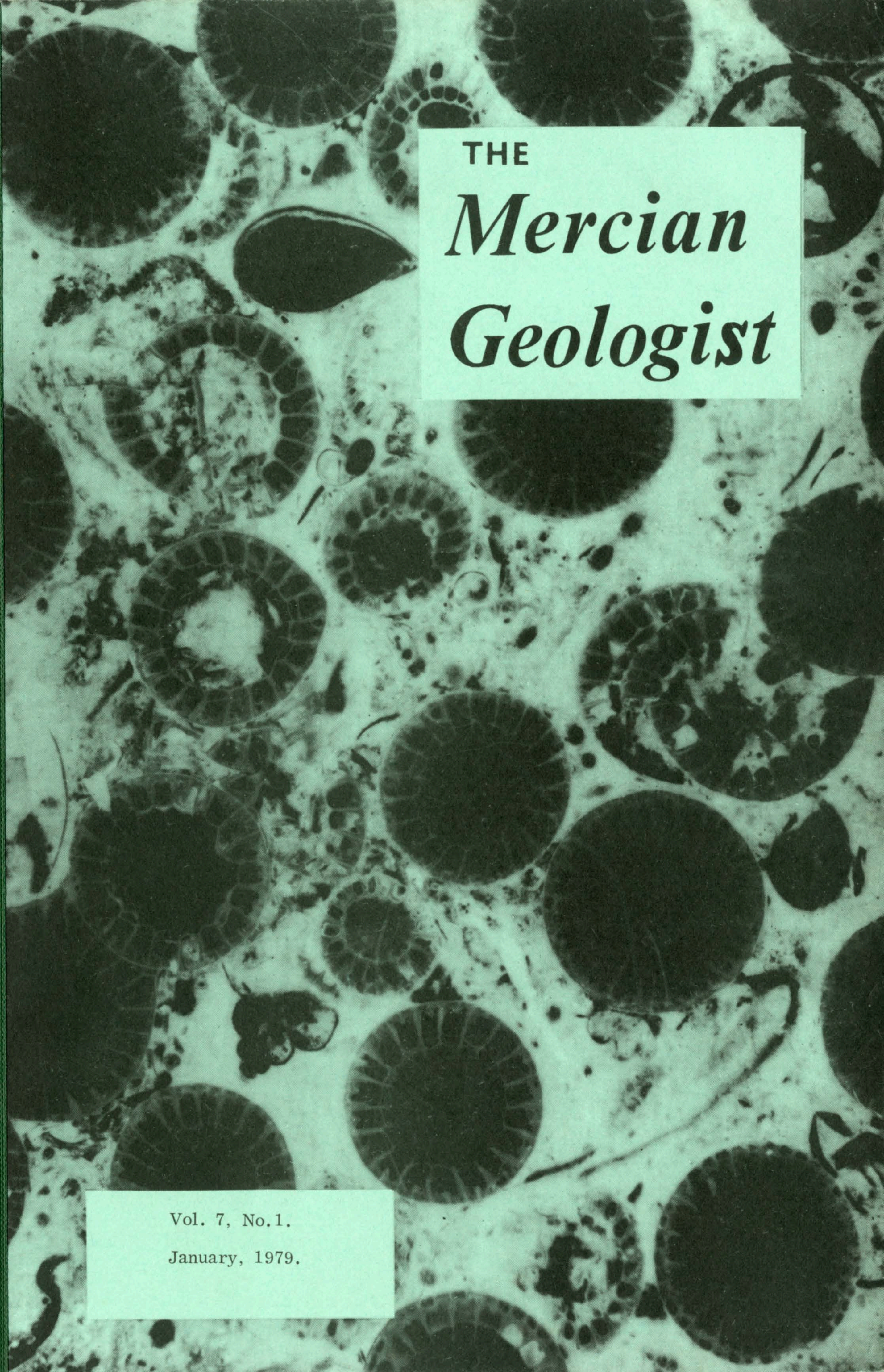

THE Mercian Geologist

Vol. 7, No.1.
January, 1979.

EAST MIDLANDS GEOLOGICAL SOCIETY

THE MERCIAN GEOLOGIST

Editor:
F. M. Taylor.

Volume 7
Number 1
January 1979

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THE EAST MIDLANDS GEOLOGICAL SOCIETY

Council 1978/79

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ISSN 0025 990X

Printed by the Nottingham University Press.

Front Cover: Transverse section of *Diphyphyllum lateseptatum* McCoy X 8 approx. Carboniferous Limestone, Zone D_2, Coombs Dale, Derbyshire.

THE ORIGIN AND SIGNIFICANCE OF DRY VALLEYS IN SOUTH EAST DERBYSHIRE

by

Peter F. Jones

Summary

Dry valleys occurring within the Triassic lowlands of south-eastern Derbyshire are described and their significance discussed. The valleys dissect patches of Wolstonian glacial drift present on the higher ground, and are frequently infilled with 'head' deposits of presumed Devensian age. Particular attention is focussed on a system of dry scarp-face valleys near Dale Abbey, and an attempt is made to trace the physiographic evolution of this restricted area. It is concluded that the valleys were formed largely by meltwater activity in a late Pleistocene periglacial environment. The suggestion is made that niveo-fluviatile processes played an important role in the evolution of south Derbyshire's landscape.

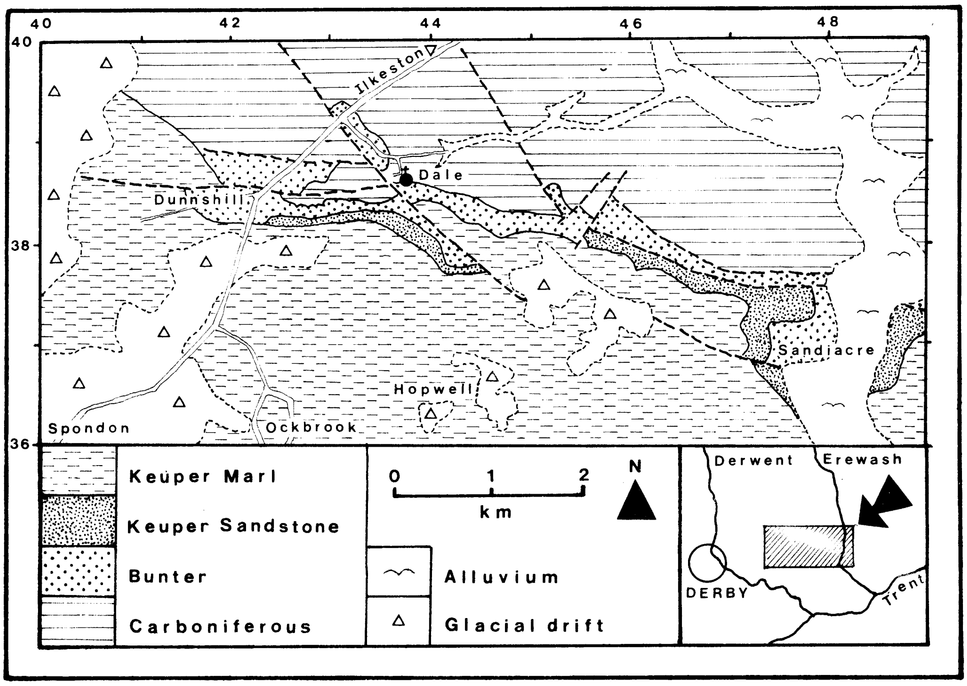
Introduction

A characteristic feature of the south Derbyshire landscape is its intensely dissected nature. Between the four major valleys occupied by the Rivers Dove, Derwent, Erewash and Trent occur extensive drainage networks which affect almost every part of the region. Many of the tributary valleys are either dry or contain misfit streams. Clearly these are indicative of changing hydrological conditions. Elucidation of the reasons for such changes is important to our understanding of the area's physiographic evolution.

In attempting to explain the extensive dry valley systems to be found on the Carboniferous Limestone outcrop of central Derbyshire, Warwick (1964) concluded that the valleys were initiated on overlying impermeable rocks and were subsequently superimposed. He further suggested that intermittent rejuvenation of the main valleys led to progressive elimination of the tributaries, many of which were left hanging. It was thought that much of this adjustment took place before the Pleistocene Period, and that only minor modification was caused by the succeeding glacial or periglacial episodes.

The results of a recent field survey (Jones 1976) of the district south of the Carboniferous outcrop appear to conflict with these views. Here, the scattered distribution of till deposits on the interfluves, as well as the frequency of minor valley incisions, suggests that the area has been subject to considerable subaerial erosion since it was last deglaciated (i.e. during post-Wolstonian time; cf. Shotton 1973). In an attempt to assess the extent and significance of post-Wolstonian dissection in south Derbyshire a sample study has been made of a relatively small area. That selected for specific analysis lies immediately to the east and north-east of Derby and forms the southern part of the Derwent-Erewash interfluve (text-fig. 1). It is an area of Triassic bedrock bounded in the north by the Carboniferous outcrop and in the south by the valley of the River Trent. Within this area attention has been focussed particularly on the system of dry valleys occurring along the Dunnshill (SK 419385) - Dale Abbey (SK 443385) escarpment (see text-figs. 2 and 3) as it is believed that these are typical of similar features developed elsewhere. The origin and significance of the dry valleys is discussed below.

Mercian Geol. Vol. 7, No. 1, 1979
pp 1 - 18, 9 Text-figs., Plates 1 and 2.



Text-fig. 1 General geology of the Dale Abbey area.

Description of the study area

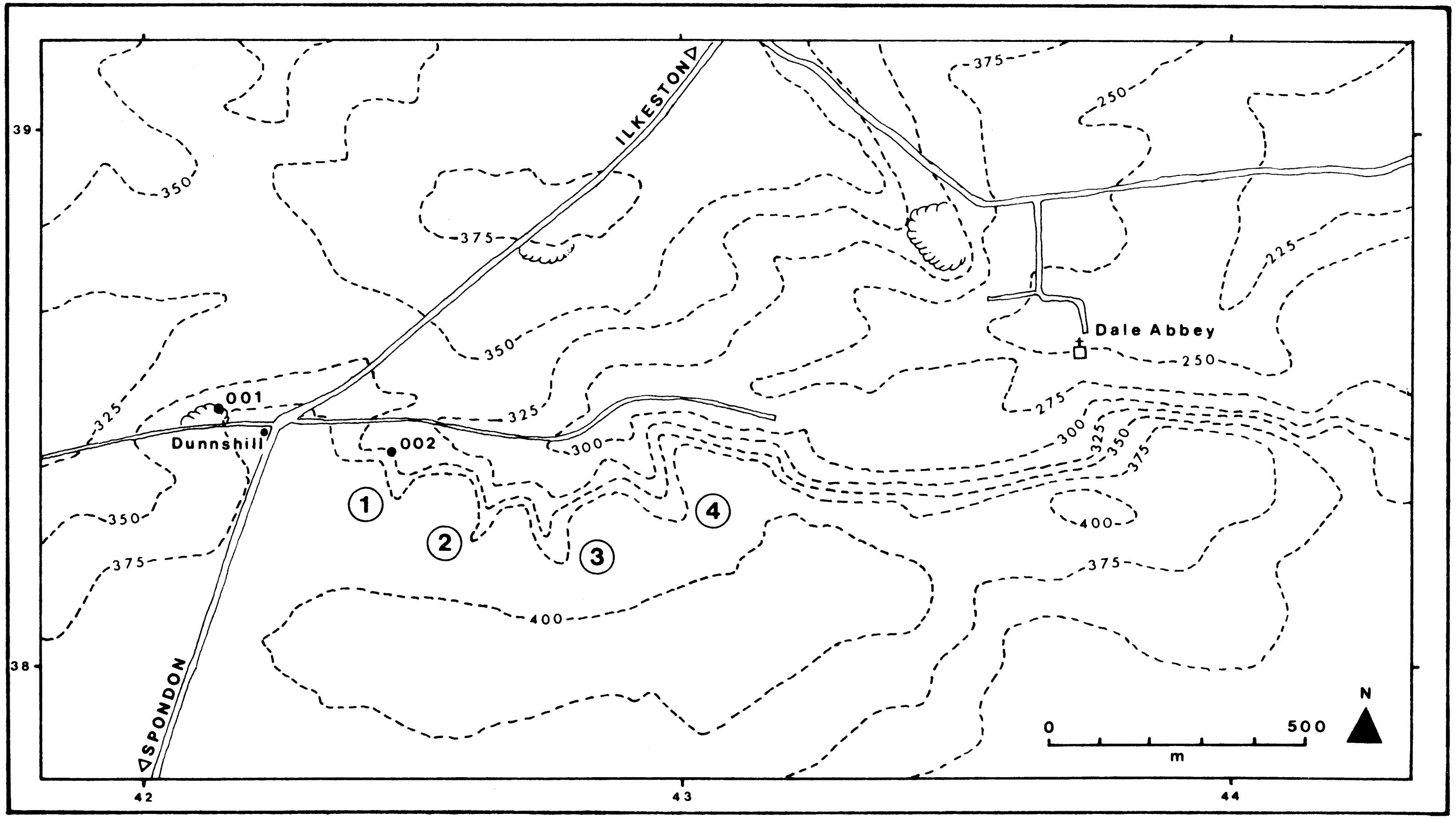
Geology

The bedrock over most of the area outlined above is Keuper Marl, but sandstones of Bunter and Keuper formations outcrop in a sinuous belt extending from Breadsall (SK 3639) in the west, to Sandiacre (SK 4736) in the east. Lithological descriptions of these rocks have been given by Gibson *et al.* (1908), Swinnerton (1948a) and Taylor (1966, 1968), and the local stratigraphical succession is illustrated in text-figure 3.

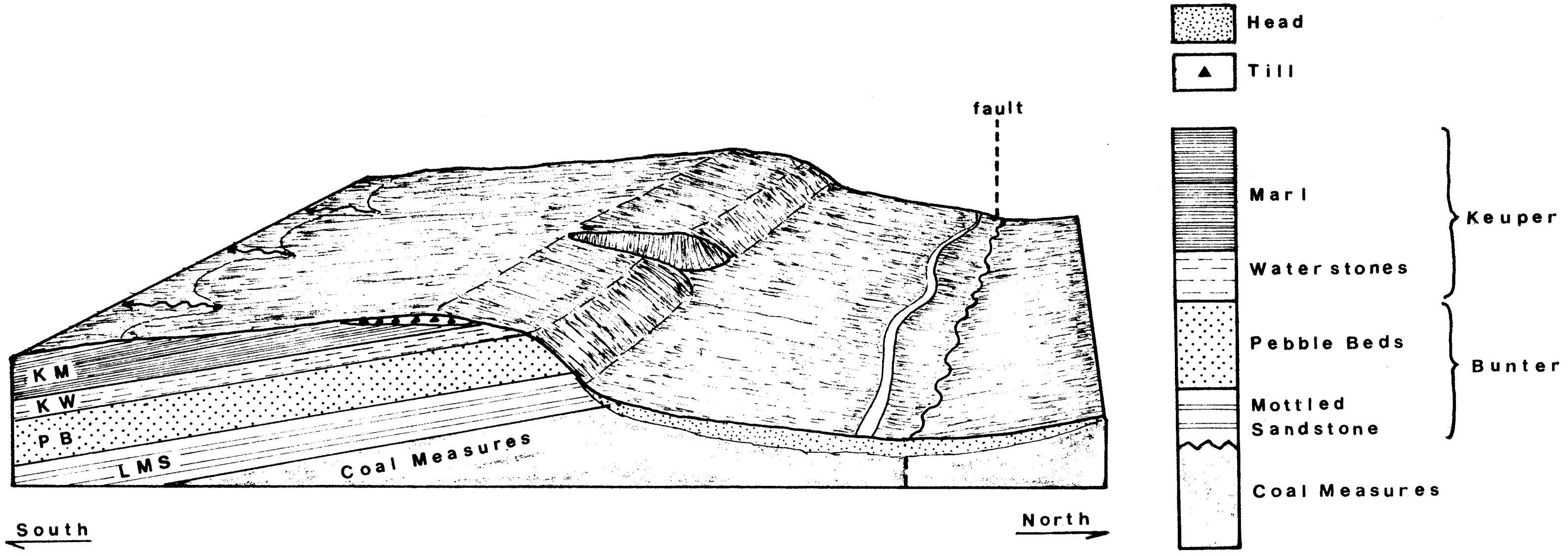
The superficial deposits include the scattered remnants of a formerly extensive till sheet of northern (Pennine) derivation (Jones 1976). Glacial deposits of eastern ('chalky') derivation which overlie Pennine tills in the Trent Valley further south (Deeley 1886, Posnansky 1960) are absent, with the exception of an isolated patch at Risley (Swinnerton 1948b). The Pennine and Chalky Tills appear to be penecontemporaneous (see, for example, Douglas 1974) and both are probably of Wolstonian age (cf. Shotton 1973). Thus the virtual absence of chalky drift on the interfluves, and its presence beneath alluvial deposits in the main river valleys (Jones 1976), must be indicative of a considerable degree of post-Wolstonian erosion on the higher ground. The occurrence of surficial flints in solifluction deposits on the interfluves may also be cited as evidence for the late-Pleistocene degradation of a former cover of chalky drift.

Physiography

The general form of the central part of the Derwent-Erewash interfluve lying between Dunnshill (SK 4238) and Borrowash (SK 4134) is that of a cuesta (text-fig. 3; Plate 1, fig. 1). It has a steep north-facing scarp face aligned E-W and a more gentle southwards-inclined dip slope. Both the Bunter Sandstone outcropping on the scarp face, and the Keuper Marl forming the dip slope, are strongly dissected by minor valleys to the extent that the overlying glacial drift is restricted to isolated patches lying mainly on the higher ground. The overall morphology of the Keuper dip slope may be seen from Ockbrook Corner (SK 418372) on the main Spondon to Ilkeston road (see text-figs. 1 and 4). The road is located on the most elevated part of the dip slope and coincides with a local watershed which separates the minor drainage basins of Chaddesden Brook in the west from the Ock-Brook in the east, (text-fig. 4). Despite the



Text-fig. 2 Location of dry valleys (numbered 1-4) on the Dunnshill-Dale Abbey escarpment.



Text-fig. 3. Schematic block diagram to illustrate the physiography of the Dale Abbey area.
LMS = Lower Mottled Sandstone PB = Bunter Pebble Beds
KW = Keuper Waterstones KM = Keuper Marl

extensive dissection of the ground, the general southward slope of the land surface towards Ockbrook (SK 4235) and Borrowash (SK 4134) is quite apparent. Further east, on the opposite (eastern) side of the Ock-Brook drainage system, the higher ground around Hopwell (SK 4336) makes a marked topographic feature. This may partly reflect the presence of a persistent and comparatively resistant skerry band in the Keuper Marl, but it is also clear that the area is a residual part of the 'initial' surface before this was extensively dissected by the Ock-Brook and more easterly Golden Brook drainage networks. Clayton (1955) depicted the elevated ground at Hopwell as representing a pre-glacial erosion surface remnant, but in view of the capping of till this suggestion must be regarded as rather unlikely.

Drainage

The impermeability of the Keuper Marl is suggested by the abundance of small streams draining the dip slope. During seasons of low rainfall, as for example the summers of 1975 and 1976, the streams become considerably diminished and often dry up completely in their upper reaches. Normally, surface water becomes noticeable between the 90 m and 105 m contour lines and, below this altitude, the streams are permanent under present climatic conditions. Above 90 m O. D. , the streams are intermittent and occupy ill-defined depressions in the ground surface. Many of these natural depressions have been utilised by farmers for the excavation of land-drainage ditches. The overall pattern presented by the system of drainage ditches and natural channels is typically dendritic (text-fig. 4), the various tributaries ultimately uniting to form a single southward flowing 'consequent' stream (i. e. accordant with the bedrock dip) which, at Borrowash, occupies a deep incision in the Keuper Marl.

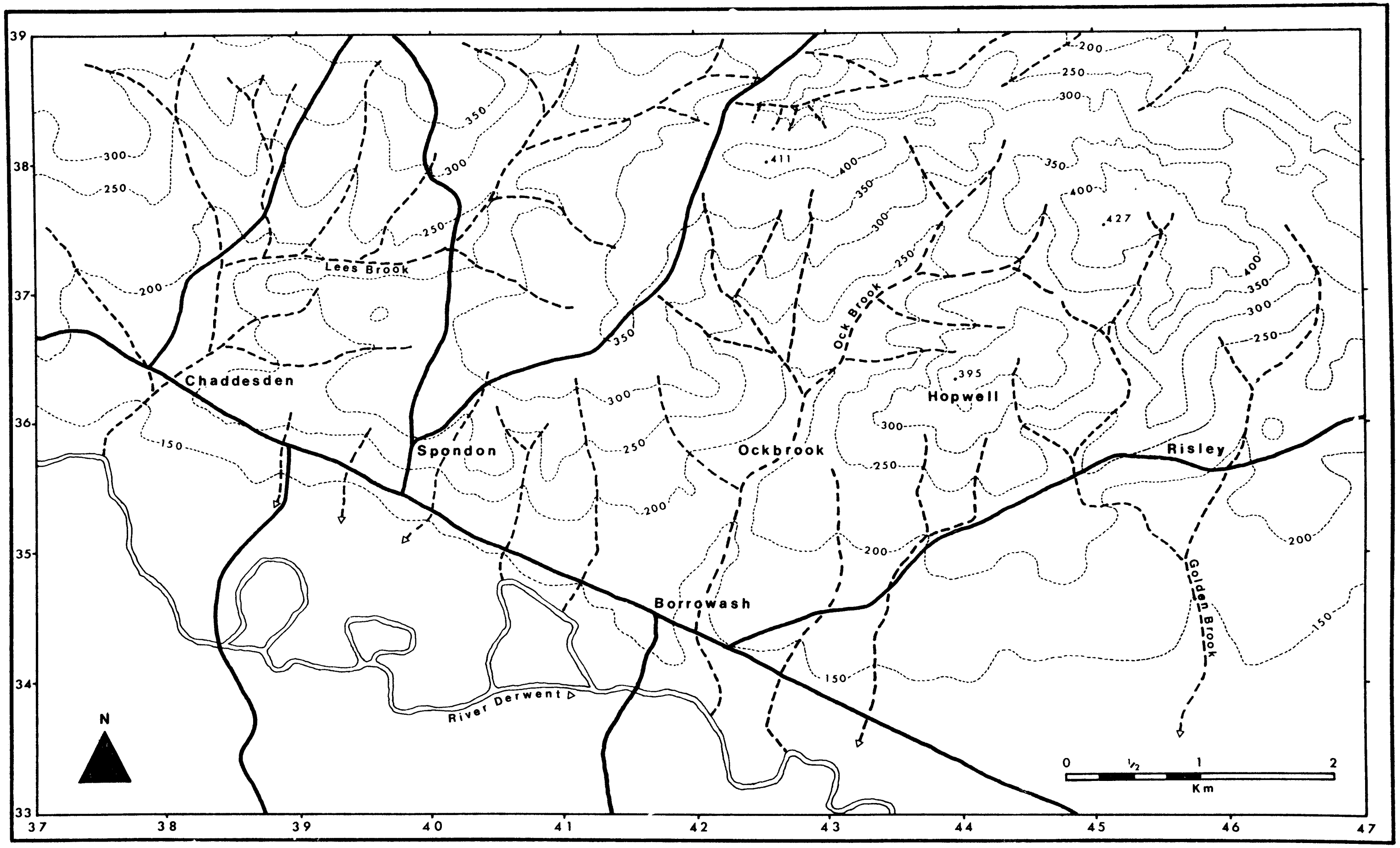
The Dunnshill-Dale Abbey escarpment and its associated dry valleys

Morphology

The road from Ockbrook Corner to Dunnshill is aligned almost parallel to the maximum slope of the ground and rises 15 m in approximately 1 km. At Dunnshill it crosses the faulted outcrop of the Bunter and passes on to the Coal Measures, (text-fig. 1). The Bunter outcrop at Dunnshill gives rise to a minor ridge, up to 7 m high, which dies out westwards in the direction of Locko Park (SK 4038). This disappearance may be due to a combination of causes including faulting, the westward thinning of the Bunter formation, and the unpenetrated blanket of till on Chaddesden Common (SK 3938). Eastwards from Dunnshill, the ridge increases progressively in amplitude, and at Dale Hills (SK 432383) becomes a prominent north-facing scarp face over 30 m high. This feature continues as far as Woodpecker Hill (SK 443384), but further east it is locally offset where the Bunter Sandstone is displaced by a series of NE - SW faults, and in some places the ridge is absent altogether.

Between Dunnshill and Dale Abbey (SK 4338) the continuity of the escarpment is broken only by the presence of four scarp-face dry valleys (text-fig. 2; Plate 1, figs. 1 and 2) and a number of less conspicuous surface depressions. The valleys are short, show incipient meanders, and look relatively youthful. They have steep sides but comparatively flat floors, particularly in their lower reaches (Plate 1, fig. 2). The longitudinal profiles show a rapid increase in gradient upwards causing the overall valleys to terminate abruptly when traced towards the escarpment crest. Morphological details of the Dale Abbey valleys are summarised in table 1 and text-figure 5. Although smaller, these dry valleys show many similarities to dry valleys described by Sparks and Lewis (1957) from the Chalk escarpment near Pegsdon in Hertfordshire.

At the base of the Dunnshill - Dale Abbey escarpment the ground surface slopes gently northwards, at approximately 4°, to a dry scarp-foot valley incised along the line of a fault (text-figs. 1 and 3). This valley increases in size eastwards, and at Dale Abbey contains the small 'misfit' stream of Sow Brook which is a tributary of the River Erewash. It seems clear that the westward extension of Sow Brook to Dunnshill and the associated scarp face tributaries formed part of a previously more extensive drainage system which, for the reasons discussed below, has considerably diminished in importance.



Text-fig. 4 Drainage pattern on the south-eastern part of the Derwent-Erewash interfluve.

Development of the escarpment

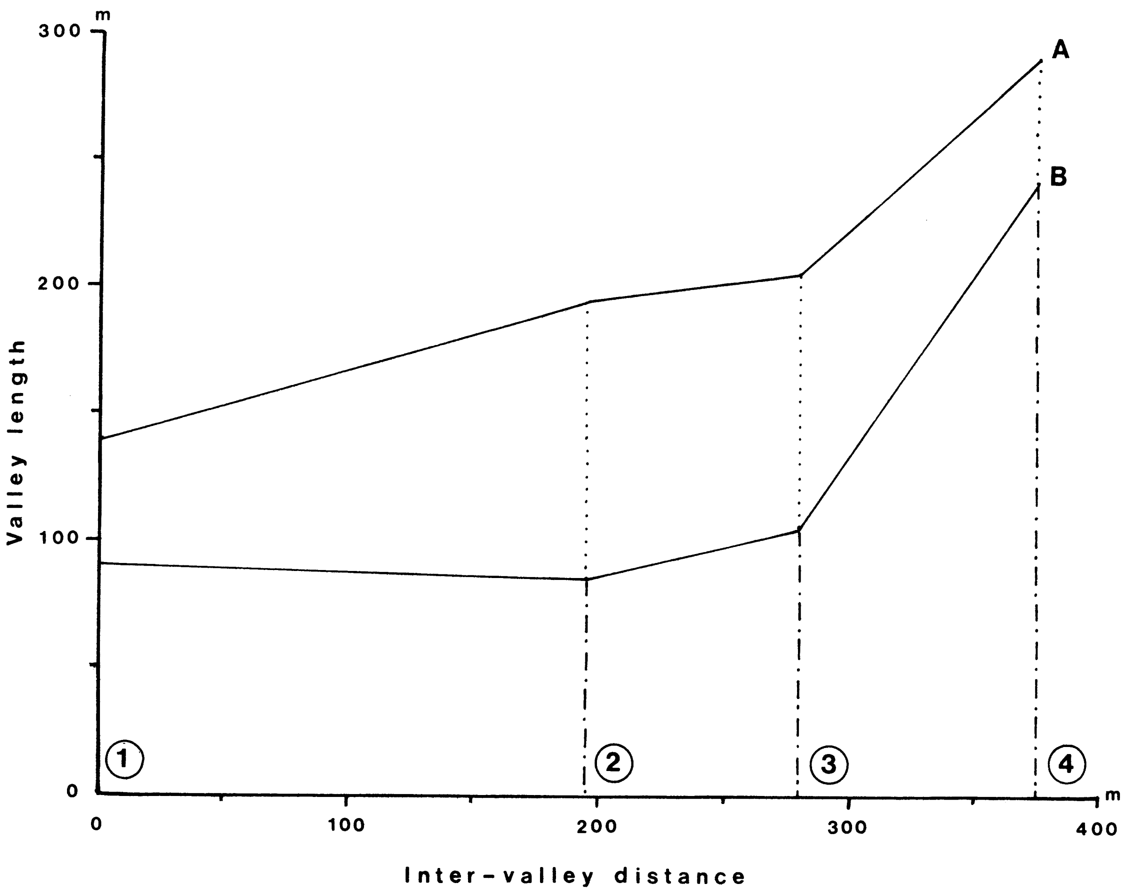
If it is conceded, as has been implied already, that the scarp-face valleys were cut by running water, it might appear that the only real problem is to explain why the streams have disappeared. However, in attempting to account for the presence of dry valleys in the Dale Abbey area it is apparent that certain other factors need to be considered. In particular, the origin of the valleys, as well as that of the escarpment itself must be established. As these features clearly result from processes no longer operative they provide an important clue to an understanding of Pleistocene landscape evolution.

With regard to the escarpment, there would seem to be a number of factors contributing to its presence. There is a close relationship between the alignment of the scarp face and the adjacent fault: both have a general E-W trend and the scarp face departs markedly from this line only at the openings of the dry valleys (text-fig. 5). However, it is unlikely that normal differential weathering of the faulted strata is the primary cause of the escarpment since west of Dunnshill, where down-faulted Keuper Marl lies abruptly against the lithologically distinct Bunter Sandstone, there is little consequent topographic expression. The progressive increase in amplitude of the escarpment east of Dunnshill is suggestive of vertical erosion by an active tributary of the River Erewash lengthening its course by headward erosion. The fault provided a line of weakness which the stream utilised, and it is notable that the present day course of Sow Brook coincides with the line of the fault for part of its length. While vertical erosion by the scarp-foot stream may have accentuated the height of the feature, the subsequent preservation of the escarpment appears to have resulted from a change in hydrological conditions coupled, perhaps, with the effects of a protective capping of impermeable Keuper Marl and Waterstones and the present day vegetation cover.

This interpretation of the initiation of the Dunnshill-Dale Abbey escarpment implies that the present scarp-foot stream, Sow Brook, was formerly more active. The ability of a vigorous subsequent stream to perform vertical, and hence headward, erosion is well known (Small, 1970 p. 236; Sparks 1972 p.140). In the case of Sow Brook, rapid erosion of this type would have been largely dependent upon a falling base-level. Its immediate base-level is, of course, the River Erewash. This river may be regarded, on a local basis, as a consequent stream for it is aligned parallel with the regional dip of the Triassic rocks, and has a discordant relationship with the underlying Coal Measures on to which it was probably superimposed. The Erewash, in turn, is dependent upon the River Trent, along which evidence of late-Pleistocene rejuvenation is to be found in the form of river terraces and buried channels (cf. Posnansky, 1960). It is known that negative movements of sea-level during the Pleistocene period were substantial and relatively rapid (cf. Mitchell 1977). Consequently, even the more distant tributaries of the Trent may have been quite quickly affected. The incision of Sow Brook and the resultant accentuation of the Dunnshill-Dale Abbey escarpment probably took place in response to such a period of rejuvenation.

Initiation of the scarp-face valleys

At some stage during its development, the Sow Brook subsequent stream appears to have acquired its own tributaries from the higher ground of the escarpment. These scarp-face valleys conform to the definition of 'obsequent' for they are inclined in a direction opposite to both that of the consequent stream (cf. Davis, 1909) and the geological dip of the beds (cf. Sparks, 1972 p.10). Originally, they may have developed as seepages in a similar manner to that described by Dury (1959, p. 21). Water percolating through the permeable Bunter Sandstone would have seeped out at the junction with the impermeable mudstone bands in the underlying Lower Mottled Sandstone (text-fig. 6). The seepages developed into springs which, by gradually washing away detritus, worked back into the scarp face to become minor streams. Once channels had been developed these would have acted as preferential routes for rivulets during periods of abundant surface run-off and hence the scarp face valleys may have become established.



Text-fig. 5. Explanation opposite, p. 9, top.

