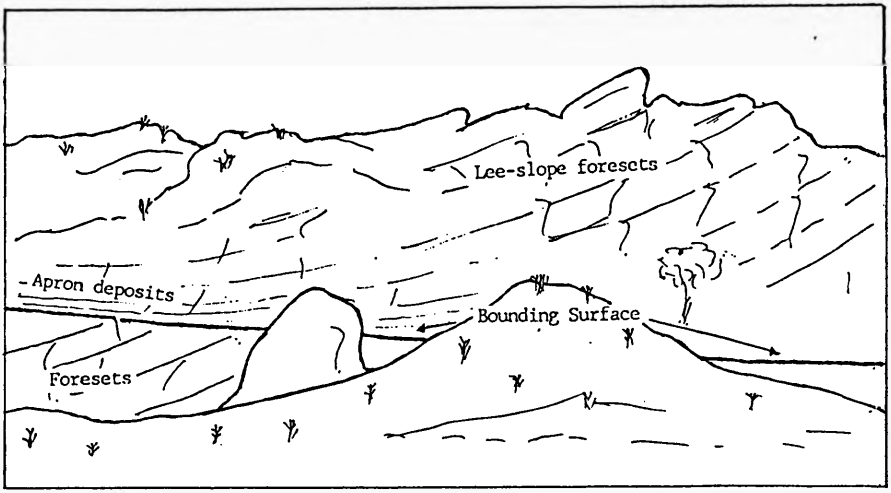


Fig. 6 Sketch of dune feature exposed in scarp face at BOWSCAR WOOD (512,345), approximately 1 km. north of Bowscar. (see explanatory diagram below)

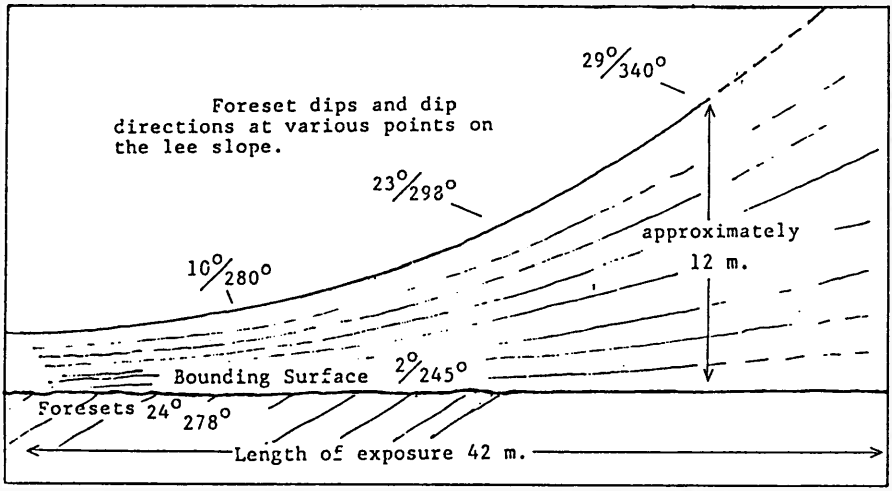


Fig. 7 Diagram constructed from measurements taken on large dune in Fig. 6 above. The orientation of the scarp face is N.N.E.-S.S.W. The foresets beneath the bounding surface are up to 400mm. thick. Note the swing of dip directions up the lee slope of the dune.

6.2. Dune size and morphology

There are a number of localities where an appreciable extent of fossilised dune foundation is exposed, so that by a process of extrapolation, an estimate can be made of the original size of the dune. One example is the exposure along the crest of Bowscar Wood (512,354), sketched in Figure 6 and with an accompanying explanatory diagram, Figure 7. Accepting a figure of about 34 degrees for the angle of repose of quartz sand in air, the steepening upward foresets forming the erstwhile lee-face of the dune can be projected upwards to give a conservative estimate of about 25 metres for the crest height of the dune. Assuming a lee-slope to stoss-slope ratio of 1 to 4 and a ripple index (length/height ratio) of say 8, this gives a dune length of about 200 metres. A similar exercise can be carried out on the dune shown in Figure 3.

Two other methods of assessing dune size have been proposed. Wilson (1972), working on modern sediments, related bed-form size to median-grain size, which is likely to be a valid concept if different bed-forms are the result of wind regimes of different strength. Secondly, foreset thickness has been related to dune height by Kocurek and Dott (1981), again based on contemporary observations. Neither of these methods is as likely to be as good an indication of dune size as the field observations already quoted, but application of these authors' findings to the Penrith Sandstone does give values consistent with the field evidence.

We can now turn to the actual type of dune that existed in the Penrith desert basin. On the basis of present-day observations, dunes are generally classified as longitudinal or transverse to the predominant wind direction. Longitudinal, or 'seif', dunes may be straight or sinuous and are associated with bimodal or seasonally varying winds. Transverse dunes can be of the individual 'barchan' type, with concave lee-side faces, or may be continuous 'akle" type dunes with a sinuous crest line; such dunes are related to predominantly unidirectional winds. More complex patterns, such as star-shaped 'rhourds' are attributed to widely varying wind directions.

Shotton (1956) was the first to show that measurements of the lee-slope dip-directions of a single dune structure could show a regular variation each side of an axis, indicating a curving crest-line. The only remaining working quarry in the Penrith Sandstone at Halfwaywell (533,352), 5 km. N.W. of Penrith, seemed in 1986 to be excavating a single dune structure and provided a unique opportunity for studying the pattern of variations in the directions of dip within a single group of foresets. This clearly indicated a crescentic lee-face to the dune, facing approximately north-west. From this type of evidence, Shotton and later Waugh (1967) concluded that the Permian sand sea in the Vale of Eden was composed of barchan dunes and this has never been challenged.

The evidence for barchan dunes in the Penrith Sandstone is based on the range of foreset dip directions recorded and the contention that certain exposures reveal the concave sweep of a barchan front slope within a single set of cross strata, as described at Halfwaywell Quarry. However, Cooke and Warren (1973) and many other authors say that barchan dunes build up on hard desert surfaces with sparse supplies of sand. In the Penrith Sandstone the thickness of the deposits, the upward sequence of dune upon dune and the apparent absence of inter-dune areas point to a continually aggrading basin with plentiful supplies of sand.

The author's observations are consistent in that the range and direction of the foreset dips indicates a unimodal palaeo-wind and hence the likelihood of some form of transverse dune. The range of dip measurements also suggests curving crest lines. The exposures indicate dune structures adjacent to each other and piled one on top of another. Cross-sections of modern barchanoid dunes, cut at right angles to the predominant wind direction, show large scale, concave-upward, trough cross-bedding (McKee 1966). Such structures are present in the Penrith Sandstone in the occasional north-south orientated quarry faces, as at Whelpdale Hill and Halfwaywell. Collinson (1985) concluded that trough-pattern bounding surfaces could reflect the sinuous nature of the crests of akle'-type transverse dunes which seem to dominate deserts with complete sand cover.

The evidence therefore seems to favour the concept of the Permian sand sea in the Vale of Eden area being composed of large transverse, sinuously crested akle'-type dunes, 25 metres or so in height and some 200 metres in length, with little or no exposed inter-dune substrate.

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ON THE LAST GLACIATION OF CUMBRIA

R.Clark

1. Introduction

Glaciation in Cumbria continues to stimulate curiosity as the considerable number of studies published in the last twenty-five years demonstrates. Recent work, both theoretical and field-based, provides new perspectives for interpreting Cumbrian circumstances. The intention of this paper is to discuss evidence and conclusions from Cumbrian studies in the light of that recent work and of current fieldwork. One particular aim is to develop a time framework which conveys some sense of dynamic progression as the local glaciation waxed and waned. A second aim is to recognise some of the still unresolved questions posed by the local glaciation.

Much of the relevant data about glaciation are from surface materials and landforms not everywhere of the same age but largely relating to late stages in the glaciation. Significant stratigraphic data are sparse and localised; long distance correlations between sites are few. Consequently any review of the glaciation is necessarily provisional as the variety of outstanding problems shows. Cumbria's regional setting was fundamental to the course of its glaciation and is discussed first, followed by consideration of the growth and movements of ice to the glacial maximum. Then aspects of glaciation during and after drumlin formation are examined.

2. The last Cumbrian glaciation

The last glaciation, the subject of the present discussion, is now securely placed in the Upper or Late part of the Devensian, itself the last division of the Upper Pleistocene. That glaciation is thought to have commenced 26,000 - 25,000 years ago, with Cumbria being

substantially free of ice by 14,500 years ago (Pennington 1977); hence the local glaciation took place within a span of 11,000 -12,000 years. No later events are to be considered. The amount of erosion and deposition in that time testifies to the vigour of glacial processes but the rates of landscape change were not uniform either in time or from place to place. Nor was the glaciation 'symmetrical' - margins of ice streams did not retreat in orderly fashion to the hill areas where glaciation was initiated and ice flow directions were not constant everywhere. Dissipation of the ice cover took less time than its accumulation and the work of meltwater was, in all probability, of greater significance in the shorter waning than in the waxing stage.

Problems in interpreting field evidence have long been apparent and the drumlins of the Vale of Eden provide a case in point. Drumlins are elongated ridges, usually of glacial sediments and often grouped in tracts or 'fields' of ridges. They are shown in so many instances to lie parallel with ice movement at the time of formation that they are used as ice direction indicators. Hollingworth (1931) noted evidence for more than one movement direction from drumlins in the Vale of Eden, explaining the circumstance by postulating complex contemporary ice motion rather than successive movements as the conditions in the ice cover changed (Hollingworth 1931, King 1976). Mapped compilations of ice movements have also conjoined diachronous data (Greig et al 1971, Taylor et al 1971, IGS 1977), for example in showing quite parallel ice flows for 40 km. along the Solway lowlands, to the east on the northern side but to the west on the southern. Parallel movements in opposite directions are also shown for the Vale of Eden (Taylor et al 1971), an inheritance from Hollingworth's work.

Glacifluvial deposits (sands and gravels) with form and structure largely unmodified can occur in sheet form (e.g. kame terraces) or in mounds (as kames and eskers), in places superimposed on drumlins and elsewhere outside the drumlin tracts. These could be accumulated beyond ice margins (as proglacial outwash), at ice margins or on, within and under immobile, unerosive ice. Such deposits can tell of changing conditions at times of deposition, as in the Brampton area (Huddart 1981) where thinning ice influenced manner and height of

accumulation as and after meltwater ceased to flow east through the Tyne Gap.

Middle Sands, i.e. beds of sand and gravel between upper and lower boulder clays or tills, have been central to interpretations of glacial events. They are widely exposed in south Lancashire and Cheshire where initially the tripartite succession was related to a single glaciation but later persistently regarded as evidence for two glaciations. Quite recently the succession, now termed the Stockport Formation, has been reattributed wholly to the Late Devensian glaciation (Worsley 1967). Middle Sands also occur in north Cumbria where they have been interpreted as indicating multiple glaciation but correspondence of age and environmental conditions between these and those further south have not been established. The upper till of the tripartite succession on the Solway plain was attributed by Trotter (1929) to a late ice incursion into north Cumbria from southern Scotland (Scottish Readvance). The lower till was regarded as the product of the Main Lake District glaciation, local representative of the Newer Drift Glaciation (now Late Devensian). Penny (1964), reviewing the history of the last glaciation, remarked that the upper till in north Cumbria contained local and reincorporated material but otherwise only southern Scottish erratics; its origin in a Scottish readvance being proved by the fact that it contorted the underlying Middle Sands. He added that "it is worth noting that there is no terminal moraine marking the limit of this Scottish readvance", the till sheet only having a feather-edge. He thus signified a short duration or lack of vigour in the proposed readvance. Huddart (1971,1972 and Huddart et al 1977) presented a variety of evidence in advocating that a readvance had affected the coastlands of west and north Cumbria though Pennington (1978) and Thomas (1985) doubted the reality of such a readvance. Thus the status of a readvance remains problematic and the north Cumbrian Middle Sands have received little attention since their description in Geological Survey memoirs (Dixon et al 1926, Eastwood et al 1968).

Another unresolved question centres on the corries or cirques of the higher fells. Corries are recesses more or less deeply eroded into ridges and summits; some hold tarns and most are located on north or

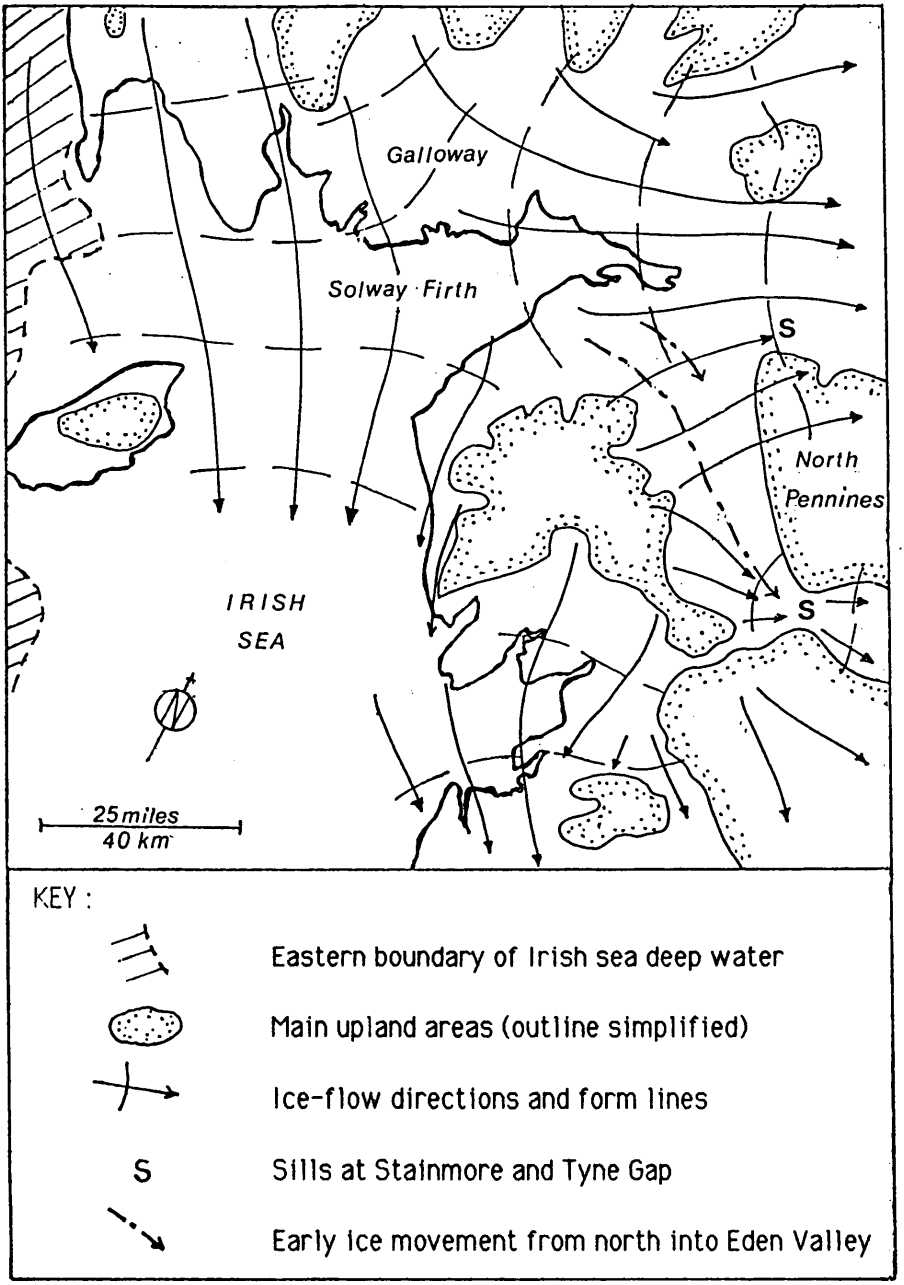


FIGURE 1 Ice movements at maximum glaciation in Cumbria.

east facing fellsides. Their development and distribution has recently been analysed (Clough 1977): their significance in the progression of a full glaciation, still unclear, is discussed below.

3. Regional and local contexts

During the waxing and waning of the glaciation in Cumbria the regional pattern of upland and lowland influenced the disposition and mobility of ice masses, illustrated in Figure 1. It determined the complexity of events in north Cumbria and their relative simplicity in the south. The outstanding facility of the southwest Scottish Highlands to generate snowfall gave an early dominance to ice entering the Irish Sea basin through the North Channel. Loading due to accumulated ice depressed the crust according to ice thickness and crust characteristics. At the same time global accumulation of ice lowered sea level. The direction of the two processes were one to keep the sea bed in contact with the sea, the other to dry out the sea bed. The most ice loaded area within the region was the north Irish Sea basin. Nevertheless Irish Sea ice is generally regarded as being mainly land-based during its fuller development. That circumstance would be achieved earlier and sustained longer in the shallow eastern part of the basin than in the lower western, with potentially significant consequences for the manner and rate of deglaciation.