<table>
<tr><th rowspan="2">MORPHOLOGICAL DETAILS</th><th colspan="8">VALLEY NUMBER (see text-fig. 2)</th></tr>
<tr><th colspan="2">1</th><th colspan="2">2</th><th colspan="2">3</th><th colspan="2">4</th></tr>
<tr><td>Length to scarp face (m)</td><td colspan="2">85</td><td colspan="2">84</td><td colspan="2">103</td><td colspan="2">242</td></tr>
<tr><td>Length to scarp-foot stream (m)</td><td colspan="2">137</td><td colspan="2">191</td><td colspan="2">203</td><td colspan="2">290</td></tr>
<tr><td>Distance : scarp to stream (m)</td><td colspan="2">52</td><td colspan="2">107</td><td colspan="2">100</td><td colspan="2">48</td></tr>
<tr><td>Distance between valleys (m)</td><td>▨</td><td colspan="2">194</td><td colspan="2">84</td><td colspan="2">101</td><td>▨</td></tr>
<tr><td>General alignment</td><td colspan="2">168°</td><td colspan="2">200°</td><td colspan="2">187°</td><td colspan="2">142°</td></tr>
<tr><td>Average inclination of valley floors</td><td colspan="2">8°</td><td colspan="2">8°</td><td colspan="2">7°</td><td colspan="2">7°</td></tr>
<tr><td>Average slope of valley sides at mouth of valley</td><td colspan="2">22°</td><td colspan="2">23°</td><td colspan="2">26°</td><td colspan="2">25°</td></tr>
<tr><td>Average slope of valley sides at mid-point of valley</td><td colspan="2">24°</td><td colspan="2">25°</td><td colspan="2">28°</td><td colspan="2">25°</td></tr>
</table>

Table 1. Morphological details of four scarp-face dry valleys occurring along the western part of the Dunnshill-Dale Abbey escarpment.

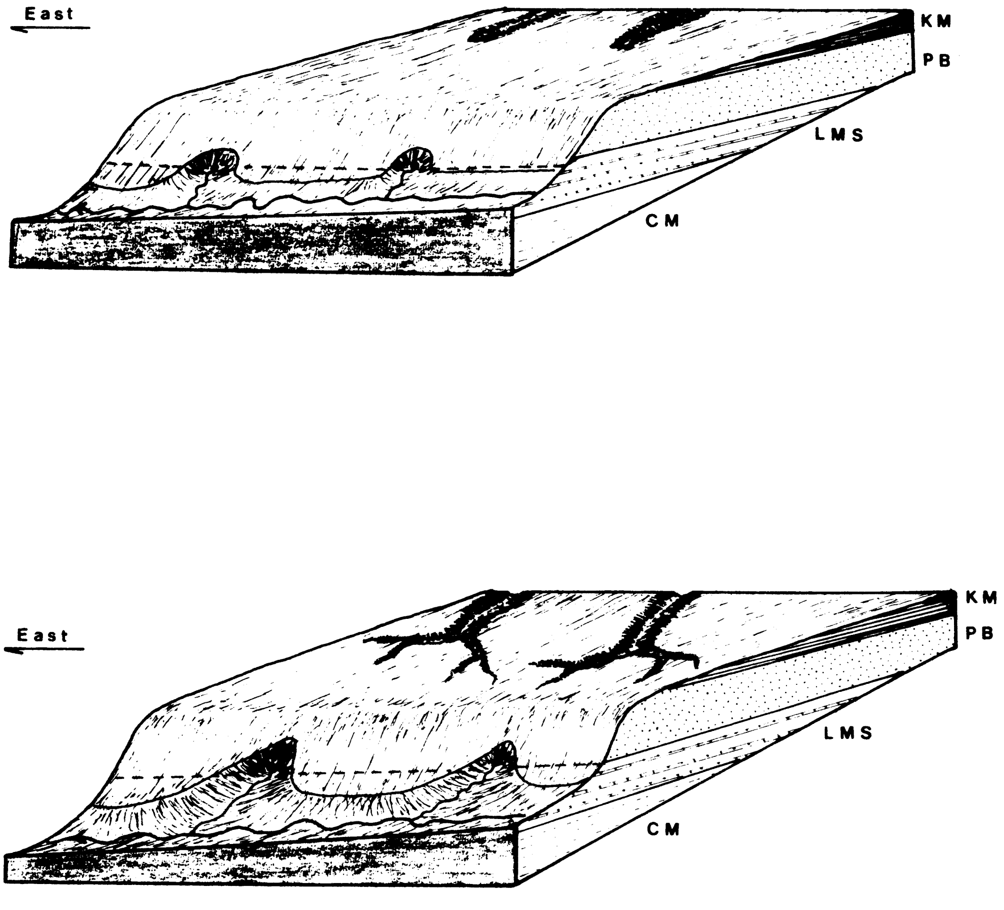
Text-fig. 5 (opposite p. 8, top.)

Graph relating the horizontal spacing of the four dry scarpface valleys to their individual lengths (valley meanders are not indicated).

A - Fault-line stream (Sow Brook continuation) which probably represents the original position of the scarp face.

B - Present scarp-face position.

1, 2, 3, 4. Lengths of dry valleys to stream (A) and present scarp face (B) shown in sequence from west to east.



Text-fig. 6

Possible origin of the scarp-face valleys as groundwater seepages. (for explanation see text)

CM = Coal Measures,
LMS = Lower Mottled Sandstone,
PB = Bunter Pebble Beds,
KM = Keuper Marl and Waterstones.

EXPLANATION FOR PLATE 1

figure 1	General view of the Dunnshill-Dale Abbey escarpment. The marked indentation in the scarp face is caused by emergence of dry valley No. 3 (see text-fig. 2). Terracettes and solifluction hummocks are visible on the lower slopes of the main scarp face.
figure 2	Dry valley in the Dunnshill-Dale Abbey escarpment. This is the most westerly of the four scarp-valleys shown in text-fig. 2. The valley floor is infilled by solifluction material.

fig. 1

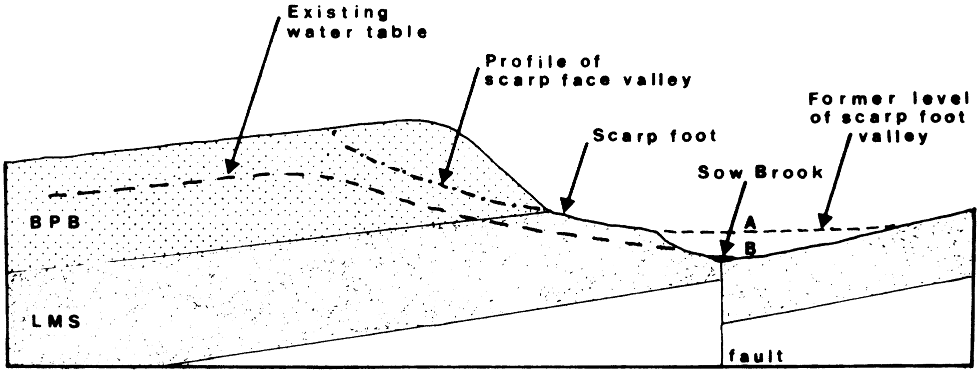
fig. 2

The weakly cemented Bunter Sandstone would have been particularly susceptible to spring 'sapping' and stream erosion of this type. In the absence of a protective cover of vegetation the overall process would have been even more effective. This is apparent at the present day in places where the vegetation cover has been removed as a result of human activity. The loose surficial layers are quite rapidly eroded leaving a bare-rock surface which is subject to direct subaerial activity. Continued rainwash causes erosion channels and gullies to be cut into the sandstone, and during heavy rainfall, sand-flows occur which result in the deposition of detritus as fan-shaped masses at the base of the affected slopes. Examples of recent activity of this nature have been seen in the old quarry at Dunnshill (text-fig. 2, Loc. 001) and adjacent to the most westerly scarp face valley (text-fig. 2, Loc. 002) where the present rate of erosion seems to be remarkably rapid. As soil is washed away, the vegetation of immediately adjacent areas is undermined and eventually removed, thus progressively increasing the size of the affected parts.

Reasons for the valleys becoming dry

The theory outlined above accounts for the presence of the scarp-face valleys and explains why they increase in size towards the east: as Sow Brook lengthened its course by headward recession towards Dunnshill, the most easterly valley would be formed first followed, in sequence from east to west, by the other three. The more easterly streams were thus in existence for a greater period of time and consequently their valleys are longer and deeper. There remains, however, the problem of the present state of dryness.

Of the several hypotheses that have been advanced to explain why valleys run dry during the normal course of erosion (see, for example, discussion by Sparks 1972, pp. 206-14), that developed from the ideas of Chandler (1909) and Fagg (1923, 1939) is the most appropriate for the Dale Abbey situation. The essence of this hypothesis is that as valleys bordering an escarpment are deepened following rejuvenation, the ground water table is progressively lowered so that springs on the scarp face run dry and the emergent streams cease to flow (text-fig. 7) This hypothesis appears to demand that the escarpment remained stationary throughout the process and that down-cutting by the scarp-foot stream was relatively rapid (cf. Sparks 1972, p. 207). It also implies that the valleys became progressively dry from their heads downwards.



Text-fig. 7 Possible application of Fagg's hypothesis to the dry valleys occuring along the Dunnshill-Dale Abbey escarpment. Before rejuvenation the scarp face valley was graded to the scarp-foot stream occuring at level A. After rejuvenation the scarp-foot stream lowered its floor to position B. Consequent lowering of the water table has resulted in the scarp-foot valley being left dry. (cf. Sparks, 1972 p. 207).
LMS = Lower Mottled Sandstone, BPB = Bunter Pebble Beds

Various factors suggest that the Chandler-Fagg hypothesis is not applicable to the valleys described in this paper. These are located near the source of a former scarp-foot stream where any down-cutting resulting from rejuvenation would have been negligible. Although very little recession of the Dale Abbey escarpment seems to have taken place (text-fig. 5), there is no evidence in the longitudinal profiles of the valleys to show that a succession of springheads migrated down-valley. It is notable that the River Erewash is itself a misfit stream and does not appear to have been responsible for the desiccation of its tributaries. Moreover, the fact that misfit streams are common throughout the entire area suggests that there must have been a much greater volume of surface run-off during the recent past and a subsequent change in hydrological conditions. An alternative explanation for the Dale Abbey dry valleys is thus required.

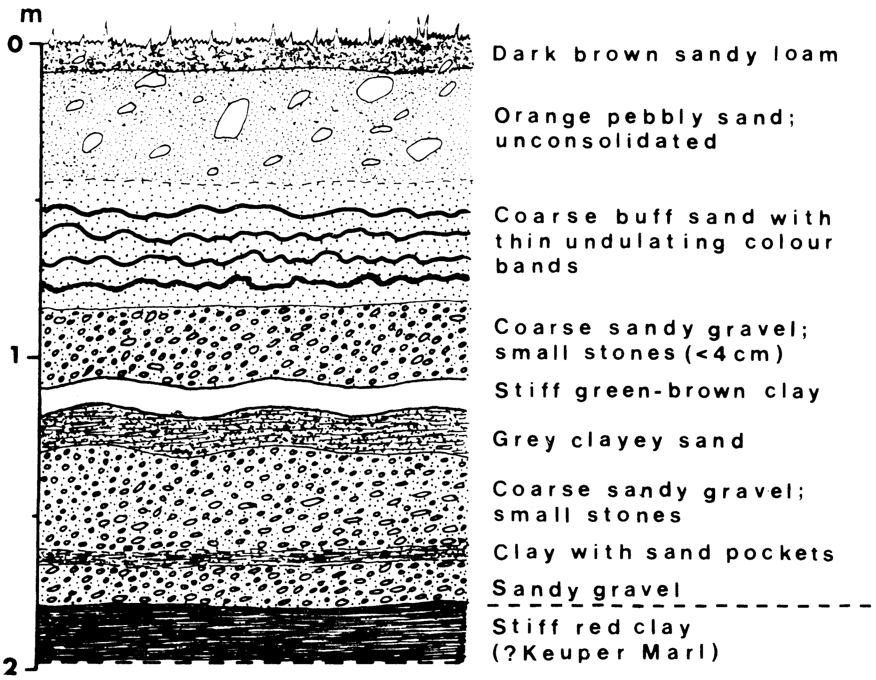
Reid (1887) suggested that certain dry valleys in the permeable Chalk of southern England could be attributed to meltwater flow during the Pleistocene Period. It was envisaged that the valleys formed under permafrost conditions when the bedrock was rendered impermeable through being frozen and seasonal run-off was restricted to the surface. Like southern England, South Derbyshire was not directly glaciated during the last (Devensian) cold stage and the area was subject to a prolonged period of periglacial activity. It seems likely that surface run-off resulting from seasonal melting of the upper ground layers would have been extremely effective in eroding the weakly cemented Bunter Sandstone. Such a mechanism would certainly account for the steep longitudinal profiles of the valleys which the theory of receding spring fails to do. It is possible, of course, that the valleys were initiated by spring sapping, but were subsequently modified by meltwater in the way described.

Against the meltwater hypothesis of dry valley formation must be set the conflicting views of Dury (1959 p.33, 1965) who suggested that many dry valleys in the Chalk were formed during periods of greater precipitation. Dury's evidence for this contention is derived from the existence of surface streams in some of the valleys following unusually heavy rainfall. It is notable that the most easterly dry valley at Dale Abbey also occasionally contains a trickle of water, and a small winding gully is present along its floor. Thus some consideration must be given to the possibility that the Dale Abbey dry valleys developed under a much wetter climatic regime and are not related to permafrost. There is, however, little independent evidence for a prolonged period of increased rainfall. By contrast, the effects of Pleistocene periglaciation are moderately well known (Worsley 1977). Furthermore, there are other positive reasons for believing that the Dale Abbey dry valleys are largely periglacial in origin and these are discussed below.

Evidence of periglacial action

Temporary exposures in the vicinity of Dale Abbey and elsewhere have revealed superficial deposits of periglacial type which are much more extensively developed than might be assumed from the limited amounts of 'head' portrayed on existing geological maps. Without such exposures, interpretation of these deposits would have been difficult. Because of their lithological resemblance to the subjacent bedrock, they have often been regarded as *in situ* regolith. However, the superficial deposits are normally crudely stratified and display other structural differences. At many localities in the region the deposits have been dissected by streams and they do not appear to be forming at the present time. There can be little doubt that these deposits are the product of mass-wasting under a climatic regime different from that operative today, and it is inferred that they are related to periglacial activity during the last (Devensian) cold stage of the Pleistocene period (Waters 1969, Jones 1976).

The subdued ridge of Bunter Sandstone at Dunnshill is capped by a thin layer of unconsolidated sand and gravel comprising a concentrate of rounded quartzite pebbles. Similar pebbles occur in the underlying bedrock but are sparsely distributed. The deposit may be examined in the old quarry alongside the bridle-road to Locko Park, 100 m west of Dunnshill (text-fig. 2, Loc. 001). A more extensive exposure was revealed during road and pipe trench excavations in July-August 1972. These excavations displayed a varied sequence of superficial materials dipping sympathetically with the slope of the ground surface. A measured section taken on the north side of the ridge is shown in text-fig. 8. Almost 2 m of unconsolidated pebbly sands with intercalated sandy clays completely obscured the scarp face and rested on a stiff red clay which possibly represented downfaulted Keuper Marl (cf. text-fig. 1). The volume of superficial material present seemed incompatible with the existing low relief of the Bunter Sandstone ridge at this point, and yet the absence of obvious 'erratic' pebbles did not indicate derivation from a former glacial cover. It is suggested that the Keuper Waterstones formation, which overlies the Bunter Sandstone to the south (text-fig. 1), formerly extended north-



Text-fig. 8 Section in solifluction deposits at the foot of the Bunter Sandstone ridge near Dunnshill.

wards to Dunnshill to make a more conspicuous morphological feature. Extensive solifluction on the scarp face would have resulted in the degradation of this feature, and the deposition of the resultant detritus on the lower slopes. Whereas this activity would have been particularly effective in a periglacial environment, subsequent subaerial modification probably continued for some time. It is likely that slopewash assisted in the formation of the existing subdued relief until the process was arrested by vegetation growth.

Although the solifluction deposits and subjacent bedrock on the scarp face and valley sides east of Dunnshill are poorly exposed, they are occasionally revealed where quarrying and recent farming operations have removed the protective vegetation cover. The most extensive exposure currently available occurs on either side of the most easterly dry valley where there is a 200 m section along the scarp face. Here, the bedrock immediately adjacent to the dry valley is affected by minor cambering (Locality 002, text-fig. 2). In the space of only 10.0 m the Bunter Sandstone is downwarped to the extent of at least 1.0 m, and at the same time is strongly affected by a series of downward tapering joints and fissures (Plate 2, fig. 1). Complete detachment and valley-ward tilting of sandstone blocks has taken place along some of the joints (text-fig. 9; Plate 2, fig. 2) so that the arrangement closely resembles a small-scale version of the dip-and-fault structures described by Hollingworth *et al.*, (1944).

The fissures are infilled with unconsolidated pebbly sand which constitutes disintegrated bedrock. No faunal remains have so far been discovered in the sand, but an apparently similar sand-filled fissure at Stapleford, 7 km to the east, was reported by Swinnerton (1945) to contain numerous bones. The bones represented small rodents, birds, frogs and toads, and included a species of lemming and four voles now extinct. A comparable fauna has been recovered from Langwith Cave, 16 km north of Stapleford, where it was associated with late Palaeolithic implements (Mullins, 1913).

The joints at Dunnshill have a general NNE-SSW alignment, parallel to the trend of the scarp face valley. They are clearly related to the cambering and presumably developed contemporaneously with it. Any suggestion that the valley alignment was predetermined by the joints seems unrealistic in view of their sparse development away from the valley. Moreover, the alignment of these structures does not conform to the pattern of tectonic joints identified in this area by Weaver (1974). Since it is clear that both cambering and fissuring

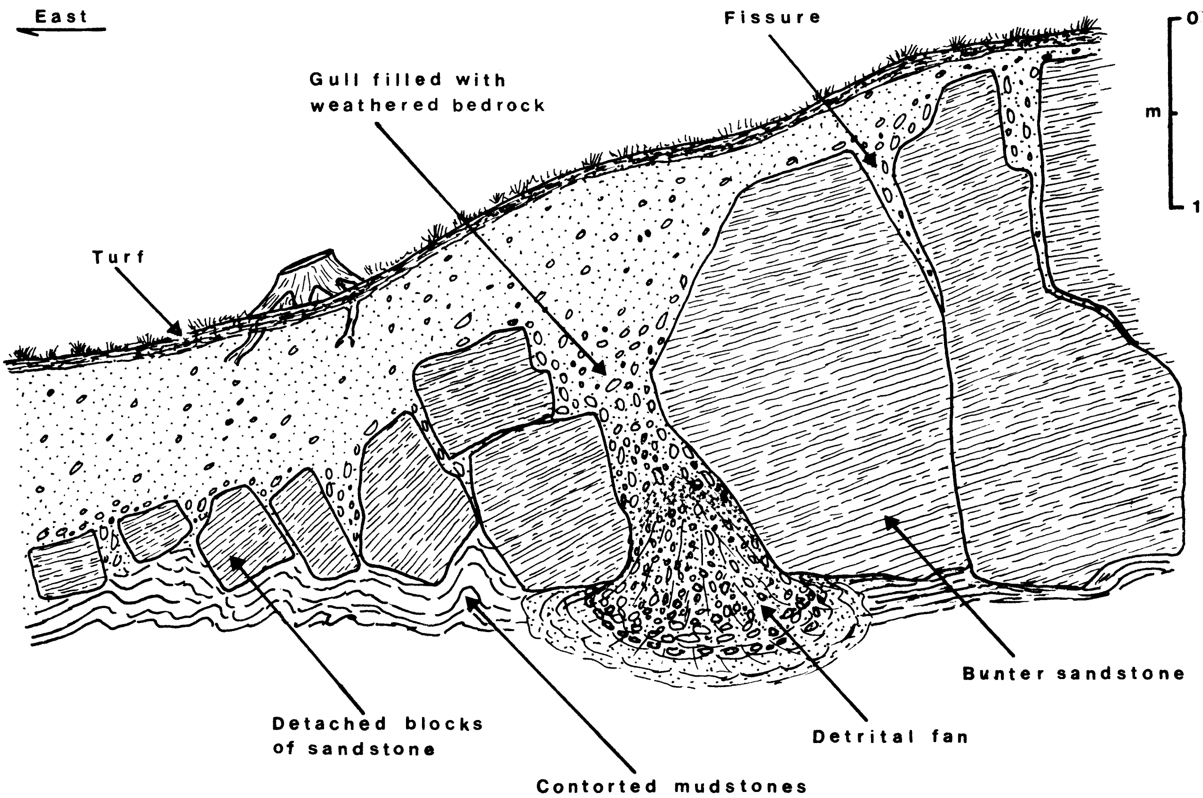
EXPLANATION FOR PLATE 2

figure 1	Cambering and gulling of Triassic bedrock near Dale Abbey (cf. text-fig. 9). Bunter Pebble Beds resting on subhorizontal Lower Mottled Sandstone (centre right) are downwarped into a scarp-face valley (extreme left). Immediately adjacent to the valley the Pebble Beds are strongly fractured and some sandstone blocks have become completely detached.
figure 2	Close-up view of detached blocks of Bunter Sandstone associated with cambering near Dale Abbey (see above). The blocks have a valley-ward tilt indicating their displacement downslope. The gull (centre right) has been infilled with weathered detritus from above (cf. text-fig. 9).

fig. 1

fig. 2

JONES, P. F. - Dry valleys in S. E. Derbyshire



Text-fig. 9 Cambering and gulling of Bunter Sandstone near Dale Abbey (c.f. Plate 2, fig. 2).

occurred in response to instability created by the valley incision, the following sequence of events is postulated:

(1) Initiation of the valley;

(2) Cambering and associated fracturing of bedrock on the valley sides; eventual widening of joints and gravitational collapse;

(3) Weathering and disintegration of surface horizons;

(4) Solifluction or downwash of weathered detritus to infill joints and valley floor;

(5) Initiation of vegetation cover; progressive reduction in the effectiveness of weathering and mass-wasting.

These events are separated only for convenience and it must be emphasised that certain processes probably operated simultaneously.

Although the origin of cambering and associated non-diastrophic structures has been the subject of considerable debate, it is now generally accepted that the majority of documented examples formed in a periglacial environment. Such phenomena appear to be widespread in Derbyshire, and in some cases they occur on a large scale (Buist and Jones 1977). It is not possible to place absolute dates on the sequence of events identified at Dale Abbey. However it seems likely that the processes of valley incision, cambering and solifluction were all intimately associated during the last (Devensian) cold stage of the Pleistocene period.

Interpretation and regional significance

The Dale Abbey valleys may have been initiated as springs during a warmer (interstadial) phase of the Devensian stage, and were subsequently deepened and lengthened by seasonal meltwater flow under permafrost conditions. Eventual melting of the permafrost, and the resultant saturation of the bedrock, would have greatly facilitated cambering and assisted in the mass movement of loose material on the valley sides. As climatic fluctuations since the last (Ipswichian) interglacial are thought to have been considerable (Coope 1975) it is quite likely that suitable conditions for activity of this type occurred on more than one occasion. In north Derbyshire, Waters (1969) has distinguished several 'head' deposits on slopes below 'Gritstone' escarpments, and at Burbage Brook, an intercalated soil horizon (dated 11,590 ± 360 years BP) indicated that periglacial modification of the landscape continued well into Late Glacial times.

Elsewhere in South Derbyshire, dry valleys are found on all types of bedrock in almost every part of the region. Particularly notable are the dry valleys which occur on the Keuper Marl since this rock is sufficiently impermeable to support surface streams under present climatic conditions. Many of these latter valleys are small, shallow features which conform to the definition of 'dells' (Washburn 1973, p. 215-6). Examples may be found on the western side of the Derwent Valley north of Derby (e.g. SK 349388) and between Breadsall (SK 3739) and Chaddesden (SK 3837) on the Derwent-Erewash interfluve. They are invariably floored by solifluction material, and many of the larger features are represented on the geological map (Sheet 125, 1972) as trails of 'head'. Some of the valleys are almost entirely infilled with solifluction detritus and are recognised in the field as shallow linear depressions towards which the ground surface slopes gently from either side over a wide area.

It has already been mentioned that the degree of post-Wolstonian erosion in South Derbyshire must have been considerable. Wolstonian glacial deposits on the interfluves have been deeply dissected and so severely degraded that only where they occupy former depressions are they preserved in any quantity. It is suggested here that much of this erosion took place during the Devensian cold stage and that the increased volume of surface run-off was provided by snowmelt and the seasonal thaw of permafrost. The valley incisions probably increased soliflual activity by creating steeper slopes. They would also have acted as gutters in transporting the eroded detritus to the main river valleys to augment the fluviatile sediments (see Jones *et al.* 1979). Solifluction and slopewash would have continued after the streams became inactive, thus creating the valley infills and the present slope morphology, until these processes were retarded by vegetation growth during Holocene times. Such activity probably played a much more significant role in the evolution of Derbyshire's landscape than has been recognised hitherto.

Conclusions

South Derbyshire has been subjected to considerable erosion in post-Wolstonian times. This is reflected in the scattered distribution of tills on the major interfluves and in the frequency of minor valley incisions. The abundance of dry valleys and misfit streams testifies to changing hydrological conditions. Whereas dry valleys on the Carboniferous Limestone outcrop of Central Derbyshire have been attributed to superimposition from an impermeable cover, those on the Permo-Triassic rocks further south are less easily explained in this way and appear to be indicative of a fluctuating climate. The available evidence suggests that the dry valleys of South Derbyshire are largely a legacy of the Devensian periglacial environment, and that niveo-fluviatile processes played an important part in the evolution of the landscape.

Acknowledgements

The author wishes to thank Dr. E. Derbyshire for his comments on the manuscript, Mr. P. Holdcroft for photographic help, and Mrs. Jean Rivett for typing the draft.

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SURFACE FEATURES OF QUARTZ SAND GRAINS FROM THE BRASSINGTON FORMATION

by

Peter Wilson

Summary

Quartz grain surface features from the sands of the Tertiary Brassington Formation have been examined with the scanning electron microscope. Two groups of features can be distinguished, those inherited from the parent deposit (Bunter Sandstone) and those associated with the weathering, deposition, and/or diagenesis of these Tertiary sediments. These features are described and discussed and representative photomicrographs shown.

Introduction

The examination of the surface features occurring on quartz sand grains as an aid to the interpretation of sedimentary environments is now a well-established technique in the geological sciences. Sorby (1880) was the first to systemmatically classify grain surface markings and to relate them to the different kinds of mechanical and chemical changes affecting the grains. This and similar early studies were limited, however, by the resolving power of the optical microscope and it is only within the last ten years with the advent of the scanning electron microscope (SEM) that the surface features of quartz sand grains have been extensively analysed.

The SEM is an ideal instrument by which to study these grain surface features, and has a number of advantages over the formerly used transmission electron microscope (TEM). With the SEM, grain surfaces are observed directly without the need for replication and thus distortions and artifacts are eliminated. The three-dimensional specimen image is displayed on a television screen and a camera attachment enables representative photomicrographs of the surface features to be taken. More detail is observable with the SEM and the wide range of magnifications (20X to 100,000X) has made this a very valuable research instrument in many scientific fields.

During the weathering, transportation, deposition, and compaction of sediments the detrital quartz sand grains may be mechanically abraded and/or chemically altered. The resulting surface features will tend to be preserved unless a subsequent sedimentary episode intervenes during which modification of existing features or the formation of new ones may take place. Consequently much useful information regarding the history of a deposit is carried by the grain surface features. Through careful examination of these surface features it has been found possible to distinguish between grains from several different environments. These are as follows: (i) source material (i.e. bedrock), (ii) diagenetic, (iii) glacial, (iv) subaqueous, (v) glacial and subaqueous combined, (vi) aeolian, and (vii) high-energy chemical environments (i.e. tropical). The most comprehensive work to-date dealing with these environments and their characteristic grain surface features is by Krinsley and Doornkamp (1973). Despite a number of recent advances in this field and numerous post-1973 papers, the publication by Krinsley and Doornkamp is still regarded as the standard reference for all SEM/quartz sand surface analysis.

The quartz grains discussed in this paper come from the sediments of the Tertiary Brassington Formation now found preserved in solution cavities within the Carboniferous Limestone of Derbyshire and Staffordshire.

Mercian Geologist, Vo.7. No.1.
1979. pp. 19-30, 2 Tables, Plates 3 - 6

The Brassington Formation

The name Brassington Formation was formally proposed by Boulter *et al.* (1971) as an alternative to the previously ill-defined term 'pocket-deposits' applied to the bodies of gravel, sand, and clay to be found in limestone solution hollows in the southern Pennines. This new stratgraphical terminology is based on detailed analyses of the sections in the area around Brassington and Friden. Essentially the formation can be divided into three distinct units:

BRASSINGTON FORMATION
3 KENSLOW MEMBER - blue-grey clays with plant remains.
2 BEES NEST MEMBER - coloured clays and thin sands, unfossiliferous.
1 KIRKHAM MEMBER - coloured sands and gravels, unfossiliferous.

The entire succession is regarded as being conformable and on palaeobotanical evidence the formation is of Lower Pliocene - Upper Miocene (Late Tertiary) age (Boulter and Chaloner, 1970; Boulter, 1971). Detailed maps showing the distribution of the Brassington Formation sediments have already been presented by Ford and King (1969), Boulter (1971), and Walsh *et al.* (1972) and for this reason a similar diagram here was thought to be unnecessary.

Several sedimentological studies have been conducted in an attempt to determine the provenance and environment of deposition for the Brassington Formation. Early workers favoured the Millstone Grit and/or Triassic sandstones as the most likely source materials (Howe, 1897; Bemrose, 1906; Scott, 1927; Fearnsides, 1932;Hughes, 1952. More recently detailed work by Ijtaba (1973) has shown that both these rock types did indeed contribute the bulk of the sediments. In particular it was shown that the main source for the Kirkham Member sands and gravels is almost certainly the Bunter Sandstone, the nearest exposure of which is now 9 km. to the south of the Brassington area at Hulland. A fluvial environment of deposition from a retreating Triassic escarpment is envisaged on the basis of current-bedding, shallow channelling, and other sedimentary structures displayed by certain exposures (Ford and King, 1969; Ijtaba, 1973; Thompson, 1977 pers. comm.).

Aims and Purposes of the Study

Three main aims and purposes for this study of the surface features on quartz grains from the sands of the Brassington Formation can be distinguished:

(i) Other sedimentological evidence has indicated a fluvial environment of deposition. It was hoped therefore to identify surface features of subaqueous origin in order to support this evidence.

(ii) The nearest present-day outcrops of source material suggest that the distance of fluvial transport was relatively short (< 9 km.). Surface features inherited from the parent deposit may not have been totally obliterated and some relic features indicating sand grain provenance may remain.

(iii) Intense chemical weathering of Tertiary age has been widely proven by many workers and the presence of gibbsite in the clays of the Kenslow Member reveals that the Brassington Formation was subjected to such weathering (Boulter, 1971; Boulter *et al.* (1971). Traces of this weathering may be present on the quartz sand grains thus providing additional evidence of a Tertiary age and weathering regime.

Fieldwork and Laboratory Techniques

Samples of sand were collected from four silica-sand pits in which the sediments of the Brassington Formation are preserved. These were, the Bees Nest Pit (SK 241546), Kirkham's Pit (SK 217540), and the Green Clay Pit (SK 241548), all of which are near Brassington, and the Kenslow Top Pit (SK 182616) near Friden. It was only intended to sample the sand-rich Kirkham Member of the succession and at the three pits near Brassington this was achieved on the basis of comparison of the individual beds with the stratigraphic details of the formation given by Ijtaba (1973). Those samples collected from the Kenslow Top Pit cannot be easily placed in the succession and it is not known whether they represent strata of the Bees Nest or Kirkham Member. At the time of sampling (July 1977) the available exposures were in very poor condition due to a temporary shut-down in sand excavation; consequently much slumping of the sediments had resulted and it was not possible to locate and sample all the Kirkham Member strata. Fragments of Bunter Sandstone from Hulland Quarry (SK 280455) were also collected (Hughes, 1952; Walsh *et al.* 1972; Ijtaba, 1973). This sandstone is now regarded as the main source of the Kirkham Member sediments and the locality represents the nearest present-day outcrop to the Brassington area. Quartz grains from the Bunter Sandstone were examined to determine the changes, if any, between the surface features of grains from the source rock and those of the Kirkham Member. Details of these sites and the material taken for analysis are outlined in Table 1.

TABLE 1. Details of samples analysed with the SEM.

LOCALITY	SAMPLE CODE.	STRATIGRAPHIC NUMBER (After Ijtaba, 1973).	DESCRIPTION
BEES NEST PIT (SK 241546)	A	12	Orange-brown sand with green clay flakes.
	B	8	Buff sand with orange streaks
	C	7	Buff sand and pebbles.
KIRKHAM'S PIT (SK 217540)	A	3	Red, yellow, and buff sand.
	B	1	Buff sand with orange streaks.
GREEN CLAY PIT (SK 241548)	-	1 (?)	Buff sand with pebbles and orange streaks. At least 5.0 m. thick.
KENSLOW TOP PIT	A	?	White sand and pebbles. At least 0.75 m. thick.
	B	?	Purplish-grey sand with red streaks, 0.22 m. thick.
	C	?	Orange sand with purple streaks. At least 0.5 m. thick.
HULLAND QUARRY (SK 280455)	HBS	-	Fragments of Bunter Sandstone.