The extent and form of high ground in Galloway and the central Southern Uplands made those areas more effective in accumulating ice than the Lake District hills. In the fullness of glaciation ice entering the Irish Sea basin from Galloway was restricted by Highland ice to courses east of a Luce Bay - Isle of Man line, with ice east of an ice culmination possibly in the Castle Douglas area forced to move east across the Nith, Annan, Esk and Liddel valleys and over the Solway Plain. It entered the catchment of the Tyne along a front from the north edge of the Pennines to the head of the North Tyne valley, where it was contiguous with ice over Teviotdale.

Position and height ensured high precipitation in the Lake District although the upland area is small with closely spaced exits and few high, broad-backed ridges; in contrast the north Pennines have a more

efficient topography for ice accumulation though lower precipitation and both areas nourished ice covers. Ice leaving the western side of the Lake District was held close to the coastlands of Cumbria by powerful off-shore ice masses in their passage south while ice from southern Scotland impinged on the western and northern fringes of the Lake District but did not extend into the upland (Eastwood et al 1968, Huddart 1971, Huddart et al 1977). Some ice from Scotland and the Lake District, diverted by the east-directed Scottish ice in the Solway plain, passed up the Eden Valley over Stainmore and into the Tees catchment. The contribution of ice from the eastern Lake District is shown by the distribution of Shap erratics as far as the east coast of England. The sill at Stainmore, a rise of some 300 m., would have made eastward passage of ice particularly susceptible to change in ice thickness and surface slope of ice in the Eden Valley. Initially the sill would have facilitated accumulation of ice in the vale. The question whether lower but narrower exits south from the vale to the Rawthey valley and Wensleydale were much used as iceways requires further study.

At the time of maximum ice accumulation, eastern Lake District ice, constrained by Scottish ice to the north and ice congestion in the upper Eden valley, moved across the Vale and over the Pennines north of Hartside, as well as through the Tyne gap. The general height of ice bed over hillcrests north of Hartside, in terms of present sea level, was 550 - 600 m. with the ice surface over the ridge at least 650 m., probably more. Ice surface height over the Vale of Eden would not have been significantly less than 700 m., with ice thickness of about 600 m. On the north eastern fringes of the Lake District ice surface heights would have been well over 700 m. with consequent implications for surface heights over the central fells. Ice surfaces heights and thicknesses in the western Solway area would have been greater than in the Eden valley and those in the northern Irish Sea basin still greater. It becomes necessary to suppose that snow accumulation zones would extend onto ice surfaces beyond the initiating hills as ice thicknesses built up, no longer dependent only on snowfall over the initiating mountains. Any extension of the accumulation zone beyond the uplands, if accompanied or followed by lowering of the "snow line", would favour further expansion of the ice

sheets. It would, however, leave the ice sheets very susceptible to rapid decline when the snow line eventually rose.

As ice built up round the Lake District, outflow down the dales would have been inhibited so fostering ice accumulation within the uplands. This buttressing effect, significant to vigour of ice erosion in the dales, was extended further south when south-flowing Irish Sea ice expanded on passing the St. Bees promontory. Consequently ice from the southern Lake District moved east of south impinging on the Fylde (Longworth 1985) with some ice in the Lune valley being diverted east between Pennines and Bowland Fells into the middle Ribble valley.

The development of ice cover to a maximum state produced a corresponding distribution of transported rocks, in places reflecting sequences of ice movement as ice masses coalesced and grew. Readily recognised indicator rocks, particularly igneous types - Criffel, Shap, Threlkeld, Eskdale, Armboth dyke and Carrock Fell rocks - have been used to plot the movement of ice but as well as the changes in movement during the waxing glaciation, the effects of change during decline need to be taken into account. Western Lake District granitic rocks, together with those from Galloway, are commonplace in the drifts of south Lancashire, Cheshire and Staffordshire. Shap granite boulders not only reached the coast of north-east England but were also carried into the Vale of York. Identifiable erratics, the gross volume of drift and its extent, collectively demonstrate the erosive and transporting power of ice.

Before turning to examine further some characteristics of the glaciation at 'maximum', it may be noted that times of maximum ice mass and maximum extent need not correspond precisely. Variations in speed of flow reflecting conditions at the bed of the ice sheet could separate the two. The term 'maximum glaciation' is used here in a general sense.

4. The glacial maximum and conditions in the Lake District Fells.

Discussion of regional and local contexts has shown how increase

in ice cover produced a pattern of dominant and constrained ice flows. The ice moved outwards towards areas of lower ice surface elevation. These elevations are shown schematically by form lines or contours drawn across ice-flow lines in Figure 1. The thickest ice was in the lowlands which could be at the same time zones of accumulation, storage and transit. The parts of the region with the greatest ice loading included the Permo-Triassic basins. In several instances zones of rapid lateral change in ice thickness coincided with boundary faults where reaction to differential loading may have been focussed. There is thus the possibility of subglacial neotectonic movements responding to loading, with response to unloading taking place as and after the ice cover dissipated. Some scarps in and near the Lake District aligned along major faults hold sites of recent slope failures and have a low degree of fretting and degradation (Clark 1988). The possibility of local movements and slope failures on fault scarps being related to recovery from differential ice loading is, however, apparently uninvestigated.

Changes in the pattern of ice flow directions from that established by the maximum stage were contingent upon change in the dominant ice mass. Post-maximum change would probably be related to an increase in net ablation (i.e. ice melting) on the ice surfaces. Any global increase in ablation raising sea levels would become effective in changing the state of the Irish Sea ice; here increasing depth of water in contact with the southern ice front would lead to ice dissipation by calving of ice and to acceleration of ice flow to the front. Such an effect would be greater and earlier in the western part of the basin where the sea bed is lower. If a comparable downdraw effect took place west of the Firth of Clyde, then ice over the North Channel could have developed a north-directed surface gradient. Discharge of ice to land-based margins could also be accelerated by an abundance of water under the ice, allowing a slippage along the bed to be added to movements within the ice; such an abundance of water would require widespread ice ablation.

The reconstruction of ice sheet maximum conditions presented in Figure 1 differs from two versions that have been presented for the British Isles Devensian Maximum (Boulton et al, 1977), based upon

computer modelling and an interpretation of geological evidence. Possible causes of the differences between the present regional and the British Isles reconstructions are the use of different levels of generalisations and the assumptions used in the computed reconstruction which presents ice flowlines for the region which do not correspond well with erratic distributions. It is interesting that the basal temperatures calculated for the computed model are well below pressure melting point, the temperature at which basal ice would melt for particular ice thicknesses, ice surface temperatures and heat flows. Cold-based ice masses, particularly at centres of ice dispersal where ice bed movement is slow, do not erode vigorously. Boulton et al (1977) noted the discrepancy between field evidence of intense erosion in the Lake District and the computation that Lake District ice at maximum was cold-based, frozen to its bed. They consequently suggested that "erosion was not primarily achieved during the glacial maxima, but during periods of much more limited glaciation when those areas merely maintained local ice-caps and valley glaciers and when high local ablation gradients produced rapid local glacier movements".

This raises one of several unresolved questions about Lake District glaciation. Huddart (1971) interpreted a sedimentary succession on the western edge of the fells as an accumulation of proglacial outwash from an advancing valley glacier, the deposition of till from that glacier, and then lodgement of till from Scottish ice. This interpretation implied erosion by Lake District valley glaciers before arrival of Scottish ice, that is before maximum glaciation. A further episode in which valley erosion may have occurred was during and following thinning of the surrounding, buttressing ice masses, so allowing easier discharge of accumulated ice down the dales. Severe areal erosion of bed rock surfaces occurred widely in the fells, to over 700 m., and extended close to the positions of the major icesheds. Clearly, at some stage in the glaciation, ice over the interfluves was temperate (wet-based), mobile and erosive. The distribution of erratics from the eastern side of the upland, from the Shap granite exposure for example, shows that at least some of that erosion took place before major changes were imposed upon the flow patterns established by the glacial maximum.

The radial elements in the Lake District valley pattern, attributed to rivers superimposed on older rocks from a domed cover of younger sedimentary rocks (Marr 1916), was alternatively explained by Linton (1957) as the consequence of vigorous erosion by ice flowing from a central ice-cap. That latter proposition has been doubted on grounds of lack of field evidence for modification of an earlier valley system and the low probability of a central ice dome (Clark 1988). The combined distribution of corries, especially well-developed ones, valley trough-heads, valley glaciers and snow patches formed in the Loch Lomond stadial (Sissons 1980), and the present distribution of precipitation suggest that there would be a compound ice cap with elongated western and lesser eastern culminations. Possible positions of ice crests on the Lake District ice cover are shown in Figure 1.

It is possible that the largest features of glacial erosion developed over several glaciations and that there have been several minor glaciations in the fells (cp Loch Lomond Stadial 10,000 - 11,000 years BP) as well as major glaciations. While it may be presumed that Lake District corries would attract snow in a minor or marginal glaciation, as many but not all did in the short glacial period within the Loch Lomond Stadial, their part in a major glaciation is unclear. Corries form only a small part of the uplands and Vincent (1985) has discussed their enigmatic nature. Snow catchment areas and corrie capacities may have been sufficient to nourish valley glaciers but growth of ice covers that concealed uplands and lowlands alike requires snow accumulation over large parts of the uplands in the early stages and thus a lowering of snowlines below those associated with snow accumulation mainly in corries.

Determination of whether snow accumulation in corries had any part in sustaining ice movement in the dales or initiating valley glaciers during a generally waning glaciation involves interpretation of moraines in some of the dales. Pennington (1978) appeared to link those valley moraines to the question of whether there had been a Scottish readvance during the latter stages of the glaciation. Doubting the reality of such a readvance, she indicated that there would be no need to postulate the occurrence of valley glacier

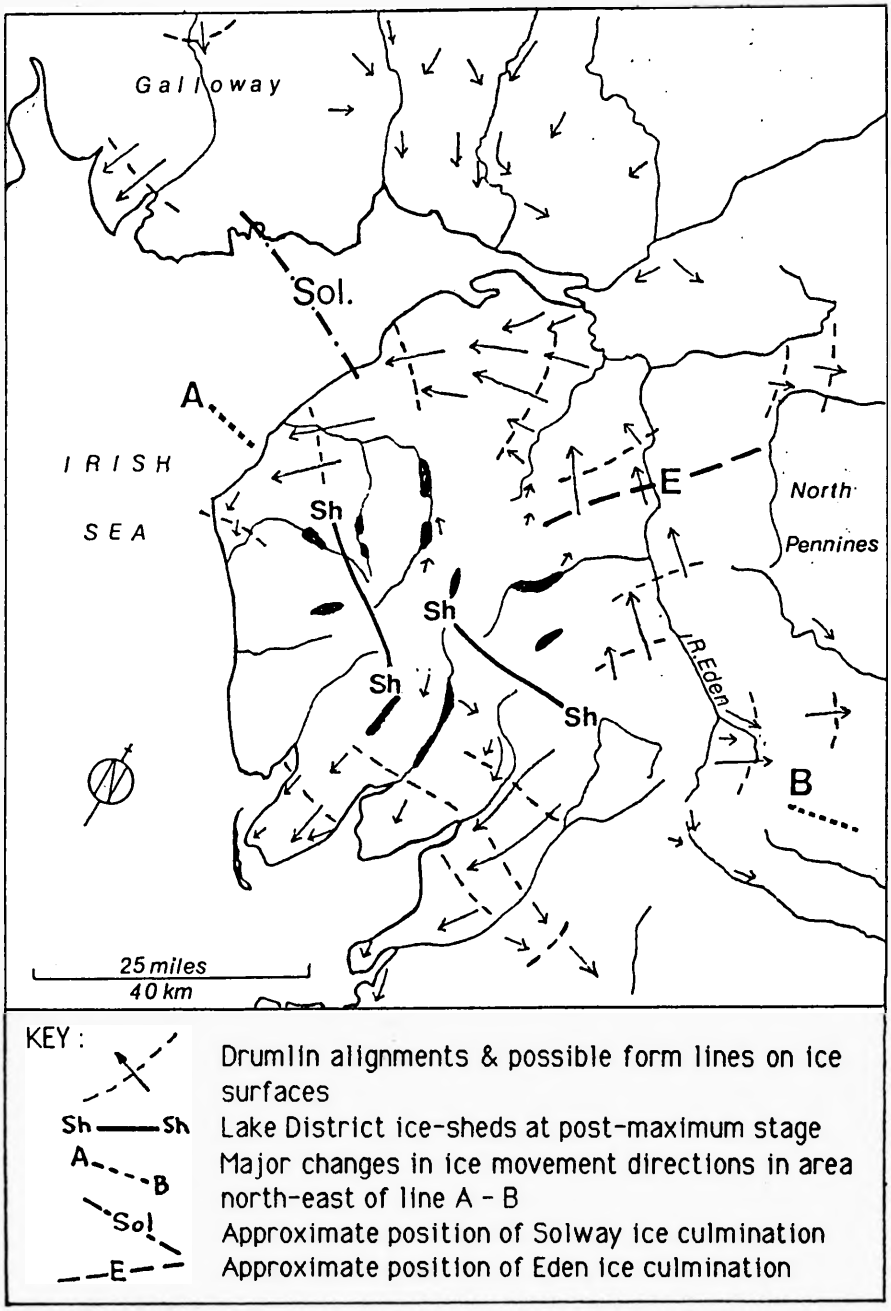


FIGURE 2 Ice movements following maximum glaciation.

standstills or readvances in Lake District valleys. Nevertheless there are morainic ridges in the dales such as Rosthwaite in Borrowdale (Wilson 1977): some, including the Rosthwaite examples and others at Wythburn, Patterdale and at the northern foot of the Kirkstone Pass, are only short distances down-valley from outermost Loch Lomond Stadial moraines (Sissons 1980).

A major characteristic of many dales is the relatively straight and unindented nature of their sides. That has been explained as due to the trimming back of projecting salients, so opening up the valleys (Marr 1916). Many dales follow structural alignments and, while valley widening is not to be doubted, the previous existence of projecting spurs and valley sinuosity is difficult to demonstrate. Marr chose the steep southern slopes of Blencathra to exemplify truncation but little there confirms that spurs once projected, that ice has greatly cut back and steepened the hillside or that the considerable width of the valley south to Threlkeld Knotts is mainly due to ice erosion.

5. The stages of drumlin formation.

The pattern of ice movement, thickness and surface form attained at the glacial maximum was not sustained. New movements were established, rather little changed in the south of Cumbria but substantially so in the north, as demonstrated by Figure 2. Much of the evidence for change is from alignment of drumlins, the contents of drift and, to a degree, paths taken by meltwater. Some drumlin tracts were formed before major changes in ice movement occurred. There are two closely linked questions - what development led to changes in directions of ice movement ? What circumstances were conducive to the making of drumlins ?

Despite considerable recent research (Habbe 1988, Tikannen 1988, Menzies 1989, Moers 1989) conditions and processes in drumlin formation are only partially understood. Indications are that ice masses involved in large scale drumlin construction were thick and coherent with vigorous movement. Other studies, for example from Ireland and Canada (Dardis 1981, Shaw 1983, McCabe & Dardis 1989),

have shown that large volumes of meltwater passed beneath ice in many drumlin areas before drumlin construction commenced. The possibility of water moving in large volumes beneath ice, moulding the underside, with sediments then filling the moulds has been discussed (Shaw 1983). The presence of more than thin films of water beneath ice at pressure melting point is required. According to some studies the ice bed became saturated and deformable, the ice base being partly supported by water. When and where such conditions prevailed ice velocities increased, leading to moulding of deformable bed. McCabe and Dardis (1989) described instances from Northern Ireland of water-borne sediments having been incorporated into lee parts of drumlins without deformation, that is, not by erosion or moulding of pre-existing sediments but by new deposition. Evidence for large scale movement of water subglacially before and during drumlin construction implies that considerable areas of the ice surface had been transferred to the ablation zone, and that, at least at times, rainfall superseded snowfall.

Arthurton and Wadge (1981) noted that many drumlins in the Penrith area incorporated glacifluvial sediments. In the Milnthorpe area there are drumlins with leeside sands and gravels. Near Ambleside a moulded rock core exposed by removal of till cover has surface forms suggestive of erosion by abrasive slurry prior to emplacement of the cover. Evidence from both north and south Cumbria thus indicates that water was present under ice at some local sites of drumlin construction.