In the laboratory the samples were cleaned of adhering materials by a similar method to that described by Krinsley and Doorkamp (1973). This involved boiling the grains in dilute hydrochloric acid solution followed by a thorough washing with distilled water before a further boiling in a stannous chloride solution. A final washing in distilled water was carried out beofre oven drying at 125°C. The isolation of quartz grains from other constituent minerals was achieved using techniques of heavy-liquid separation and the grains were then sieved to the five standard sand size fractions (i.e. 2000 - 1000 - 500 - 250 - 125 - 63 μm). Ten quartz grains from each size fraction within each sample were selected for SEM analysis. Theoretically this gives a total of 50 grains per sample but the absence of certain size fractions resulted in the examination of 50 grains from two samples, 40 grains from six samples, and 30 grains from one sample. A total of 370 quartz grains have therefore been examined. Individual grains were mounted on small aluminium specimen stubs, using double-sided adhesive tape, and coated with gold in readiness for the SEM work. The fragments of Bunter Sandstone were cleaned in the same way as the individual grains but mounted with Durofix adhesive. Using these small rock particles 'tens' of grains could be examined but it was not possible to accurately determine their size.

Surface feature identification was based on a comparison with previously published photomicrographs of such features and the degree of grain roundness was determined using the Powers' Roundness Scale (Powers, 1953). Basic analysis took the form of recording the presence or absence of 15 diagnostic surface features on each grain and then calculating the percentage frequency occurrence of these features for each grain size fraction (Table 2). This list of features is a modification of that suggested by Margolis and Kennett (1971) and Margolis and Krinsley (1974) and arises from the grouping together of certain features of similar origin and the recent recognition of additional diagnostic features. Categories 1 and 2 refer to the grain's outline, 3-10 are features of mechanical origin, 11-14 are features of chemical origin, and category 15 is a feature of unknown origin although both mechanical and chemical influences may be involved.

Quartz Grain Surface Features

The surface features observed and their percentage frequency of occurrence in all the Brassington Formation samples analysed are summarized in Table 2.

Most of the quartz grains examined display angular or sub-angular outlines. Rounded and sub-rounded grains are also present but are only found frequently in the Kenslow Top Pit and Kirkham's Pit samples. In many cases it can be seen that the angular grain outline is not necessarily an original characteristic; modification of the detrital grains having taken place through the development of euhedral quartz overgrowths (Plate 3, fig.1).

Features of chemical origin on these quartz grains comprise crystallographic overgrowths, diagenetic etching, and oriented etch pits. The secondary growths of quartz are very similar to those recorded on grains in the Penrith Sandstone (Waugh, 1970) and Millstone Grit sandstones (Wilson, 1978 *In Press*). Growth stages in the quartz euhedra are clearly visible. Initial oriented crystalline projections marking the first stage of development followed by merging and overlap to produce larger crystal faces are shown in Plate 3, figs. 2 and 3). With continued growth large areas of the detrital grain surface become masked by well-developed prism (*m* and/or rhombohedral (*r* and *z*) faces (Plate 3, fig.4). Overgrowths are of common occurrence on grains from the Bees Nest, Kirkham's, and Green Clay Pits but are much less frequent in the Kenslow Top Pit samples (Table 2). Grain surface areas lacking in overgrowths tend to be very rough and irregular in appearance and have therefore been referred to as diagenetic etching (Krinsley and Doornkamp, 1973). Such surfaces are visible where overgrowths are absent as shown in Plate 3, figs. 1-4. Subdued relief and a lack of prominent features characterize these surfaces and a solution and/or precipitation cycle in the history of the grains is indicated. A very distinct cycle of quartz solution is displayed by the presence of crystallographically oriented etch pits and two morphological types of pit can be distinguished. Firstly, oriented V-shaped pits on the overgrowth surfaces of grains from the Bees Nest, Kirkham's, and Green Clay Pits (Plate 4, figs. 1 and 2) and secondly, etch pits that form 'zip-like' trails across the surfaces of grains from the Kenslow Top Pit samples (Plate 4, figs. 3 and 4). The length of these trails varies between 8 and 40 μm with a mean of 18.5 μm, while their width has a range of from 1 to 7 μm with a mean of 3 μm. These two morphologically distinct etch features may be related in that small V-shaped pits can be seen to occur in close proximity to the trails in Plate 4, fig. 3 and 4. The trail patterns may result from the merging of the individual V-shaped pits.

SURFACE FEATURE CATEGORIES

		1 ANGULAR OUTLINE	2 ROUNDED OUTLINE	3 BREAKAGE BLOCKS	4 CONCHOIDAL FRACTURES	5 STEP-LIKE FRACTURES	6 ARC-SHAPED STEPS	7 GROOVES AND SCRATCHES	8 STRIATIONS	9 CLEAVAGE PLATES	10 IMPACT PITS	11 OVERGROWTHS	12 DIAGENETIC ETCHING	13 ETCH PITS	14 HIGH-ENERGY CHEMICAL	15 CRACK PATTERNS
BEES NEST PIT A (12)																
GRAIN SIZE	1000 - 500 μm	100	0	0	10	0	10	0	0	50	0	100	100	30	0	0
	500 - 250 μm	100	0	0	10	10	10	0	0	30	0	100	100	30	0	0
	250 - 125 μm	100	0	0	0	0	0	0	0	30	0	100	100	10	0	0
	125 - 63 μm	100	0	0	0	0	0	0	0	30	0	100	100	0	0	0
BEES NEST PIT B (8)																
GRAIN SIZE	1000 - 500 μm	100	0	0	0	0	0	0	0	10	0	100	100	50	0	0
	500 - 250 μm	100	0	0	0	0	0	0	0	40	0	100	100	20	0	0
	250 - 125 μm	100	0	0	0	0	0	0	0	20	0	100	100	0	0	0
	125 - 63 μm	80	20	0	0	0	0	0	0	20	0	100	100	10	0	0
BEES NEST PIT C (7)																
GRAIN SIZE	1000 - 500 μm	60	40	0	0	20	0	0	0	40	0	100	100	30	0	0
	500 - 250 μm	100	0	0	0	10	10	0	0	40	0	100	100	30	0	0
	250 - 125 μm	100	0	0	10	10	0	0	0	10	0	100	100	0	0	0
	125 - 63 μm	100	0	0	0	20	0	0	0	40	0	100	100	0	0	0
KIRKHAM'S PIT A (3)																
GRAIN SIZE	500 - 250 μm	40	60	0	0	0	0	0	0	30	0	40	100	0	0	60
	250 - 125 μm	30	70	0	20	30	0	0	0	90	0	10	100	0	0	50
	125 - 63 μm	40	60	0	10	10	0	0	0	70	0	40	100	0	0	60
KIRKHAM'S PIT B (1)																
GRAIN SIZE	1000 - 500 μm	60	40	0	0	0	0	0	0	30	0	10	100	50	0	0
	500 - 250 μm	100	0	0	0	0	0	0	0	30	0	100	100	20	0	0
	250 - 125 μm	80	20	0	10	20	0	0	0	30	0	100	100	0	0	0
	125 - 63 μm	100	0	0	0	20	30	0	0	20	0	100	100	20	0	0
GREEN CLAY PIT (1?)																
GRAIN SIZE	1000 - 500 μm	100	0	0	0	0	0	0	0	30	0	100	100	30	0	0
	500 - 250 μm	100	0	0	0	0	0	0	0	30	0	100	100	30	0	0
	250 - 125 μm	80	20	0	0	0	0	0	0	20	0	100	100	30	0	0
	125 - 63 μm	100	0	0	0	0	0	0	0	20	0	100	100	0	0	0
KENSLOW TOP PIT A (?)																
GRAIN SIZE	1000 - 500 μm	70	30	0	0	0	0	0	0	40	0	0	100	20	0	0
	500 - 250 μm	80	20	0	0	0	0	0	0	50	0	0	100	60	0	0
	250 - 125 μm	100	30	0	0	0	0	0	0	30	0	0	100	40	0	0
	125 - 63 μm	70	30	0	0	0	0	0	0	30	0	20	100	20	0	0
KENSLOW TOP PIT B (?)																
GRAIN SIZE	2000 - 1000 μm	70	30	0	0	20	0	0	0	50	0	20	100	50	0	0
	1000 - 500 μm	80	20	0	0	0	0	0	0	30	0	0	100	50	0	0
	500 - 250 μm	60	40	0	0	0	0	0	0	50	0	10	100	80	0	0
	250 - 125 μm	80	20	0	0	0	0	0	0	20	0	10	100	40	0	0
	125 - 63 μm	80	20	0	0	0	0	0	0	40	0	20	100	40	0	0
KENSLOW TOP PIT C (?)																
GRAIN SIZE	2000 - 1000 μm	90	10	0	0	0	0	0	0	30	0	0	100	80	0	10
	1000 - 500 μm	60	40	0	0	0	20	0	0	40	0	0	100	60	0	0
	500 - 250 μm	80	20	0	10	0	0	0	0	20	0	0	100	30	0	0
	250 - 125 μm	70	30	0	0	0	0	0	0	30	0	10	100	20	0	0
	125 - 63 μm	60	40	0	0	0	0	0	0	40	0	0	100	40	0	0

TABLE 2 Percentage frequency occurence of selected surface features on quartz sand grains from the Brassington formation. Based on samples of ten grains per grain size fraction.

Explanation for Plates 3 & 4

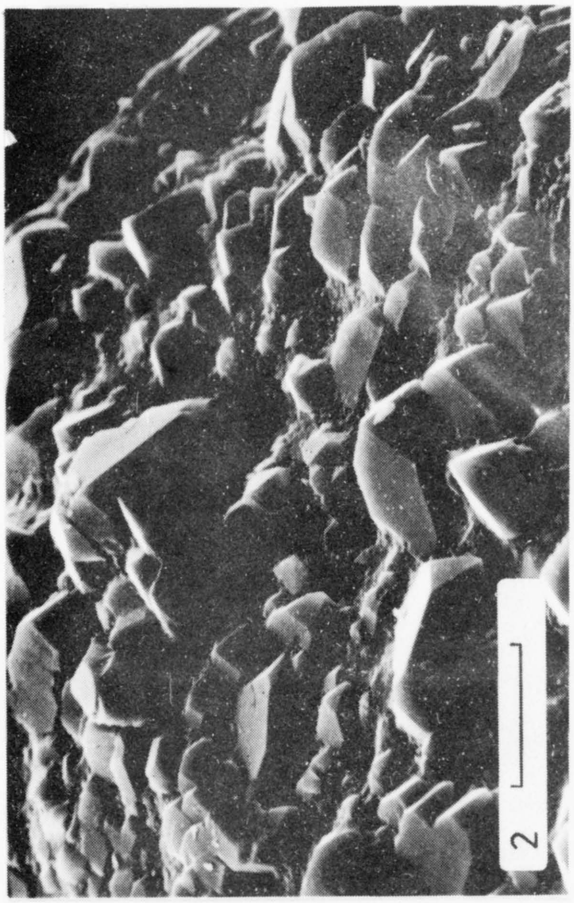
Plate 3

Figure 1	Angular quartz sand grain. Modification of the original detrital grain caused by overgrowths. Scale bar = 100 μm
Figure 2	Initial oriented crystalline projections marking the first stage of overgrowth development. Scale bar = 4 μm.
Figure 3	Merging and overlap of overgrowths to produce larger crystal faces. Scale bar = 40 μm.
Figure 4	The result of continued growth of the quartz euhedra. Large areas of the detrital grain are masked by well-developed prism (*m*) and rhombohedral (*r* and *z*) faces. Scale bar = 20 μm.

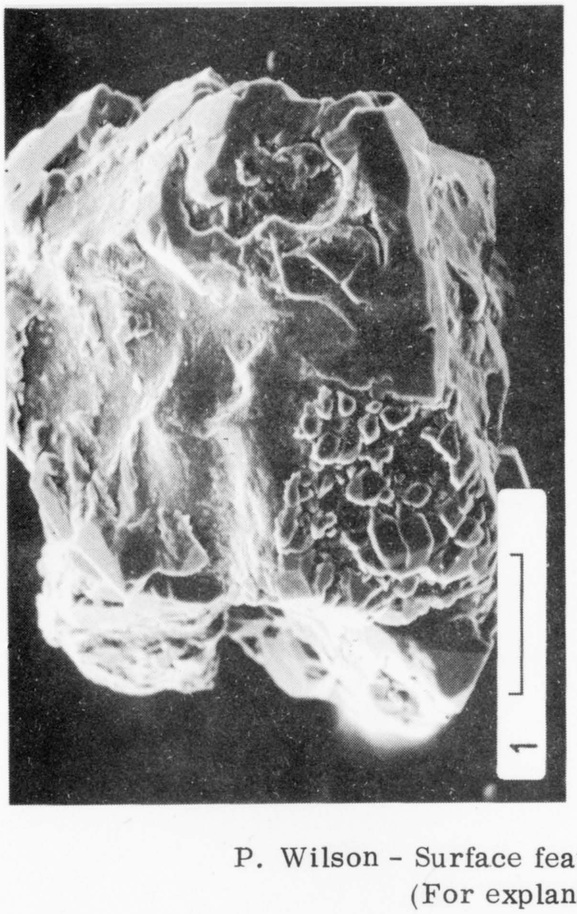
Plate 4

Figure 1	Oriented V-shaped etch pits. Scale bar = 20 μm.
Figure 2	Oriented V-shaped etch pits. Scale bar = 4 μm.
Figure 3	'Zip-like' trails and small oriented V-shaped pits. Note the divergent orientations of the trails. Scale bar = 4 μm.

Plate 3


2


4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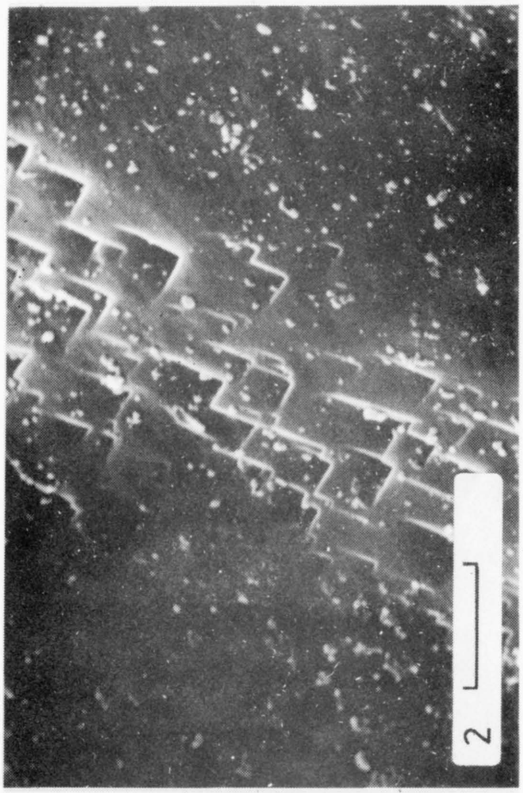
3

P. Wilson - Surface features of quartz sand grains
(For explanation see p. 24.

Plate 4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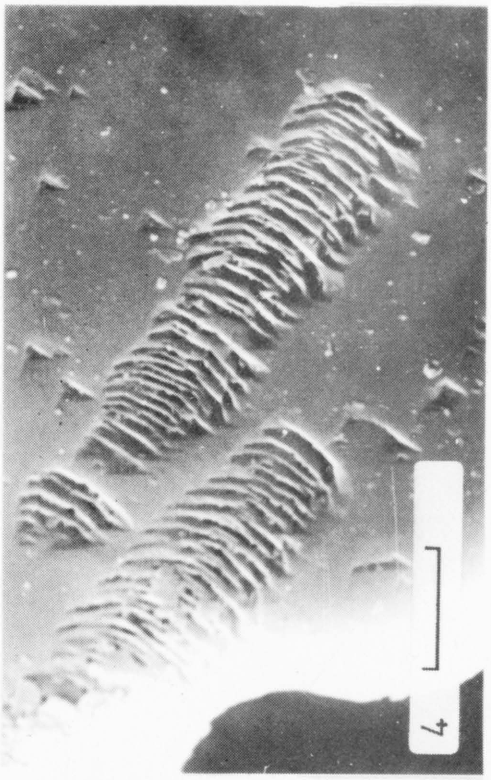
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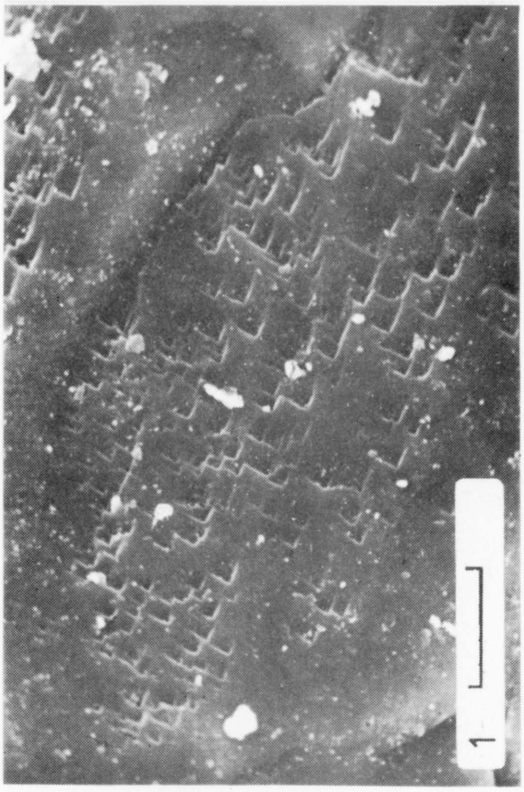
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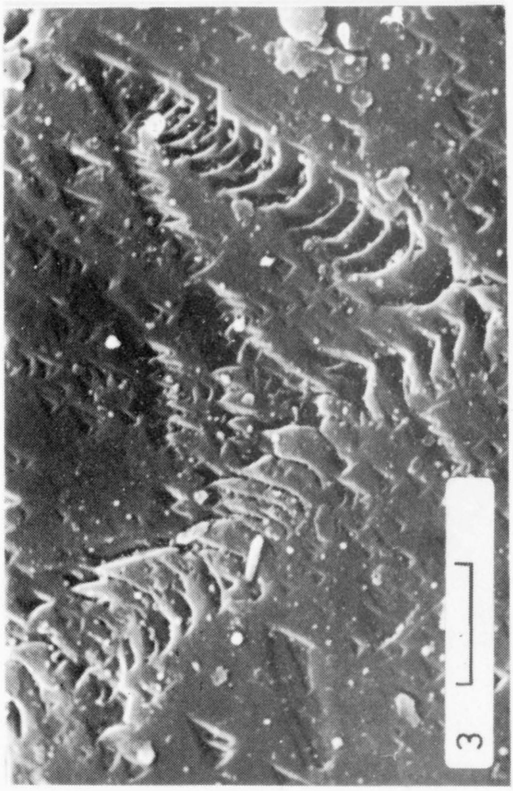
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P. Wilson - Surface features of quartz sand grains
(For explanation see p. 24.

The only mechanically produced surface features observed on these quartz grains are cleavage plates and fracture patterns. Of these the former are the most common although in many cases their presence is related to the degree of overgrowth development. They have only been recorded on those grain surfaces lacking overgrowths. Where overgrowths completely mask the detrital grain cleavage plates have not been observed. Two forms of cleavage are recognized by Krinsley and Doornkamp (1973) and in both cases these features may be described as a series of thin parallel plates, either continuous or discontinuous, running across the grain surface. One type of cleavage may be identified at the grain edges by a stepped or overlapping sequence of plates. These are the grain-edge expression of internal cleavage surfaces. The second type of cleavage is usually observed on those faces unaffected by diagenesis. Here, parallel plates, with depressions between them, oriented at some angle to the grain surface have been termed 'upturned plates'. The cleavage plates observed in this study are of the former type, the 'grain-edge' plates (Plate 5, figs. 1 - 3). A solution/precipitation phase is suggested by the subdued nature of the plates; jagged and sharply irregular plate outlines similar to those produced under experimental conditions have not been found (Margolis and Krinsley, 1974). Although many of the plates are short and discontinuous some can be traced for distances of up to 80 μm across the grain surface. Mechanical fracture patterns comprise conchoidal fractures, semi-parallel step-like fractures, and arc-shaped steps and are rare throughout all the samples examined. In some samples they have not been recorded at all (Table 2). Where they are present, however, they are clearly defined and sharp and fresh in appearance with no sign of subsequent modification (Plate 5, figs. 4 and 5).

An additional surface feature that has only been observed on grains from Kirkham's Pit and also on one grain from the Kenslow Top Pit, is arcuate, circular or polygonal cracks (Plate 6, figs 1 and 2). The cracks vary in width from between 0.1 and 1.0 μm with a mean of 0.35 μm while the maximum dimension of the 'polygons' varies between 13 and 35 μm with a mean of 20 μm. These crack patterns have only been recorded on grain surfaces devoid of overgrowths.

The samples of Bunter Sandstone examined for comparative purposes, show quartz grains of angular outline with crystallographic overgrowths and areas of diagenetic etching, some of which display grain-edge cleavage plates (Plate 6, figs 3 and 4).

Discussion

One of the aims in undertaking this study was to identify surface features of subaqueous origin in order to support a fluvial environment of deposition as proposed by previous workers. But such features have not been observed and this could possibly cast doubt on a subaqueous (fluvial) depositional regime. However, of all the subaqueous environments studied, in terms of their quartz grain surface features, the fluvial environment is the one in which grains are not usually exposed to sufficient abrasion levels to remove pre-existing features (Margolis and Kennett, 1971). These authors have stated that.

> "River sands show few diagnostic features, except that most exhibit rounded outlines, low relief, irregularly finely pitted surfaces, overgrowths and diagenetic etch patterns. Furthermore, large variability in the surface features occurs in sands from different rivers, hence the combining of several samples from different rivers results in a more or less random display of features. This distribution of features is not indicative of events occurring in the river cycle but rather reflects previous transportational and diagenetic history."

Although fluvial sands do not display any characteristic assemblage of surface features, randomly oriented V-shaped impact pits of low density occurrence have been observed on some grains by Margolis and Kennett (1971). This impact pitting is not as prevalent as on beach sands but is an indication of the amount of subaqueous abrasion experienced by a sand grain. The total lack of such subaqueous transport features on quartz grains from the Brassington Formation is suggestive of a short-distance, low-energy fluvial regime. In this respect the evidence for fluvial transport is of an indirect nature.

If the environment of deposition cannot account for the observed surface features on these quartz grains then the features must be regarded as reflecting pre-depositional and/or post-depositional events. It has been shown by Ijtaba (1973) that the sands of the Kirkham Member of the Brassington Formation were derived from the Bunter Sandstone now found to outcrop a short distance to the south of the Brassington area. A comparison of the quartz grains examined from the Brassington Formation with those of the Bunter Sandstone reveals those features of an inherited nature or representing pre-depositional events. These comprise overgrowths, diagenetic surfaces, and cleavage plates. All are present in the sands and the sandstone samples examined. The fact that the overgrowths pre-date the deposition of the Brassington Formation is significant in terms of the short-distance, low-energy fluvial regime suggested. These smooth crystal faces would be ideal sites for the recognition of subaqueous impact features as opposed to the rough irregular diagenetic surfaces. However, no sign of mechanical abrasion has been found on these overgrowths. The arcuate, circular or polygonal cracks seen on grains from Kirkham's Pit and Kenslow Top Pit are also thought to pre-date the deposition of the Brassington Formation. Although they have not been observed on grains examined from the Bunter Sandstone it is considered that the environment under which this sandstone was deposited is more likely to have produced these crack patterns than that associated with the Brassington Formation. Lucchi and Casa (1968), Lucchi (1970), Krinsley and Doornkamp (1973), Baker (1976), and Krinsley *et al.* (1976) consider such cracks patterns to be a principal feature of quartz sand grains from the hot desert environment; an environment associated with Bunter Sandstone formation (Edwards and Trotter, 1954).

The remaining surface features observed on quartz grains from the Brassington Formation, and whose origin has yet to be accounted for, are the mechanical fracture patterns and the crystallographically oriented pits. These features have not been recorded on grains from the Bunter Sandstone and are not considered to be inherited features.

The mechanical fracture patterns may result from the breakdown of the Bunter Sandstone prior to its erosion and deposition as a fluvial sediment or may be due to movement of the grains in the sediment after deposition. It is not thought likely that these fractures are a product of fluvial transport. A fluvial regime incapable of producing V-shaped impact pits is hardly likely to cause grain fracturing. All the observed fractures are sharp and fresh in appearance and do not exhibit signs of subsequent weathering. If these fractures do relate to the breakdown of the Bunter Sandstone then further evidence for short-distance, low-energy fluvial transport maybe implied by the lack of fracture modification through subaqueous abrasion. The possibility that grain fracturing has been induced through movement of the sediment after deposition must also be considered. On the basis of field evidence and laboratory tests a mechanism of solution subsidence has been proposed for the preservation of the Brassington Formation (Walsh *et al.* 1972). It is not known whether a gradual sag of these sediments could have produced the grain fractures but the possibility exists. Although the origin of these fracture patterns is open to question their low frequency of occurrence (Table 2) suggests that their formative mechanism was not of major importance in the history of the deposit.

In the case of the oriented etch pits these features are possibly contemporaneous with sand deposition or may post-date such events. Of the two etch forms recorded, the 'zip-like' trails are probably the more unusual. Pits of the V-shaped variety have been observed by many previous workers but to the author's knowledge etch pits forming 'zip-like' trails across the grain surface have only been recorded on two previous occasions (Bull, 1976, 1977; Friedman *et al.* 1976). Bull (1976, 1977) reported the presence of these trails on the euhedral crystal faces of quartz grains incorporated in cave sediments. The relevant point here for the Brassington Formation is that a fluvial environment was also involved in the deposition of these cave sediments. Whether such trails result from solution by fluvial waters is not known due to a lack of data concerning the chemistry of these streams. If the etch pits are not associated with fluvial quartz dissolution then a post-depositional phase of chemical weathering possibly caused by migrating pore space waters would seem likely.

In addition to the aims of determining the environment of deposition and the source material for the sands of the Brassington Formation it was also hoped to identify surface features associated with chemical weathering under Tertiary climatic conditions; the formation having been shown, on palaeobotanical evidence, to be of Lower Pliocene - Upper Miocene age and gibbsite having been discovered in the clays of the Kenslow Member (Boulter and Chaloner, 1970; Boulter, 1971; Boulter *et al.* 1971). The floristic evidence indicates a warm, oceanic climate whereas the presence of gibbsite suggests extreme tropical weathering. This aspect of the study has been somewhat unfruitful, however, for there is no evidence to support any form of post-depositional high-energy chemical weathering except for the etch pits already described and discussed. Features associated with intense surface disintegration, recorded by Doornkamp and Krinsley (1971), have not been recorded here. Indeed, almost all the grains examined display clear and fresh surface features that are either related to the diagenetic history of the source rock or to the breakdown, deposition, and subsequent modification of these sediments. The lack of intense chemically weathered quartz grain surfaces may be regarded as due to deep burial and therefore protection of the Kirkham Member sands from Tertiary climatic influences. It is only in the uppermost (Kenslow) member of the succession that such weathering has been identified by the presence of gibbsite.

Although these sands display little by way of direct evidence to support the findings of previous researchers they do show some remarkably clear and fresh surface features that testify to a negligible amount of post-depositional diagenetic alteration. Surface features of Tertiary and pre-Tertiary age are well preserved on the majority of the grains examined.

Conclusions

The application of scanning electron microscopy to quartz grain surface features from the sands of the Brassington Formation has resulted in the formulation of a number of conclusions regarding their derivation, mode of deposition, and post-depositional alteration. These conclusions are listed below:

(i) A number of the observed grain surface features, i.e. overgrowths, diagenetic surfaces, cleavage plates, and crack patterns, are considered to be inherited from the Bunter Sandstone.

(ii) A fluvial environment of deposition has previously been proposed for the sediments of the Brassington Formation. The inherited surface features lack any sign of subaqueous abrasion and a short-distance, low-energy fluvial regime is therefore suggested.

(iii) Surface features not inherited from the Bunter Sandstone reveal phases of mechanical and chemical weathering. The very low frequency occurrence of mechanical fracture patterns indicates that this form of weathering was of very limited extent. The oriented etch pits suggest a more prolonged phase of chemical weathering during which many grains have been affected. These two cycles of weathering are not easily dated in terms of Brassington Formation deposition.

(iv) Despite the proven Tertiary age for the Brassington Formation no evidence of high-energy chemical weathering, to support such an age, has been found on the quartz grains.

Acknowledgements

The author is indebted to Mr. H. Beedham of Hoben Quarries Ltd. and Mr. T.W. Andrews of DSF Refractories Ltd. for permission to visit their excavations and take samples of sand. He would also like to thank Drs. P.J. Vincent and J.V. Clarke for supervising the research and Drs. P.T. Walsh and D.B. Thompson and Mr. M. Ijtaba for their interest in the project and their help. A research studentship at the University of Salford is also gratefully acknowledged.

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Explanation for Plates 5 & 6

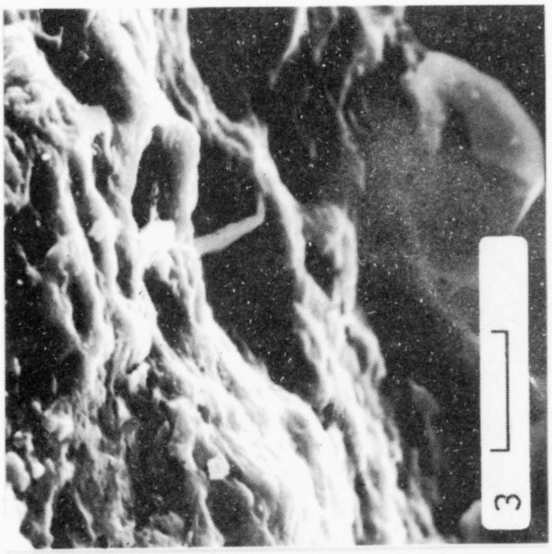
Plate 5

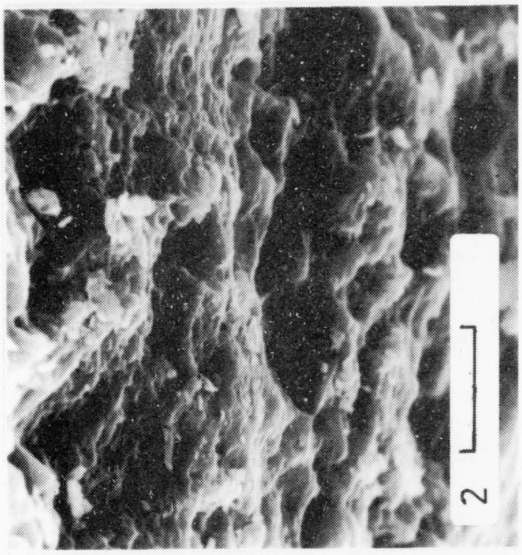
Figure 1	Grain-edge cleavage plates. Scale bar = 10 μm.
Figure 2	Grain-edge cleavage plates. Scale bar = 10 μm.
Figure 3	Grain-edge cleavage plates. Scale bar = 2 μm.
Figure 4	Conchoidal fracture with semi-parallel step-like patterns. Note the sharp and fresh appearance of the fracture surface. Scale bar = 4 μm.

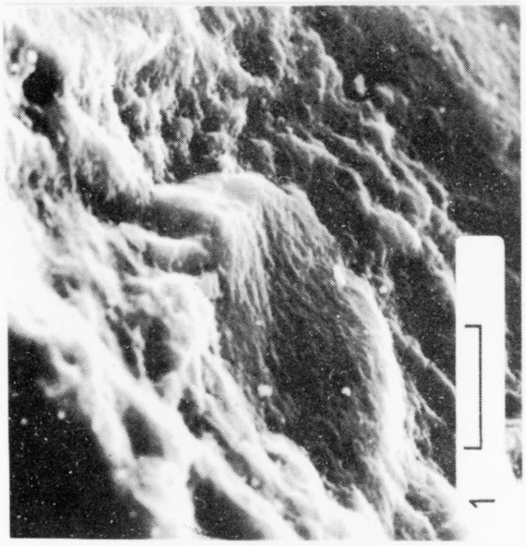
Plate 6

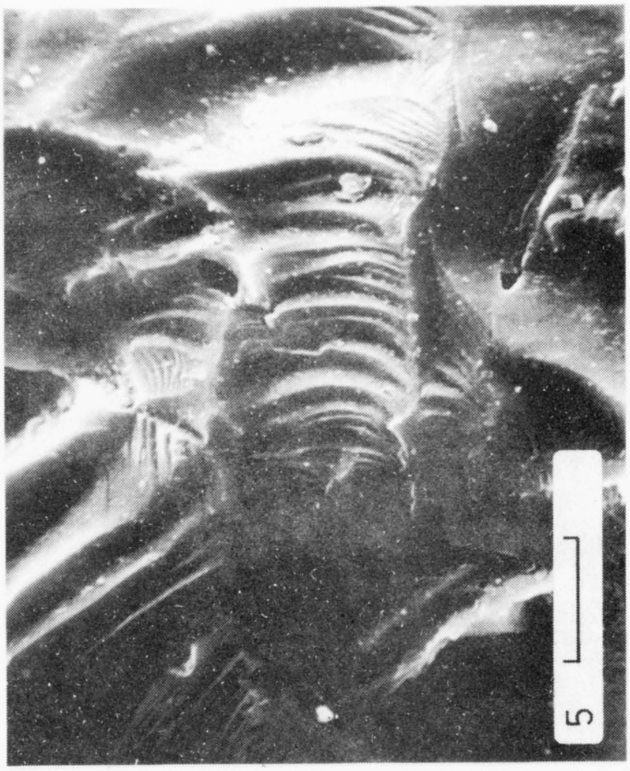
Figure 1	Arcuate, circular or polygonal cracks on grains from Kirkham's Pit Scale bar = 10 μm.
Figure 2	Arcuate, circular or polygonal cracks on grains from Kirkham's Pit. Scale bar = 10 μm.
Figure 3	Bunter Sandstone fragment. Grains display angular outlines, overgrowths, diagenetic surfaces, and cleavage plates. Scale bar = 200 μm.
Figure 4	Bunter Sandstone fragment as above. Scale bar = 200 μm.

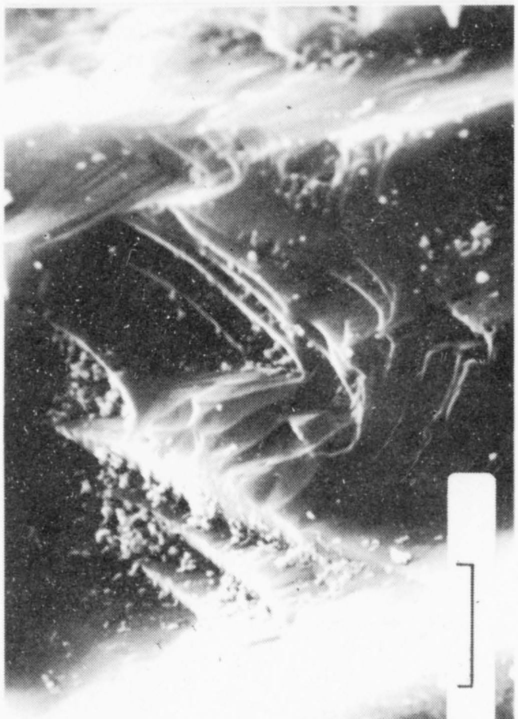
Plate 5





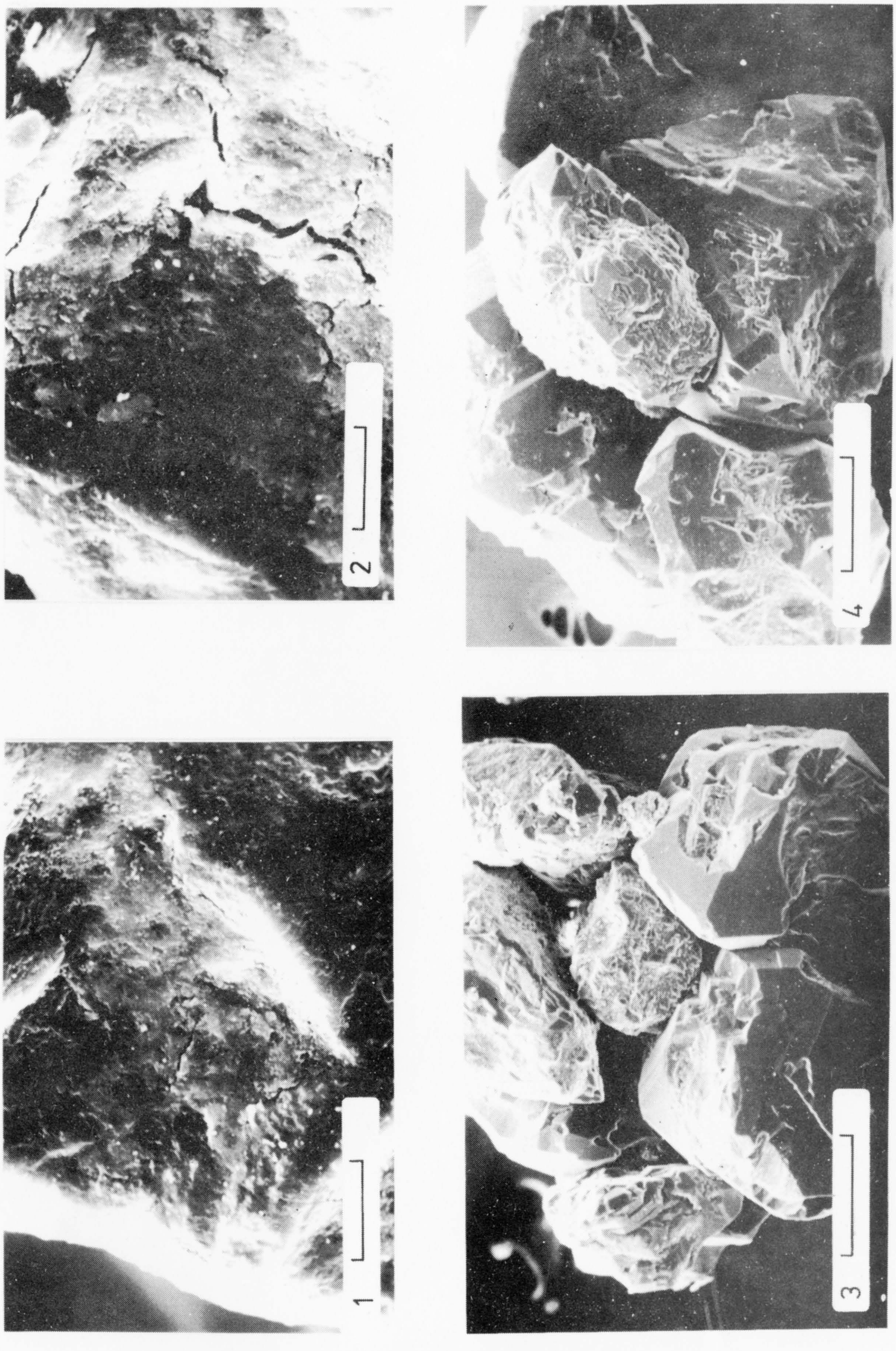






P. Wilson - Surface features of quartz sand grains
(For explanation see p. 30.

Plate 6



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(For explanation see p. 30.

THE EAST MIDLANDS AULACOGEN OF CALEDONIAN AGE

by

A. M. Evans

Summary

It is suggested that a late Precambrian to Lower Palaeozoic aulacogen is present in the basement of much of eastern England. In this structure Pre-Cambrian, Cambrian, Ordovician and Silurian rocks of geosynclinal-type were laid down. It is suggested that the supposed westerly trending ridge of Pre-Cambrian rocks stretching from the Wash to the Melton Mowbray-Market Harborough area represents Lower Palaeozoic extrusive and intrusive rocks belonging to the East Midlands aulacogen which is here described.

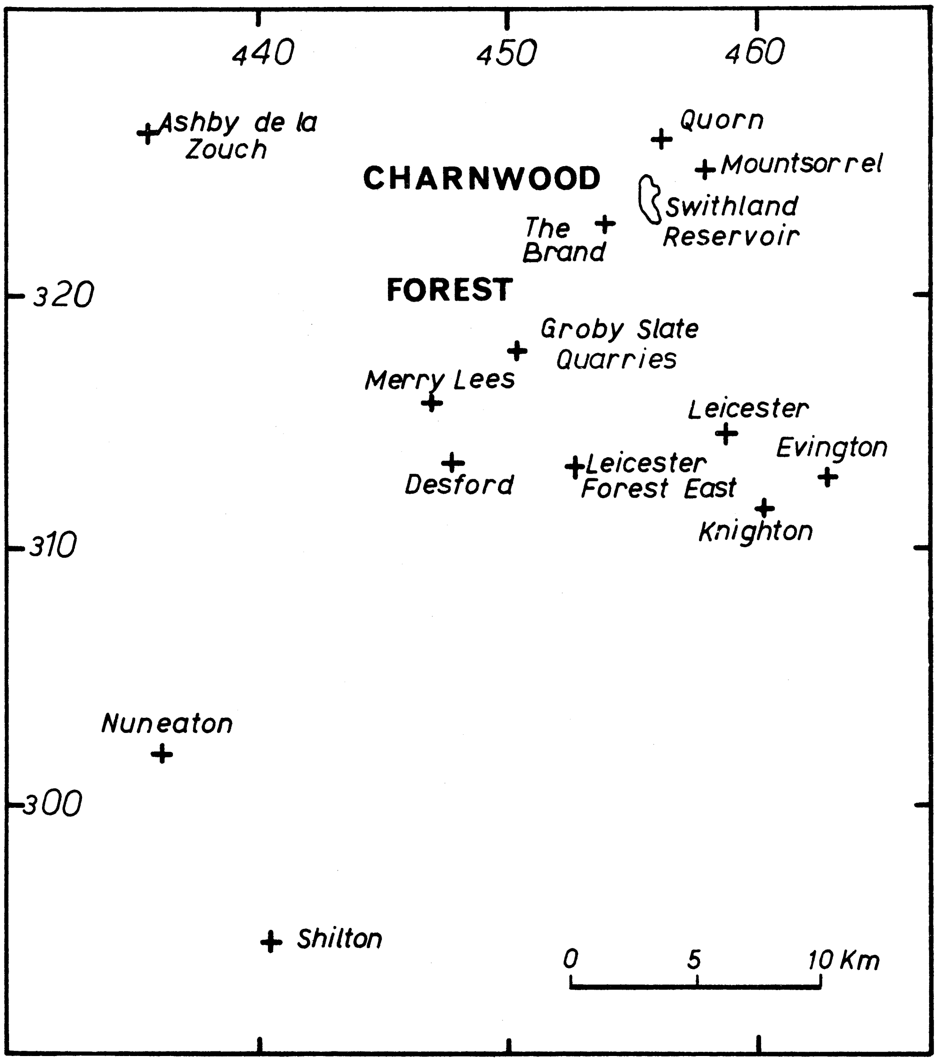
Evidence is presented for the widespread occurrence of deformed Lower Palaeozoic rocks in this aulacogen including hitherto unpublished data for the presence of such rocks in the general area between Charnwood Forest and Nuneaton. The rocks in the trough suffered at least two phases of folding, one late Pre-Cambrian and one late Caledonian. The former was about north-west to south-east and the latter about westerly axes, with the accompanying development of a penetrative cleavage which affects both the Pre-Cambrian of Charnwood Forest and the Cambrian and Ordovician. Radiometric dating and field evidence suggest that this cleavage is Caledonian in age.

Subduction occurred beneath the aulacogen over a long period giving rise to a widespread suite of plutonic and volcanic rocks in both the Pre-Cambrian and the Lower Palaeozoic. The available data suggests a north-easterly dipping subduction zone.

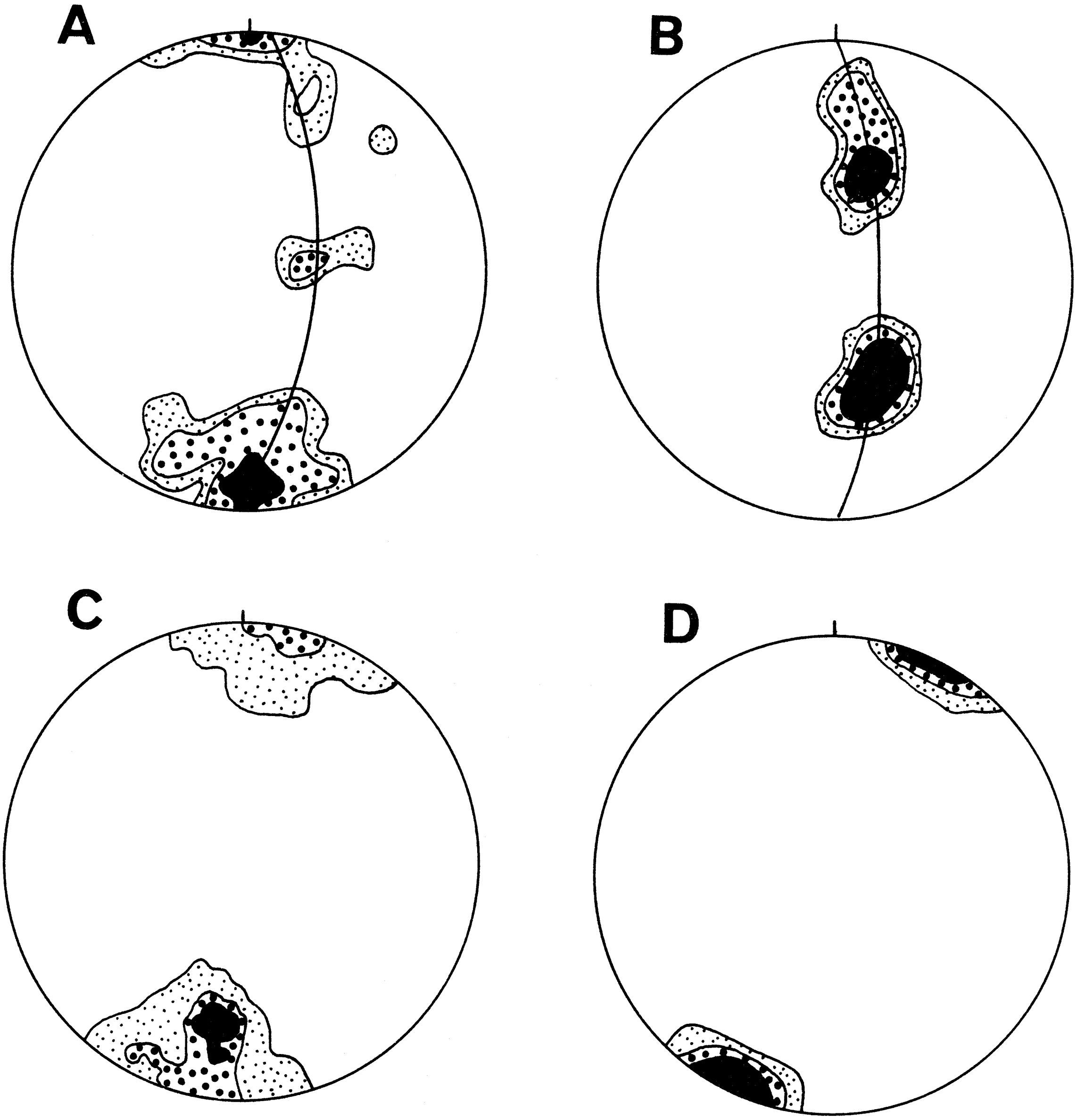
Introduction

There are many intriguing and unanswered questions concerning the Pre-Cambrian and Lower Palaeozoic history of the East Midlands. Some of these have already been discussed (Evans, 1963, 1968, Evans and Maroof 1976, Le Bas 1968, 1972) and the suggestion put foward that an important igneous-tectonic north-westerly trending Caledonian province is present beneath the East Midlands. Most of the evidence comes from the exposed Pre-Cambrian and Lower Palaeozoic rocks of Charnwood Forest and the Nuneaton area, boreholes and geophysical data. The Cambrian and Ordovician of Nuneaton have for long been interpreted as representing shelf sedimentation laid down on a stable Pre-Cambrian craton flanking the Anglo-Welsh Lower Palaeozoic mobile belt (Owen 1976). A question which must, however, be considered is whether another mobile belt of the same age is present underlying much of England to the east of Nuneaton. Just such a possibility was put forward by Turner in 1949 in a brilliantly deductive paper in which he suggested that the Anglo-Welsh Caledonides form an arc, convex northwards, which surrounds the Midland Massif and passes south-eastwards into central Europe. Turner considered the folding of these Lower Palaeozoic rocks to be Hercynian in age and their south-western margin to lie just east of the Pre-Cambrian Charnian rocks of Charnwood Forest. Wills (1951) proposed a similar hypothesis but invoked a Caledonian age for the folding. Recently Dewey and Kidd (1974) suggested that the Leicestershire area may represent a late Pre-cambrian aulacogen. Such a structure is a large, long-lived, fault-bounded trench which intersects the Caledonian geosyncline at a high angle and extends into the neighbouring craton, (text-fig.3), p. 37, (Smith, 1976). The trench becomes filled with sediments and volcanics. Despite the work of the above mentioned and other authors,

Mercian Geol. Vol. 7, No. 1,
1979, pp. 31-42, 4 text-figs.



Text-fig. 1. Locality map showing the position of Charnwood Forest and other nearby places mentioned in the text.



Text-fig. 2. Stereograms of: A - 78 poles to bedding for the Ordovician of the Merry Lees Drifts; B - 68 poles to bedding for the Cambrian of Swithland Reservoir (contours for A and B at 3, 4 and 10 per cent of one per cent areas); C - 500 poles to cleavage for the Charnian of Charnwood Forest (contours at 2,8 and 12 per cent); D - 100 poles to cleavage for the Cambrian at Swithland Reservoir (contours at 4, 12 and 20 per cent). All data projected on the lower hemisphere of a Lambert equal-area net.

the general tendency in most stratigraphical reconstructions is to treat the whole of the English Midlands as a stable Pre-Cambrian cratonic block which carries only a thin veneer of Lower Palaeozoic sediment (Read and Watson 1975, Owen 1976, Lovell 1977).

In this paper further evidence for the existence of a Lower Palaeozoic mobile belt underlying much of the East Midlands is put forward. It is postulated that the south-western margin of this belt may be close to Nuneaton and that the Pre-Cambrian of Charnwood Forest was considerably deformed during the Caledonian orogeny. It is further postulated that this orogeny was responsible for the principal cleavage in the Charnian Supergroup and the folding and cleavage in the adjacent Palaeozoic rocks.

Folding in the Charnian and Lower Palaeozoic rocks

The notion that the Charnoid trend is north-westerly was challenged by Evans (1963) who pointed out that although the major anticline followed this trend over much of Charnwood Forest it is flexed round to an easterly trend in the south-east of the Forest, whilst the cleavage is generally oblique to the fold and, over the Forest as a whole, has an average strike of 280°. A model was proposed in which it was suggested that the folding and oblique cleavage, though broadly non-parallel, were coeval and probably Pre-Cambrian in age. Later data suggested that this hypothesis must be abandoned and that the Charnian rocks have suffered at least two important phases of deformation (Evans *et al.* 1968).

Minor folds in the Charnian

A number of minor folds are present and these have wavelengths ranging from 0.3 - 1.6 km. Some of these trend parallel to the major anticline and possess a similar plunge. Another group have horizontal or subhorizontal axes which trend obliquely across the main fold with a trend concentrated about 280°. They are concentric folds with north-south slickensides developed on those bedding surfaces which accommodated much of the bedding slip. Localities of many of these folds are given in Evans (1963).

Minor folds in the Lower Palaeozoic

The only extensive exposures of unequivocal Lower Palaeozoic rocks east of the Nuneaton area were those exposed in the Merry Lees Drift (text-fig. 1) near Desford. (Butterley and Mitchell 1945). These rocks lie beneath the Trias and fossil evidence indicates a Tremadoc age. Butterley and Mitchell recorded dips approaching the vertical with a westerly strike. Le Bas (1972) noted the presence of a weak cleavage.

Recently structural mapping of these rocks has revealed the presence of minor folding and a stereogram of poles to bedding is given in text-fig. 2. This indicates a plunge of about 22° in a direction 275° essentially parallel to the second group of minor folds in the Charnian and with a very different plunge from that of the probably late Precambrian folding.

On the east side of Charnwood Forest, but not in contact with the Charnian rocks, the Caledonian Mountsorrel igneous complex crops out. Associated with it are a number of small outcrops of pelitic hornfelses. Some of these appear to be rafts within the granodiorite. On the south-western shore of Swithland Reservoir (text-fig. 1), however, low grade hornfelses, about half a kilometre from the contact, are well exposed in and around a small hill. Le Bas (1972) has made out a good case for regarding these as hornfelsed Cambrian rocks. Bedding is clearly visible and this has been folded into an upright syncline with a vertical axial surface. The axis plunges at about 16° in a direction 269° (text-fig. 2). This is about 45° anticlockwise to the principal axial trend in the nearby Charnian rocks. The joint controlled shapes of the xenoliths in the Mountsorrel complex indicate that it is probably a permitted intrusion emplaced by stoping, in which case it is unlikely that this fold has been tilted by the intrusion.

It can hardly be a coincidence that the folds at Merry Lees and Swithland Reservoir have similar trends which are parallel to the second group of folds in the Charnian rocks.

The conclusion is drawn that all these folds are probably Lower or mid-Palaeozoic in age.

Cleavage in the Charnian and Lower Palaeozoic rocks

The presence of cleavage in the Ordovician rocks of the Merry Lees Drifts was noted by Le Bas (1972). Unfortunately this was not detected underground, but only in thin sections. A weak cleavage is also present in the Ordovician rocks intersected by the Evington and Knighton boreholes in Leicester City, specimens of which can be seen in the Leicester Museum. Relict cleavage is readily apparent in the hornfelses at Swithland Reservoir. Here it has an average trend of 296° (text-fig. 2) essentially parallel to that in the nearest outcrops of Charnian rocks, but oblique to the minor fold described above.

It must be borne in mind, however, that although the average cleavage trend in the Charnian rocks is 280° (text-fig. 2) it does show some variety in its trend and intensity. It actually follows a somewhat sinuous course across the Forest from approximately westerly in the south to more north-westerly in the centre and west-north-westerly in the north-west (Evans 1963).

Recently boreholes by the Institute of Geological Sciences and National Coal Board have penetrated Upper Cambrian (not Tremadoc) rocks between Charnwood Forest and Nuneaton (details in press). These are at Rotherwood (SK345155) just south of Ashby-de-la-Zouch, Dadlington (SP399991) 7. 5 km north-east of Nuneaton and Leicester Forest East (SK525029) 6. 5 km west-south-west of Leicester (text-fig. 1). The rocks are dominantly well cleaved slates and the bedding frequently shows high dips. These discoveries together with the evidence from the Merry Lees Drifts and the Leicester boreholes indicate the presence of considerably deformed and folded Cambrian and Ordovician sediments *to the west* of Charnwood Forest. To these may probably be added the non-fossiliferous slates of the Lindridge borehole near Desford.

In the broad region of basement rocks running north-westwards beneath the Mesozoic rocks from near Dover through the Home Counties, Bedfordshire, Northamptonshire, Leicestershire, Nottinghamshire and the adjoining areas there is now abundant evidence of the development of Cambrian to Silurian rocks (Le Bas 1972). Cleavage has been reported as being present in some of these and there is an extensive development of calc-alkaline intrusions (Le Bas 1972; Evans and Maroof 1976).

A number of authors (e. g. Kent 1968; Le Bas 1972) have postulated the presence of a westerly to west-south-westerly trending ridge of Pre-Cambrian rocks which cuts across this Palaeozoic belt. The area of the ridge includes Melton Mowbray, where a recent National Coal Board borehole has proved the presence of granite which probably represents part of the Mountsorrel batholith (Evans and Maroof 1976), and the Thorpe-by-Water borehole near Uppingham from which Bath (*vide* Richardson and Oxburgh 1978) obtained an Ordovician Rb-Sr whole rock isochron. Speculation that this ridge was Pre-Cambrian was quite reasonably based in part on the presence of volcanics in the Glinton, North Creake and Orton boreholes which were compared with the Pre-Cambrian volcanics of the Charnian Supergroup. Evans (1964), however, considered that the Orton rock was not directly comparable with any Charnian volcanics. It is suggested by the present writer that these may be volcanic equivalents of the extensive Caledonian plutonic intrusions of this belt and that the existence of this E.-W. Pre-Cambrian ridge must be in some doubt.

Kent (1968) has noted the geosynclinal nature of the Lower Palaeozoics intersected by boreholes at Twyford and Calvert (Bucks.) and Huntingdon. These show high dips but no metamorphic features have been reported. He considered that the basement rocks further north encountered in deep boreholes might be Pre-Cambrian or Lower Palaeozoic. These include sheared (cleaved?) mudstone at Eakring (Notts.), vertical and steeply dipping quartzites at Nocton, Bardney and Stixwould (Lincs.), purple phyllites at Foston (Lincs.), phyllitic shales at Sproxton (Leices.) and schistose rocks at Dukes Wood (Notts.). In view of the above evidence the probabilities now favour the interpretation of these rocks as being

Lower Palaeozoic in age particularly in view of the fact that cleaved Arenig or Llanvirn pelites were found in the Eyam (Derbys.) borehole (Dunham, 1973). This implies the presence of mildly metamorphosed and considerably deformed and cleaved rocks of this age in this region.

Further west in Derbyshire the Woo Dale borehole near Buxton penetrated altered lavas and pyroclastics which lie north-west of the Lower Palaeozoic area between Nuneaton and Charnwood Forest (Cope, 1973). These rocks have been shown from their micro-fauna to be Ordovician (Downie pers. comm.).

Age of the cleavage

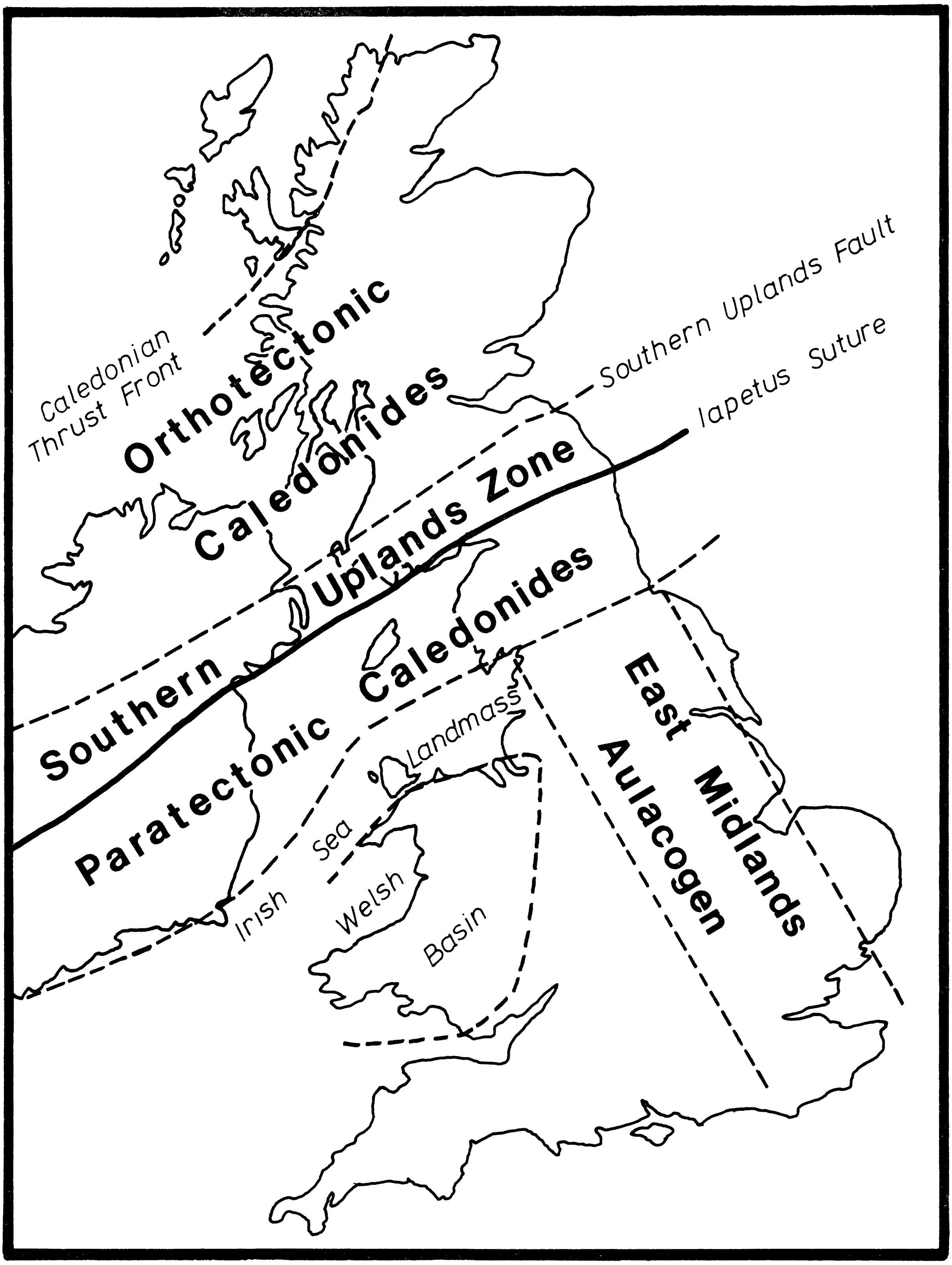
Cleavage has now been reported from Cambrian and Ordovian rocks over an extensive part of the Midlands. Its orientation has only been measured at Mountsorrel, where it is clearly parallel with that in the nearby Charnian rocks suggesting a Palaeozoic age for the latter cleavage. This possibility is supported by the parallelism of the Charnian cleavage with the second group of minor folds which are colinear with those in the Palaeozoic rocks.

An important point of evidence is the development of a rude cleavage in the Lubcloud dyke (SK478164). This rock, which intrudes the Charnian, is petrographically similar to the Mountsorrel granodiorite and most previous workers have considered that it is of the same age and derived from the same magma. Meneisy and Miller (1963) obtained a K-Ar age of 374±13 Ma on a sample of biotite from the dyke. This compares closely with the K-Ar age of 403±18 Ma Miller and Padmore (1961) obtained for the Mountsorrel granodiorite. This evidence can be interpreted as showing that the main Charnian cleavage was developed after the intrusion of the Lubcloud dyke and is therefore probably a Caledonian feature, a conclusion which was reached by Jones (1927) who noted its parallelism with shear zones in the Mountsorrel granodiorite.

In order to test the above hypothesis samples of Charnian Swithland Slate from the Brand (SK538137) and the Groby Slate Quarries (SK508082) (text-fig. 1) were kindly analysed by Dr. D.C. Rex at Leeds University. These gave K-Ar whole rock ages of 417±16 Ma and 398±16 Ma respectively, ($\lambda_\beta = 4.962 \times 10^{-10}\ y^{-1}$, $\lambda_\epsilon = 0.581 \times 10^{-10}\ y^{-1}$, $^{40}K = 0.01167$ atom %). A sample of well cleaved Upper Cambrian from the Rotherwood borehole gave 477±19. The ages for the Charnian material taken on their own would merely indicate a minimum age for the formation of the cleavage, (Dodson and Rex, 1971). However, taken in conjunction with the structural and stratigraphical evidence they strongly support a Caledonian age for the impression of the cleavage. The older date for the Upper Cambrian sample may well be due to the fact that this sample is not as well recrystallized as the Charnian samples, it probably contains some non-recrystallized detrital 2M muscovite and it would not meet all the criteria listed by Dodson and Rex (1971). It is possible that this Cambrian rock sample was at a higher structural level than the Charnian samples when the cleavage was formed. The results obtained from the Swithland Slate are in close agreement with the date of 413 ± 19 Ma which Meneisy and Miller (1963) obtained on tuffs from Beacon Hill a date which they considered might be that of the formation of the cleavage.

Regional Implications

There seems to be little doubt in view of the evidence cited above that an important Caledonian mobile belt exists beneath the East Midlands. Turner (1949) considered that this was a continuation of the exposed Anglo-Welsh Caledonian geosyncline which he termed the Central European Branch. Le Bas (1972) also strongly argued the case for a development of thick geosynclinal Lower Palaeozoic rocks in this belt and he emphasized the important development of a calcalkaline intrusive suite typical of orogenic belts. Evans and Maroof (1976) demonstrated that these intrusions are of batholithic dimensions and because of the common development of magnetite they can be outlined by following the magnetic anomalies they produce. The data put forward in this paper shows that these Lower Palaeozoic rocks and the Pre-Cambrian basement (in the form of the Charnian rocks) underwent an important and widespread Caledonian deformation.



Text-fig. 3. Position of the East Midlands aulacogen relative to the Caledonide orogenic belt based in part on the reconstruction in Phillips *et al.* (1976).

In considering the regional picture there appear to be a number of possibilities. First is that put forward by Turner and by Le Bas, second is an extension of the hypothesis of Dewey and Kidd (1974) that we are dealing with the development of an aulacogen and third that the Anglo-Welsh geosyncline is much wider than was previously thought and that the south-eastern front lies much further to the south-east than has previously been considered to be the case.