The larger drumlin tracts of northern England, southern Scotland and parts of Northern Ireland lie well behind the outer limits reached by Late Devensian ice. Unless those limits were reached by accelerated ice flow in a drumlin construction stage, and that has not yet been demonstrated to be the case, the drumlins round the north Irish Sea basin would have been formed after ice reached its maximum extent. Earlier discussion has touched on the potential influence of rising sea levels on ice flow rates and decline of the ice cover. Not all accelerated ice streams within the ice sheets could have been reached by rising seas and been subject to the downdraw effect of rapid carving of icebergs from ice fronts.

New ice movement directions were in several places associated with drumlin construction. They were necessarily contingent on thinning of the Irish Sea ice sheet, the breakdown of the ice-shed separating eastward flow in the Solway area from southward flow in the west of Cumbria, and the cessation of the flow, largely of Lake District ice, across the northernmost Pennines into the Tyne catchment. Those breakdowns must have progressed eastwards with ice surfaces becoming higher in the east than those in the Irish Sea basin where a major reduction of ice mass was involved. The decline of the ice sheet may have been associated with drumlin construction and incision of channels by meltwater under the land-based part of the ice with calving into the sea concentrated in the western side of the Irish Sea basin during the earlier part of deglaciaton (Pantin 1977, Thomas 1985). One consequence of such deglaciation was the de-buttressing of once constrained ice masses including not only the Solway-Eden ice but also ice in the Lake district. Inner margins of main drumlin tracts are at the edges of the uplands so that the effects of accelerated ice flow on the low ground would be transferred quickly to the Lake District upland. Wet-based ice moving rapidly down the dales may have then effected significant erosion.

The evidence for more than one direction of ice movement in part of the Vale of Eden (Hollingworth 1931) has been re-examined by Letzer (1981), who described from the southern part of the Vale superimposition of north-directed drumlins on those aligned east towards the Stainmore sill. Only in the part of the Vale around Brough is there absence of the north-directed grain. From the upper Lune valley near Tebay north to about 8 km. south of Carlisle there is a continuous tract of north-directed drumlins and moulded rock ridges. The relation of drumlins immediately east and west of Carlisle to that tract is not clear. In the Penrith area drumlins are aligned across the earlier direction ofice flow northeast across the Vale to the Tyne catchment. The drumlin tract in the northern part of the Eden area lies west of a line from Gamblesby by Kirkoswald to Ainstable; drumlins there are aligned north of west although there is a little evidence, though not equivocal, that drift moulding by north-flowing ice took place near Croglin. It may be that early drumlin formation by north-flowing ice took place while ice still moved east through the

Tyne Gap, but that the main development was later. There is also a drumlin tract with grain west of north from the lower reaches of Ullswater north to the lower Caldew valley, south of Dalston, where it appears to join a drumlin tract west of Carlisle. That tract can be followed to the coast north of Maryport and then across the lowland of West Cumbria to the vicinity of Whitehaven, where the grain is north-south. The Carlisle-Whitehaven tract represents a movement of ice to the Irish Sea basin, but showing in the south the remaining influence of thick ice west of the present coast. It would be useful to examine that drumlin tract for evidence of development in more than one stage, earlier perhaps in the southern part of the tract.

South of the central fells where change in ice-flow direction was quite small, drumlin alignments from Duddon to Lune valleys are west of south, telling of a relaxation of the constraint earlier imposed by Irish Sea ice. Round Lancaster the grain is more nearly north-south. East-directed drumlins in the Greta and Wenning valleys may tell of ice congestion along the Lune, if contemporary with those in and west of the Lune valley; if somewhat earlier they may predate the effects of the thinning of the Irish Sea ice mass.

Drumlin formation in association with east-moving ice took place elsewhere so showing that conditions conducive to drumlin formation occurred, locally at least, before the full effects on ice movement of the decline of the Irish Sea ice were felt. East-directed drumlins occupy an area in the Tyne Gap while drumlins and moulded till between Canonbie and Gilsland may show that some convergence of ice-flow into the Tyne Gap replaced the earlier, more direct flow east over the Bewcastle- Spadeadam fells as the ice thickness declined. Similarly, drumlins east of the Stainmore sill are accompanied by eastward converging drumlins around Brough towards the sill (Burgess and Halliday 1979, Letzer 1981).

In the southern Solway plain meltwater incisions in the drumlin tract have exposed the local Middle Sands in places (Dixon et al 1926, Eastwood et al 1968); the Middle Sands imply deposition of water-borne sediments in lowlands before the last tills were emplaced. The extent or continuity of sand beds away from channel

sides is not well known but any significant lateral extension would take them beneath the drumlins. It would be useful to know if the beds form a widespread sheet or are related to specific meltwater input points.

6. The end of the glaciation.

Relationships between high-level meltwater discharge routes, direction of ice surface slope and drumlin formation are also important to describing later stages in the glaciation. Although an abundance of subglacial water before and during drumlin construction has been implied by a number of studies, little is known about water entry into and movement under vigorously moving ice nor about the impact of ice flow acceleration on water movement, particularly at the transitions between vigorous and weak flows.

On the north side of the Stainmore sill there are high level meltwater channels (c 490 m.) leading in the same direction as the aligned till ridges. The Butt Hill channel (c 490 m.) across the Pennine crest north of Hartside is clearly related to an eastward fall in ice surface. Further north on the Pennines and the Bewcastle Fells there are smaller and lower channels (380-410 m.) which also conducted water eastward. Considerable ablation must have occurred in these localities while ice surfaces were high and the ice still thick.

On West Fell south of Caldbeck, west-directed meltwater channels occur up to 510 m. forming part of an extensive melt discharge system. This implies a west-directed ice surface slope, so post-dating the the local ice surface slope reversal and hence the collapse of the Irish Sea ice sheet. Thus they show that the contemporary local ice surface was high and adjacent ice still thick. Many lower channels to the north show that the controls exercised on meltwater routes by ice thickness and surface slope persisted even as water found lower paths and after drumlin formation in parts of the lowland south of the Solway Firth. Time relation between formation of the higher channels and episodes of drumlin formation and rock-head moulding have not been determined. Reversal of meltwater discharge through the Tyne Gap, 30 km. east of West Fell,

took place when the last level of eastward water movement was c.140 m. Meltwater flows west across the Solway plain at the time of reversal were below 120 m., which gives some measure of falls in levels of water movement in that part of Cumbria between formation of West Fell channels and the Tyne Gap reversal.

Two examples from south Cumbria will serve to illustrate the question of relationships between movement of meltwater and processes in adjacent thick ice. Firstly, was the ice in the valleys and over the lower parts of the plateau still mobile and erosive when meltwater was constrained to flow south across cols (250-300 m.) on the higher northern parts of the plateau between Lune and Kent valleys ? In the west there is a complex of meltwater discharge channels between Gosforth and the Ennerdale-Kirkland area (Huddart 1967). When water passed through the cols at Black Pots (410 m.), Heckbarley (over 350 m.) and Murton Fell (c.300 m.), was thick ice over the lower ground to south and west still mobile and actively shaping the underlying land surface ?

In several Cumbrian drumlin tracts there are rock hills moulded parallel with drift drumlins. Structural and lithological influences do not completely explain their shapes. Many are so much larger than adjacent drumlins that the rock hills merit such a name as 'mega-rock drumlin'. Examples include hills on the Penrith Sandstone scarp, especially north of Penrith, those such as Warnell Fell between Caldbeck and Mealsgate, Watch Hill and its neighbours east of Cockermouth, and Dean Moor near Whitehaven. The drumlins in soft sediments are the products of the last glaciation. Could the erosional power of mobile ice and water during the drumlin phases of one glaciation have been sufficient to mould rockhead to such a degree or are those hills evidence of a recurrence of ice-flow patterns in successive glaciations ?

In parts of Cumbria, ice lost mobility while still thick, coherent and continuous. Many lowground meltwater channels cross drumlin tracts in courses indicative of continued ice control. Channels with uphill reaches show that flow under hydrostatic head occurred locally, as near Newton Reigny northwest of Penrith. Ice over lower

ground remained thick enough initially to deny direct passage of meltwater to the centres of valleys or offshore (Smith 1967). The patterns of several channel systems reflect the creation of successively lower paths until valley floors were accessible to meltwater. The adjacent Ellen, Derwent, Pow Beck, Ehen, Calder and Irt valleys demonstrate that point. Establishment of a chronology combining developments in those valleys is facilitated by the presence of cross-linking channels from Mealsgate south to the Gosforth area. While ice was still thick but sufficiently reduced for the high level channels (Murton Fell, Black Pots, Nannycatch, for example) to have been abandoned, a lower major discharge route developed leading from Loweswater and Pardshaw. Its persistence was due to the broad Kirkland-Arlecdon col which continued to discharge water south until channels became available to take water west by Branthwaite at levels below 130 m. The Ehen valley became a major discharge route by successively receiving the late discharge from the high-level fellside channels to the east, water from the col to the north and meltwater that entered the Keekle valley from the Moresby Parks area (east of Whitehaven). The Marron valley, which crosses the drumlin grain obliquely, is of interest. It probably replaced the Branthwaite route and carried meltwater north towards the Derwent valley axis. The disposition and extent of glacifluvial sediments on the sides of the valley suggest that the possibility of earlier flow upslope under hydrostatic head south to the Arlecdon col requires examination.

The channels in the area surrounding the Skiddaw block are significant in that they reveal the progressive development of channel systems which first drained the whole area into the Ellen valley, then to the Ellen and Caldew valleys. Later as ice in the Keswick-Bassenthwaite lowland thinned, meltwater from the southern and eastern margins gained access to the Derwent valley (Boardman 1982).

Some aspects of meltwater movement in the Solway plain have been touched upon and the implications of principal landforms are briefly outlined. At Aspatria, sand and gravel deposits (Eastwood et al 1968) lie on a ridge north of the Ellen valley. The depositional

circumstance is unresolved but may be related to meltwater escape from the Solway plain while ice was still thick, possibly mobile and moulding the underlying surface. Within the plain and post-dating drumlin construction is the Chalk Lodge-Moorthwaite system (east of Wigton) of major channels with entry levels at 100-35 m. Large volumes of water entered the channels from the east. Some was from the Caldew melt channel before it gained access to the lowest parts of the plain. Much must have been from the Petteril and Eden areas but the input levels suggest that the channels belong to the period within which the Gilsland channel (30 km. to the east) ceased to carry water east to the Tyne. The plan of the system shows that a precursor of the present Wampool channel was then in existence. That channel in its present form took water from the outskirts of Carlisle initially, at an input level c.50 m., and received further inputs near Bigland in channels traceable from Burgh by Sands near the Solway shore.

Channel sizes, locations of input points, lack of tributaries all point to large volumes of water being concentrated in the Carlisle-Burgh area but with free access into the lower Solway area being inhibited, presumably by residual or renewed ice. Discharge could have included water from Petteril to Esk, and Sark to Kirtle Water on the Scottish side. Into that pattern of westward discharge have to be fitted the southeast-directed Thursby eskers (Huddart 1973) east of the Bigland channels and north of the Wampool, and the Holme St Cuthbert sands and gravels, believed by Huddart (1972) also to have been built by water from the northwest. Huddart considered that deposit to be ice-marginal and noted the substantial proportion of Scottish material, including Criffel granites. The deposit lies across the Solway from Criffel and the mouth of Nithsdale. If the Holme St Cuthbert deposits were ice-frontal, then ice would have lain across the Solway Firth. It is evident that the final stages of glaciation in the north of Cumbria are closely tied to those in southern Scotland.

6. Conclusions.

A dynamic framework has been proposed able to accommodate a variety of evidence on the last glaciation in Cumbria. As well, the nature of gaps in current interpretation and data has been outlined in

the hope that some of the outstanding questions may be resolved in future studies of glaciation in Cumbria.

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EXCURSIONS

EXCURSION TO THE BEWCASTLE AREA, NORTH CUMBRIA

Leader : I. Gray

8th. May, 1988

The party met at Bewcastle Church (565,746), with its Anglian Cross and situated within the Roman Fort. Moving west along a minor road past Demesne Farm and the Castle, the party followed the line of the Maiden Way to Crossgreens. Old sandstone quarries in the Parkhead Sandstone on the right are said to have provided stone for the Roman Fort and the later church. After passing New House and Park Head, the cars were left at the cattle grid near the small stone shed (563,770). Turning through the farm track gate for Ashycroft and Stockastead, the track was taken to the left towards Ashy Cleugh. The party paused to view the remains of an old limekiln which had been in use until 1951. The limestone quarry adjacent to the kiln worked the Bogside Limestone at the base of the Bewcastle Beds in the Lower Border Group.

The rocks in this area on the gently dipping eastern limb of the Bewcastle anticline, first described by Garwood in the 1930's, are well exposed in most of the streams flowing west off the Bewcastle Fells, such as the Ashy Cleugh, see Figure 1. Most of the units of this Group up to the deltaic sandstones of the Middle Border Group are visible apart from the lowest Group (Lynebank Beds) which is not exposed in Ashy Cleugh. The Bewcastle Beds consist of 'Yoredale' type limestones with intervening sandstones, a more arenaceous sequence than the preceding marine Lynebank Beds.

Going upstream the rather thin Kittybeck Limestone is seen near the fence, followed by the next limestone (Holmehead). An area of tumbled rocks and waterfalls in the Ashycroft woods marks the Parkhead Sandstone (the Roman Quarry). In dry weather the watercourse is usually a succession of dry limestone areas with

surface water over the shale and sandstone, but with the early start of the Bewcastle monsoon, the going became very wet and the excursion terminated in torrential rain on the eastern edge of the woods. Thereafter further thin limestones up to the Main Algal Beds should have been seen beautifully exposed upstream in gorge sections. However the party retreated to the Lime Kiln Inn, near the church, where the meeting was abandoned.

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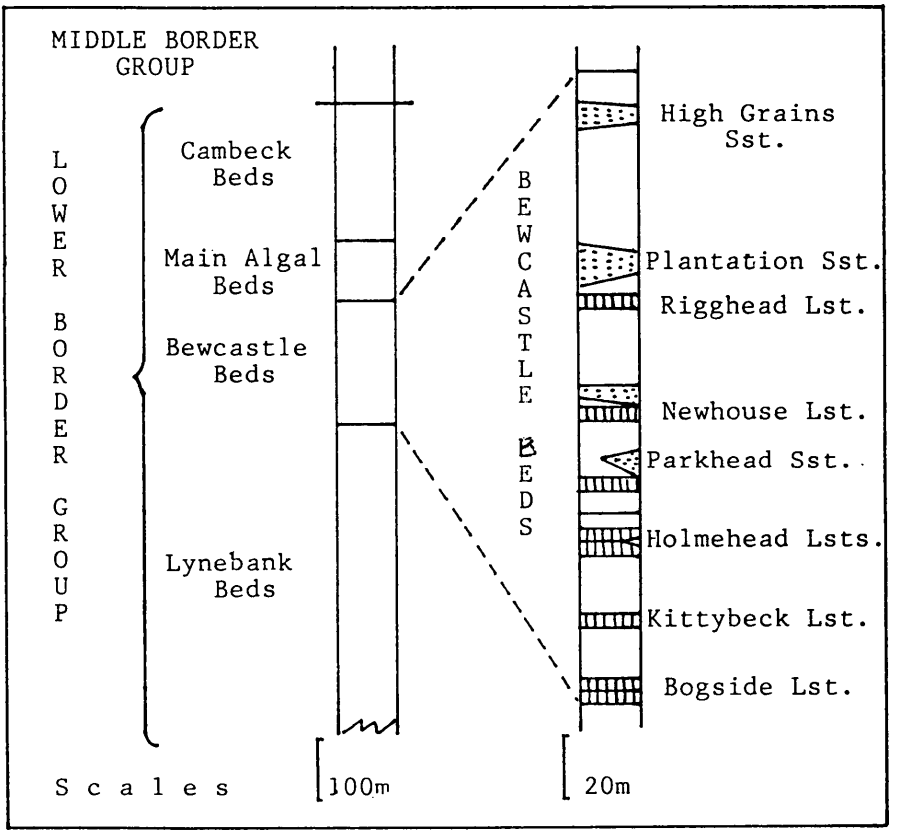


Figure 1 - Section showing the succession in Ashy Cleugh.

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GEOL. SURVEY Sheet 12 (Bewcastle) - 1:63,360 scale.

GEOL. SURVEY Special Sheet NY 57 (Bewcastle) - 1:25,000 scale.

MINERALISATION IN THE VALE OF NEWLANDS, CUMBRIA

Leader : Mr. B. Young

2nd. July, 1988 - Joint meeting with the Russell Society.