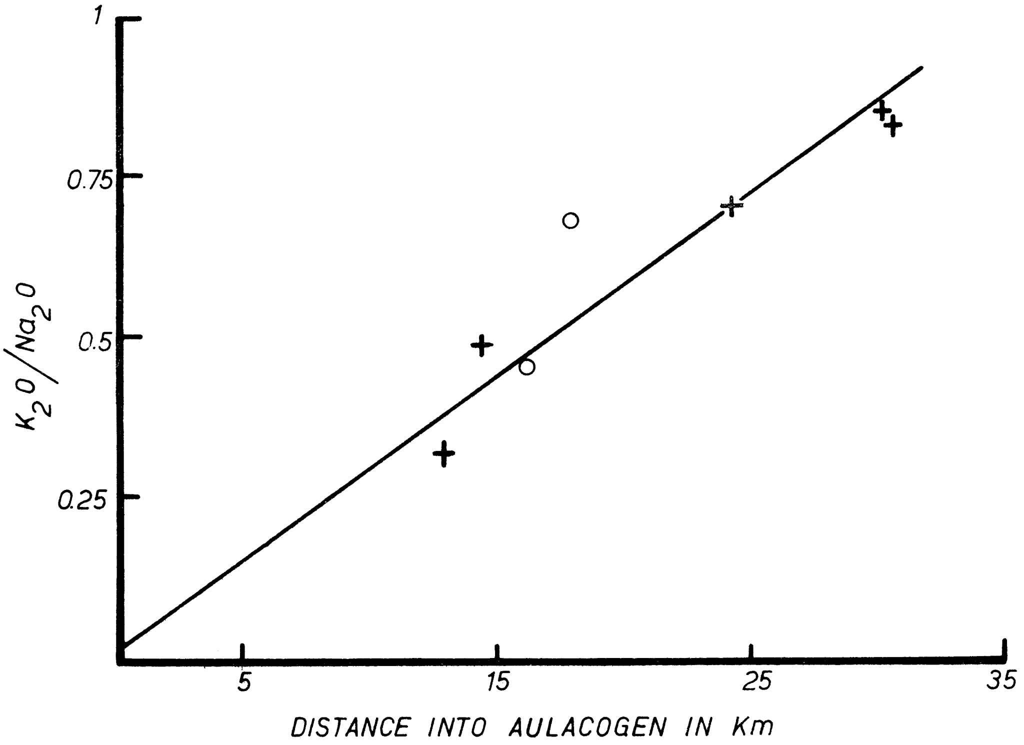
In examining these possibilities it is important to elucidate the principal lithological and tectonic trends. Unfortunately these are not unequivocal. In the Nuneaton area the lithological trends of the Cambrian are north-westerly (Allen, 1968) and the palaeocurrent direction in the Upper Cambrian was north-easterly (Rushton, 1974), perhaps indicating marginal flow into a NW trending trough. The dip is south westwards, but as there is little discordance between the Cambrian and the Coal Measures this is probably a Hercynian effect. Minor folds in the Pre-Cambrian tuffs after elimination of the Hercynian tilt have a south-easterly plunge like the major Charnian fold.

In the Charnwood Forest area and its environs the Charnian has an older north-westerly lithological and tectonic trend with a superposed westerly trend, the latter, as has been shown above, is a Caledonian effect. Le Bas (1972) considered that the surrounding Caledonian region had a west-north-westerly trend. The belt of positive magnetic anomalies which reflect the outline of the principal Caledonian intrusions (Evans and Maroof 1976) has a clearly defined north-westerly trend. Detailed examination of the individual anomalies, however, indicates a subsidiary 260° - 290° trend. Since large orogenic intrusions normally follow the principal tectonic trend it is considered that the bulk of the present evidence indicates that this is north-westerly and that the westerly trending second phase of folding and the cleavage of the Charnwood Forest area follow a less important late stage tectonic trend. This westerly trend is of course well known in central and north-east Wales (Bassett 1955; Shackleton 1954) and northern England (Moseley 1972). It seems therefore that the dominant trend is perpendicular to that of the Anglo-Welsh geosyncline and the third possibility can be ruled out, leaving a choice between deposition along a geosynclinal plate margin or in an aulacogen.

On the whole the second possibility seems more likely in the light of present plate tectonic theory. A well documented example is that of the Athapuscow aulacogen in the NW Canadian shield (Hoffman 1973). This contains sediments over 11 km thick in a trough which reaches 70 km in width where it joins the Coronation Gulf geosyncline and which narrows to less than 20 km some 200 km into the adjacent foreland. These dimensions are comparable with those with which we are dealing in the East Midlands. Volcanism is, however, dominantly basaltic but late stage tonalitic and granodioritic intrusions are present. Deformation is only by open folding and faulting and no regional metamorphism or penetrative deformation has been reported. On the other hand the Mount Isa geosyncline now regarded as an aulacogen contains strongly folded geosynclinal-type sediments with both basic and acid volcanics (Heidecker 1976; Windley 1977). The Southern Alberta aulacogen (Burke and Dewey 1973) contains dioritic intrusions some of which are granophyric like the Charnian and Nuneaton diorites. Aulacogens may be as wide as 160 km (Keweenawan trough, Windley 1977) and their calc-alkaline igneous rocks are presumably related to subduction zones as has been suggested for the Lower Benue trough where the sediments were deformed during the closure of the aulacogen (Burke and Dewey 1973). It is proposed that the mobile belt discussed in this paper be called the East Midlands aulacogen (text-fig. 3).

Polarity of the subduction zone

Igneous activity occurred throughout the known existence of the aulacogen from at least 684±29 Ma to 433±17 Ma (Meneisy and Miller 1963; Cribb 1975) and, if the Northern Diorites of the Charnian be included, as late as 311±92 Ma. Exposures of hypabyssal and plutonic rocks across the aulacogen are not common. Of those which do crop out, most belong to the diorite-tonalite clan with a fairly restricted variation in silica content and this permits an *approximate* estimate of the trend of the K_2O/Na_2O ratio across the south-western part of the aulacogen modifying the method of Hatherton and Dickinson (1969). $^{87}Sr/^{86}Sr$ ratios for these rocks from Charnwood Forest and the surrounding area suggest a subduction zone origin (Cribb 1975). Data for these rock-types falling close to a line running from Shilton (SP 405845) to Quorn (SK 562164) have been taken from Le Bas (1968 and 1972) and Thorpe (1974) and plotted on text-fig. 4. The data fall close to a straight line indicating what appears to be a significant increase in the K_2O/Na_2O ratio in a north-easterly direction. This suggests that the calc-alkaline intrusions were generated along a north-eastward dipping subduction zone. Partial closure of the aulacogen related to this subduction was probably responsible for the earlier north-westerly trending folds. The cause of the later westerly trending structures is less certain. They could have been formed by north-south stresses generated by dextral movement of the sides of the aulacogen.



Text-fig. 4. Plot of the K_2O/Na_2O ratio in diorites (crosses) and tonalites (circles) along a north-easterly line across the south-western portion of the East Midlands aulacogen.

Conclusions

1. Deformed Lower Palaeozoic rocks of geosynclinal aspect and thickness are widespread in the basement of the East Midlands.

2. A stable Pre-Cambrian Midland massif in Lower Palaeozoic times was much smaller than is envisaged in most stratigraphical reconstructions of the area.

3. The supposed presence of a westerly trending ridge of Pre-Cambrian rocks cutting across the Lower Palaeozoic rocks from the Wash to the Melton Mowbray - Market Harborough area is questioned.

4. A north-westerly trending aulacogen ran through the area between Nuneaton and the east coast during late Pre-Cambrian and Lower Palaeozoic times. The Pre-Cambrian Charnian rocks and a geosynclinal suite of Cambrian, Ordovician and Silurian rocks were laid down in the structure, here called the East Midlands Aulacogen (text-fig. 3).

5. Deposits in the aulacogen underwent at least two periods of deformation. The first was responsible for the north-westerly folding in the Charnian. The second gave rise to westerly trending folds and cleavage which affect the Pre-Cambrian, Cambrian and Ordovician. The second deformation probably occurred about the end of Lower Palaeozoic time.

6. Subduction took place beneath this aulacogen over a long period and the available data suggests a north-easterly dipping subduction zone. Melting along this subduction zone gave rise to a suite of calc-alkaline intrusions some of which are of batholithic dimensions. Deformation related to this subduction was responsible for the north-westerly folding in the Charnian. The compression which gave rise to the later structures could have been generated by dextral movement of the sides of the aulacogen.

Acknowledgements

It gives the author great pleasure to acknowledge the kindness of many Leicestershire landowners, too numerous to name here, who have granted him access to their estates. The staff of the NCB colliery at Desford were most helpful as were the staff at Leicester Museum Dr. R.A. Old of the IGS kindly supplied some borehole specimens. Dr. D.C. Rex' help is acknowledged above. Finally I would like to thank colleagues at Leicester University for helpful discussions during the progress of this work.

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THE UPPER CARBONIFEROUS ROCKS OF THE EWDEN VALLEY, SOUTH YORKSHIRE

by

Rachel V. Elliott

Summary

Re-mapping Upper Carboniferous rocks of the Ewden Valley has resulted in a revised correlation of the sandstones, in particular sandstones previously regarded as Rivelin Grit being now recognised as Heyden Rock. Structural analysis has indicated the presence and extent of the Ewden Monocline trending WNW to ENE and throwing the strata down to the south by more than 30 m.

Introduction

Lying some 12 km northwest of Sheffield, the Ewden Valley is cut through strata of Upper Carboniferous age, comprising the upper part of the Millstone Grit and the lower part of the Coal Measures Groups.

Ewden Valley lies within the bounds of two Geological Survey of Great Britain 1 inch : 1 mile sheets (No. 86, Holmfirth and Glossop and No. 87, Barnsley) and the only published descriptions of the detailed geology are in the accompanying memoirs (Bromehead *et al.* 1933 and Mitchell *et al.* 1947).

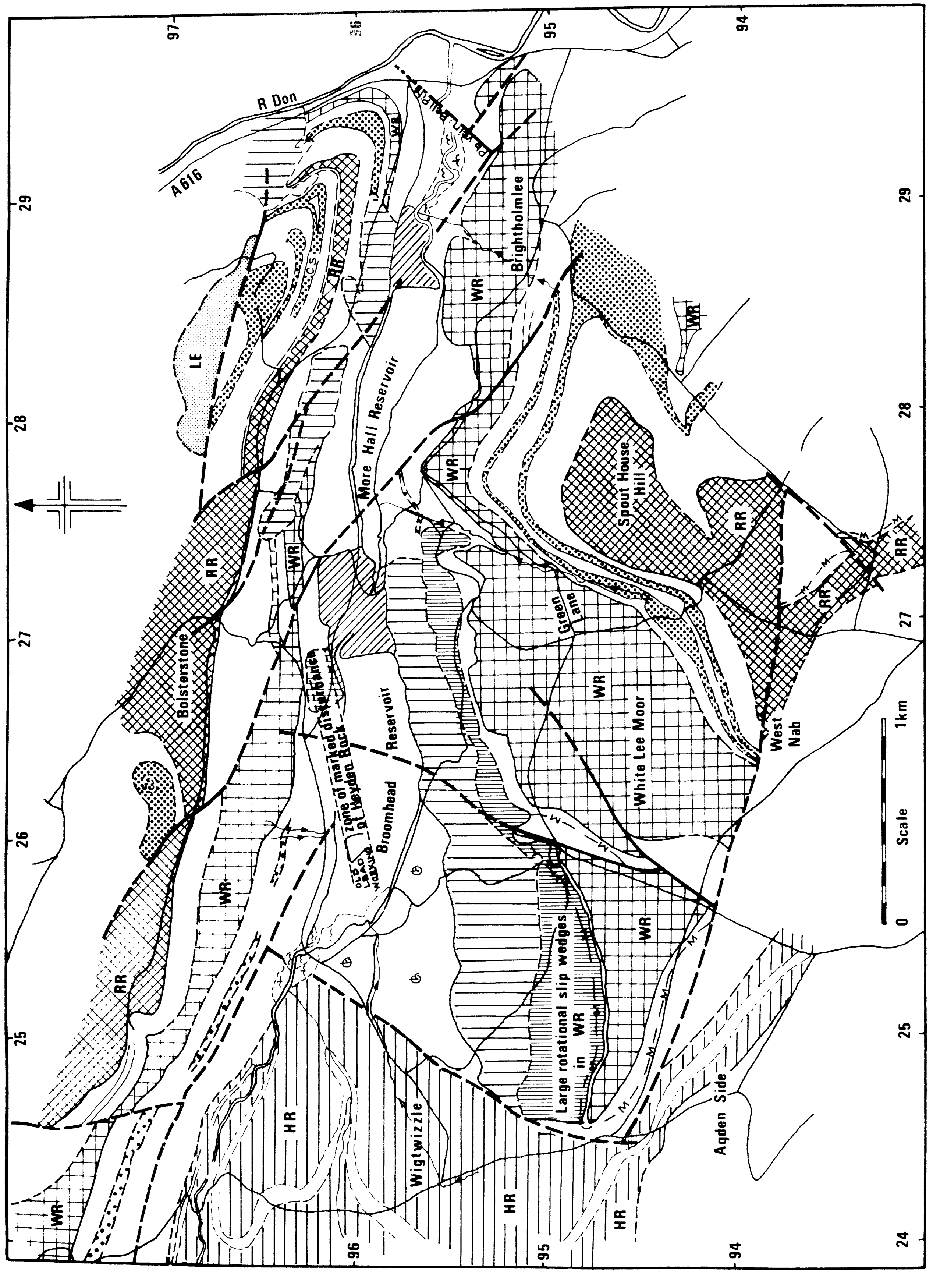
As a result of detailed field mapping for a B.Sc. thesis at the University of Leicester it became evident that previous authors had mis-correlated some of the Millstone Grit sandstones. Briefly, the outcrop of the "Rivelin Grit" on the Holmfirth Sheet could be traced directly in the field to correlate with a lower sandstone on the Barnsley Sheet. In the accompanying memoir Bromehead *et al.* (1947) commented only that the Rivelin Grit of the country to the west of the Barnsley Sheet was not the equivalent of the Rivelin Grit of the Rivelin Valley. Ramsbottom's correlation chart (1966) showed the correct situation but neither he nor Bromehead discussed the details. The correlation and map presented herein (text-fig. 1) remove the discrepancy in the regional correlation and allow a re-assessment of the faulting and the construction of a detailed structure contour map which partly fills the gap in the structural pattern left in the Barnsley Memoir where the contours were taken at much higher Coal Measure horizons and in the Holmfirth Memoir where only a generalised structure map was presented (Bromehead *et al.* 1947, p.117). Additional evidence of local structure has been obtained from the records of the geology of the foundations of the Broomhead Dam (Bendelow 1944).

Details of the Succession

The stratigraphical succession is shown in text-fig. 1, and is composed of 223 m of Millstone Grit sandstones and shales, followed by 115 m of Lower Coal Measures rocks. The latter are only seen in a small area to the north and east of Bolsterstone and, as no change from the sequence shown on the Geological Survey maps and memoirs has been detected, their stratigraphy need not be discussed further.

The lowest sandstone of the Millstone Grit succession in the Ewden Valley is identifiable as the Heyden Rock as it is overlain by shales containing the *Reticuloceras bilingue*

Mercian Geol. Vol. 7, No. 1,
1979, 3-text-figs., pp 43-49.



R Don
A616
97
96
95
94
24
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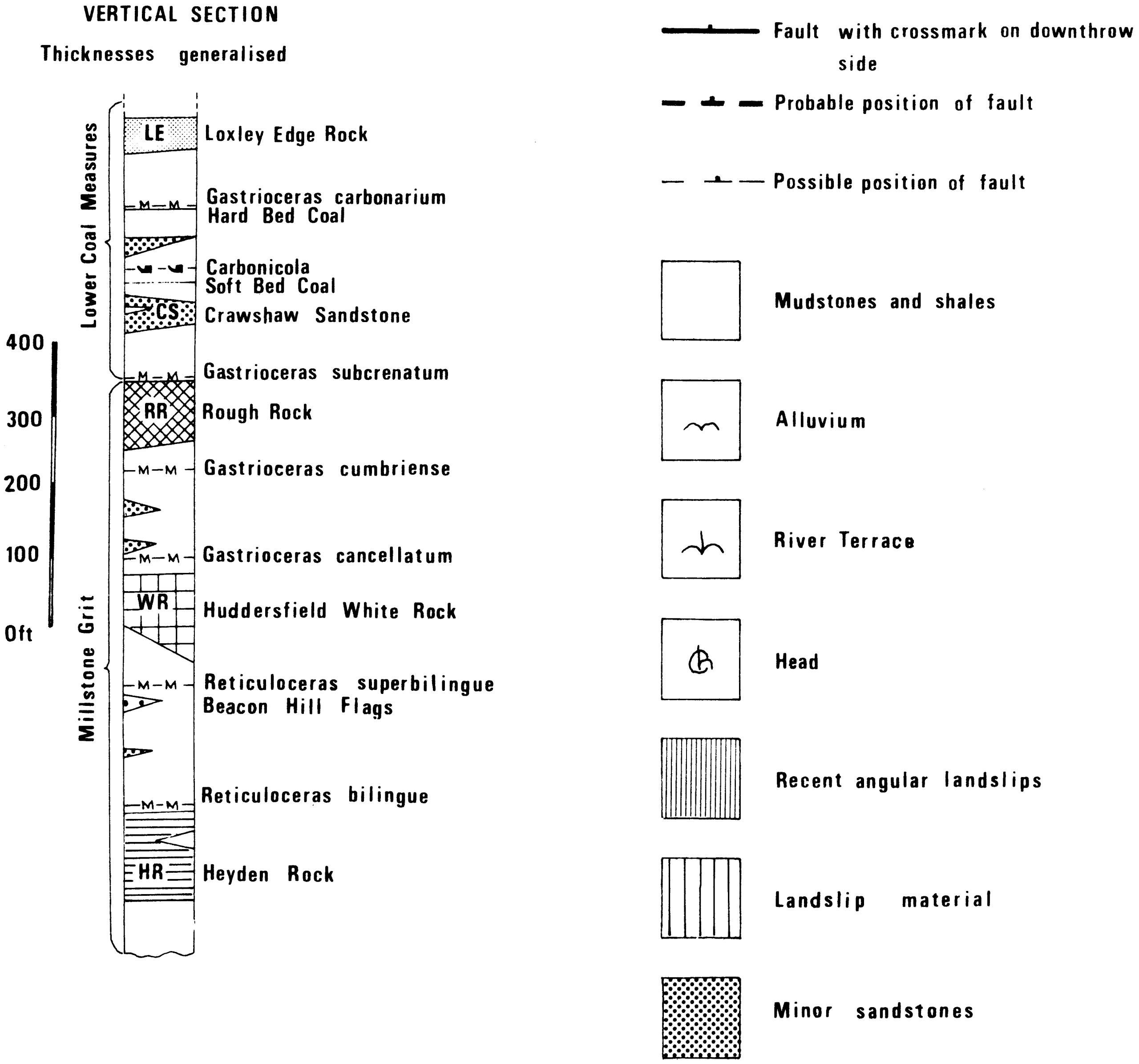
26
27
28
29
LE
RR
WR
HR
CS
Bolsterstone
More Hall Reservoir
Brightholmlee
zone of marked disturbance at Heyden Bock
OLD LEAD WORKING
Broomhead Reservoir
Wigtwizzle
Large rotational slip wedges in WR
Green Lane
White Lee Moor
Spout House Hill
West Nab
Agden Side
Scale
0
1km

marine band, [1] as noted by Bromehead *et al.* (1933), which established correlation with the Heyden Rock of the type area in southeast Lancashire. In the Ewden Valley the Heyden Rock is approximately 36 m thick and consists of two leaves of a coarse pebbly arkose separated by a shale layer 7.7 m thick. The lower arkose is 21 m thick and the upper 7.7 m thick.

Footnote: [1] Throughout this paper the goniatites and marine bands are recorded under their modern names of *Reticuloceras bilingue* and *R. superbilingue* etc. though they were originally recorded in the memoirs under the old system of *R. reticulatum* mut β and mut γ etc.

Text-fig. 1, opposite, Geological map of the Ewden Valley
below, Stratigraphical succession and key to map



The base of the Heyden Rock is seen on Agden Side (SK 249940)[2] and Hurkling Edge (239946) where it forms a steep south-facing escarpment, weathered into crags, which show small scale current bedding in the coarse grit. Northeast from Hurkling Edge, the shale parting forms a boggy hollow, which curves southwards to where Mortimer Road crosses the edge. Here the lower Heyden Rock is seen as 1 m of flaggy sandstone overlain by shale which has a dip of 10° north and a strike of 150°. This shale is seen again in the Ewden Beck at 252964 as a succession of 6 m of black shale overlying cross-bedded sandstone. Further downstream at 253963, there is a small outcrop of flaggy sandstone representing the upper leaf. The lower leaf outcrops in the Ewden Beck from 252964 to the edge of the area mapped. Upstream (west) of Ewden Bridge, at 238968, shale outcrops again with sandstone above, but this is not the base of the Heyden Rock, for 90 m further upstream there is another 6 m of cross-bedded, ripple-marked sandstone.

To the east of Wigtwizzle the Heyden Rock is thrown up against the Huddersfield White Rock and lower beds, as is seen in Lee Lane Dyke at 248955 and Allas Lane Dyke at 2512959. At Lee Lane Dyke there is a continuous outcrop of an uncorrelated sandstone possibly equivalent to the Beacon Hill Flags. A window of shale appears for a short distance at 248955 which indicates the base of the sandstone, but there is a marked steepening of dip upstream towards the shale which suggests that the Wigtwizzle fault brings the shale parting of the Heyden Rock on the upthrow side of the fault against this uncorrelated sandstone on the downthrow side. At Allas Lane Dyke (251959) there is a faulted junction showing 1.3 m of orange to grey shale steepening upstream from 20° to 72° over a distance of 4 m lying against a sandstone which flattens upstream.

The Heyden Rock should outcrop north of the Ewden Beck below the Huddersfield White Rock but there is no feature and this is taken to indicate that it has been cut out by the Ewden Fault downthrowing it 66 m to the south.

Between the Heyden Rock and the Huddersfield White Rock are 56 m of measures with the *Reticuloceras bilingue* marine band at the base, on top of the Heyden Rock. Bromehead *et al.* (1933) recorded a dark silty shale with goniatites from this horizon east of Smallfield Lane at 256939. The shales contain two thin discontinuous flaggy sandstones. The topmost sandstone, the Beacon Hill Flags, has a ganister above it. The shales are not seen at outcrop on the northern side of the valley. On the southern side of the valley in Lee Lane Dyke (252957) there is a shale exposed with slabby sandstone above and below it. The sandstone is persistent upstream until it is cut out by the Wigtwizzle Fault. The sandstone is too high to be the Heyden Rock and too low to be the Beacon Hill Flags.

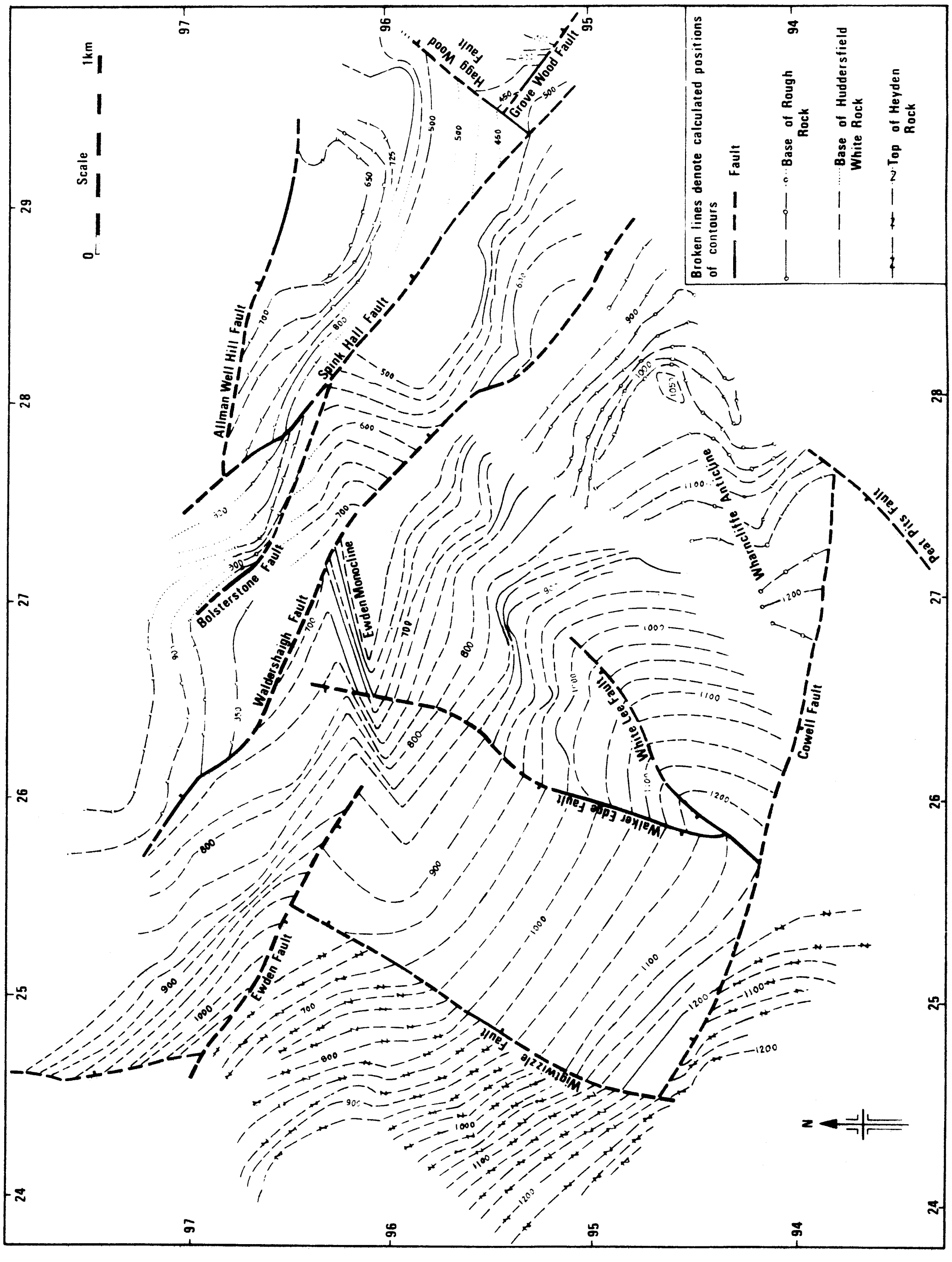
In Raynor Clough these measures between the Heyden Rock and the Huddersfield White Rock are siltstones and shales with the Beacon Hill Flags 2 m thick forming a waterfall, at 276956, about 5 m in height. The sandstone is discontinuous and forms flaggy lenses several feet in thickness further east. At the waterfall, the Beacon Hill Flags are 30 m below the Huddersfield White Rock, the succession above being of black and grey shales.

The Beacon Hill Flags north of Cowell House (2595) are totally obscured by the Canyard Landslips. To the east of the Walker Edge Fault the flags are exposed where the ganister above them has been worked for firebricks (261950).

South of the bridge in Raynor Clough (276955) Mitchell *et al.* (1947, p.6) recovered goniatites, bivalves and ostracods from the *Reticuloceras superbilingue* marine band, which consists of nodular ironstone in black sulphurous shales.

The shales pass upwards into the Huddersfield White Rock which forms cliffs 14 m high in Raynor Clough. There is a mudstone with sandy intercalations at the base followed by 14 m of sandstone and a further 7.7 m of sandstone succeeds out of the Clough.

Footnote: [2] All grid references herein are in 100 km square SK.



Text-fig. 2. Structure contour map of the Ewden Valley.

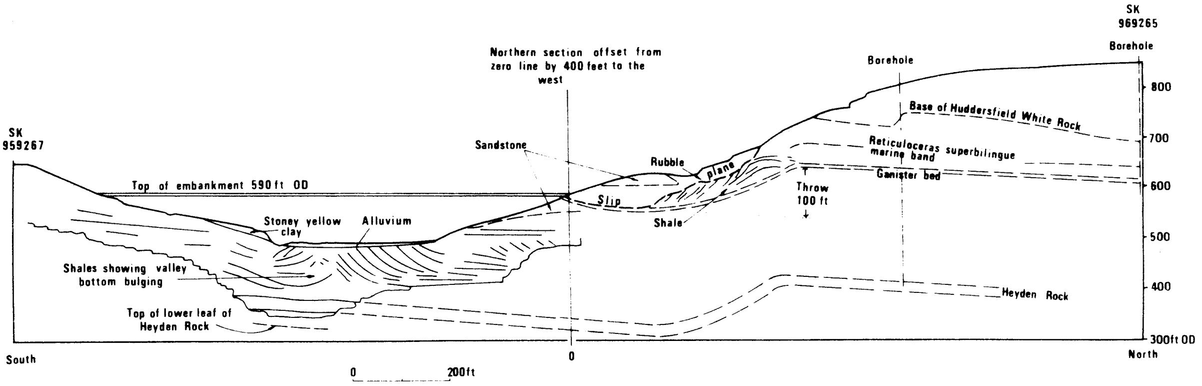
At Loadfield Quarries (258949), the Huddersfield White Rock is 12 - 15 m thick. It is a well-bedded, fine-grained, siliceous sandstone. At the top are 2 m of ganister with a flaggy sandstone above bearing the rhizoids of *Stigmaria sp.* at the horizon of the Ringinglow Coal.

On the north side of the valley, the Huddersfield White Rock forms a prominent scarp across the hillside from Yew Trees Wood at 267967 where 6 m of alternating flaggy, laminated and massive sandstone is found, to Thorpe's Brow (239976), and 3 m of flaggy cross-bedded, micaceous sandstone is seen in a quarry.

To the north of Ewden Valley, the Upper Meltham Coal is well developed above the Huddersfield White Rock and to the south similarly, as the Ringinglow Coal. This seam is only poorly developed on the ganister above the Huddersfield White Rock in Ewden and was found at one locality, Loadfield Quarries (258949).

Structure

The structure contour map (text-fig. 2) has been assembled in three parts using three stratigraphic horizons which are respectively best exposed in the southwest, centre and northeast, providing the best control from as many points of observation as possible. As exposures of the marine bands are rare, the top or bottom surfaces of the major sandstones have been used. While a single surface map could be produced by adding or subtracting thicknesses as appropriate, this procedure might well overlook variations in thickness. Both the Rough Rock and the Huddersfield White Rock are known to vary in thickness regionally, though detailed variations in the Ewden area cannot be determined. The method also has the advantage of showing a previously unrecognised structure, the Ewden Monocline, which is not mentioned in either of the memoirs, though disturbed strata were revealed in the investigations of the Waldershelf Landslip which occurred during the construction of the Broomhead Reservoir between 1924 and 1930 (Bendelow 1944 ; Section re-drawn as text-fig. 3)



Text-fig. 3. Section along the Broomhead Dam Trench showing the Ewden Monocline and the the Waldershelf Landslip of 1924 (after Bendelow 1944).

During the cutting of the bywash bay at the northern end of the embankment the removal of a sandstone "toe" caused instability allowing a large mass of sandstone and shale to slip down into the bywash bay. A series of boreholes was followed by deep drainage trenches being cut and these revealed the existence of a monoclinal flexure dropping the beds by 30 m to the south over a width of 120 m. Some evidence of earlier movement was found in these investigations suggesting that the disturbance extended westwards along strike to the site of the old Broomhead Lead Mine now submerged in the reservoir, where disturbed sandstones of the Heyden Rock could be inferred from the very limited mine records. Possible further extension of this structure to the west is suggested by the irregularities of dip in Lee Lane Dyke. However, much of the displacement caused by the monocline seems to be taken up by the Ewden Fault trending west-northwest parallel to Ewden Beck from where Wood Brook enters the Reservoir at 261962. The Ewden Fault has been inferred from the necessity of explaining the relationship of the Heyden Rock which is some 60 m lower than its expected stratigraphical position below the Huddersfield White Rock. This relationship is at variance with Bromehead *et al.* (1933) as their map infers a fault in the same position with the throw in the opposite direction, to the north.

Eastwards the Ewden Monocline is less obvious, but nevertheless still present. East of Ewden Village a flexure is visible in the structure contours on the base of the Huddersfield White Rock, and still further east these and contours on the Rough Rock show a distinct flexure down to the south though with an elongate depression on the crest at Hollin Edge. Lower Coal Measures are preserved hereabouts.

The structure contour map also shows the Wharncliffe Anticline clearly in the southeast. Although there has been little comment in the literature, this anticline is clearly displayed as a broad but gentle structure in the Wharncliffe Crags escarpment to the east of the Ewden area, where the fold plunges to the east-northeast. The present study shows that the Wharncliffe Anticline extends further west than previously noted.

Conclusions

The re-mapping of the Ewden Valley has provided field evidence for the identification of the Heyden Rock in the outcrops previously noted as Rivelin Grit by Bromehead *et al.* (1933) at the western end of the Ewden Valley. This supports the correlations shown in Ramsbottom's pictorial diagrams (1966 and 1969) wherein the Rivelin Grit is equated with the Huddersfield White Rock, and the Heyden Rock is equated with the Pule Hill Grit. The structural analysis has added considerably to knowledge of the Ewden Monocline, previously known only from the foundation studies of the Waldershelf Landslip. Although seemingly of local significance only, this flexure is parallel to and has the same sense of movement as the Don Monocline to the south.

Acknowledgements

Thanks are due to Dr. N.J.D. Butcher for his assistance in the field and to Dr. T.D. Ford for his critical reading of the manuscript.

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THE SULPHIDE MINERALOGY AND PARAGENESIS OF THE SOUTH PENNINE OREFIELD, ENGLAND

by

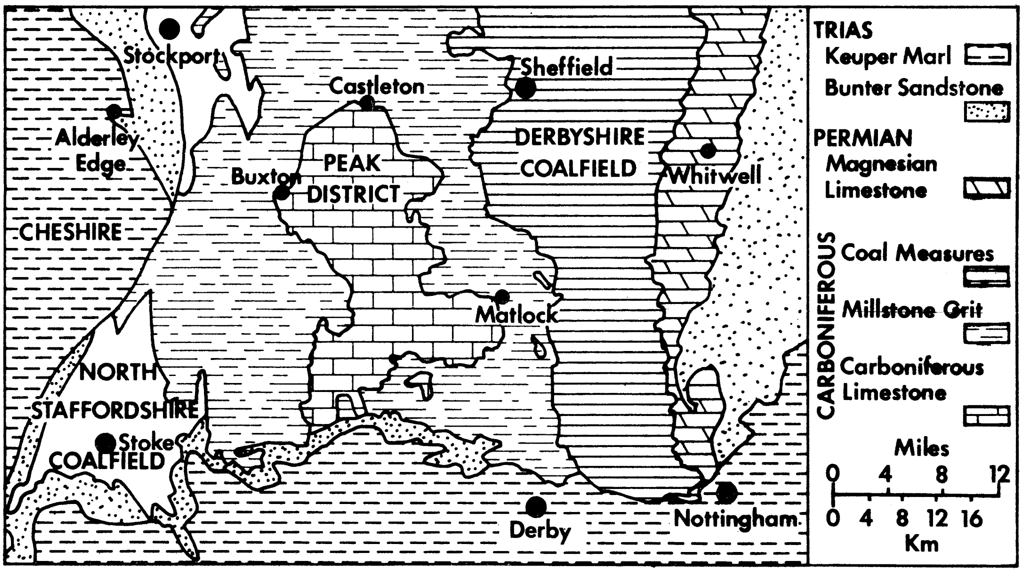
R.A. Ixer and R. Townley

Summary

Throughout the South Pennine Orefield, the primary sulphide mineralogy is remarkably uniform and simple, and displays a consistent paragenetic sequence. The main sulphides, in their generalized paragenetic sequence are bravoite, nickel-rich and nickel-poor pyrite and marcasite, chalcopyrite, galena and sphalerite. There is no systematic variation in the abundance, presence or absence of any sulphide mineral except for an overall increase towards the more mineralized eastern margin of the orefield. The widespread occurrence of nickel bearing sulphides throughout the orefield, and their appearance in similar deposits elsewhere, suggests that nickel is an important and characteristic minor element of the ore-fluids that produce Mississippi Valley style mineralization.

Introduction

The South Pennine orefield is composed of the galena-sphalerite-baryte-fluorite deposits occurring within the Carboniferous Limestone outcrop of the Derbyshire Dome and its associated outliers of Ashover and Crich, (text-fig.1). The anomalous copper-rich district of Ecton, Staffordshire is not considered to be part of the orefield. Together with the North Pennine orefields, that of the south has been classified as belonging to the Mississippi Valley type of deposit by Dunham (1966), and Worley and Ford (1977), although Emblin (1978) has proposed a new classification for them both, namely the Pennine type.



Text-fig. 1. An outline geological map of the South Pennine Orefield.

Mercian Geol. Vol. 7 No. 1
1979 pp. 51-63, 3 text-figs., Plates 7-10

The mineralization consists of small vein deposits, metasomatic replacements, or cavern infillings with a simple primary mineralogy. This comprises calcite, baryte, fluorite and occasionally quartz 'gangue' together with much smaller amounts of galena and sphalerite and very minor amounts of iron, nickel and copper sulphides. The sulphides have two main styles of crystallisation, either as small inclusions within the 'gangue', or, less commonly, as mixed sulphide ore. Subsequent oxidation has resulted in numerous secondary sulphides, carbonates, sulphates and hydrated oxides.

The orefield was formerly mined for galena with minor exploitation of cerussite, sphalerite, smithsonite, wad and ochre (Ford and Ineson, 1971). The orefield is now a major producer of fluorspar; with minor barytes, and galena often extracted as by-products of the fluorspar mining. A little calcite is also mined producing some by-product galena.

The economic importance of the fluorspar deposits has meant that recent mineralogical studies of the orefield have concentrated on the minerals excepting the sulphides, to the virtual exclusion of these sulphides. There is now, however, enough information to allow detailed discussion of the mineralogy, paragenesis and distribution of the sulphides within the orefield for the first time.

Previous mineralisation investigations in the South Pennines

General aspects of the South Pennine orefield have been discussed by Wedd and Drabble (1908), Varvill (1959), Ford and Ineson (1971) and Ford (1976).

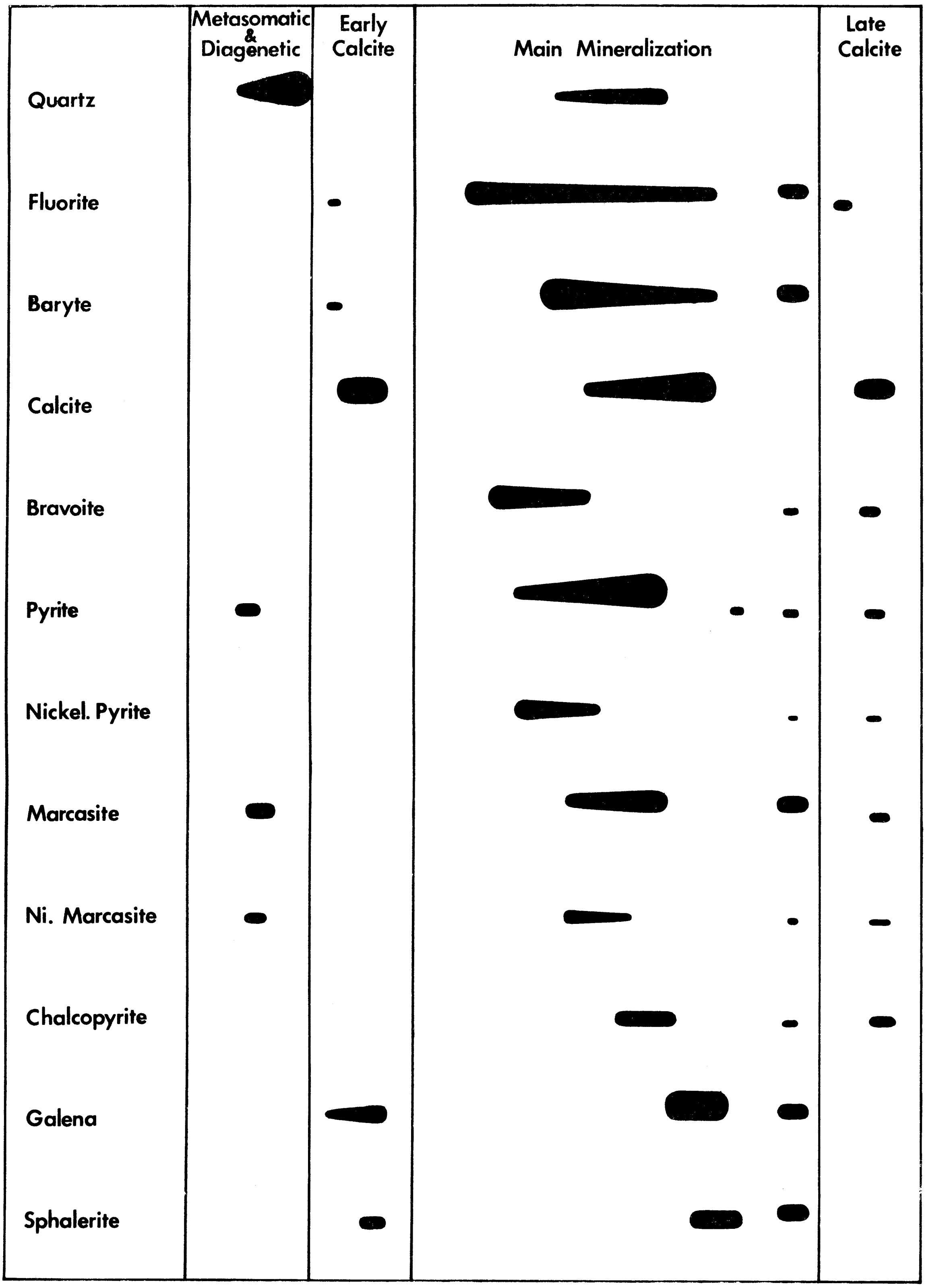
Regional zoning of the non-sulphides was described by Wedd and Drabble (1908), Dunham (1952) and especially by Mueller (1951, 1954a) who delineated three temperature controlled zones. The zones are an eastern fluorite zone passing westwards into a cooler baryte zone and finally a calcite zone. Additionally, Mueller (1951) defined a pyritic calcite zone and subdivided the fluorite zone into three, the easternmost of which consisted of fluorite with sulphide inclusions. Varvill (1959) suggested the existence of two northwest to southeast trending 'lead-belts'. More recently, Firman and Bagshaw (1974), in an extensive re-examination of both the mine records and field evidence, found that Mueller's concept of a fluorite zone succeeded westward by baryte is broadly true but the junctions are by no means clear cut.

Table 1 shows Mueller's (1954a) table for the concentration of minerals in his three zones, which is the only quantitative estimate of the amounts of sulphide present in the orefield. Ford (1976) has amended this assessment by noting the restriction of sphalerite to the eastern margin, and the more widely spread occurrence of pyrite and chalcopyrite. Townley (1976) qualitatively investigated the regional distribution of the sulphides as briefly discussed by Ixer (1978).

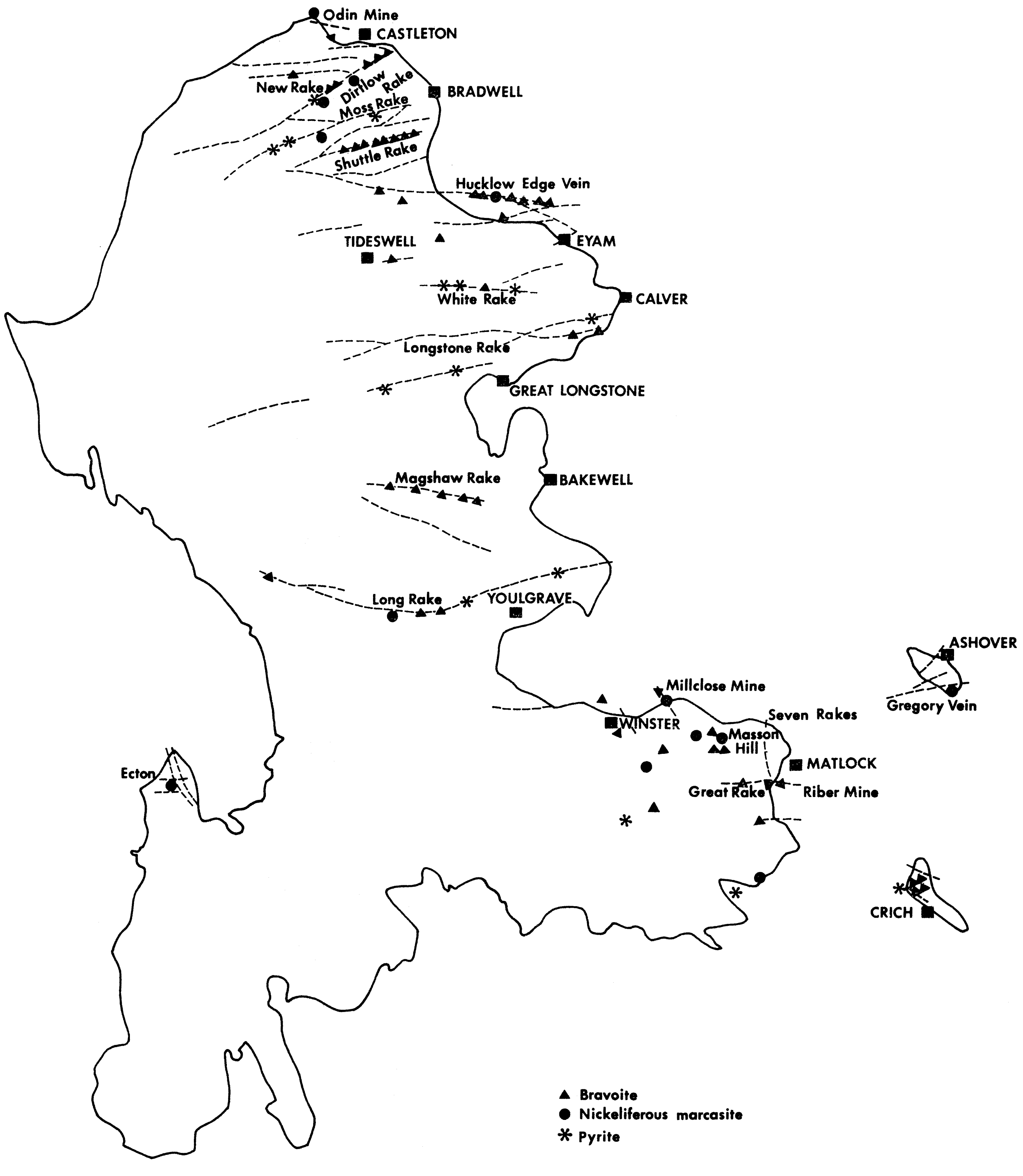
Table 1 Zonal Distribution of Hydrothermal Minerals (Mueller, 1954a)

	CaF_2	$BaSO_4$	$CaCO_3$	PbS	ZnS	$CuFeS_2$	FeS_2
Fluorite Zone	10-50%	2-10%	40-85%	1-5%	0.5-3%	0-10%	0-5%
Baryte Zone	1-10%	10-40%	55-85%	0.5-3%	0-1%	-	-
Calcite Zone	0-1%	0-10%	80-99%	0-3%	0-0.5%	-	-

Few detailed mineralogical and paragenetic studies of individual deposits have been published. These include bravoite and other sulphides from Millclose Mine, Wensley Dale (Bannister, 1940); Vaughan, 1969); the baryte-galena deposit of Golconda Mine, Brassington (Ford and King, 1965) and the fluorite-calcite-sulphide mineralization of a pipe deposit in the Blende Vein, at Sheldon near Bakewell (Worley, 1976). Ford (1976) correlated the paragenetic sequences from four other individual deposits, those of the multiphase baryte-fluorite vein



Text-fig. 2. A generalized paragenesis for the South Pennine Orefield. The diagram gives no information as to the relative proportions of the minerals; the quantities of sulphides are far less than those of the non-sulphides.



Text-fig. 3 Distribution of bravoite, nickeliferous marcasite and pyrite in the South Pennine Orefield.

deposits of Raper Mine (Ineson and Al-Kufaishi, 1970); the essentially single phase fluorite-baryte-calcite-sulphide replacement deposit of Masson Hill (Ixer, 1974) and two vein deposits from Ashover, (White, 1968). Although the four deposits show some differences in their detailed paragenesis, in that the rakes show repetitive phases of mineralization whilst the flats show a more continuous mineralization sequence, this is largely due to differences in mineralization styles. Rake deposits being more susceptible to tectonic movements than metasomatic flats will show a number of generations of mineralization as they open and close to the mineralizing fluids, whereas unmineralized portions of a flat deposit are always receptive to the fluids. However, all four deposits show an early silicification followed by an extensive phase of fluorite, baryte and sulphide mineralization and, finally, late calcite. Firman and Bagshaw (1974) proposed a similar generalized paragenesis comprising an early rhombohedral/columnar calcite phase with galena and rare sphalerite, the main flourite-barite and sulphide mineralization and a final scalenohedral calcite phase.

Similar studies have been carried out on the North Pennine Orefield which is mainly situated over two basement highs, the Alston and Askrigg Blocks. The mineralisation of the Alston Block shows clear concentric zoning of both sulphides and non-sulphides but no clear paragenetic sequence, except that fluorite is earlier than baryte (Dunham, 1948, 1966). Similarly, the Askrigg block has been shown to contain mineral zoning (Small, 1978). Recently, Hagni and Trancynger (1977) have described a detailed paragentic sequence from the Magmont Mine, Missouri, which may be used as a model for the 'typical' Mississippi Valley ores, and Ixer (1978) has noted the mineralogical similarity between the sulphides of this type of deposit and certain red-bed deposits, including that of Alderley Edge, Cheshire.

Method of study

Representative mineralized specimens were collected from 80 sites throughout the orefield, mainly from small outcrops within open old mine workings or their accompanying dumps. The location of the collecting sites and description of the material collected at them is given in Townley (1976). Three distinct types of mineralized specimen were found; large single crystals of calcite, fluorite or baryte carrying sulphide inclusions; mixed sulphide-gangue or banded sulphide and non-sulphide ores: mineralized lavas and agglomerates.

Multiple polished sections were made of material from three richly mineralized areas, namely Masson Hill-Oxclose Mine (28 sections), Millclose Mine (12 sections) and Odin Mine (12 sections), and the paragenesis established from the textural relationships of the sulphides. Additionally, 94 sections were made, one or more from each site, and these were used to investigate the full sulphide mineralogy of the ores and to search for any evidence of regional zoning.

Paragenesis

Investigation of the ore material shows the sulphides to have three distinct modes of formation. The most common mode is as bands of inclusions in fluorite, calcite, baryte or, more rarely, quartz. This has been discussed by Mueller (1954b), Ford (1976), Firman and Bagshaw (1974) and Ixer (1974). Less commonly galena, sphalerite and marcasite form a banded ore with baryte and fluorite, the good terminations of the crystals indicating that this ore-type is due to open void infilling; or even rarer, the sulphides and non-sulphides are randomly intergrown producing a non-banded mixed ore. All three modes of occurrence are largely contemporaneous.

The sulphides will now be discussed in greater detail, both as inclusions and as non-inclusions.

Sulphide inclusions

Bravoite, marcasite and chalcopyrite are the most common inclusions, with nickeliferous pyrite and marcasite, and pyrite being less common; galena, sphalerite and haematite are rare.

Typically the inclusions occur in bands which are up to 500 µm wide and are sub-parallel to each other and to the growth zones of the enclosing host mineral. The sulphides are especially concentrated at the junction of two host minerals which are usually baryte, fluorite or calcite. The bands can be mono-mineralic chalcopyrite or marcasite, or comprise bravoite with successive pyrite and marcasite overgrowths. Each band, however, is comprised of one type of inclusion with its own modal size, although adjacent bands may have different modal sizes and mineralogies. Alternating bands of chalcopyrite and marcasite are common.

Bravoite inclusions have a pentagonal dodecahedral habit and mainly belong to the zoning types II, IV and V of El Baz and Amstutz (1963) and have two modal sizes of 30 µm and 5 µm. Larger crystals up to 150 µm are rare, having been found at Crich and near Winster. They show a variety of colours from brown, purple, lilac to steel-grey and within one crystal colour zoning (up to 10 µm wide) is usual. The colouration varies between successive bands and even within one band there are variations in colour, successive order of colours, and the width and arrangement of zoning. Partial overgrowths of pyrite (10 µm) and marcasite (15 µm), both often nickeliferous, form aggregates up to 80 µm in size. The overgrowth is characteristically in the growth direction of the enclosing host mineral.

Euhedral to subhedral nickeliferous pyrite up to 50 µm is typical, as are marcasite laths, up to 200x10 µm, although small marcasite mosaics are also common. Nickel-rich zones in the marcasite are of the order of 5 µm wide. Chalcopyrite is found as anhedral or rare subhedral crystals up to 450 µm enclosing iron-nickel sulphides, or as isolated grains. Sphalerite and galena are rare forming 100 µm anhedra and 30 - 60 µm rounded grains respectively. Very fine haematite dust, far smaller than 1 µm, has been found in calcite and fluorite crystals near Bradwell and in quartz near Bonsall and Calton Hill. Fine haematite inclusions are common in vesicular calcite from the altered lavas of the orefield (Ixer, 1972).

Other inclusions are native copper and bornite identified in hand specimen within late stage calcite from Riber Mine (Varvill, 1959). Mueller (1954b) stated that the sequence of inclusion bands in crystals could be correlated between different veins and over some distance. The present study has shown that the crystallization sequences can be quite complex and therefore the correlation of crystals between veins would be difficult, using recognition of these bands as the sole criterion. In one instance a limited correlation was found to be practicable. The very rare and hence distinctive band of haematite inclusions in fluorite crystals found in the Bradwell Pipe could be correlated across the many smaller pockets of vug infilling or metasomatic replacements that make up the pipe over a distance of approximately 100 m.

Non-inclusion sulphides

These will be discussed in the order of their paragenetic sequence during the main fluorite-baryte-sulphide mineralization, as shown by their textural relationships. A more complete paragenesis is depicted in text-fig. 2.

Iron-nickel sulphides

Bravoite was identified optically by its zoning and/or deep colouration. Its occurrence in the South Pennine orefield has been reviewed in Lunn *et al.*, (1974) and Ixer (1978), who has also summarized the available chemical analyses. The two major textures, first recognised from Oxclose Mine (Ixer, 1974) are constant throughout the orefield:

(a) Brown to pink-brown euhedral to subhedral crystals (20 - 60 μm) occur and reach 1000 x 250 μm at Oxclose and approximately 3.5 cms at Millclose Mine (Bannister, 1940). Zoning is broad but sharply defined, although within each zone fainter and more diffuse zones can be seen. Colouration is usually darker towards the centre of the grain but as was shown by Vaughan (1969) there is no precise correlation between colour and nickel, cobalt and iron content of the grain. Most of the eight zoning types described by El Baz and Amstutz (1963) are represented. Pyrite and marcasite enclose or are intergrown with this bravoite (Plate 7, figs. 1 and 2). This is especially true at Oxclose Mine where a pseudomyrmekitic texture has resulted, (Plate 7, figs 3 and 4).

(b) Less common are small (5 - 10 μm) very euhedral, lilac, zoned bravoite crystals enclosed within 200 μm pyrite aggregates.

Both types of bravoite, as with all the iron-nickel sulphides, are oxidized to goethite.

Nickeliferous pyrite was distinguished from pyrite by its pink/brown colour, faint zoning and lower reflectivity. It has a more restricted paragenesis than pyrite as it only crystallized within the main mineralization phase.

The earliest pyrite is present as diagenetic framboids (10 - 35 μm) within shale or limestone host rocks, and is associated with poorly crystalline 'spongy' pyrite, (Plate 8, figs 1 and 2). It is surrounded by radiating twinned marcasite euhedral crystals (Plate 8, fig. 3) or enclosed by chalcopyrite anhedral crystals as at Odin Mine, (Plate 8, fig. 4). The pyrite and nickeliferous pyrite found with the main mineralization typically form rims and overgrowths (10 - 20 μm) around bravoite, or as independent anhedral or subhedral crystals (70 μm) associated with it, (Plate 9, figs. 1 and 2). At the end of the main mineralization, large pyrite grains, often up to several centimetres in size, are associated with baryte and calcite crystals. These pyrite crystals contain relict marcasite laths (400 x 100 μm). The final phase comprises veinlets (5 - 10 μm) of poorly polished brown pyrite that forms as rims around all other primary sulphides. This is the 'melnickovite' pyrite of Ramdohr, (1969) although a less mineralogically ambiguous term would be colloidal pyrite.

Within the altered and mineralized lavas, there are also a number of pyrite generations. The earliest is a poorly crystalline spongy pyrite that encloses and replaces titanium dioxide pseudomorphs after ilmenite laths. Later subhedral pyrite, introduced with calcite veining, contains thin (2 - 5 μm) faint nickeliferous zones. Angular pyrrhotite (up to 70 μm), showing patchy alteration to marcasite and haematite, is found within this pyrite but is not associated with the nickeliferous areas. Pyrrhotite is recorded elsewhere, notably from the Tansley borehole (Ramsbottom *et al.*, 1962) and from Eyam and Longstone Edge (Ford and Sarjeant, 1964).

Nickeliferous marcasite was distinguished optically from marcasite by its green-white to blue-lilac reflection pleochroism and lower reflectivity. Its presence was first recorded from the orefield by Townley (1976) and Ixer (1978) who also gave confirmatory electron microprobe analyses.

As with the other iron-nickel sulphides, there are a number of generations of marcasite crystallization. The earliest marcasites are the small framboids and radiating marcasite crystals associated with the pyrite framboids (Plate 10, fig. 1) and of diagenetic origin. Some of the marcasite is nickeliferous. Most of the nickeliferous marcasite is associated with the main mineralization. The marcasite crystallizes both as overgrowths and intergrowths with bravoite and pyrite. Small (10 μm) nickeliferous areas and 1-3 μm lilac-brown nickeliferous zones can be observed throughout the marcasite in a similar manner to zoning type VIII of El Baz and Amstutz (1963). The zones are sinuous, cutting anisotropic boundaries and are often the continuations of zoning within enclosed bravoite. Euhedral bladed marcasites commonly carry repeated 5 μm wide nickeliferous zones parallel to the crystal outline, (Plate 9, figs. 3 and 4). Later generations of marcasite include the relict marcasite crystals within large pyrite crystals and large radiating marcasites associated with alternating baryte, fluorite, sphalerite and galena within the rhythmically banded ore.

Chalcopyrite

Chalcopyrite is less common than the iron-nickel sulphides but encloses them, seen especially in samples from Odin Mine (Plate 7, fig. 4). Chalcopyrite forms twinned anhedral crystals reaching a maximum size of approximately 0.5 cms. Smaller euhedral to subhedral crystals (30 - 50 μm) occur along growth zones in sphalerite and are concentrated at the margins (Plate 10, fig. 3). Chalcopyrite is variously altered to bornite, covellite, chalcocite, cuprite and green carbonates; the alteration forming successive concentric rims 4 - 10 μm wide (Plate 10, fig. 4).

Galena

Galena is seen as subhedral to euhedral crystals up to 10 cms across independently of whether it is an early or late generation. It commonly encloses other sulphides, and rarely, native silver at Odin Mine. Native silver has been reported from the South Pennine orefield (Ford and Sarjeant, 1964) but generally the galena has a low silver content (< 50 ppm) (El Shazly *et al.*, 1957). Alteration to cerussite and anglesite is common, accompanied by framboidal secondary galena.

Sphalerite

Sphalerite is rarer than galena. It is iron-poor with less than 5 mol % Fe, normally with yellow-white internal reflections, but some have a deep purple colouration along twin planes. El Shazly, *et al.*, (1957) report a restricted trace element content for sphalerites from Millclose Mine with only cadmium (3000 ppm) and silver (200 ppm) being of significant concentration. Higher cadmium values, up to 0.88 wt % are reported from Oxclose Mine (Ixer, 1972). Smithsonite containing relict sphalerite and greenockite flecks (5 μm) also suggests locally higher cadmium at Oxclose Mine. Alteration of sphalerite is extensive, with rims of covellite and goethite and replacement by hemimorphite, smithsonite and goethite, (Plate 10, fig. 2).

Other sulphides

A few other sulphides have been reported, including millerite, in association with pyrite; pyrrhotite, chalcopyrite and sphalerite from the Tansley borehole (Ramsbottom *et al.*, 1962) and from Ashover; cinnabar in smithsonite from the Matlock area (Braithwaite and Greenland, 1963); tetrahedrite from Millclose Mine (Bannister, 1940) and arsenopyrite from Eyam and Longstone Edge (Ford and Sarjeant, 1964).

Regional distribution

The iron-nickel sulphides were the most widely distributed phases within the ore material studied, generally as bravoite accompanied by pyrite and marcasite. Both zoned and unzoned bravoites were identified from 50 sites, unzoned bravoite from a further 8 and optically identifiable nickeliferous marcasite from 12. The distribution of bravoite and nickeliferous marcasite is shown in text-fig. 3. At only 16 sites were no nickeliferous minerals recognised and these sites were randomly distributed and often closely surrounded by others containing a high proportion of nickel minerals. Galena was almost as widespread but chalcopyrite and sphalerite were less common.

There was a general increase in the occurrence of sulphides towards the more intensely mineralized eastern margin of the orefield, but no other regional variation could be detected in either the presence or absence of phases or their relative amounts.

Insufficient material was collected for any quantitative discussion on the relative abundances of the sulphides, but a visual estimation, using all the sections, gave the following relative modal proportions: galena very much greater than sphalerite > marcasite > pyrite very much greater than bravoite > chalcopyrite.

Summary and conclusions

The primary sulphide mineralogy of the ores is simple and uniform throughout the orefield and consists of iron-nickel sulphides, chalcopyrite, galena and sphalerite. Most of the sulphide mineralization is epigenetic, the contribution of diagenetic processes is slight, unlike at Magmont Mine, Missouri where much of the sulphide is thought to be diagenetic, especially the early generations of pyrite and bravoite (Hagni and Trancynger, 1977). Both field and textural evidence suggests that the ores have resulted from the three main episodes of mineralization discussed by Firman and Bagshaw (1974); namely an early calcite-galena-sphalerite episode followed by the main fluorite-baryte-calcite-sulphide mineralization, and finally calcite again. The repetitive nature of the mineralization, which is typical of the Mississippi Valley type of deposit, is most clearly demonstrated by the sulphide inclusions. Throughout the mineralization there is a constant sequence of bravoite with successive pyrite, marcasite and chalcopyrite overgrowths. This repetition is also shown by the three major generations of galena and sphalerite, and the two generations of pyrite and marcasite. However, significant bravoite and chalcopyrite crystallization only occurred once and this within the main mineralization. This restricted paragenetic position largely accounts for the relative paucity of the two sulphides.

The distribution of the sulphides shows no evidence for any east-west variation or zoning, other than a general increase of all sulphides eastwards, nor is there any evidence of lead-belts. This supports the view of Firman and Bagshaw (1974) who found a similar increase in intensity of mineralization towards the eastern margin of the orefield. However a more subtle regional zoning may be present, and if so, a systematic investigation of the trace element contents of sphalerite, galena and chalcopyrite would seem to provide the best chance in recognising it. At present, there are virtually no published trace element analyses for the sulphides from the South Pennines Orefield. Chemical analyses of the nickel minerals show that the nickel content within one specimen can vary widely, for example between 0.15 - 6.40 wt.% in nickeliferous marcasite and between 8.0 - 24.0 wt.% for bravoite, Ixer (1978). This variation in chemistry, which reflects the extreme sensitivity of the iron-nickel sulphides to changes in the local mineralizing fluids, means that its use would be severely limited in investigating regional variations.

The widespread distribution of sulphides containing nickel in the South Pennine orefield is indicative of nickel as a characteristic minor element in the ore fluids, of at least equal importance with copper. Bravoite is known from similar deposits found in the Askrigg Block and Mendip Hills in England (Townley, 1976) and accompanies vaesite in the Cave-in-Rock district Illinois (Park, 1967) and siegenite from Magmont Mine, Missouri (Hagni and Trancynger, 1977) suggesting that nickel is characteristic of Mississippi Valley style deposits generally.

Acknowledgements

Some of this work represents a M.Sc. thesis presented to the University of Aston in Birmingham (R. Townley), who wishes to thank Mrs. Nona Martin for her constant help and encouragement. Dr. D.J. Vaughan is thanked for his reading of the manuscript, as is Dr. T.D. Ford for his suggestions; and the technical staff of Aston for the execution of the typing, photographs and diagrams.