In indifferent weather, the party assembled near the hamlet of Little Town (233,194) in the Vale of Newlands. Outlining the day's programme, the leader mentioned that there appeared to be two major phases of mineralisation. The older phase along lines of weakness trending roughly E-W, appeared to be of Devonian age and composed of Cu/S/As compounds. This older suite of mineral veins was clearly offset by a series of N-S trending faults running along the Newlands valley. This younger suite carrying Pb/Zn/S minerals was thought to have been emplaced in the late Carboniferous.

The party walked 2.5 km. south along a cart track to the site of the Castle Nook mine (228,170), where a series of copper-bearing "trial" adits showed the presence of chalcopyrite, arsenopyrite, galena, sphalerite and some hemimorphite and pyromorphite. Conspicuous streaks of vein quartz were observed in the crags above the mined area.

After crossing Newlands Beck, the party visited the site of the Dale Head mines (224,155) which, along with Goldscope mine, are

thought to have been worked intermittently since medieval times. The stoped-out E-W copper veins of St. Thomas's Work and the Long Work could be clearly traced. The mineral assemblages on these veins, which included chalcopyrite, pyrite, marcasite and pyrrhotite, were thought from fluid inclusion studies to have crystallised from saline solutions at about 350° C. They clearly are minerals of primary origin. The party then ascended to the ruined buildings and dressing floors perched on the shoulder of a coombe nearly 1000 ft. above the valley floor, while a few robust members scrambled a few hundred feet higher to look at the adit and spoil heap on the Dale Head South Vein, just 300 ft. short of the summit of the mountain. The minerals here, malachite, bornite and chrysocolla, contrasted strongly with those found 1000 ft. below and were of supergene origin.

Returning northwards Goldscope Mine (231,184) was visited, which again showed copper mineralisation on the E-W trending Goldscope Lode, offset by the lead-bearing Sealby's and Scope End Lodes. While the name might raise hopes of "striking it rich", only traces of gold had been established with no reliable record of the precious metal having been extracted. However, considerable amounts of silver had been obtained from the lead ore mined there. A recent find on the spoil heaps had been a large block of massive silver-bearing copper mineral tetrahedrite. Following refreshments at Low Snab Farm, the visit concluded with a vote of thanks from Mr. Dodd.

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VOLCANIC ROCKS IN WASDALE

Leader : Dr. M. Petterson

31st. July, 1988 - Joint meeting with Westmorland Geological Soc.

The visit commenced at an ice-scoured exposure on the lakeside (158,060) below Water Crag.The leader explained the significance of texture and mineralogy in modern interpretations of volcanic rock types. The rock at the lakeside, andesite, probably formed by crystallisation in the interior of an effusive lava flow in an early stage of the activity of the Borrowdale volcano. The next stop at the roadside, 100ft. higher, showed flow brecciation - a chaotic, blocky texture with a sandy matrix. This exposure might be interpreted as the top of a lava flow, or possibly a sill with brecciation caused when magma intruded wet volcanic sediments.

Above the road, a prominent tuff horizon was investigated. Fine and coarse laminations showed sedimentary features,such as false

bedding, and small scale contemporary faulting. The largely parallel nature of the bedding suggested that this was an air-fall (rather than a water-worked) deposit. The next two exposures on Water Crag and at a small quarry (155,061) showed a feldspathic, slightly porphyritic andesite. Flow-brecciation seen at the base passed upwards into massive lava with flat-lying flow joints. On the hillside north of this quarry a prominent linear, pink-weathering dyke was investigated; it consisted of a fine-grained silicic rock chemically resembling the Ennerdale Granophyre.

Lunch was taken at an ice-scoured knoll between Water Crag and Long Crag where bedded ash-fall tuff mantled the brecciated top of a thick andesite flow. Again opinions were divided on whether this exposure might alternatively be of tuff disrupted and partly fluidised by a sill-like intrusion.

Higher still on top of Goat Crag, thick layers of ignimbrite were seen. Dr. Petterson explained how a gassy, explosive eruption of silica-rich magma could lead to the formation of three characteristic rock types - a ground surge deposit followed by the main ignimbrite sequence and an air-fall tuff. The ignimbrite is distinguished by its eutaxitic fiamme texture, caused by squashed pumice fragments rewelded at residual rock temperatures of 500-700^{0}C. To the north-east across Goat Gill, the massive ignimbrites could be seen at a lower level, indicating substantial down-faulting. A similar fault could be discerned, apparently dipping at 45^{0} E below Illgill Head across Wastwater. The leader explained that recent ideas had thrown up the possibility of a large caldera having formed as a result of the rapid emptying of the magma chamber of the Borrowdale Volcano, and that this faulting marked the edges of such a caldera. Before leaving Goat Crag, the party viewed a basal surge deposit composed of spherical lapilli 5-15 mm. in diameter, and showing low-angle sedimentary structures.

A precarious descent was made down Goat Gill to the cars at the roadside, where Dr. Petterson was thanked for providing an interesting and informative excursion for the members of the two Societies.

GLACIAL AND FLUVIAL SEDIMENTS AND SOILS IN MOSEDALE, SOUTH OF THRELKELD

Leader : Dr. J. Boardman

2nd. April, 1989

The party met at High Row, Dockray (380,220) and walked west along the old coach road to Barbary Rigg. This feature is a superb example of a terminal moraine fringing the site of a former corrie glacier, which developed the north-facing basin of Mother Sike, adjacent to Wolf Crags (grid square 3522). Dr. Boardman showed where the moraine had been breached by Mother Sike and explained that the basin contained up to 2 metres of peat which had accumulated during the 10,000 years since the Loch Lomond Stadial at the end of the Devensian Stage. The relative effects of glacier erosion and deposition caused by major valley glaciers versus "snowpatch" or corrie glaciers was discussed and Dr. Boardman suggested that the effects of the Wolf Crags corrie glacier would have been relatively small and very local. Little active down-cutting occurred over its estimated time-span of 500 years, merely a flushing out of accumulated debris together with frost-riven material from the headwall. He stressed that the general features of the terrain developed over several stadials and interstadials, supporting this by showing the presence of boulders of Threlkeld Micro-granite in the moraine ridge, which must have been transported uphill some 3 km.

The group then traversed 2 km. of Mosedale Beck, running north from Wolf Crags from the breach in the moraine, where Dr. Boardman had established a succession of three tills;

Wolf Crag Till -	suggested age	10,000 years	Devensian Stage
Threlkeld Till -	" "	20,000 years	" "
Thornsgill Till -	" "	200,000 years	Unknown Stage

The Thornsgill Till was pointed out at several sites where prolonged weathering had penetrated up to 4 metres, altering the colour from greenish-blue to yellow and allowing the development of irregular

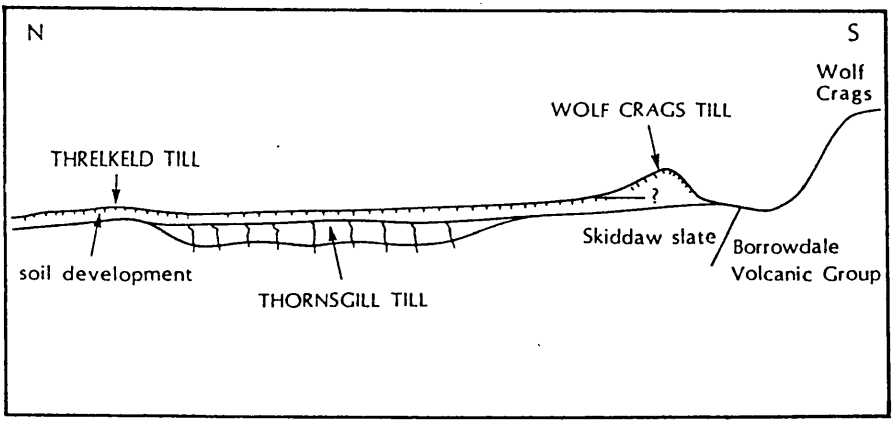


Figure 1 - Profile of the Mosedale Beck Valley, showing the relationship between the three tills.

black, manganese-rich bands. At one site the occurrence of a thin band of highly compressed peat containing recognisable willow twigs, together with pollen of various grasses and sedges, suggested the former existence of a temperate climate. Radiometric analyses of this peat layer gave a ^{14}C age greater than 50,000 years and a U-dating of between 75,000 and 105,000 years, that is older than the Devensian. The overlying Threlkeld Till, attributed to the Devensian, is in a much less weathered condition and contains wisps of the yellow Thornsgill material incorporated into it, presumably scooped up by the passage of Devensian ice. The clear differentiation between the Threlkeld and Thornsgill Tills was best seen just west of the confluence of Caral Beck (355,238).

Dr. Boardman drew attention to river terraces evident on the sides of Matterdale Beck and noted that soil investigations had led to the conclusion that the lowest terrace, typically 4 metres above the stream bed, was of comparatively recent origin. The chance finding of a 1749 local letter had revealed a catastrophic cloudburst in July of that year, which had damaged buildings in the Vale of St. John and had caused considerable erosion in local valleys. The power of the torrent in Mosedale Beck caused by this 1749 cloudburst must have been

awesome as it trundles masses of boulders up to 1 metre in diameter, leaving them in several places as massive boulder bars. The scouring by the floodwaters had lowered the channel by several metres and undercut the valley walls, leading to subsequent landslips. Just south of the ruins of High Hollows (356,249), remains were found of an ancient water leat truncated by the 1749 flood, although the reasons for its construction were obscure. The excursion was then completed by returning uphill to the old coach road and to High Row.

[Note : Persons wishing to investigate the enclosed land to the east of Mosedale Beck must obtain the prior permission of the landowner at Lobbs Farm (360,247)].

LIMESTONE QUARRIES NEAR MOOTA

Leaders : F. Lawton & M. Sanderson

24th. May, 1989

The party met on a minor road to Blindcrake, south of the Moota Motel (154,357), and headed N.W. to three disused limestone quarries, known as Clintz A, B and C (believed to have closed in the late 60's). The limestones of the Lower Carboniferous in West Cumbria are traditionally numbered from top to bottom, First to Seventh Limestones, and separated by shales and sandstones. These quarries are in the Fourth and Fifth Limestones, of Asbian age.

Quarry A (162,363) in the lower part of the Fifth Limestone was briefly visited. Part of the quarry appeared to have been back-filled, possibly no longer viable when mineralised limestone was encountered; evidence for this was seen in the presence of barite veining.

Quarry B (162,359) is the largest of the three, with beds dipping at 10^0 in a westerly direction. The quarry faces show the top 4 m. of

the Fifth Limestone, succeeded by the Fourth Shale (sandy mudstones with thin coal bands) and then the Fourth Limestone, which continues to the top and is split by thin shale bands. Notable features on the north and west faces of the quarry are the presence of pothole solution features at the top of the Fifth Limestone. They tend to be conical in shape with smooth walls and are generally filled with argillaceous sandstone. Such solution is a karst feature and the potholes may represent a palaeokarst surface (where the limestone was dissolved while exposed on a land surface during a period of uplift). The south face provided a similar aspect although the potholes here were much smaller. A possible fault was noted on the west face, apparently parallel with the larger Blindcrake fault nearby. Fossiliferous horizons were present at several horizons with fine bands of *Lithostotion Junceum* and *Martini* seen at the quarry top in the Fourth Limestone. Shaly material in waste tips contained many brachiopods, corals and crinoids.

Finally a very quick visit was made to Quarry C (158,364) to see three coal seams, ranging from 120 mm. to 230 mm. in thickness, within the Fourth Limestone. Please note that the quarries are on private land and prior permission had been sought for this visit.

GLACIAL LANDFORMS AND SEDIMENTS IN SWINDALE

Leader : Dr.A.G.Lunn

4 June, 1989 - Joint meeting with Westmorland Geological Society

The party assembled near Swindale Foot (522,142) where the leader and Tom Shipp outlined the geological and glacial settings of the excursion. All the glacial landforms and sediments seen resulted from the Devensian glaciation (maximum extent 18,000BP), the subsequent warming followed by a temporary re-advance in the Loch Lomond stadial (11,000-10,000BP). The Lake District landform studies of Sissons (1980) did not include Swindale and the objective of the excursion was to see if his conclusions were applicable to this

area. He had concluded that most of the valley glaciers had retreated during deglaciation and their remnants then decayed in situ; he identified the extent of the Loch Lomond phase by the down-valleys limits of 'hummocky moraine' - irregular mounds of drift varying from accumulations of boulders to a sandy till with clasts.

The party walked 2 km up the valley and noted the glacial striated cobbles, incorporated in the walls, and the roche moutonées high on the sides of the valley, indicating a considerable thickness of ice at some time. Close to Swindale Head farm (504,125) the hanging valley of Mosedale was recognised. Lodgement till, exposed where cut by a tributary stream close to a footbridge (503,119), was characterised by poorly-sorted, matrix-supported clasts while hillocks further up the valley were identified as typical hummocky moraine.

The party climbed into the hanging valley of Mosedale (508,112), where Mosedale was seen to be a through valley connecting Swindale with Long Sleddale with evidence of ice transport in both directions. This valley may have resulted from the period of the main glaciation, with linear streaming within the ice sheet causing the ice divide to migrate at different stages. Looking down into the head of the Swindale valley, it was suggested that the snout of the final advance rested against a very large roche moutonée at about the limit of the hummocky moraine. The flat, swampy area above the hummocky moraine was variously suggested to be an infilled ice-dammed lake, a piedmont fan formed by ice emerging from Mosedale or the result of a late ice surge with the clasts washed away in a melt accelerated by rising temperatures.The final stop was in the north-east facing corrie lying above Hobgrumble Gill (500,110) which could have formed a significant icefield, backed by a considerable ice-gathering ground on the plateau beyond, and provided the Loch Lomond surge in Swindale.

The party returned along the western ridge via High Birkin Knott and Hare Shaw to reach the cars by a steep descent near Swindale Foot Crag. Thanks were expressed to the leader on behalf of both Societies for an excursion to a little visited valley which all enjoyed.

Reference:
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LANDSCAPES & ROCKS IN THE EYCOTT HILL & CARROCK FELL AREAS

Leader : Dr. R. Clark

25th. June,1989.

The party met on the roadside 1.5 km. north of Berrier (388,306) where small roadside quarries, probably in the Seventh Limestone, are near the local base of the Carboniferous. East of the road the Sixth and Fifth limestones form the higher part of the escarpment.

The first stop at Eycott Hill (387,294) afforded extensive views, allowing appreciation of landscape characteristics. Here andesite lavas and tuffs of the Eycott Volcanic Group dip north-east at slightly steeper angles than the limestones. The bench-like hilltop corresponds well with the unconformity beneath the limestone. Its surface is ice-scoured with the variously resistant rocks differentially eroded into parallel ridges and troughs. To the south the easterly dips of the Mell Fell conglomerates prompted the question whether the country rising west towards Dockray and Matterdale Common may represent the base of the conglomerate. It extends north to within 1 km. of Eycott Hill; its further extension may be limited by non-deposition or erosion prior to the deposition of the Seventh Limestone. To the south-west, the generally smooth, rounded outlines of the northern parts of the Helvellyn Fells are a reminder of diminished glacial erosion compared with the southern fells. It was suggested that the considerable breadth of the Troutbeck-Threlkeld valley was not sufficiently explained in terms of river and ice erosion and that it may also be thought of as a 'moat' which almost completely isolates the Skiddaw block.

On route to the second stop a good example of planation facet at Hutton Roof was crossed and also an ice-eroded rock bar of Eycott lavas across the River Caldew (361,342). To the north the river followed a melt-water channel; to the south it flowed across an aggraded basin, depth to rock-head unknown.

A stop on the Carrock Beck (355,352) showed that the beck had cut through thick tills (350,350) above an old ore smelting site, had reworked glacial deposits to leave very coarse lag deposits containing large boulders carried by ice from Carrock Fell. Nearby pits on the Long Hill esker (352,345) were examined and pebbles in a suite of southern origin included those from Armboth Dyke and some pink granites of undetermined origin. The esker was considered with the Caldew channel and adjacent features to tell of late meltwater discharge north in a waning ice sheet with a suite of meltwater channels from 510 m. on West Fell (334,355) to the Caldew at 210 m. reflecting falling discharge levels in thinning ice. Adjacent to the esker, the rock-cored Bannest Hill drumlin is a reminder of an earlier phase of active northward iceflow which had also shaped the composite Howe drumlin west of Eycott Hill.