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EXPLANATION FOR PLATES 7 and 8

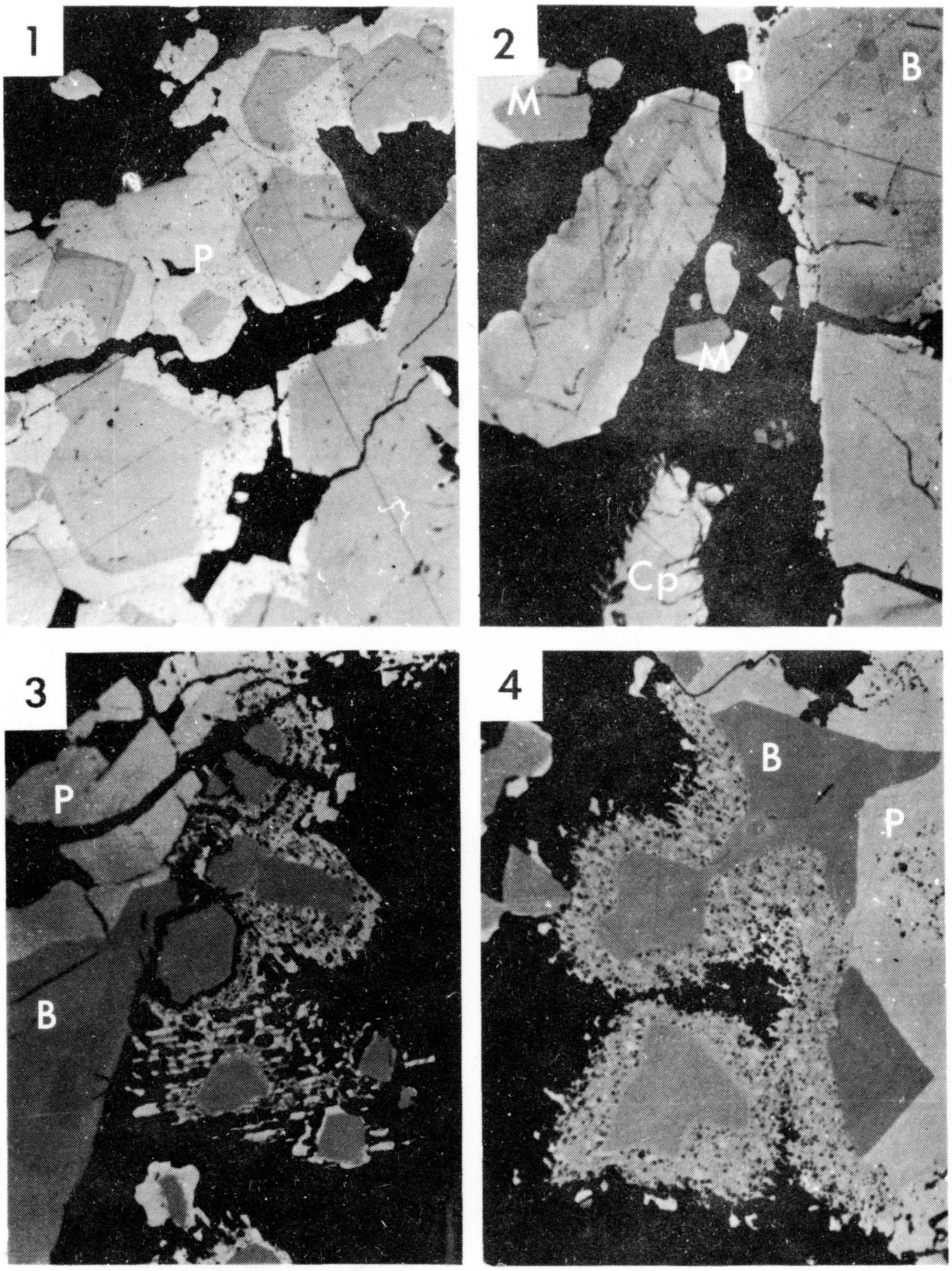
Plate 7, fig. 1	Euhedral zoned bravoite (dark greys) enclosed in zoned nickeliferous pyrite (P) and pyrite. Later fracturing of the sulphides have been healed by fluorite x 180.
Plate 7, fig. 2	Irregularly zoned bravoite is associated with unzoned bravoite (B) which is enclosed in pyrite (P) and euhedral marcasite (M). Chalcopyrite (Cp) that is altering along its cleavage to covellite and geothite is also present x 250.
Plate 7, fig. 3	Zoned bravoite (B) is enclosed in nickeliferous pyrite (P). Bravoite and pyrite are intergrown in a coarse pseudomyrmekitic texture. Later fracturing is cemented by fluorite x 400.
Plate 7, fig. 4	Zoned bravoite (B) is partially enclosed by nickeliferous pyrite (P) and partially surrounded by a fine-grained pseudomyrmekitic intergrowth of bravoite and pyrite x 500.

All specimens are from Oxclose Mine, reflected light, plane polarised light, oil immersion.

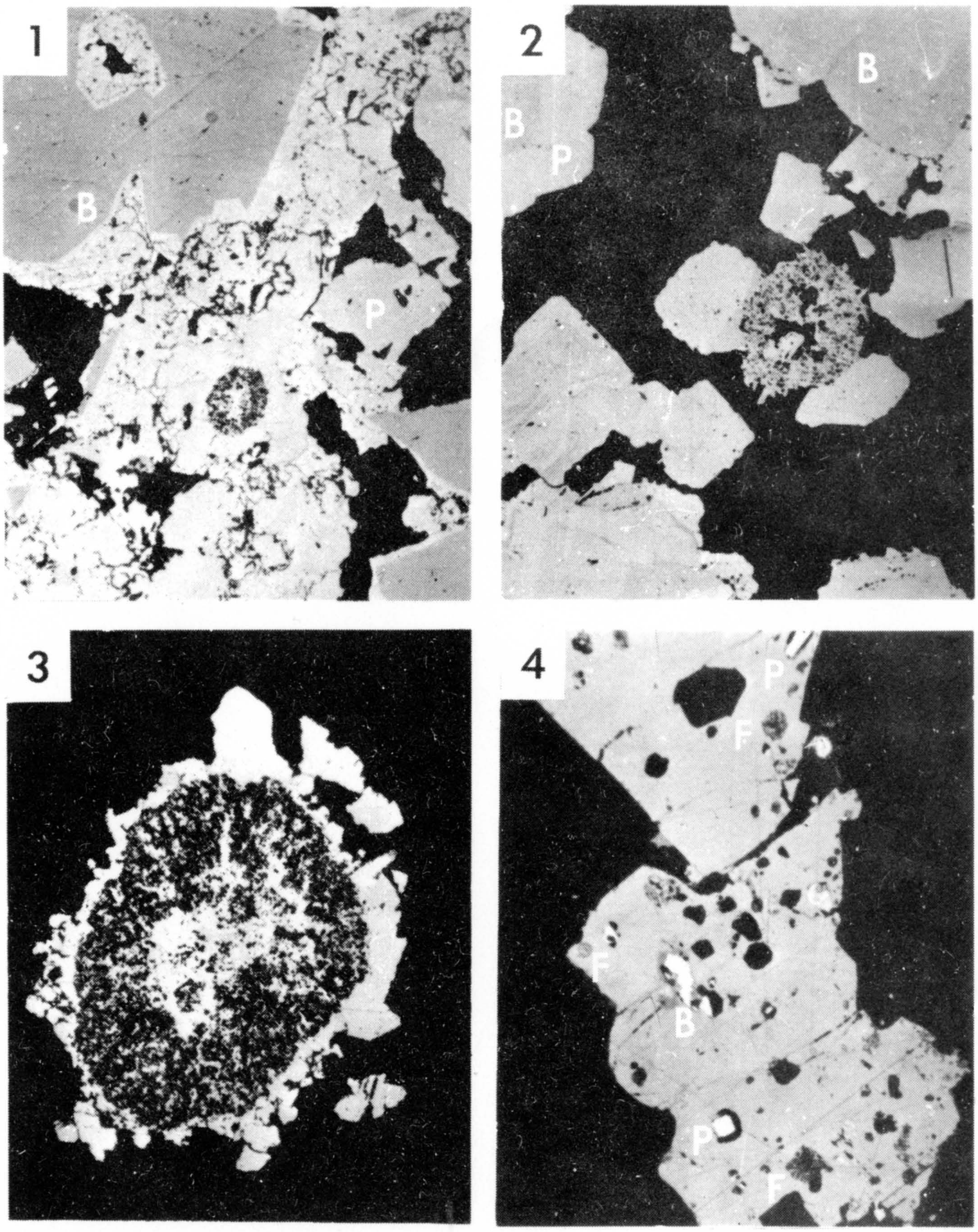
Plate 8, fig. 1	Rounded spongy pyrite (centre) enclosed in nickeliferous pyrite (P), marcasite and coarse-grained bravoite (B). Oxclose Mine, x 120.
Plate 8, fig. 2	Diagenetic pyrite (centre) with a rim of acicular marcasite. Irregularly zoned and unzoned bravoite (B) and pyrite (P) is associated. Oxclose Mine, x 250.
Plate 8, fig. 3	Oxidized spongy diagenetic pyrite together with the accompanying rim of radiating euhedral marcasite crystals. Oxclose Mine, x 600.
Plate 8, fig. 4	Anhedral chalcopyrite enclosing euhedral pyrite cubes (P), bladed marcasite (white) with bornite rims (B) and oxidized pyrite framboids (F). Odin Mine, x 150.

All photomicrographs are in reflected light, plane polarised light and oil immersion.

Plate 7



Ixer & Townley S. Pennine Sulphate mineralogy (For explanation see p. **62**).



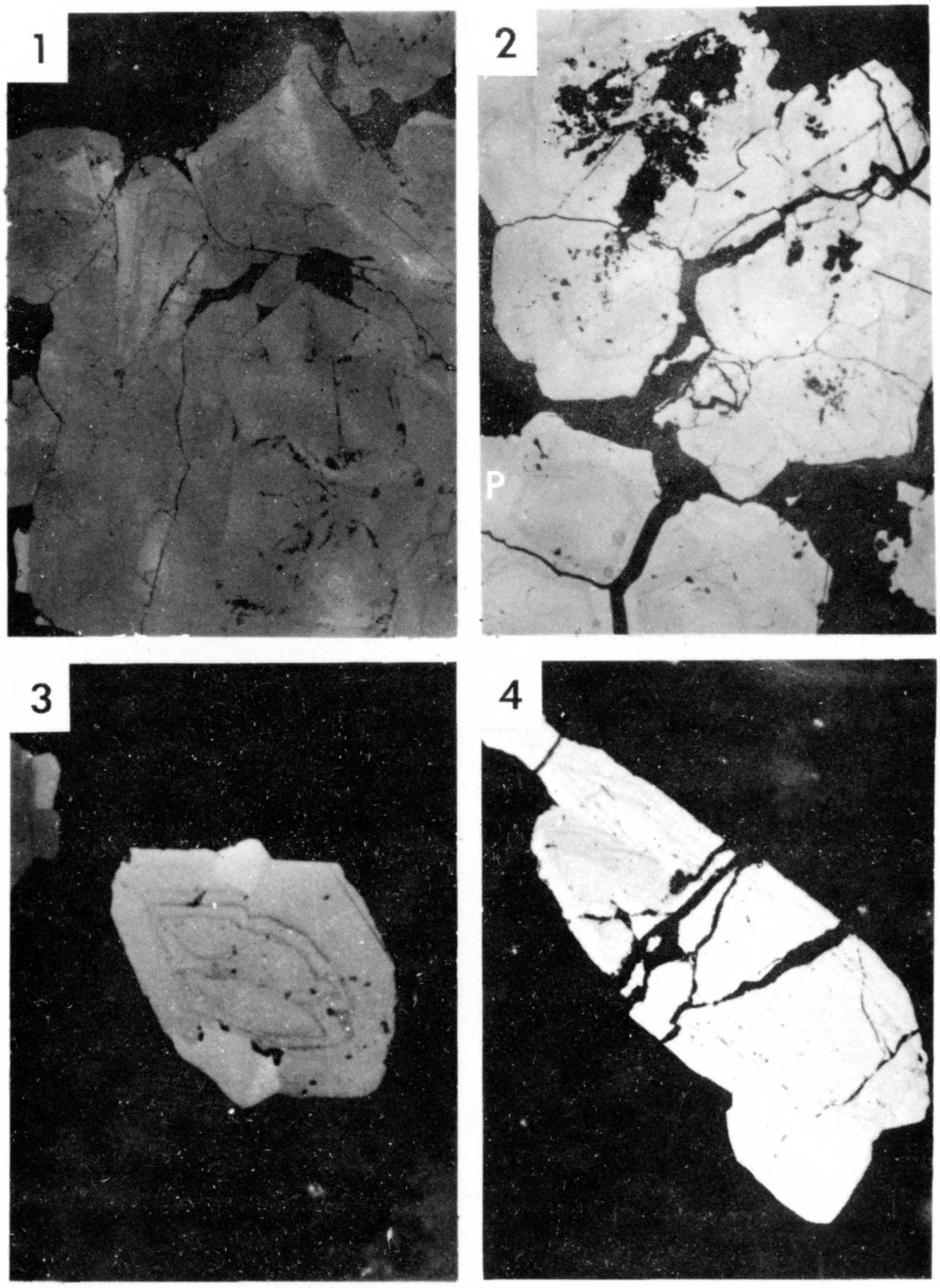
Ixer & Townley S. Pennine sulphide mineralisation
(For explanation see p. **62**).

EXPLANATIONS FOR PLATES 9 and 10

Plate 9, fig. 1	Nickeliferous pyrite showing faint but characteristic zoning. Oxclose Mine, x 200, plane polarised light.
Plate 9, fig. 2	Zoned nickeliferous pyrite subhedra with pyrite overgrowths (P). Later fracturing is cemented by calcite. Hucklow Edge, x 300, plane polarised light.
Plate 9, fig. 3	Euhedral and zoned marcasites. Different shades of grey show grains in slightly different anisotropy positions. Odin Mine, x 1400, partially crossed polars.
Plate 9, fig. 4	Fractured zoned marcasite. Odin Mine, x 400, partially crossed polars.

All photomicrographs are in reflected light, oil immersion.

Plate 10, fig. 1	Radiating aggregate of late generation marcasite with crystals in differing anistropy positions. Odin Mine, x 450, partially crossed polars, oil immersion.
Plate 10, fig. 2	Euhedral marcasite (M) associated with sphalerite. The sphalerite shows the characteristically lighter coloured rim associated with its alteration. Oxclose Mine, x 250, plane polarised light, oil immersion.
Plate 10, fig. 3	Anhedral sphalerite (S) with triangular chalcopyrite inclusions and a chalcopyrite rim. The chalcopyrite encloses pyrite cubes (P) and framboidal pyrite. Calcite (dark grey) and baryte (lighter grey B) are the 'gangue' minerals. Odin Mine, x 100, plane polarised light.
Plate 10, fig. 4	Anhedral chalcopyrite altering to covellite about its margin and along its cleavage. Oxclose Mine, x 250, plane polarised light, oil immersion.



Ixer & Townley S. Pennine sulphide mineralisation
(For explanation see p. **64**).

Plate 10

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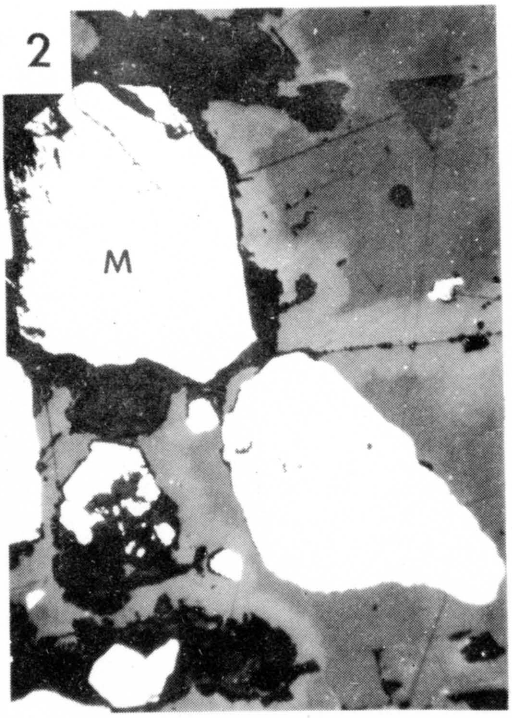
2



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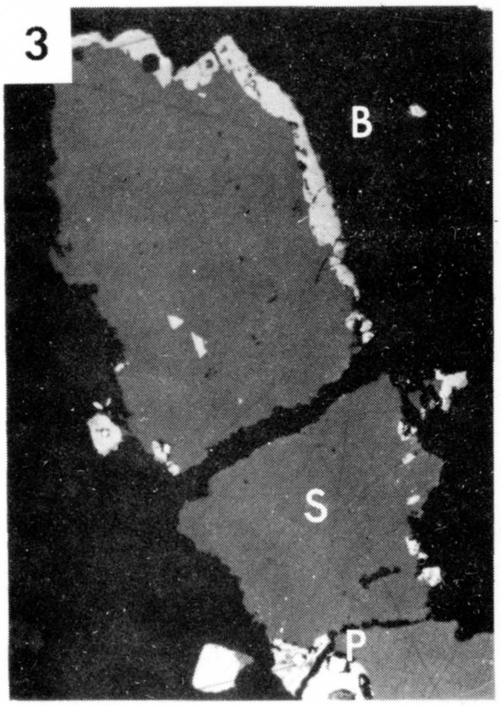

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Ixer & Townley S. Pennine sulphide mineralisation
(For explanation see p. **64**).

A NEW COLOBODONT FISH FROM THE TRIAS OF SPAIN

by

J.H. Sykes and O.J. Simon

Summary

Andalusias ewerti a new genus and species of fish, based on teeth from Triassic carbonate rocks of the Betic Cordilleras, southern Spain, is described and its affinities discussed. The stratigraphic position, and associated fauna of the rock samples, are considered.

Introduction

Examination of Triassic rocks from southern Spain for their microfossil content has revealed a wide range of fish remains, details of which are being prepared for publication. A preliminary faunal list is given in Table 1, p. 68. A type of fish teeth, quite common in some samples, is considered to differ sufficiently widely from known genera to be allocated to a new genus, which forms the subject of the present paper. The specimens have been obtained from a number of horizons of carbonate rocks in the provinces of Jaén and Almeria.

Within the Betic Cordilleras, the Alpine Fold-Belt of southern Spain, two major tectonic zones can be distinguished: The External, and Internal or Betic, Zones (text-fig. 1). The former is made up of the Prebetic and Subbetic Zones, consisting of non-metamorphic rocks of Permo-Triassic and younger age. In the Betic Zone, four super-imposed tectonic complexes are recognised, which in ascending order are:

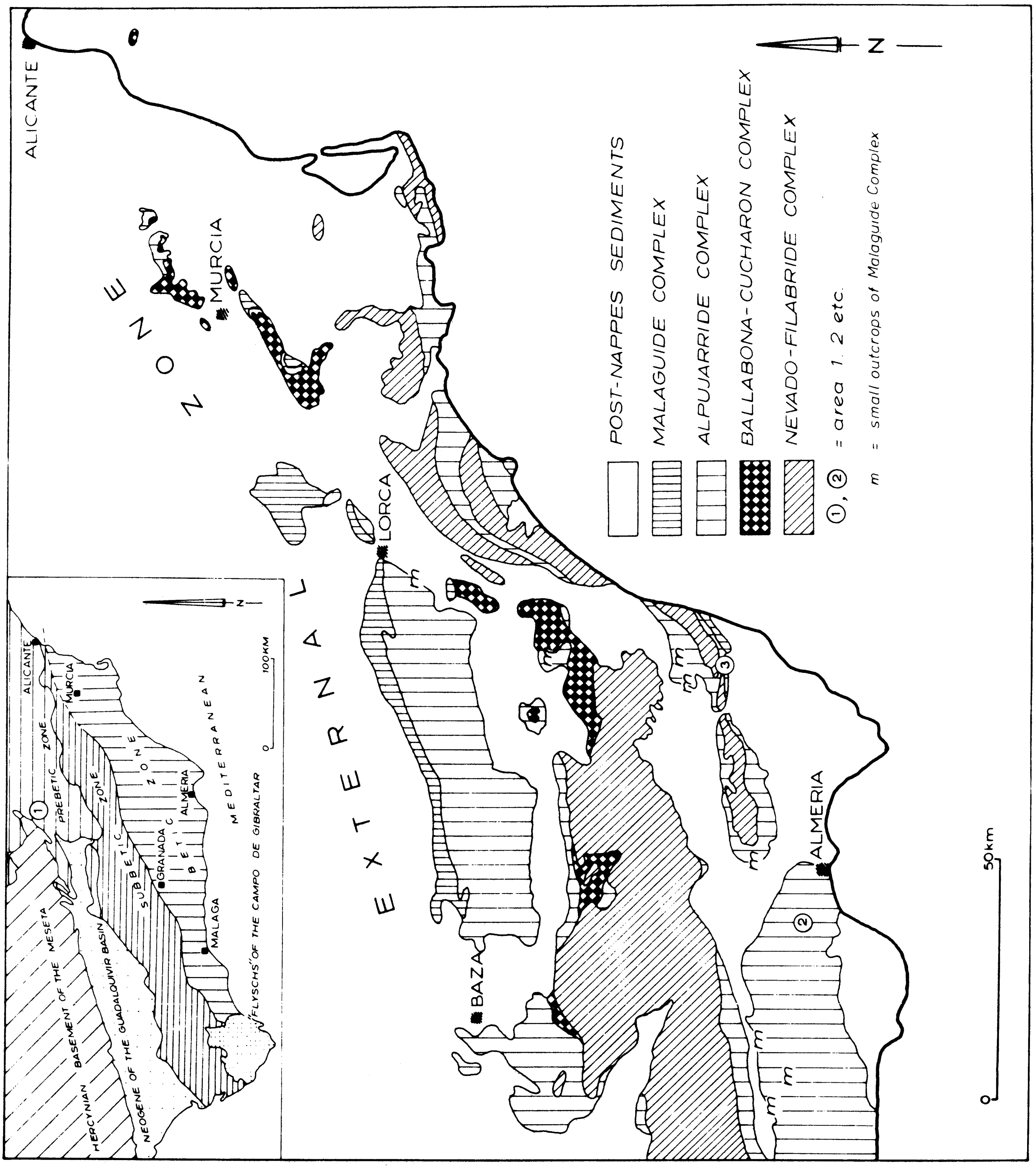
(4) Malaguide Complex
(3) Alpujarride Complex
(2) Ballabona-Cucharón Complex
(1) Nevado-Filabride Complex

after Egeler & Simon, (1969), Egeler & Fontboté, (1976).

The complexes listed above consist mainly of Triassic and older sediments, which generally show the effects of Alpine metamorphism. The Triassic is the only system with representative strata in all four complexes of the Betic Cordilleras. Consequently only Triassic rock sequences can be used for correlation between the various tectonic complexes and therefore are of essential importance for a reconstruction of the palaeogeography of the Betic Realm. Until recently dating was essentially based on lithological criteria and on some well preserved macrofaunas, but in the last few years also with the aid of microfossils. Early results using these fossils have been published by Van den Boogaard (1966), Simon (1966), Kozur & Simon (1972), Van den Boogaard & Simon (1973), Kozur *et al.* (1974) and Simon & Kozur (1977).

All the fish fragments are minute, almost entirely being less than 2 mm in size. The specimens were extracted by dissolving the carbonate rocks in dilute formic acid. This method of preparation along with the original metamorphism of the specimens from the Betic Zone (EW 76/007, Si 71/042 and 24-42/Ka/292), have combined to affect their state of preservation. Instead of the usual black and brownish colours of the phosphatic remains, they are nearly all white or yellowish-brown. Many are very fragile and appear to have a granular or corroded surface. Specimens from the External Zone (Si 77/016), which have not been metamorphosed, are better preserved.

Mercian Geol. Vol. 7, No. 1. 1979
pp. 65-74, 3 text-figs., Plate 11.



Text-fig. 1 - Tectonic sketch-map of the Betic Cordilleras.

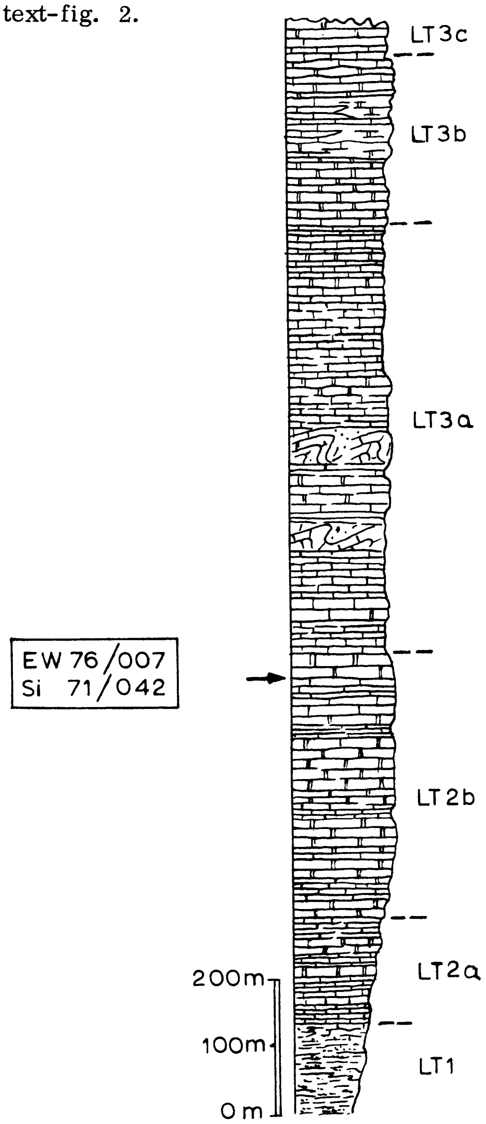
Prebetic Zone (area 1 of text-fig. 1.)

The holotype and paratypes, nos. 5, 6, and 7, have been obtained from sample Si 77/016, collected approximately 16 m above the base of a carbonate sequence in the 'Hornos-Siles Formation' (Lopez Garrido, 1971). From carbonate rocks in this Formation, bivalves and cephalopods have been reported by Schmidt (1935), Lopez Garrido (1971) and Hirsch (1977).

Micro-fossils from sample Si/77/016 comprise gastropod and echinoderm fragments with conodonts (*Pseudofurnishius murcianus* Van den Boogaard, 1966). Hirsch (1977) considers that the carbonate sequence of the 'Hornos-Siles Formation' can be correlated with the higher part of the Fuente-Aledo Formation of the Ballabona-Cucharón Complex (Betic Zone), as both have comparable bivalve and conodont faunas. He attributes a late Ladinian age to this part of the Fuente-Aledo Formation and to the carbonate sequence of the 'Hornos-Siles Formation'. Kozur *et al.* (1974) suggests a Cordevolian (early Carnian) age to the higher part of the Fuente-Aledo Formation on the evidence of ostracodes and holothurian sclerites. Preliminary determinations of the microflora indicate a late Ladinian to early Carnian age for the carbonate sequence of the 'Hornos-Siles Formation'.

Betic Zone (Alpujarride Complex, area 2 of text-fig. 1).

Paratypes nos. 1, 2, 4, and 8 from samples EW 76/007 and Si 71/042 come from the Alpujarride Lujar Unit (*sensu* Ewert, 1976) in the Sierra de Gador. A simplified vertical section of the Lujar Unit with the indicated stratigraphical position of the samples is given in text-fig. 2.



LT3c - Limestones

LT3b - Dolomites

LT3a - Limestones with intercalations of calcschists and dolomites.

slumped sediments and breccias.

LT2b - Dolomites

LT2a - Calcschists with intercalations of dolomites and gypsum.

LT2a - LT3c middle to late Trias

LT1 - Permo-Triassic phyllites, quartzites and calcschists; gypsum.

Text-fig. 2. Composite columnar section of the Lujar Unit, area 2 of text-fig. 1. after Ewert, 1976).

The macrofossil, *Nautilus* sp., has been described from the same bed as that from which sample Si 71/042 was taken (Jacquin, 1970). The microfauna contains conodonts and ostracodes. The conodonts, *Pseudofurnishius murcianus* Van den Boogaard 1966, and *Epigondolella mungoensis* (Diebel 1956), indicate a late Landinian age (Simon & Kozur, 1977; Hirsch, 1977), but a preliminary determination of the ostracodes from sample EW 76/007 suggest an early Carnian age. Samples from the lower part of sequence LT 3a (text-fig. 2) contain holothurian sclerites pertaining to the early Carnian. Approximately 150 m below samples EW 76/007 and Si 71/042 carbonate rocks occur which contain ostracodes indicating a late Ladinian age. Awaiting definite determinations of the ostracodes of sample EW 76/007, a late Ladinian to early Carnian age is tentatively assigned to samples EW 76/007 and Si 71/042.

Betic Zone (Alpujarride Complex, area 3 of text-fig. 1.)

Paratype 3 obtained from sample 24-42/Ka/292 was collected from the lower-most part of an Alpujarride carbonate sequence overlying a phyllite-quartzite sequence. The exact stratigraphical position of the sample in the sequence could not be established due to the fact that the contact between the two sequences at that place is of a tectonic nature. The macrofauna consists of indeterminate bivalves. The microfauna includes ostracodes, foraminifera and echinoid fragments. Preliminary dating of the ostracodes suggests a late Ladinian age.

Table 1 - Fossil frequency in relation to the Betic Zone Complexes and to the External Zone.

Fossils	Nevado-Filabride Complex	Ballabona-Cucharón Complex	Alpujarride Complex	Subbetic	Prebetic
Hybodus plicatilis		1	2		
Acrodus sp. 'A'			1		
Acrodus sp. 'B'		2			
Acrodus sp. 'C'		1			5
Indet. Dermal Denticles	3	2	164	1	164
Saurichthys longidens					17
Saurichthys apicalis		1	17		2
Birgeria mougeoti			3		1
Gyrolepis albertii			6		2
Gyrolepis sp.		1	52	18	302
Andalusias ewerti.		1	55	4	44
Indet. teeth 'A'		2	12		19
Indet. teeth 'B'			4		2
Indet. teeth 'C'	10				
Indet. scales			18		43
Indet. fish		1	9		10

All Specimens are deposited in the Geological Institute, University of Amsterdam.

Systematic Palaeontology

Outline classification based on Andrews *et al.* (1967).

Sub-class	Actinopterygii
Order	Perleidiformes
Family	Colobodontidae Stensiö 1916
Genus	*Andalusias* gen. nov.

Name:- derived from the area in which the specimens were found.

Diagnosis

Sub-circular and oval crushing teeth with multi-tuberculate caps which may or may not cover the upper surface of the pedicel. The caps are depressed and have major and minor tubercles. They vary between types which are flattened and seated horizontally on the pedicel and others which are inclined and rather spikey.

Type species, *Andalusias ewerti* sp. nov., pl.11, figs. 1 - 9.

Name:- in honour of Mr. Klaus Ewert of Adra, Spain.

Diagnosis

Variable crushing teeth, oval and sub-circular in upper view with a barrel-shaped pedicel. Depressed tooth caps, seated horizontally on the pedicel, have one major and several minor tubercles peripherally on the upper face; these caps usually equal the pedicel in length and width. Teeth having a cap which is smaller in length and width than the pedicel, generally have fewer minor tubercles. Many of the caps are seated at an angle on the pedicel; the more steeply inclined the cap, the more significant is the major tubercle and the less significant are the minor tubercles. Teeth which are oval in upper view have a depressed cap with one major tubercle, one minor tubercle and, at times, minor tubercles on the upper face.

Holotype: Specimen no. K3337, sample Si 77/016, pl. 11, fig. 1.
Type locality: Siles, province of Jaen, Universal Transverse Mercator Coordinates (U. T. M. C.), 05.35.47/42.48.60, Prebetic Zone, Hornos-Siles Formation, late Ladinian - early Carnian.
Paratypes 1 and 4: P58853 and P59290, sample EW 76/007, pl. 11, figs. 2 and 5.
Type locality: Sierra de Gador, province of Almeria, U. T. M. C., 05.21.85/40.75.97., Alpujarride Complex, late Ladinian-early Carnian.

Paratypes 2 and 8: P58854 and P592294, sample Si 71/042, pl. 11, figs. 3 and 9.
Type locality: Sierra de Gador, province of Almeria, U. T. M. C. 05.32.82/40.82.20., Alpujarride Complex, late Ladinian - early Carnian.

Paratype 3: P58855, sample 24 - 42/Ka/292, pl. 11, fig. 4.
Type locality: Sierra Gabrera, province of Almeria, U. T. M. C., 05.92.10/40.99.52., Alpujarride Complex, late Ladinian.

Paratypes 5, 6, 7: P59291, P59292 and P59293, sample Si 77/016, pl. 11, figs. 6, 7, and 8.
Type Locality: Siles, province of Jaen, U. T. M. C., 05.35.47/42.48.60. Prebetic Zone, 'Hornos-Siles Formation', late Ladinian - early Carnian.

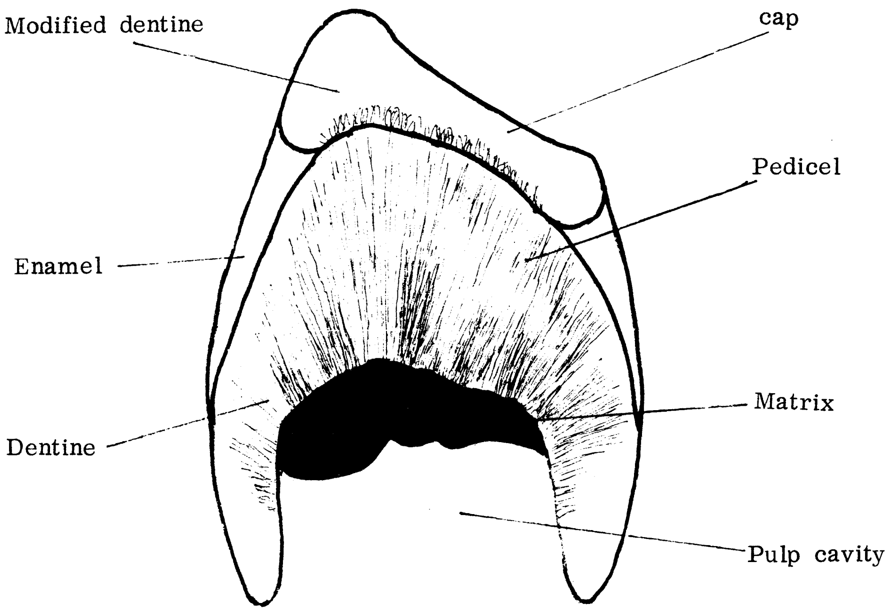
The holotype numbered K 3337 and the thin section, text-fig. 3, are deposited in the Geological Institute, University of Amsterdam; the paratypes, in the British Museum of Natural History, London.