The last stop was at the foot of the fault-aligned Carrock Fell scarp (354,334) where attention centred on the morphology of scarp-foot deposits. The gabbros had produced an extremely coarse steep block scree while the longer granophyre slopes east from the summit of Carrock Fell gave rise to a greater variety of clast sizes including silt. Debris accumulated at the foot of Scurth Crags had moved east to build an apron with shallow-concave surface mantled in granophyre blocks with a few gabbro boulders, presumably from reworked tills. The apron extends some 300 m. east, terminating in a steep, somewhat lobate frontal slope. Stone pits in the proximal part of the apron showed several metres of debris with large clasts dispersed in a finer matrix, with some concentration of large clasts near the surface. Such a matrix may flow and support big boulders, depending upon bed slope, proportion of silt and clay in the matrix, thickness and water content. The short but severe Loch Lomond stadial was suggested as the most probable occasion for formation. The meeting closed with thanks from Tom Shipp.

THE SKIDDAW GROUP OF CAUSEY PIKE

Leader : Dr. A. H. Cooper

9th. July, 1989.

Recent work by the British Geological Survey has shown that three distinctive Ordovician stratigraphical belts (northern, central and southern) are recognisable in the Lake District and northern England, illustrated in Figure 1. These well-defined belts, probably separated by ENE trending faults, controlled thickness and facies variations from the Tremadoc to early Caradoc. The stratigraphical differences between the belts progressively diminished later in the Lower Palaeozoic succession before their faulted boundaries were reactivated in the early Devonian Acadian orogeny (Soper et al., 1987).

The Skiddaw Group (late Tremadoc-early Llanvirn) in the northern belt comprises at least 5 km. of turbiditic greywacke and siltstone, in which five Formations are recognised in Figure 2:

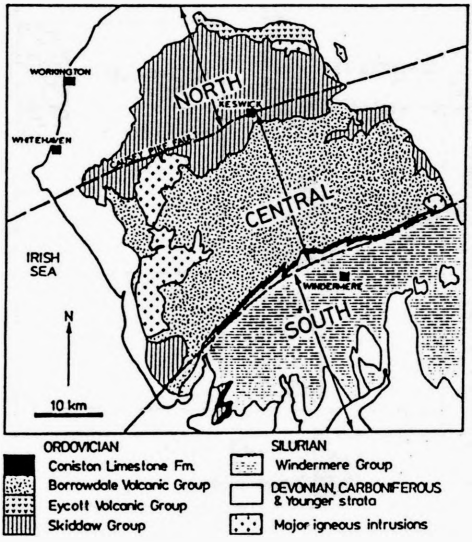
Formations	Previous Classifications
Kirkstile	Loweswater Flags & Kirk Stile Slates
Loweswater	(Jackson, 1978)
Hope Beck	Hope Beck Slates (Jackson, 1961 & 1978)
Watch Hill	Bitter Beck & Watch Hill Grits
Bitter Beck	(Molyneux & Rushton, 1988)

These formations were folded in the early Llanvirn by major slump folds overturned to the SE (Webb & Cooper, 1988) and are now recognised to be unconformably overlain (Millward, pers. comm. 1989) by the Eycott Volcanic Group, dated as Llanvirn (Downie & Soper, 1972).

The Causey Pike Fault and the Crummock Water aureole separates the northern and central belts (Cooper et al., 1988). This metamorphic aureole corresponds to a geophysical anomaly interpreted to be related to an elongate granitic body emplaced into a shear zone. The metamorphic recrystallisation within the aureole has been dated at 401+/-3 Ma. Both the bedding and the upright cleavages (N.N.E. & N.N.W. trending) in the metamorphosed Skiddaw Group rocks are sygmoidally

deformed across the aureole, suggesting post-cleavage sinistral movement along the Causey Pike Fault Zone. The aureole is post-dated by southerly directed thrusting seen on Causey Pike.

The Skiddaw Group of the central belt is at least 4 km. thick with the lower part composed of Arenig turbiditic mudstones (Molyneux & Rushton, 1989) and overlain by an early Llanvirn olistostrome, the Buttermere Formation (Webb & Cooper, 1988). This Formation of Tremadoc and Arenig sediments, about 1.5 km. thick, was redeposited and slumped to the N.W. Blocks of turbiditic greywacke sequence in the olistostrome reach 1.5 km. in length and display complex folding; these are the Robinson Member, well exposed on the south flank of Causey Pike. A sedimentary breccia sequence, the Goats Gill Member, also occurs in the Buttermere Formation on the west side of Robinson (Webb & Cooper, op. cit.) while above the Formation occur some 1.5km. of Llanvirn mudstones with beds of volcani-clastic rocks (Tarn Moor Formation), perhaps corresponding to part of the Eycott Volcanic Group. The Skiddaw Group is unconformably overlain in the central belt by 4-6 km. Borrowdale Volcanic Group, of ?Llandeilo-Caradoc age (Moseley & Millward, 1982).



The Excursion - localities visited are shown on Figure 3, which shows the structure of the district. Grid references refer to Grid Letters NY.

Figure 1 Simplified geological map of the Lake District showing the three early Ordovician stratigraphical belts. The Causey Pike Fault separates the northern and central belts and a major monoclinal structure approximately follows the junction between the central and southern belts.

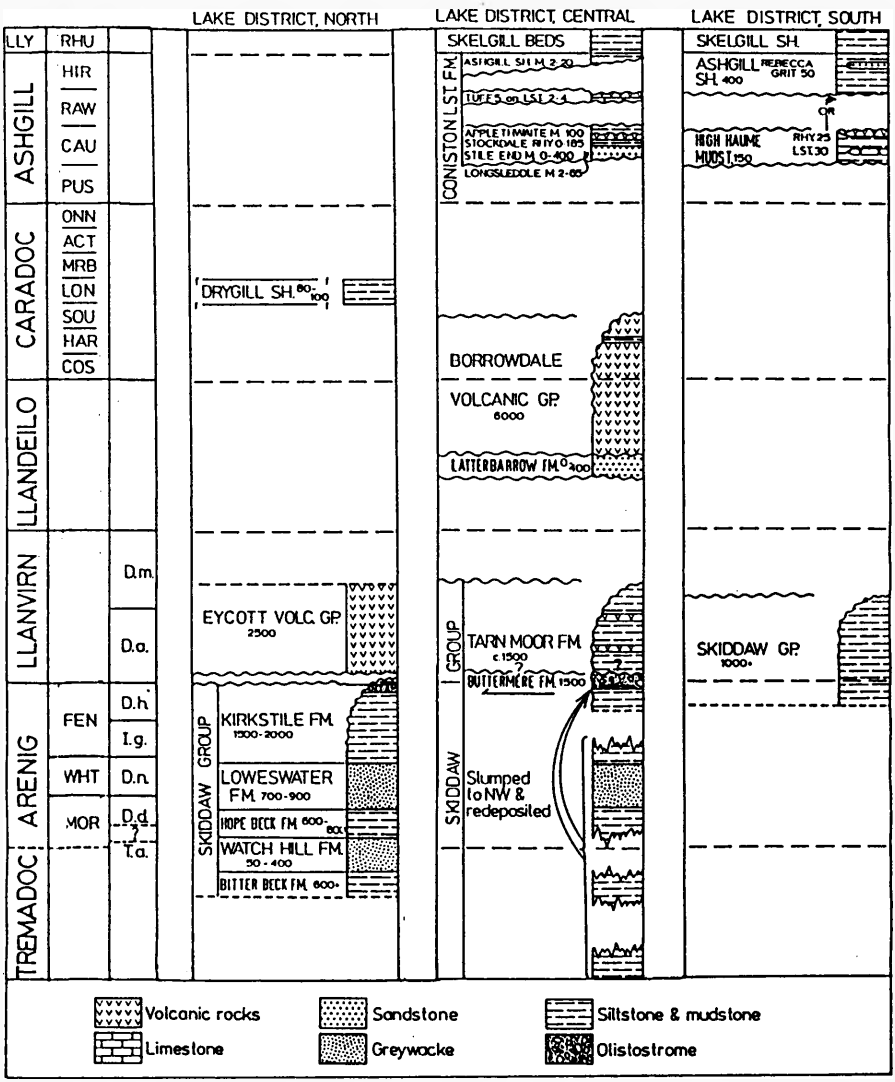


Figure 2 Chronostratigraphical sections with lithological ornament for the three stratigraphical belts in the Lake District. (approximate thicknesses in metres).
Early Ordovician biozones have the following abbreviations: Ta-Tetragraptus approximatus, Dd-Didymograptus deflexus, Dn-D. nitridus, Ig-Isograptus gibberulus, Dh-D. hirundo, Da-D. artus (formerly D. bifidus) Dm-D. murchisoni.

The principal objectives of the excursion are :

a) to look at the faulted junction between the northern and central belts of the Skiddaw Group, separated by the Causey Pike Fault. In Stonycroft Gill (2290 2125) the Kirkstile Formation is thrust over the Buttermere Formation olistostrome. Further west on Causey Pike (2192 2084) the Kirkstile Formation is hornfelsed by the Crummock Water aureole (Cooper et al, 1988). The eastern end of the aureole is thrust southwards by Causey Pike Fault, here over a sandstone raft (Robinson Member) within the Buttermere Formation.

b) To study the hornsfels and tourmaline veins on Causey Pike and the gradational northern margin of the Crummock Water aureole.

c) To look at breccia beds in the Kirkstile Formation on Outerside and search for graptolites.

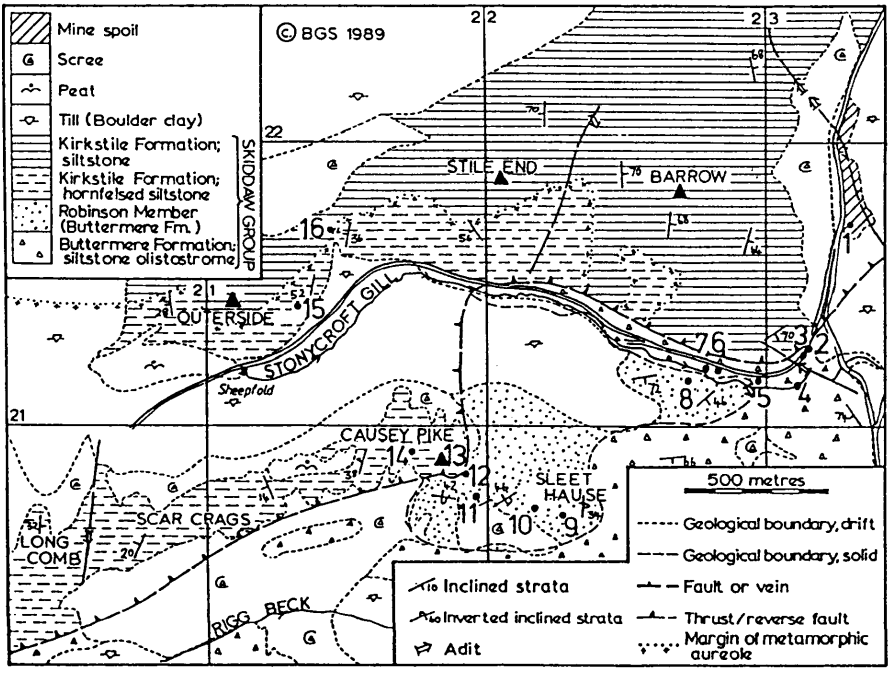


Figure 3 Geological sketch map of the Causey Pike area, showing localities 1 - 16 visited during the excursion.

Locality 1 The excursion starts at the the car park on the old spoil heaps at the disused Barrow Mine, Uzzicar (2332 2173). Numerous adits followed the vein on the face of Barrow with further workings extending from a shaft 100 m. east of the car park (Adams, 1989).

Locality 2 Old Mine Road (2313 2126) - dislocated and brecciated slump folds in the dominantly siltstone facies of the Buttermere Formation are exposed in the bed of the track. Sporadic sandstone laminae pick out tight angular folds with shear planes; two fracture cleavages are also present.

Locality 3 Stonycroft Mine (2310 2122) - a straight section of Stonycroft Gill runs ESE, following an excavated quartz/galena vein. A shaft to deeper levels was once open in the stream bed, with the water diverted by a dam and watercourse (Eastwood, 1921).

Locality 4 Old dam in Stonycroft Gill (2311 2114) - the watercourse meets the stream, with water-worn surfaces showing complex slump-folded and disrupted siltstone and mudstone, typical of the Buttermere Formation.

Locality 5 Viewpoint (2296 2115) to WSW of Causey Pike; the craggy upper part of the peak consists of hornsfelsed siltstone of the Kirkstile Formation (Cooper et al., 1988) which is thrust southwards over the Robinson Member, the fault running along the foot of the crags above the heather-covered slopes.

Locality 6 A recently landscaped site of former smelt works and reservoir (2283 2120) - littered with slag and unused cobalt ore from an unsuccessful cobalt mine (Adams, op.sit.) in Long Comb (2058 2060). Viewed from south of the Gill, the line of the old smeltworks flue can be seen running north up the hill.

Locality 7 Quarry (2277 2120) - in the edge of a quartz-rich greywacke raft forming part of the Robinson Member (Buttermere Formation) here comprising about 60% siltstone and 40% quartz-rich greywacke in thin to medium beds displaying classic Bouma "B" and "C" turbidite units (Bouma, 1962). The sandstone sequence strikes

northwards directly into complex folded siltstones and mudstones; to the south it can be traced onto the north flank of Sleet Hause.

Locality 8 North flank of Sleet Hause (2270 2115) - mainly fine-grained greywackes in thin to medium beds, part of a large sandstone raft (Robinson Member). The order of the Bouma cross-laminated units and the truncated cross-laminations in the "C" units show these beds to be the correct way up and dipping to the SE.

Locality 9 Sleet Hause (2119 2067) - Catbells (Buttermere Formation) and the extensive spoil heaps of Yewthwaite Mine (240 194) are visible to SE. Southwards the U-shaped glacial valley of Newlands Beck leads up to Dale Head. The valley bottom is covered by glacial till and extensive screes occur along the valley sides. Several hanging valleys are visible between Dale Head and High Spy, Hindscarth and Robinson. The unconformity (Skiddaw Group/B.V.G.) crosses Maiden Moor and below High Spy and Dale Head but the junction is not very obvious.

Locality 10 West of Sleet Hause (2114 2070) - on the footpath and the south flank of the ridge inverted quartz-rich greywacke beds are exposed. Examples of truncated cross-laminations, convolute laminations and the "B"-"C" progression of turbidite units affords good evidence of inversion. The sandstone beds from here to the south side of Causey Pike are mainly inverted, complexly folded and sheared, forming a large (c.1 km. across) sandstone raft (Robinson Member) with a discordant relationship to all the surrounding brecciated and slump-folded siltstones and mudstones.

Locality 11 SE of Causey Pike (2196 2074) - complex folding in the same sandstone raft as Locality 10 exposes a tight, inverted syncline which appears to be discordant with surrounding sandstones. Way-up evidence is difficult to find but once located conclusively shows the beds to be inverted.

Locality 12 Footpath at the foot of the crags which top Causey Pike (2192 2084) - the Causey Pike Thrust is exposed here with the southerly-directed overthrust block composed of bleached and

hornfelsed Kirkstile Formation with tourmaline veins (Cooper et al., 1988, Fortey & Cooper, 1986), part of the Crummock Water aureole. The footwall rock comprises the Robinson Member sandstones with sheared siltstone present next to the thrust surface. The aureole post-dates the two upright cleavages which are sygmoidally deformed across it suggesting sinistral displacement along the Causey Pike Fault, approximately synchronous with the hornfelsing and prior to the thrusting.

Locality 13 Causey Pike summit (2188 2086) - hornfelsed Kirkstile Formation with slump folds and cross-cutting dark grey tourmaline veins are present. The panoramic view shows the features noted at Locality 9 and the geomorphology of the Keswick area. The lithological contrasts between rocks of the Skiddaw and Borrowdale Volcanic Groups are apparent, with the dominantly siltstone lithologies of the Skiddaw Group giving smooth rounded fells on Skiddaw and the more contrasting lithologies of the Borrowdale Volcanic Group forming rugged crags, such as Walla Crag east of Derwentwater. The resistant hornfels of the Crummock Water aureole also form craggy topography, extending west along the ridge from Causey Pike; the trace of the Causey Pike Fault is visible at the foot of the crags to the south of this ridge.

Locality 14 West of Causey Pike summit (2180 2089) - disharmonic slump folds exposed adjacent to the path are preserved in hornfelsed Kirkstile Formation siltstones. The folds are cut by E-W trending tourmaline veins, up to 1 cm. wide, with hard, tourmalinised wall rock selvages and central core of crystalline tourmaline and quartz. The veins developed by alteration after cleavage which is re-orientated by shear across the aureole and fault.

Locality 15 Sheepfold on SE flank of Outerside (2114 2120) - Here the upper part of the Kirkstile Formation is exposed - a dominantly hornfelsed siltstone with abundant slump folds and also hornfelsed intraformational breccias or debris flows, comprising sandstone or siltstone clasts in a mudstone matrix; these are the fluxoturbidites of Jackson (1978).

Locality 16 East of Outerside (2144 2169) - hornfelsing decreases after the previous locality as the gradational northern boundary of the aureole is approached. The Kirkstile Formation here comprises siltstones and mudstones with slump folds. Graptolites indicative of the basal *Didymograptus artus* (formerly *D. bifidus*) zone occur here; on the north face of Outerside nearby graptolites of the *D. hirundo* zone occur, suggesting that the Arenig—Llanvirn boundary occurs between these localities. From Locality 16 the starting point may be regained via the Cobalt Mine track along Stonycroft Gill.

Acknowledgements The leader would like to thank Drs.B.C.Webb, A.W.A.Rushton & S.G.Molyneux for discussions in the preparation of this guide; Drs.D.Millward, R.Hughes, E.W.Johnson, A.J.Reedman & Mr.B.Young for critically reading and Mrs.J.Dunkley for typing the manuscript. The report is published by permission of the Director of the British Geological Survey (NERC).

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GEOLOGY OF THE HELSINGTON/LEVENS AREA

Leader : Mr. A. S. Day

10th. September, 1989.

The party assembled at Helsington Church (488,889) where the leader introduced the geology. The Helsington/Levens area is situated on the southern section of a broad ridge of Carboniferous rocks disappearing south of Levens but extending northwards to Scout Scar and Cunswick Scar. A sharp escarpment forms the western edge and a gentle dip slope descends to the east. Glacial scouring and deposition has modified the dip slope with a patchy cover of drumlins and moraines frequently obscuring the solid geology beneath. Three of the Lower Carboniferous stratigraphic stages may be recognised :

1) Chadian - at the base, represented by the dark, fine-grained micritic Martin Limestone.
2) Arundian - Red Hill Oolite and the overlying Dalton Beds.
3) Holkerian - the uppermost stage, represented by the light- coloured Park Limestone.

The area was originally mapped by Garwood, using the fossil content of the limestones and Vaughan's well-known zonal divisions of the Lower Carboniferous. His map gives a broad overall picture of the area but in detail there are some inaccuracies, especially south of Helsington which the leader had remapped.

The Park Limestone formed the high ground where the party met, extending northwards along the ridge and down the dip slope to the east. It was a light-coloured, platy-weathering limestone, giving a distinct ring when hammered. It contained much comminuted shell debris with a few larger fossils in this basal section of the Holkerian stage. Walking north, the party crossed the Brigsteer road and followed the Scout Scar footpath upward through the Park Limestone. The various units of this limestone show as low ridges undulating in gentle anticlines and synclines with intervening slacks, probably representing thin shales between the limestone units. The coral *Lithostrotian vorticale* enters the

limestone about 10 m. above its base and is then found throughout. Higher the larger cerioid coral *Lithostrotian aranea* occurs as a marker band at 485,896. A low cliff further north formed a prominent feature; a continuation of this trend (NNW-SSE) linked with the off-set section of the southern end of Scout Scar appeared to indicate a fault with downthrow to the west, bringing basal Park Limestone and the higher horizons with *Lithostrotian aranea* together. The party then returned to Helsington Church and continued to Berry Holme Farm (493,888) where nearby a tabular, cross-bedded sandstone was exposed in a small, overgrown quarry. This horizon, in the Upper Dalton Beds, was unusual in that the contact with overlying Park Limestone was an erosion surface.

After lunch the party met at the southern entrance to Brigsteer Park (489,876), where the road crossed the Dalton Beds to the underlying Red Hill Oolite which formed a prominent step in the lower part of the scarp. The normal Red Hill Oolite is a pseudo-oolite composed of small pellets, that is a pelletal micrite in a rather sparry matrix, and the exposure seen in the woods (487,876) was not typical. An outcrop of the Dalton Beds, proven by the presence of a distinctive brachiopod *Delepinea carinata,* was seen east of the road and some 400 m. north of the parking place (489,877). A short distance to the south a fault trending from Cinderbarrow to Sizergh Castle had a downthrow to the south.

Cars were then parked near Quaggs House Farm (485,867) where, low down on the scarp face, was exposed a section of Upper Dalton and Lower Park Beds. A bed of dolomitic limestone, of widespread occurrence throughout the area, provided a distinctive marker virtually on the boundary between the two beds and therefore between the Arundian and Holkerian stages.

Later stops at roadside exposures on the A.6 examined the Helsington/Sizergh fault, firstly near the roundabout (504,869) where over some 200 m. rocks of the Park Limestone were thrown against Middle and Lower Dalton Beds. The final locality near the motorway link road bridge over the River Kent was reached down a small road on the east of the A.6. (507,865). Here a further extension of the Helsington/Sizergh fault was demonstrated in the river bank south of the bridge.

LECTURES

SEDIMENTOLOGY OF THE St.BEES SANDSTONE IN WEST CUMBRIA

A lecture given on 7 September 1988 at
Lorton Road School, Cockermouth, by
Dr. Brian Waugh of Poroperm-Geochem Ltd.

The discovery in 1974 by Hydrocarbons GB of the Morecambe Bay gas field marked the end of the first phase of the exploration of the Irish Sea. Prior to this exploration, little was known about off-shore Triassic rocks. The Morecambe Bay field is a very large one (5 trillion cu.ft. of gas reserves) with the gas sourced from the underlying Upper Carboniferous (Westphalian) strata.

The St. Bees Sandstones (Lower Triassic age) comprise the lower part of the Sherwood Sandstone Group, previously known as the Bunter Sandstone. The Sherwood Sandstone Group consists of fluvial sediments, varying from the very coarse, proximal, red conglomerates of the south coast (Exeter district) to sand-dominated distal sediments in N.E. and N.W. England. Vast thicknesses (1000m.) of the St. Bees Sandstone were deposited in the rapidly subsiding Irish Sea or Manx-Furness basin, a half-graben with an early Permian boundary fault on its eastern flank. The basin was separated from the Solway-Carlisle basin to the north by the Ramsey-Whitehaven ridge and the initiation of these structures coincided with the evolution of the major North Sea graben and rifting further west associated with the opening of the Atlantic Ocean.

Commencing in the north, the lowest horizons of the St. Bees Shales at Saltom Bay consist of silty mudstones & siltstones containing small calcite concretions, representing calcrete or caliche soils formed during the drying out of mudflats. Inland these horizons consist of brockram wedges interbedded with the siltstones, representing local outwash fans from the high ground of the Lake District which periodically transgressed on to the mudflat margins.

Further south sandstone begins to form an increasing proportion, as seen at Hutbank Quarry, with the units consisting of flat, evenly bedded sandstones which can be traced for long distances and are full of evidence of sub-aerial exposure-soils, dessication cracks, etc. Ascending the sequence, sandstones become thicker and are increasingly dominant until shale horizons are missing. Most of these sandstones are massive and structureless but may have traces of ripple bedding at the top.

At Aikband Quarry, some 200m. above the base of the formation, the sandstones begin to lose their parallelism with the development of undulating erosion surfaces, with mudclast conglomerates immediately above these surfaces. The next 800-1000m. around the Lighthouse cliffs is devoid of shales and is composed of a complex series of sandstones. The most obvious features are the erosion surfaces but different styles of cross bedding and deformation structures can be seen. The palaeocurrent directions here are principally from the south. McKays Quarry shows very thick sandstones with examples of erosion surfaces where the scour contacts cut into underlying beds, representing old river channels scouring into the soft sediments over which the rivers were flowing. Mudclast conglomerates show that muds were being deposited in quiet water localities, then ripped up and reworked to form mudclast conglomerates at the base of the channel. The lower part of the bed above this erosion surface is dominated by planar bedding, produced by rivers flowing at peak flood discharge capacity. These floods spread out vast sheets of sediments with grain size orientated on bedding planes, giving "primary current lineation" (P.C.L.). Above these planar beds, cross bedding indicated different bed forms such as tabular cross bedding representing sandbanks and trough cross bedding formed by crescentic aqueous sand waves, all indicating declining flood conditions. At the top of these units are ripple laminations formed in shallow water, finishing with tiny trough-shaped ripples. Striking parallels with the disastrous floods in Bangladesh were drawn.

In interpreting these deposits it should be remembered that there are two principal fluviatile depositional environments - meandering

river and braided river systems. The St. Bees Sandstone is a classic example of a braided system with its enormous lateral extent covered by a myriad of channels forming an anastomysing network and depositing a wide varieties of sand bars. The British Triassic sequence represents a major river flowing northwards out of the mountains of N. and W. France - forming proximal deposits in S.Devon with the distal phases found principally in the North and Irish Seas.

ASPECTS OF CARBONATE DEPOSITION IN THE CARIBBEAN

A lecture given on 5 October 1988
at the West Cumbria College, Workington, by
Dr. Dan Bosence of the Department of Geology,
Royal Holloway and Bedford New College, University of London.

Dr. Bosence spoke about aspects of his extensive research into contemporary carbonate deposition processes in the Bahamas. Detailed studies of these processes is of great importance in understanding the formation of limestones in the past. It is customary to interpret past environments in terms of the present and this is particularly true of limestone formations but Dr. Bosance explained that his research had indicated that sometimes, especially in the more remote geological past, this principle could be misleading. Carbonate deposition is largely a function of the plant and animal communities present in the environment. These however have varied considerably in the successive periods of geological time, even though the physical parameters may have been much the same. For example, present day reefs are dominated by corals but at different times in the past stromatoporoids, algae, sponges and large bivalves have all been dominant reef builders.

Dr. Bosence explained that he and his team had made detailed studies of carbonate deposition over a number of years in the reef areas of the Florida Keys. The methods used included underwater

mapping, not only of the topography but also of the distribution of the plant and animal communities. Quantitative assessment of carbonate production and transport were made in a sequence of environments from fore-reef to back-reef and near-shore areas. The sedimentary history of each locality was investigated by core sampling; the laboratory back-up included electron microscope studies and measurements of mineral composition.

Much of their time had been devoted to investigating the nature of shallow water carbonate 'mounds' that are a feature in back-reef lagoonal areas. These are elongated structures which vary in size, depending upon their location, but may be hundreds or metres in length and up to three metres in height. The mounds seem to be related to depressions in the pre-transgressive Holocene basement. They build up mainly as a result of increased production by carbonate-forming organisms in comparison with inter-mound areas. In addition there is a component of carbonate sediment transported from elsewhere by wave and storm action.

Finally the speaker suggested that mound structures of this nature could be identified in ancient reef limestones and he cited possible examples in Derbyshire, Devon and in Belgium.

In thanking Dr. Bosence for coming to talk to the Society the President, Tom Shipp, commented on the way that a rather difficult and unfamiliar subject had been presented and how, with the help of excellent slides, the speaker had introduced the complexities of fieldwork in a submarine environment.

ASPECTS OF THE GEOLOGY OF THE ISLE OF MAN

A lecture given on 11 January, 1989,
at Lorton Road School, Cockermouth by Mr. E. Skipsey

Threequarters of the Isle is occupied by the Manx Slates with only small patches of Devonian and Carboniferous rocks present on the west and south coasts respectively while drift covers most of the northern plain. A useful excursion guide to the Isle of Man was published by Ford (1987), to which reference should be made for further information

The Manx Slate Group is of Upper Cambrian - Lower Ordovician age and includes a variety of rock types - slates grading into phyllites, flagstones, turbidites and slump breccias. The rocks are folded into a major syncline trending N.E.-S.W. along the axis of the island with secondary cross folds and faults and are well exposed at Kallow Point, Bradda Head and Peel Castle.

The Foxdale granite quarries are cut into the roof of a large intrusion, the Foxdale-Archallan granite of probable Devonian age, which is intruded into the Manx Slate Group. Late stage hyrothermal fluids resulted in kaolinisation within the granite and the intrusion of pegmatite veins around its margins.

The Peel Sandstone outcrops for about 1.5 km. along the coast north of Peel and consists of fluvial sediments laid down in a semi-arid climate, probably during the Devonian Period. A good place to see the sandstones is at Traie Fogog bay where they dip seaward at about 30-40 degrees. Channel and overbank spill sandstones alternate with mudstone clast breccias. Further north at Whitesands Bay, the sandstones show much evidence of aridity and complex structures with load casts and slump folds.

The Lower Carboniferous rocks and associated volcanics outcrop along the shoreline for a distance of 10 km. on either side of

Castletown. The lowest sequence, the Langness conglomerate, is best seen in low cliffs adjacent to the Langness lighthouse where it rests with marked unconformity upon steeply dipping and reddened Manx Slates. Better known is the limestone succession of Arundian to Brigantian age and exposed at Port St.Mary where there are some 50m. of thinly bedded dark limestones with large algal mounds, often with corals and bryozoa colonies.

A series of volcanic rocks of Brigantian age, principally agglomerates and tuffs, occur at Scarlet Point. The presence of pillow lavas and of tuffs interbedded with black, porcellaneous limestone show the rocks to be submarine deposits. The conical mound of basalt at the Stack of Scarlet is believed to be a vent. In addition to the volcanic rocks, a series of olivene dolerite dykes cuts the rocks of the Isle. They are of Tertiary age and "focus" on the Mourne Mountains Tertiary igneous centre in N. Ireland.

The northern end of the island is notable for the 'drift plain' which is entirely of Pleistocene or Recent deposits and some fine glacial sequences can be seen along these northern coasts.

Following this talk, Tom Shipp completed the meeting with a brief account of the principles of seismic surveys, with illustrations to show marine and land seismic arrays, recording procedures and examples of seismic reflection profiles.

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Geology Today 3 (2), 64-69.

B.G.S. 1:50,000 Geological Map - Isle of Man Special Sheet.

GEOLOGY IN COLD CLIMATES

Presidential lecture given by M. T. Shipp
on 8 March, 1989, at the
West Cumbria College, Whitehaven.

Introduction

On moving to Cumbria 30 years ago to take up a teaching post in Workington, Mr. Shipp had met and joined the small group of enthusiasts led by Edgar Shackleton and Charles Edmonds who, in 1962, formed the Cumberland Geological Society. He told how through his interest in mountaineering and skiing he had been drawn, like Agassiz and Buckland in the 1830's, into speculations on the evidence that the climate of Britain had in the recent past been able to support glaciation. The imprint of the glaciation was recognisably stamped upon the landscape although the moving ice streams had now gone.

The President's active re-acquaintance with glaciers came later, starting in the 1970's with family holidays in Austria, Montana and the Canadian Rockies. During the past seven years he has to the Bugaboo Range in British Columbia, Iceland, the Lyngen Alps in Arctic Norway and twice to Spitsbergen in the Barents Sea, far to the north of Norway. It is from these travels that the subject matter of this lecture has been drawn.

What is meant by "Cold Climates"

This is difficult to define, but he considered regions which are unquestionably Arctic in terms of both climate and location within the Arctic Circle, as well as regions outside the Arctic Circle which, due to their situation or mountainous terrain, attract perennial snow and ice cover, and may be described as sub-Arctic.

The Arctic Circle is defined by the inclination of the Earth's axis of spin at $66^1/_2{}^0$ to the plane of its orbit around the sun, marking the zone of summer appearance of the midnight sun. Thus a person standing at 80^0 N on the northern tip of Spitsbergen at midsummer

would find the sun describing an ellipse in the sky, rising from 13^0 above the northern horizon at midnight to 35^0 above the southern horizon at midday. The observer would find that the "midnight sun" occurs at this northern latitude for 135 "nights" between 15 April and 27 August, with the "polar nights" during which the sun stays below the horizon lasting for 122 days from 22 October to 20 February. However, maps showing extent of ice cover on the shallow North Polar seas (Figure 1) show irregular variations from winter to summer, variations dependent upon factors other than latitude and sun's angle. One of the most significant of these factors is the circulation of water in the world's oceans and seas. The driving forces for these

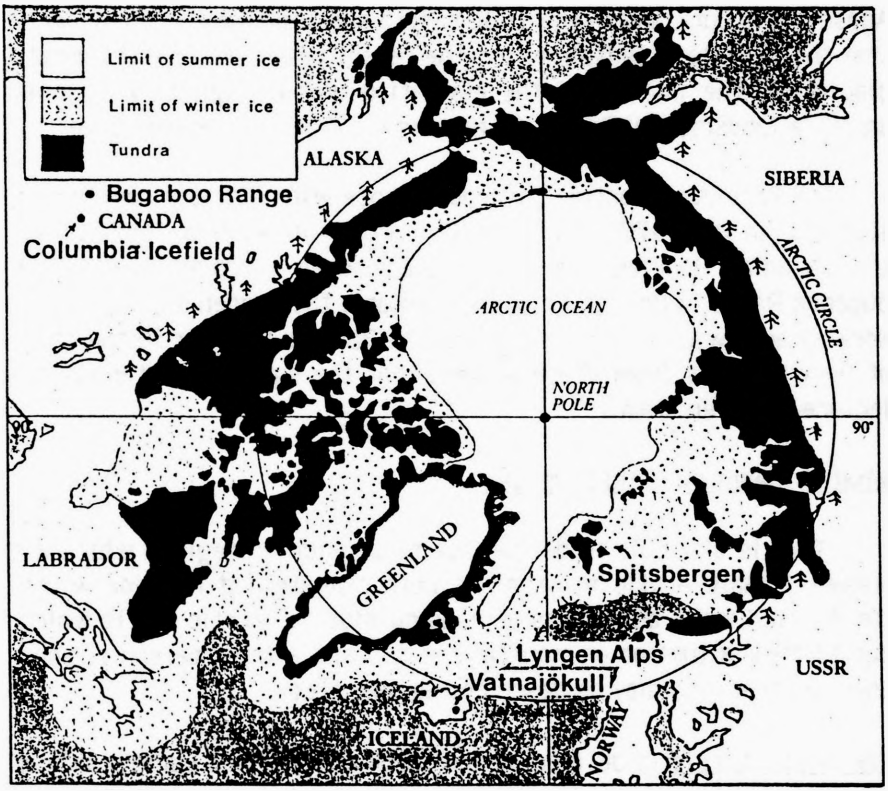


Figure 1. Map showing Arctic regions and extent of ice.
(adapted from 'Kingdom of the Ice Bear', p.8.)

currents is very complex, involving factors such as Earth's rotation, lunar gravitational effects and density variations due to temperature and salinity. The currents vary with depth and are directed by continental and sea-floor rugosities. Nowhere is this effect more strikingly shown than in the Norwegian Sea between Svalbard and Greenland where the relatively warm Norwegian Current (the northern extremity of the Gulf Stream) keeps the west coast of Svalbard ice-free for much of the year, while the cold East Greenland Current swilling southwards from the land-locked Arctic Ocean results in year-round icing up of Greenland's east coast. Thus the distribution of today's climatic zones is dependent upon a plethora of conditions of heat supply and flow, of mass movements of oceanic water and atmospheric gases channelled by irregularities on the Earth's surface.

Outside the Arctic Circle those regions that undergo marked seasonal freezing and thawing are considered as "cold climates". Thus a mountainous area well to the south of the Arctic Circle may be described, due to its altitude, as having a "sub-Arctic" climate.

Snow and ice cover : ice-caps and glaciers.

Even within the Arctic, snow and ice cover is by no means universal; it has been estimated that only 60% of the land area of the Svalbard group of islands, just over 1000 km. from the North Pole, is permanently covered by snow and ice. The amount of snow and ice cover will depend on a number of factors, including :

Precipitation, which is relatively low in the high Arctic and in the lee of mountain ranges but much higher on mountains lying athwart the path of moist prevailing winds, for example in Scandinavia.

Wind patterns, leading to snow drift formation in favourable locations.

Annual mean temperature, dependent upon latitude and altitude.

In cold areas with high precipitation a build-up of snow layers leads to general blanketing of the terrain and the eventual formation of a snow dome or ice cap which may completely mask the underlying topography. An example of this is Iceland's Vatnajokull snowfield. On a much smaller scale are the snow domes only a few kilometres in

extent blanketing individual mountain summits, such as Balgesvarri (a 1622 m. plateau summit) in the Lyngen Alps of northern Norway. Investigation of the area by successive B.S.E.S. (British Schools Exploring Society) expeditions has shed light on the relations between plateau snow domes and the valley glaciers nourished from them. Topographic surveys of Balgesvarri summit were undertaken in 1979 and 1984, with a further expedition to visit the area this summer (Gellatly,A.F. et al. 1986).

A snow pit 15 m. deep was dug in 1984 to sample annual deposition over the five year period and it was found that the snow became increasingly granular and compacted with depth. Each annual layer could be clearly distinguished by its relatively clean base and dirtier top. Irregular lenses of clear ice had developed parallel to the snow surface; it appeared that the fresh powder snow at the surface changed to granular firn or neve by partial melting and re-freezing. The increasing effects of pressure lead to compaction in at least two ways. Firstly by pressure melting at points of contact between snow crystals, with re-freezing taking place in the interstices. Secondly by the squeezing out of entrapped air. Redeposition of ice within the firn also takes place by sublimation, the evaporation of water vapour from small, irregular grains followed by condensation on larger, more rounded grains, a process probably brought about by small variations of partial pressure of water vapour over surfaces of varying texture. The ice lenses may grow partly by this mechanism , which is not too difficult to appreciate in the context of hoar frost, which is effectively "ice-dew" that has crystallised directly from water vapour at temperatures below zero Celsius. Hoar frost is exceedingly common and can result in quite spectacular ice excrescences in mountainous and glaciated regions.

The combined effects of annual deposition, pressure, annual temperature variation, melting, re-freezing and sublimation result in great variations in the rate at which fresh snow is converted to glacier ice. In a temperate, mountainous area where the winter snowfall may be as much as 10 m., the transformation from granular snow may be complete in fewer than ten years, with glacier ice present at less than 20 m. below the fresh snow surface. However in a

very cold Arctic region such as Nordaustlandet in Svalbard, with very low annual precipitation, conversion to ice may take thousands of years and occur at much greater depths.

What can happen to that ice once it has formed ? It depends ultimately on the fundamental structure of the water molecule, which consists of two hydrogen atoms strongly bonded to an oxygen atom. However the three atoms do not form a straight line; the molecule is "bent" at an angle of about 105°. In addition, the electrons forming the bonds are not evenly distributed but are drawn slightly towards the central oxygen atom, with the result that the molecule is polarised. This polarisation causes the molecules to "clump together" in groups of four or five in liquid water (which is a reason why water exists in the liquid rather than gaseous form). However the molecular structure in ice is a theoretically infinite three-dimensional pattern, forming a super-molecule in which the forces holding the H_2O units are about 10 times weaker than the internal H-O bonds. Thus a mass of ice, loaded from above, will begin to deform once the load is great enough to overcome weak intermolecular forces within the ice. This deformation begins at a depth of about 50 m., varying with temperature. If the ice is perched on a sloping surface the deformation pattern becomes more complex as shearing stresses add to static loading; the different possibilities are shown in Figure 3.

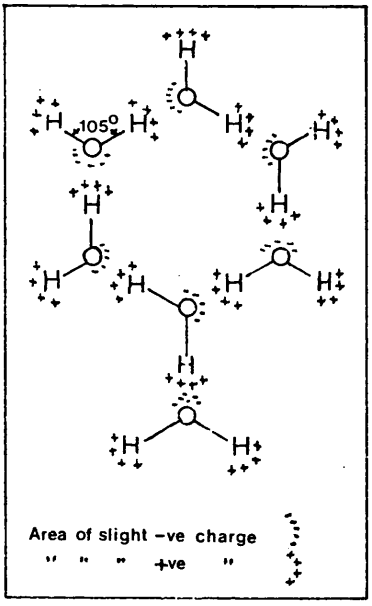
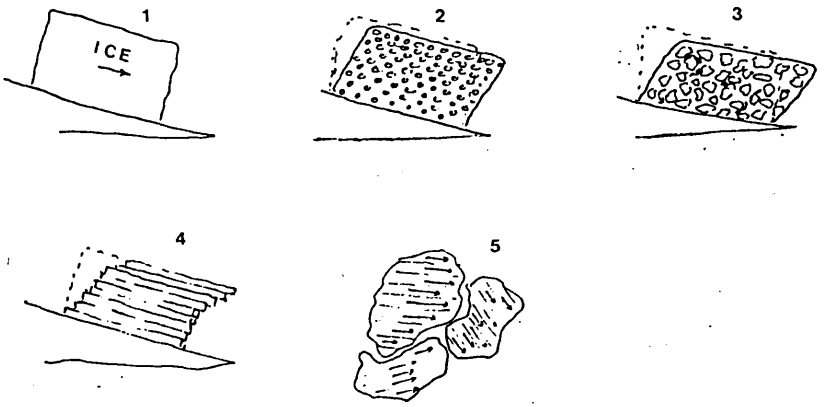


Figure 2. Structure of the water "supermolecule". (note the hexagonal pattern, as in snowflakes)

Glacier flow

Balance between accumulation, movement downslope and eventual wastage by melting can be summarised by the diagram of



Key : (1) single block of ice sliding downslope - contributory factor in the movement of some temperate glaciers where ice base is lubricated by water film.

(2) intergranular adjustments - effect like squeezing bag of dried peas; unlikely as ice crystals tend to be irregular and interlocking.

(3) melting and re-freezing along grain boundaries. (ice $\rightleftharpoons$ water)

(4) internal slip planes - likely when stresses are high, resulting in bands of ice crystals shearing over each other; seen near snout of temperate valley glacier where "active" ice overrides stagnant.

(5) Intergranular gliding & recrystallisation - deformation or "creep" when under sufficient stress, like metal fatigue.

Figure 3. Possible movement & deformation of ice within a glacier. (after Sharp, 1960)

a longitudinal sector through a glacier (Figure 4), which may be matched against a typical valley glacier. In summer the virgin upper slopes of fresh snow contrast with the dirty, fractured lower regions. The surface of the glacier is often supplemented by snow and ice avalanching down from the sides, often forming avalanche cones. In regions where the ice is under tension, as in the upper slopes or above irregularities in the bedrock profile, the ice is crevassed. These crevasses, while forming hazards to people moving on the glacier, seldom extend more than 50 m. into the ice. In the lower reached they often form moulins, open channels into which the surface meltwater cascades before reappearing from under the glacier snout. The crevasse patterns can often be used to indicate the stress fields within glaciers.

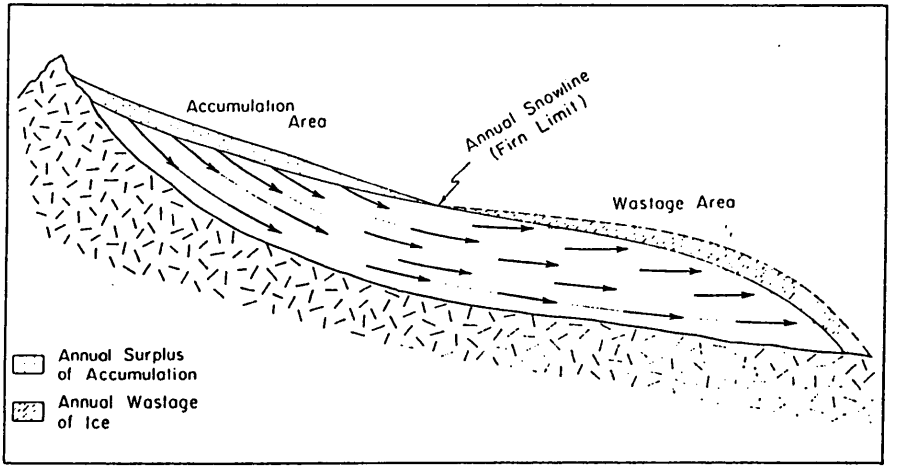


Figure 4. Longitudinal section through valley glacier, showing the accumulation-wastage area relationship and deduced longitudinal flowlines.

The surfaces of most glaciers show a series of distinctly patterned features, of which crevasses form a part. The behaviour of ice under stress may be compared with metamorphism in rocks in terms of plastic deformation involving folding, shearing and recrystallisation.This is particularly evident at two typical locations

in glaciers. Firstly at ice-falls or valley constrictions where the ice-flow "pulsates" during the year, leading to the development of ogives which show on the lower glacier surface as arcuate transverse bands. These should not be confused with the annual depositional layers seen in the high snowfields. The second, and much more obvious feature indicating major flow patterns in glaciers appear as a series of stripes of rock waste on the glacier surface which extend down into the ice. These bands of debris are particularly striking in glaciers fed by a number of tributaries. They can usually be traced back to individual landforms such as aretes or prominent cliffs. Ribbons of debris are often used by mountaineers as a relatively safe highway along the glacier.

The rate of flow of a glacier depends on several factors, including the temperature and state of crystallinity of the ice, lubrication by meltwater, ground slope and the presence or absence of obstructions. Generally speaking, the speed of ice is slower at the edges of the glacier and faster in the centre. The speed can be measured by placing a row of poles in a straight line across the glacier and periodically noting their positions relative to survey points on solid ground at opposite sides of the glacier. Many glaciers flow at an overall speed of less than 1 m. per day, although speeds may increase locally by a factor of ten or more in steep reaches or over icefalls.

Glacier Surges

Glacier surges are interesting phenomena which amply illustrate the delicate balance between formation, transport and melting-away of glacier ice. Surges are perhaps more commonplace than has generally been realised. Meier and Post (1969) identified 204 surging glaciers in western North America alone, while Olav Leistol (1969) reported 10 surges in Spitsbergen and hinted at many more having occurred.Glaciers that surge seem to do so repeatedly, the surge typically taking place over 2 to 3 years with a quiescent interval of 20 or more years. During the quiescent phase an "ice bulge" appears to form in the middle reaches of the glacier, dammed by stagnant ice lower down. There comes a time when the pressure of the dammed ice cannot be contained and the glacier then surges rapidly and

chaotically, shoving a mass of splintered, crevassed ice forward by anything up to 10 km. The pressure release has been attributed to the physical failure of relatively dry ice (Nielsen 1969) or to processes aided by water lubrication within the glacier (Robin and Weertman 1973). Lubrication-aided surges of glaciers, well known and feared in Iceland and known there as "Jökulhlaups", are usually initiated by volcanic eruption and consequent melting beneath an ice-cap.

Mr. Shipp's interest in glacier surges was initiated by a paper by Lamplugh (1911) in which he describes the effects of a surge of the Sefström Glacier in Spitsbergen, a glacier visited in 1986 and 1987. Work on the glacier and its moraines was undertaken by groups of young members of the B.S.E.S in 1987; preliminary findings of this work have been published by that Society (Gellatly,A.F.,1987).

De Geer reported that between successive visits in 1882 and 1896 he had found that the Sefström Glacier had surged 6 km. forward across the head of Ekmanfjord, coming to rest across the two small islands of Coraholm and Flintholm. Lamplugh's photographs in 1910 showed a stranded lobe of heavily crevassed, decaying ice on Coraholm, some twenty years after the inferred date of the surge. The approximate positions of the ice front has subsequently been mapped periodically. The periodicity of this glacier is unknown but it would be interesting to see if and when it surges again. A recent surge in the nearby Holström Glacier appears to have gone largely unrecorded.

The glacier margins : meltwater and moraine.

The effects of meltwater and the persistence of morainic and other glacier deposits are the two aspects of former glaciation that are most evident today in many parts of Britain. Such features are often difficult to interpret, due to agriculture, quarrying, construction and the growth of vegetation. Thus visits to the snouts of active glaciers to see processes of glacial deposition and glacifluvial transport and sorting in action helps one to gain an insight into those same processes that happened ten thousand years ago in Cumbria.

A glacier collects, quarries and transports huge quantities of debris, fragments ranging from house-sizes blocks to the finest pulverised rock dust. In Iceland a glacier's load is frequently supplemented by considerable quantities of volcanic ash. The fate of all glacier ice is ultimate melting, either directly by the sun's energy, by contact with warm air or by contact with seawater. Huge quantities of meltwater, the volumes of which vary both diurnally and annually, transport the finer rock flour for long distances while the heavier boulders and cobbles pile at the glacier margin as moraine. Some of the rock waste may only be transported a short distance to infill the valley bottom and give rise to areas of braided rivers. In areas where the ice margin is receding, the annual recession may be marked by low moraine ridges, sub-parallel to the edge of the glacier.

Investigation of the moraines around the head of Ekmanfjord presented considerable problems. The Sefström and Holmström glacier catchment areas lay largely on red sandstones with only the top metre or two thawing during the summer. Consequently the moraine formed ice-cored ridges from which oozed evil-looking streams of red sludge, making access on to the glacier extremely hazardous - even the fjord water was red.

Some effects of arctic cold on the landscape.

Finally a number of features induced by the extreme of temperature in the arctic were described.

Frost-shattering is ubiquitous in Arctic and sub-Arctic regions and often spectacular. On one occasion in Spitsbergen, white gypsum frost-shattered boulders looming through the mist gave members of the 1987 B.S.E.S. expedition a considerable "polar bear" scare! Freezing and thawing of the surface of permafrosted level ground moves stones upwards and outwards into polygons which trail off on slopes to form stone stripes. The gravelly centres of these stone polygons proved convenient tent sites provided that they were well drained, with the surrounding stones used to anchor the tent valances. Ground ice shrinkage during the winter, followed by partial melting during summers, leads to the development of polygonal shrinkage

cracks and associated frost boils. On steeper slopes, partial summer thawing leads to the development of solifluction lobes while individual boulders on slopes sometimes showed evidence of ploughing. Sedimentary strata are often severely frost-riven, with open crevasse-like gulls developing parallel to escarpment edges.

Plant growth is limited in the Arctic by the generally low temperatures, the searingly cold winds and the lack of liquid water due to low rainfall and low temperatures. Consequently tundra plants often show preferential growth along frost cracks which allow both root penetration and a modicum of shelter. Since decay in the Arctic is minimal, subsequent plant growth develops on cushions of dead vegetation leading to the formation of turf hummocks.

Despite their rigour, the world's cold places can be breathtakingly beautiful as the President's many colour slides taken in Arctic and cold mountainous regions demonstrate.

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GEOLOGICAL MAPPING FROM SPACE

A lecture given on 20 September, 1989
at Lorton Road School, Cockermouth, by
Dr. J. McM. Moore of Imperial College.

Dr. Moore said that his talk would review a range of remote sensing techniques used for geological applications and he defined remote sensing as "the science of studying objects out of reach", which involves the use of planes and satellites. The area of the spectrum of most interest was the visible spectrum and longer wavelengths. Most radiations of shorter wavelengths than the visible spectrum were absorbed by carbon dioxide and water vapour in the atmosphere.

When using aerial photographs, geologists analyse them for two broad aspects - texture for structures and colour/shade for rock types. The cheapest applications are with panchromatic black and white film producing patterns of composite greys (30-40 shades) from the natural colours. This type of photography is used for stereo-interpretations. True colour film has some applications as 600-700 shades can be perceived by the human eye but is much more expensive. Film can cover infrared radiations where there is much greater reflectance from vegetation, resulting in more detail being provided in differentiating vegetation types. False colour film is used for infrared filming where the colour bands are shifted for interpretation purposes. The quality of photographs depends upon the resolution of the film, of which there are several different aspects :

a) spatial resolution is a measure of the magnification.
b) spectral resolution concerns the width of the spectral band used as a smaller section will give a finer resolution.
c) temporal resolution - many geological subjects may have crude resolution as the film takes several days to expose.

While photographic images produce the best results, the film has to be returned to earth for processing. TV images are slower to record but can be transmitted by TV camera to base and so are much cheaper to operate. Line scanner techniques are used, as in the

Landsat satellites, where large numbers of discrete picture elements (60m. across) are taken, recognising 256 levels of shade between black and white. These images are then computer processed for interpretation. Landsat multi-spectral scanning provides images in four different colours for each plot of ground. These are then combined for different analyses and give great variety of interpretations. Recent Landsat pictures have picture elements down to 30m. cross-section and up to seven colour bands, giving scope for enlargement, although the complex colours are difficult to interpret. Many examples were given showing different resolutions, such as vegetation cover, textural and structure studies. The individual composites can be combined to form large mosaics covering very large regions, thus revealing broad regional structures.

There are other application which make use of longer wave lengths. Higher infrared radiations can be used to record heat emissions, known as Thermal Imagery. Radar waves are another method; they require a signal to be transmitted by the operator which is recorded on its return. The roughness of the object's surface determines the scatter of the beam and hence the amount of energy returned to the source recorder. The major advantage of radar is that it can be used through fog and cloud cover and at both high and low angles to the subject.

THE WORK OF B.G.S. IN NORTHERN ENGLAND

A lecture given on 11 October, 1989, at
West Cumbria College, Workington, by
Mr. B. Young of British Geological Survey.

The B.G.S. was set up to make and keep up to date the geological maps of Britain. 470 staff are employed at the Survey's main base at Keyworth in Nottinghamshire, with more at other stations. Several specialist units are based at Keyworth, including the Geophysics and the Overseas Divisions, the Survey Library and a Rock Store holding

several million specimens. Other specialist units, such as the Palaeontology, Seismology and Hydrocarbons Divisions are based at Edinburgh, which is also the base for field survey geologists covering Northern Britain.

The twelve field staff based at the Newcastle District Office are responsible for mapping the area extending from the Scottish border to a line between the River Tees and Morecambe Bay. In the Newcastle Office are kept all the 6inch to 1mile field slips of the Primary Survey and the slips of more recent work, together with collections of geological records such as measured and borehole sections; these are available for the public to consult.

While limited work is proceeding in remapping the southern part of the Northumberland coalfield, most effort is devoted to work in Cumbria. The remapping the Cumberland coalfield has commenced in the north but it is intended to cover the whole coalfield, including consideration of the sandstones high in the Coal Measures exposed by the Keekle river diversion.

Other staff are revising the Cockermouth sheet, working on the north Caldbeck Fells and the outcrops of the Eycott Lavas where previous interpretations are likely to be revised. The Eycott Group appears to rest unconformably upon the Skiddaw Group and supposed interdigitations are now thought to be sills. The Eycotts are almost vertical and represent a thickness of 2 - 3 km. Some features are similar to the Borrowdale Group with which they may be coeval.

Other work on the Skiddaw Group has been undertaken around Buttermere and the Vale of Lorton and the 1:25 000 map NY12 is in the Drawing Office. The Skiddaw Group has been divided into northern and southern units, separated by the Causey Pike thrust which may east to Cross Fell. In the Crummock Water area, a band of metasomatised and bleached Blakefell Mudstone has been related to a long underlying intrusion, which may have provided a mechanism to leach and concentrate many Lake District minerals.

Much attention has been given to the Borrowdale Group, in the area

of the old Ambleside sheet where Leeds and Sheffield Universities have research contracts to remap the district around Scafell. It is now believed that the whole of the Borrowdale lavas could be the product of one small strata volcano, with active faulting and slumping occurring during eruption. Marine mudstones found in Holehouse Gill, Ulpha, were found to contain datable microfossils, showing that the B.V.G. here is Caradocian. Since most of the sequence is above this horizon, the B.V.G. is not as old as originally thought.

After discussing a particularly interesting series of exposures around Devoke Water, Mr. Young drew attention to the need to protect such exposures from damage by excessive hammering and urged the Society to take an active interest in the conservation of geological sites in the Lake District.

THE DALRADIAN ROCKS OF DONEGAL

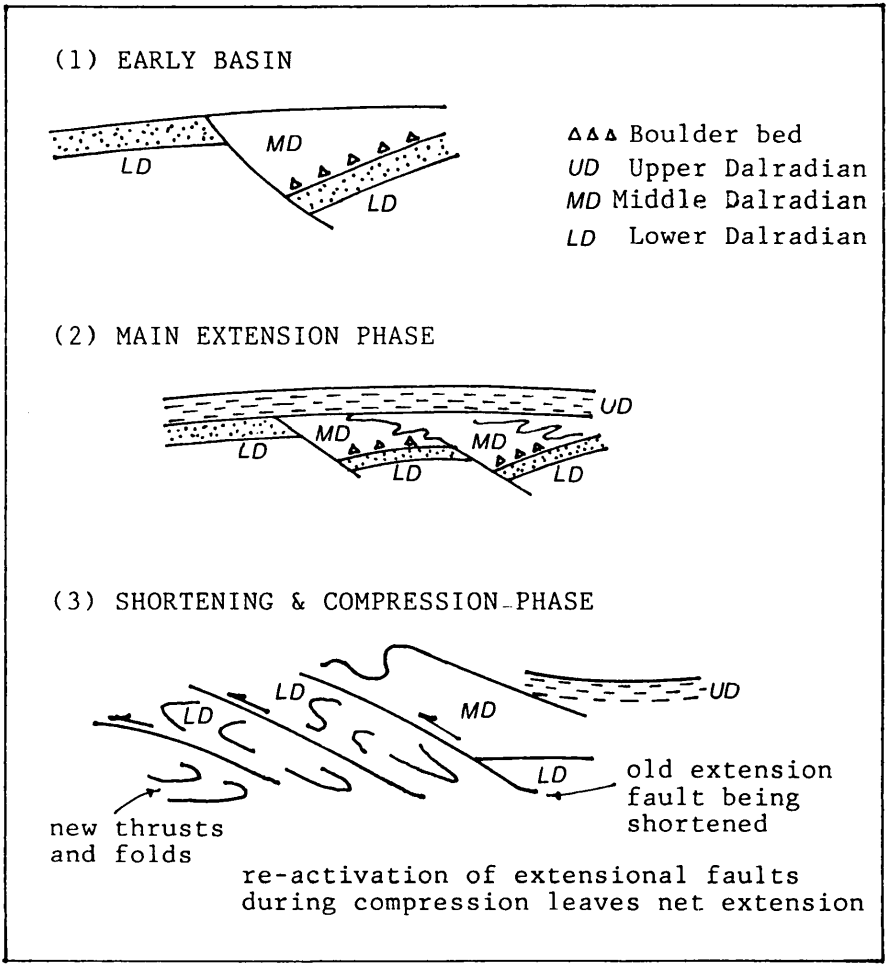
A lecture given on 8 November, 1989,
at West Cumbria College, Whitehaven, by Dr. A. Bell

Dr. Bell introduced his talk by referring to the period when he worked for the Irish Geological Survey when he was introduced to Donegal. He first outlined the geology of Ireland which he characterised as a coastal rim of mountains with a largely flat interior, essentially of Carboniferous rocks.

Donegal is situated in the far north-west of Ireland, extending from Derry and Lough Foyle in the north to Donegal Bay in the south-west. It is a mountainous county with a rugged coastline, containing the highest sea cliffs in Europe, and a large granite complex running NE-SW through its centre. The granite is intruded into rocks of late Pre-Cambrian and early Lower Palaeozoic age which are highly folded and cleaved although of a lower metamorphic grade (chlorite grade) than the Dalradian rocks of Scotland, to which they

show a close resemblance. The Donegal Dalradian sequence is divided into three divisions :

Upper division - similar to the Southern Highlands
Middle division - similar to the Argyll district
Lower division - similar to the Appin district



Diagrams illustrating extensional and compressional phases in the development of the Dalradian structures.

Lower Dalradian consists of fining-up sequences of siltstones and silty mudstones with sandstone horizons, in which sedimentary structures are still preserved, and carbonate horizons of the cementstone type. This succession was laid down in shallow water, probably marine, but no fossils have been found to define the environment or to date them.

Middle Dalradian has a significant marker near the base formed by a boulder bed which is equivalent to the Port Askaig Boulder Beds of Scotland. This unit has been interpreted as a glacial tillite and is associated with other cold climate features such as "varved" siltstones and large, sand-filled polygons. A notable feature is the widespread occurrence of granite boulders of unknown origin. Above and below are further shallow water deposits with prominent quartzite units and minor cementstone carbonate beds.

Upper Dalradian rocks are poorly exposed and consist of greywackes, like those of the Southern Uplands. These are fine-grained turbidites formed in a deep water basin in contrast to the lower parts of the Dalradian which were shallow water deposits.

Late in the Lower Palaeozoic a major orogenic episode occurred - Caledonian Orogeny - which resulted in repeated phases of folding and cleavage. It is not possible to sort out the stratigraphic story until the structures are deciphered. The Dalradian rocks suffered intense deformation with the carbonates in particular being highly mobilised and show very complex deformation structures. These structures are marked by major fractures - not only thrusting where older rocks mount on top of younger ones but slides formed by deep normal faulting. During the shortening and compression the early basin growth faults were re-activated, along with new thrusts and folds, as illustrated. These deep normal faults reached down into the lower crust, tapping the sources of intrusive rocks occurring in Donegal of late Silurian-early Devonian age.

OBITUARIES

Dr. F.H. Day, F.I.C., F.G.S.

Many of the older members of the Society were saddened when they learned that Dr. F. H. Day had passed away on 29th. April 1988. Harry Day was a Founder Member of the Cumberland Geological Society and provided our young organisation with a great deal of support and advice in the early years. Later, after retirement, he went to live in Lindale, in south Cumbria, and immediately became instrumental in stimulating local interest in geology so that in 1973 the Westmorland Geological Society was formed, with Dr. Day as its first President. He held this post for ten years, leading many excursions and each year giving a Presidential Address. Dr. Day retained his interest and relationship with our Society and it is largely due to his influence that the two Societies relate so well together, sharing many members and organising joint activities.

Like so many amateur geologists Dr. Day was a chemist by profession. He was associated all his life with Carlisle and particularly with the Technical College. As a student there, he took an external degree in chemistry at London University and went on to his doctorate. He became a member of the college staff and eventually Head of the Science Department and Vice-Principal. He was introduced to the natural world at an early age by his father, who was President of the Carlisle Natural History Society and together they visited the old mines of Cumbria, especially in the Carrock Fell area. The study of minerals there, together with his training as a chemist, inspired the writing and publication of an authoritative work entitled "The Chemical Elements in Nature", and also developed a deep interest in geology which he taught to degree standard. Dr. Day also taught the subject in Evening Classes at a number of Adult Education Centres in the district. There can be little doubt that his enthusiasm as a teacher and excursion leader brought a life-long interest in the subject to many of his students, not least to the writer of these lines.

Harry Day will be remembered not only for his infectious

enthusiasm for geology, but also for his wide knowledge of all things in the natural world - a well rounded man whom it was a privilege to know. It is especially sad to have to record that on 18th. September, 1989, his wife Mary, who so often accompanied him, also passed away, but not before she had been able to read and approve this notice.

F. J. C.

Mrs Phoebe Roach Wallace.

The Society is sorry to record the death of Phoebe Wallace in 1988. She was one of the first members of the West Cumberland Geology Group and was a staunch attender during the early years of the Cumberland Geological Society. She also extended her interest in geology by attending W.E.A. and University Extramural classes over a number of years.

By profession a talented photographer and an active member of the Workington Camera Club, she and her late husband were widely travelled, undertaking cycling expeditions to Austria and Switzerland. Following his death, she continued travelling to such locations as Egypt, the Far East and to North and South America, until overtaken by increasing infirmity.

T.S.

ANNUAL DINNERS - 1988 and 1989

The 1988 Annual Dinner was held on16th April at Mockerkin Hall Hotel when members and guests enjoyed a very pleasant evening. The guests of honour were Mr. and Mrs. Russell Arthurton of the British Geological Survey.

The 1989 Annual Dinner was also held at Mockerkin Hall Hotel on 1st. April when Dr. and Mrs. John Boardman of Brighton Polytechnic were the guests of honour.

THE 28TH ANNUAL GENERAL MEETING of the Society was held at the Workington College on 22st February, 1989, with the President, Mr. T. Shipp in the chair. The accounts for the year ended 31st. December 1988 were presented and accepted. The Officers and Council were elected for the following year as detailed below.

THE 29TH ANNUAL GENERAL MEETING of the Society was held at the Workington College on 21st February, 1990, with the President, Mr. T. Shipp in the chair. The accounts for the year ended 31st. December 1989 were presented and accepted. The Officers and Council were elected for the following year as detailed below. The President in thanking Officers for their efforts made particular reference to the many years of service provided by two retiring members: Derek Allen as Membership Secretary and to Denys Hinde as a long serving member of Council.

OFFICERS ELECTED FOR :	1989	1990
President	T.Shipp	I.Gray
Senior Vice-President	Dr.F.J.Cockersole	T.Shipp
Vice-President	I.Gray	K.W.Bond
General Secretary	K.W.Bond	K.W.Bond
Hon.Treasurer	F.T.B.Lawton	F.T.B.Lawton
Membership Secretary	B.D.Allen	Mrs.M.Fox
Excursion Secretary	M.B.Dodd	M.B.Dodd
Publications Secretary	T.P.Loftus	T.P.Loftus
Editor	E.Skipsey	E.Skipsey
Librarian	M.F.Burton	M.F.Burton
Council Members	J.D.Hinde	G.E.Curzon
	G.E.Curzon	Dr.T.F.Dias
	Mrs.M.Fox	D.R.E.Dickens
		Dr.R.A.Smith
(co-opted)		D.A.Kelly
		Dr.R.Clark
Honorary Auditor	J.F.Yearsley	J.F.Yearsley

AN EXTRA-ORDINARY MEETING held immediately before the A.G.M. on 21st February, 1990, approved the following rule changes :

1. Rule 20 was amended by the following sentence - 'Past Presidents of the Society will be ex-officio members of the Council'.
2. Rule 27 was amended by the following sentence - 'The President shall hold office for not more than two years. In the event of a successor not being available at the time of the A.G.M., Council may re-elect the present incumbent for a maximum of one further year'.

After the A.G.M. on 21st February, Mr. Chris. Lane described the founding of the Lakeland Mines and Quarries Trust, of which he is the Secretary. The Trust had gained 2000 members in the first three years of its existence and is planning a museum of Lakeland mining and quarrying at the old Threlkeld Quarry. After many questions from members, Mr. Shipp expressed the Society's thanks for an interesting account of an important development.

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