Description of the teeth.

The isolated teeth consist of single crowns having a depressed cap on a short pedicel which has a pulp-cavity in the lower part. In well preserved specimens the pedicel is generally barrel-shaped, narrowing slightly at the top and down to the base. The pulp-cavity is wide making the walls of the pedicel thin, the lowest part of the pedicel being rarely preserved.

The teeth vary between types which are oval in upper view (pl. 11, fig. 8 & 9 , and those which are sub-circular to slightly oval in upper view (pl. 11, figs 1 to 7). In the latter type, which are by far the most common, the tooth cap is depressed and has a series of tubercles on the upper surface (pl. 11, figs. 1 to 6). These consist of one major tubercle and up to twelve minor tubercles, arranged peripherally around the upper surface of the cap. Most of this type are found as isolated tooth caps (pl. 11, fig. 3). In some more complete specimens the cap is as wide and as long as the pedicel, covering the upper area of the tooth (pl. 11, fig. 2). In others the cap is less than the length and width of the pedicel (pl. 11, figs. 1, 4 & 5) and, generally, these have less minor tubercles. Most of the latter type have the cap seated at an angle on the pedicel. When the inclination is steep, minor tubercles are insignificant or absent and the major tubercles prominent and off-centre (pl. 11, figs. 6 & 7).

The type which is oval in upper view has a rather different cap; there are two dominant tubercles on the upper surface, one larger than the other (pl. 11, figs. 8 & 9). Nearly all the specimens of this type are found as isolated tooth caps, though, when found attached to the pedicel, they are seated horizontally. Only one oval specimen shows a tooth cap which is smaller in its width and length than the pedicel.



Text-fig. 3. Median, lateral section of tooth from sample Si 77/016, X 140.

In thin section (text-fig. 3) the teeth are shown to comprise of an independant cap of modified dentine seated on a pedicel made up of dentine with a surrounding layer of enamel. Radiating through the dentine there are very fine tubules which are dense in parts. They penetrate a short way into the cap at an angle which makes a dense tangled mat of tubules.

Discussion

The Families Colobodontidae and Lepidotidae Owen 1860, and the Order Pycnodontiformes have species which possess crushing teeth. These teeth are low, rounded and knob-like and are placed in rows or patches on the jaws and roof of the mouth. Each tooth has a cap of enamel or modified dentine on a stem-like pedicel which has a pulp-cavity in the lower part. The pedicel is usually short and is fixed at its base to the supporting bone. The new species is of this general morphological tooth type.

In the Family Colobodontidae there are four genera with fishes which have simple, tuberculate crushing teeth. The genus *Colobodus* Agassiz 1844, has several such species which, in general, have teeth with a centrally placed apical tubercle on a cap which is striate (Woodward, 1895; p. 68, Dames, 1888, pls. 2 & 3).

Nephrotus chorzowensis H. v. Meyer 1851, has some non-striate crushing teeth many of which have a simple, low, round tubercle on the upper face (Sykes, 1979, pl.13 fig. 7). Some have a more pointed, off-centre tubercle and there are much larger marginal teeth which are cylindrical in shape (Schmidt, 1928, p.36

The genus *Aetheodontus* Brough (Brough, 1939, p. 51) is described as having crushing dentition of cylindrical, blunt, teeth, varying in size and having a rounded crown with a small central pappilla. The species *Aetheodontus besanensis* Brough (Brough, 1939, p. 52) has short cylindrical teeth which have an off-centre tubercle on the apex; some of the teeth having a rather spikey appearance.

The genus *Perleidus* Stensiö 1932 has fishes with teeth similar to those in the genus *Aetheodontus*. *Perleidus stoschiensis* Stensiö (Stensiö, p. 218) has several rows of minute teeth some of which are oval and most of which are round in upper view. They have an apical tubercle which is either central or off-centre.

Teeth are one of the hard remains of fishes which are most likely to be preserved and become fossilised. Many species have been founded on isolated teeth and in considering their affinities both anatomical and stratigraphical considerations must be taken into account The Colobodonts are important and well known throughout the Trias and they have proved extremely useful in zoning the Triassic rocks of Bulgaria (Stefanov, 1977). Of the four genera mentioned *Colobodus* and *Nephrotus* range from the early to the late Triassic whilst *Aetheodontus* is Middle Triassic and *Perleidus* early Triassic.

Fish crushing teeth with tuberculate caps are quite common amongst the species of the Family Colobodontidae and there are several in which the apical tubercles are placed off-centre. The more spikey types of the Spanish specimens compare quite closely with many teeth of the fishes in the genera *Aetheodontus* and *Perleidus*. Species of the latter genus also show examples of a similar combination of both oval and rounded teeth. However, none of these species has teeth which compare with the multi-tuberculate nature of the Spanish specimens and so it is concluded that the new species should be placed in a new genus within the Family Colobodontidae.

In the Family Lepidotidae there are also fishes with small crushing teeth, though teeth from the Triassic Lepidotids are not very well known. One specimen from the Middle Triassic of Northern Italy (British Museum of Natural History No. P19355) has minute oval teeth with a small, rather pointed, off-centre, apical tubercle. The teeth vary in size but all have the same form.

Teeth of the genus *Lepidotes* Agassiz, 1833, are rare in the Triassic rocks and they generally have a smooth, round cap. Specimens of *Lepidotes* teeth from The Wealden (Patterson, 1966, p. 258; Sykes, 1979, pl. 12 fig. 9) have a single apical tubercle some of which are placed off-centre. Other exceptions being a small sample of specimens from the Rhaetic at Watchet (Woodward, 1895, p. 122) which are tentatively placed in the genus *Lepidotes* and some similar teeth from the Rhaetic at Barnstone (Sykes, 1979).

Although some of the Spanish specimens have features in common with Lepidotids there is not sufficient evidence to show that they belong to that family.

Many fossil fishes in the Order Pycnodontiformes have smooth, rounded, knob-like teeth many with an apical indentation. In the genus *Mesturus* Wagner, 1862, fish teeth have a crimped and indentented coronal apex.

One species from the Upper Jurassic *Mesturus leedsi* Woodward, 1895, has teeth which have varying patterns of minute tubercles on an indented upper surface (Lehman, 1966, fig. 166). A pycnodont from the Lower Jurassic, *Eomesodon liassicus* Edgerton has smooth teeth with a single apical tubercle but generally the pycnodont teeth are of a non-tuberculate form. *Eomesodon hoefori* from the late Triassic of Austria, is the only Triassic pycnodont species known and its teeth are not determined.

An approach to the depressed nature of the horizontally seated caps of the Spanish specimens is made in the pycnodont *Mesturus* though major and dominant, off-centre tubercles are not present in that genus. The stratigraphical horizon of *Andalusias ewerti* is considered to be of Ladinian-Carnian age, near the Middle and late Triassic boundary (Kozur, *et al.*, 1974). To consider the species as a pycnodont would extend the range of that Order on a species of isolated teeth that show affinities to a family which is much better known and more significant in the Trias.

Acknowledgements

The authors are greatly indebted to Dr. C. Patterson (British Museum Nat. Hist.) for his assistance with the palaeontology. Thanks are due to Mrs. C. Mulder-Blanken for taking the scanning electron micrographs, to Miss M. Rietveld, Mrs. C. Mulder-Blanken and Mr. J. Kaper for the preparation and sorting of the residues. Thanks are also due to Mr. Klaus Ewert for information on the geology of the Sierra de Gador, to Dr. H. Visscher and Mr. R. Besems for information on the geology of the Siles region and for their help during the collection of the samples. One of the authors (O.J. Simon) wishes to thank the Netherlands Organisation for the Advancement of Pure Research (Z.W.O.) for financial support.

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EXPLANATION FOR PLATE 11

Fig. 1. Tooth with inclined tuberculate cap, holotype, oblique lateral view. No. K 3337.

Fig. 2. Tooth with horizontal tuberculate cap (detached from pedicel in transit), Paratype no. 1, oblique upper view, no. P58853.

Fig. 3. Detached tooth cap, paratype no. 2 upper view, no. P58854.

Fig. 4. Tooth with broken pedicel, paratype no. 3, lateral view, no. P58855.

Fig. 5. Tooth with slightly inclined, tuberculate cap, paratype no. 4 oblique, lateral view, no. P59290.

Fig. 6. Worn tooth with inclined tuberculate cap, paratype no. 5, oblique lateral view no. P59291.

Fig. 7. Tooth with steeply inclined cap, paratype no. 6, lateral view, no. P59292.

Fig. 8. Oval tooth with major tubercles, paratype no. 7, oblique upper view, no. P592293.

Fig. 9. Oval tooth with major and minor tubercles, paratype no. 8, transverse, oblique, upper view, no. P59294.

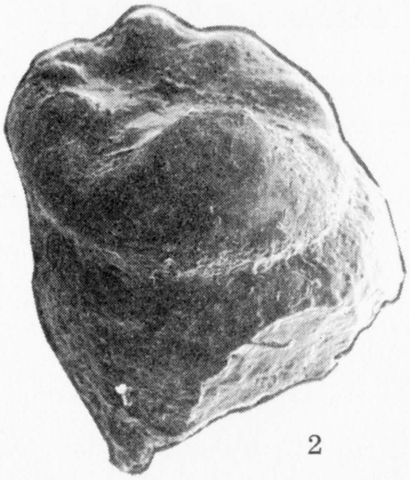
P numbers refer to specimens in the British Museum (Nat. Hist.)

Dimensions (in mm)

Fig. no.	Length	Depth	Height
1	0.4	0.4	0.7
2	0.3	0.3	0.4
3	0.6	0.7	0.3
4	0.4	0.5	0.4
5	0.4	0.4	0.7
6	0.3	0.3	0.4
7	0.3	0.3	0.6
8	0.3	0.4	0.5
9	0.5	0.7	0.3

Plate 11


1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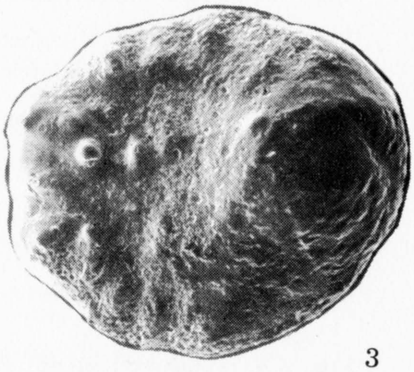
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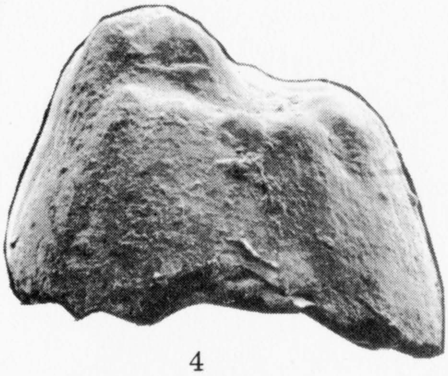
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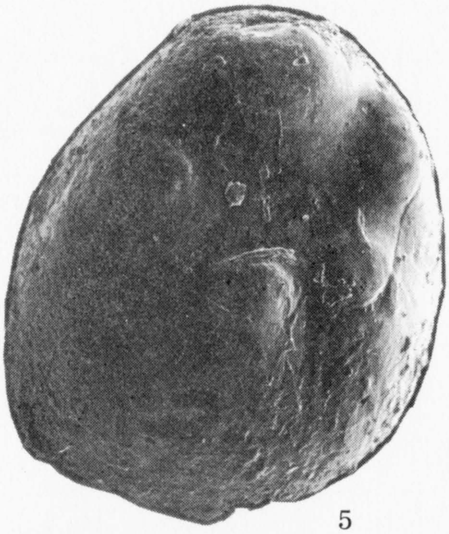
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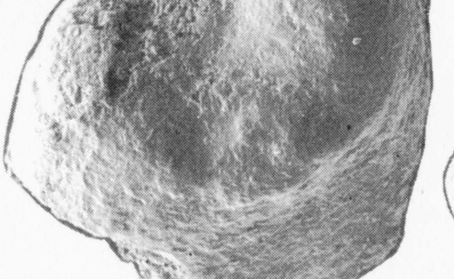
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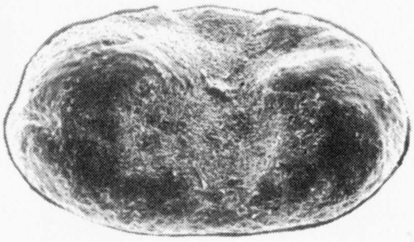
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EXCURSION REPORT: LATHKILL DALE AND PILSBURY, DERBYSHIRE.

Leaders: N. Aitkenhead and J. I. Chisholm Sunday, 12th June, 1977.

The purpose of the excursion was to examine the youngest (D_1-D_2) [1] part of the Derbyshire Carboniferous Limestone, first on the relatively stable 'shelf' area of deposition, and then at the boundary between the shelf and a more rapidly subsiding 'basin' that lay to the west. On the shelf there is an unbroken sequence of shallow-water limestones but at the shelf margin the sequence is interrupted by erosional breaks and diversified by the presence of apron-reef limestones dipping outwards towards the basin. The sequences in the two areas are compared in Table 1. The recently published 1:25000 geological sheet SK16 (Monyash) covers both areas (Chisholm, Aitkenhead and Price 1977).

The party assembled at Monyash and were taken by coach to Haddon Grove, SK 180 662, where the morning's leader, J. I. Chisholm, pointed out that the surrounding gentle slopes of the limestone plateau approximate to dip-slopes, developed on surfaces near the top of the Monsal Dale Limestones.

The strata lie in a broad shallow syncline, the Monyash Syncline; Lathkill Dale is incised into the axial region of this structure. The party descended into Lathkill Dale by a track, the sides of which provide a good section through the Monsal Dale Limestones. Pale limestones predominate in the top part of the sequence, and contain a cherty shell bed (Lathkill Shell Bed) full of brachiopods in a concave -up position. The band has proved to be an excellent marker horizon, invaluable in the mapping of this area. It is recognisable throughout Lathkill Dale and extends eastwards into Bradford Dale, SK 202 639, and northwards to Bole Hill, SK 184 676.

At Carters' Mill, SK 1839 6579, the party paused to examine thinly bedded dark limestones that underlie the pale limestones, and the possible reasons for the colour difference were discussed. The greater concentration of hydrocarbons in the dark facies is traditionally regarded as the main factor, but the higher clay content of dark limestones may also play a part. The party then walked up the dale towards Monyash, stopping briefly to collect silicified corals in the dark limestone, and to examine the source of the River Lathkill. At Ricklow Quarry SK 165 661 the party climbed up the dale side to see a reef knoll in the Eyam Limestones, which rest sharply on the Monsal Dale Limestones. Most time was spent examining the limestones that abut against the knoll, including a coarsely crinoidal limestone (the quarried bed) and dark well-bedded limestones.

After lunch at Monyash the party proceeded across the outcrop of the older Bee Low Limestones to the vicinity of Pilsbury where an outlier of Monsal Dale Limestones forms a shallow syncline on the shelf (north-east) side of the discontinuous apron-reef. A brief stop was made on the north-eastern flank of this syncline, SK 1207 6364, to examine small exposures showing the contrast between the pale Bee Low Limestones and the overlying dark shelly basal beds of the Monsal Dale Limestones. The apron-reef itself was then examined where it crops out on the steep hillside, 1192 6325, overlooking the road between Pilsbury and Parks Barn. The reef-limestones here are typical of the facies, being pale grey, poorly bedded, and fine-grained or micritic, with a rich brachiopod fauna. At the top of the hillside, SK 1200 6329, at a level estimated to be c. 9.0 m stratigraphically above the top of the apron-reef, the party examined a 10 m crag of pale limestone made up largely of brachiopod shells. The fauna has been examined by M. Mitchell of the Institute of Geological Sciences who reports that it consists mainly of gigantoproductoids including *G. edelburgensis* and *Linoprotonia hemisphaerica*, and is probably of D_2 age. A nearby gully, SK 1209 6325, immediately above Parks Barn exposes some 18.7 m of limestone below the level of these shelly beds. Near the base of this sequence members examined a 1.6 m bed of conglomerate in which rounded pebbles of micrite are dispersed in a coarsely crinoidal matrix.

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[1] The coral-brachiopod zones D_1 and D_2 are broadly equivalent to the Dinantian stages Asbian and Brigantian of George *et al.* 1976

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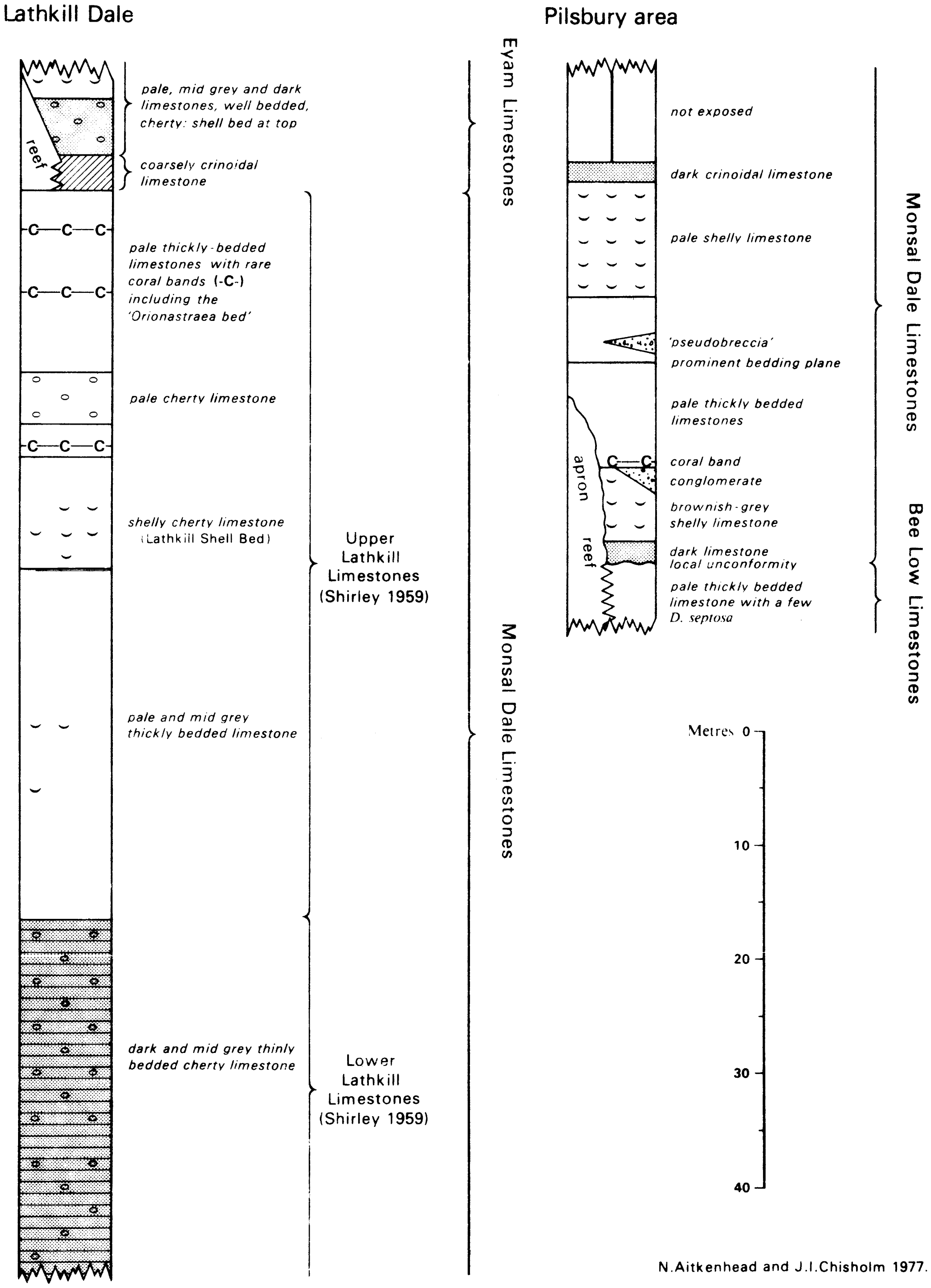
pp 75-77, Table 1.

Table I – Generalised sequences



N.Aitkenhead and J.I.Chisholm 1977.

There seemed to be general agreement that the conglomerate may have resulted from the erosion of the D_1 apron-reef during exposure to wave action in early D_2 times (see also Ludford, Madgett and Sadler 1973). A coral band immediately overlying the conglomerate has yielded *Lonsdaleia* ?, and a specimen of *Palaeosmilia regia* has been collected from a continuation of the band 110 m to the south-east. These identifications, made by M. Mitchell, indicate the D_2 age of this sequence. The overlying beds in the gully were seen to consist mainly of pale fine-grained limestones except for one bed of pseudobreccia 0.3 to 1.4 m thick. Various possible origins for this bed were discussed including palaeokarstic and bioturbation processes.

Leaving the limestone, the party traversed the valley of the River Dove cut in Namurian shales of mainly E_2 to R_1 age, and ascended, in a last burst of exercise, the escarpment below Sheen Hill formed by the Sheen Sandstones. The lowest leaf of these sandstones is closely underlain by the *Reticuloceras bilingue* (early form) Marine Band and when crossing the outcrop of this band one of the leaders suddenly plunged into a bramble-infested gully to emerge some minutes later clutching a few specimens of the eponymous goniatite as confirmation of the mapped line at this point SK 1128 6302, (Chisholm, Aitkenhead and Price 1977).

The coach was waiting at the top of the slope for the party's return to Nottingham.

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BOOK REVIEWS.

RAUP, D.M. and STANLEY, S.M. *Principles of Palaeontology* Freeman & Co., Reading, England. 2nd Edition, July 1978. SBN 0-7167-0022-0. £12.20.

The first edition of this book was well received in 1971 the main criticism at that time being centred on the need for a basic course in fossil morphology before appreciating some of the auguments presented in the book. This still applies and is particularly relevant in the chapter on Identification of Fossils and that on Descriptions, the latter commencing on p. 27 and not on p. 25 as listed in the Contents page. Given such a background, the book is an excellent précis of palaeontological work on preservation and the fossil record, life-history as illustrated by fossils, population studies, and taxonomic principles all included in Part I of the book. Part II is concerned with the use of palaeontological data in stratigraphy, palaeoecology, evolution and palaeogeography.

Readers may wish to know if it is worth renewing their first editions. A comparison of the two editions shows small changes in Part I. There appears to be some change in headings, some lost, some gained but the overall content is much the same. There is the inclusion of a section on multivariant analysis in chapter 4, Populations as a Unit. More drastic changes are to be found in Part II, particularly the final Chapter 12, Biogeography. In the second edition, the chapter is expanded considerably with a full treatment of climate, dispersal of organisms, plate tectonics and mass extinction. I think that the re-arrangement of the chapters in Part II improves the sequence of material in the text. The improvement in Part II should help in the decision to renew one's old copy.

The revision is completed by an updated bibliography and now there are two indexes - author and subject. The edition is well produced and pleasing in appearance, as was the first edition. The book is recommended for Palaeontologists and Biologists at College or University with a reasonable background in morphology.

F. M. Taylor

WARWICK, G.T., and WHITE, D.E., *Geological Handbook for the Wren's Nest Nature Reserve* - R.J.O. Hamblin, Nature Conservancy Council. 1978. 16pp, 8pls. 20p.

Despite the prevalence of local youths armed with catapults and aggressive intentions, the Wren's Nest Hill continues to be a popular haunt of professional and amateur geologists and of geology students. The main attraction is, of course, the abundant, diverse and beautifully preserved fossils that have gained the Wenlock Limestone of Dudley an international reputation. Superb specimens, particularly of trilobites and crinoids, may be seen in many museums, including the Natural History Museum in Wollaton Hall. These were mainly collected in the last century, when the limestone was being actively quarried, but brachiopods, trilobites, corals, crinoids, bryozoa and other fossils are still to be seen in abundance at the Wren's Nest Reserve (where indiscriminate fossil collecting is now rightly discouraged). In addition, the reserve provides a useful field example of a simple geological structure and an excellent demonstration of the importance of the limestone in the economic development of the area.

The current handbook is the third that has been produced for the area, and repeats some of the content of the previous publications. Visitors to the reserve, however, will still need to obtain a copy of the guide to the geological trail (published in 1975, price 15p), which includes a geological map, a feature which is sadly lacking from the new handbook. The major contribution of the new guide is to describe some new exposures that have been cut through the rocks at Wren's Nest by the Nature Conservancy. By digging a 40 m long trench on the eastern side of the hill and cleaning and extending a nearby quarry face, a complete sequence of strata from the uppermost Wenlock Shale, through the limestone, to the basal "Lower Ludlow Shales" has been exposed. A detailed measured section is included as a fold-out at the end of the guide, providing an invaluable basis for teaching and research. Photographs of the excavations are among the plates, which also include a view of the western side of Wren's Nest Hill in 1921, before housing estates replaced the open fields.

Copies of the trail guide and the handbook are obtainable from:

Geology and Physiography Section,
Nature Conservancy Council,
Foxhold House,
Thornford Road,
Crookham Common,
Newbury,
Berkshire. RG15 8EL.

R. J. Aldridge.

LETTER TO THE EDITOR

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Dear Editor,

Concerning the book review in Mercian Geologist vol. 6, no. 4., p. 311. I would like to defend the inclusion of a chapter on Pleistocene Geology within a book entitled *Limestones and caves of the Peak District* As the author of this particular chapter, I felt it was extremely important to document the sedimentology and climatology of this underestimated *geological* time period. All too often Pleistocene Geology is "relegated" to the geographical sphere, whereas in many cases it is the Pleistocene geological processes which are responsible for uncovering geological structures or for altering the hydrological system which could give rise to speleogenesis.

In N. Derbyshire the Pleistocene period was responsible for modifying the landscape and altering cave formation. Of course many of the caves are also clogged with Pleistocene sediments.

I would, however, like to apologize for omitting the acknowledgement of Fig. 25 to my earlier paper published in Mercian Geologist vol. 6, no. 2, with slight amendments.

Lastly I would like to say how easy it is to have little control over what is printed as no doubt Dr. Taylor you will agree. In paragraph 4, line 9 of your review you cite letter S on fig. 41 as being on p. 111, whereas it really appears on p. 168. Difficult isn't it.

Cynthia V. Burek, May 25th 1978.

Secretary's Report for 1976

The Society's thirteenth year commenced with a steady membership of over 500, and a record of enthusiastic support for meetings both indoor and in the field which was sustained and even exceeded during 1976.

Fourteen meetings were held during the year. The indoor meetings comprised four lectures from visiting speakers, a joint meeting with the Matlock Field Club, a Presidential Address, an Annual Dinner and an Annual General Meeting which was followed by a Collectors' Meeting. Four day field meetings were held, a weekend field meeting and, for the first time, a field meeting which lasted a whole week, an experiment which was crowned with success.

The Annual General Meeting in March opened in the presence of well over a hundred members. There was no Society business on this occasion other than the election of officers and Members of Council for 1976. It was proposed from the Chair that the serving Secretary, Treasurer and Editor should remain in office for a further year, but the President, Mr. H.R. Potter, had held office for three successive years, and was due to retire and become Vice-President. In his place, the Society was pleased to elect Dr. W.A. Cummins to hold the office of President for 1976.

At this time also, five Ordinary Members of Council were due to retire, and five new Members were duly elected to fill their places.

A vote of thanks was proposed to the retiring President and Ordinary Members.

At the close of the business meeting, the large assembly was free to visit the exhibitions and collections which spread over many benches in the great Swinnerton Laboratory.

A list of exhibits follows:

1.	A.E.G. Allsop	(a) Derbyshire spar. (b) E.M.G.S. excursion slides.
2.	P.G. Baker	Cotswold faunas of Jurassic age.
3.	F.J. Beasley	Sandstone with calcite and small vughs of quartz from Lower Keuper Sandstone, mid-Staffordshire.
4.	M. & D. Beaumont	Creatures great and small.
5.	E. Berry	Photographs of the Ingleton weekend field excursion.
6.	P. Binks	Crystals and rocks.
7.	M. Boneham	The geology of ocean floors.
8.	H.B. Briggs	Equipment display, boots, anoraks, rucksacks etc.
9.	B. Cantrill	Rocks from Anglesey and Llyn Peninsular.
10.	I. Clulow	Problematica and miscellaneous minerals.
11.	Derby Museum	18th Century geological inlaid marble tablets by White Watson.
12.	N. & J. Ellor	Fossils collected in Derbyshire and Norfolk.
13.	H. G. Fryer & J.H. Sykes	Specimens from Rhaetic Bone Beds and other sites.
14.	A. Gillespie	Zambia lead, zinc and copper ore.
15.	R.C. Gratton	Silurian and Carboniferous fossils.
16.	D. Hanford	Agates from Scarborough.

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17.	R.O. Harris	Rocks, fossils and minerals
18.	R.J. Hawkins	(a) Photographs of stone axes. (b) Skull of a "beaker man".
19.	I.M.E. Horne	Various examples of crystals.
20.	N. Leiter	Rock collection.
21.	R. Meinerzhagen	Fossil collection.
22.	J.C.G. Millar	File of geological localities.
23.	R.W. Morrell	Collection of Mollusca.
24.	R.W. Morrell & P. Spencer	(a) Pliocene mollusca from Walton-on-Naze, Essex. (b) Minerals from Folkstone.
25.	D.M. Morrow	Metamorphic rocks from Sinen Ghyll, Lake District.
26.	E. Ramsell	Slides, maps, etc. of Olduvai Gorge, Tanzania.
27.	P. Spencer	Out of print geological survey memoirs.
28.	F.M. Taylor	Brachiopod populations from three localities in the East Midlands.
29.	The Editor	"Mercian Geologist" series.

A most successful adjunct to the meeting was the customary sale of rock, fossil and mineral specimens, held by Mr. J.H. Sykes. His profit on this occasion reached the sum of £26, which he donated to the Society's Trust Fund, thereby bringing his total contributions to £125.10. The Society very greatly appreciates this splendid effort on the part of Mr. Sykes.

We were privileged in April to have for our last indoor meeting of the season so eminent an authority as Professor R.F. Peel of the University of Bristol to speak on the subject of Saharan landscapes, and to give to his enormous audience such a masterly account of the geography of the arid regions of the earth.

The weekend excursion, held in May, was led by Dr. P.G. Baker of Derby College of Technology, and centred in the Cotswold Hills. Attention was focussed on Bajocian deposits and the Middle Jurassic marine transgression, and the party followed the succession during the two days.

Early in June, Dr. F. M. Taylor of Nottingham University led a large party to the Castleton area, and members spent a rewarding day which began at Windy Knoll and continued by Mam Tor and Treak Cliff to the rich goniatite locality at Cow Low Nick. It is always appreciated when a leader can dispense with the coach for the entire day.

The July excursion took place at the height of a heat wave, when Dr. R.B. Elliott of Nottingham University led a party of 52, under tropical conditions, to examine igneous rocks in Charnwood Forest. Certainly, never before had members encountered such fiercely hot intrusions, baking in the torrid sun, radiating heat and scarcely to be touched by hand, an unexpectedly graphic illustration of the extremely interesting rocks of Charnwood.

For the first time in the Society's history, a full week excursion was planned, to take place late in July, and based in Edinburgh. We were very much indebted to Mr.M.A.E.Browne of the Institute of Geological Sciences in Edinburgh, a member of the Society, who most kindly arranged six daily excursions, and found leaders from amongst his colleagues of the I.G.S. and from the Grant Institute of Geology, University of Edinburgh. The Society party found excellent accommodation in the Pollock Halls of Residence, Edinburgh University, and enjoyed an unsurpassed succession of field meetings amongst the Bathgate Hills, the Pentland Hills, to the Arthur's Seat complex, to the Eastern Berwickshire coast, culminating in the classic locality at Siccar Point, to the oilshales of South Queensferry and the enormous opencast

Westfield Colliery, described to us as the largest coal-hole in Europe, and finally to the coast traverse in East Fife from Elie to Pittenween. The Society gives grateful thanks to Dr. E.N.K. Clarkson and Mr. P. Aspin of the Grant Institute of Geology, to Mr.S.K.Munro, Mr. D.C. Grieg, Mr. W. Tulloch, Dr. M. Armstrong, and Mr. I.H. Forsyth of the Institute of Geological Sciences in Edinburgh, and most of all to Mr. M.A.E. Browne who made the excursion possible.

In September, Mrs. M. Beaumont, B.A., led a coach excursion to the Huddersfield area, a day crowded with interest, for in between the many localities which were visited, a commentary was given through loud speakers as the coach travelled. After a most comprehensive survey of the geology of the area, Mrs. Beaumont added gilt to the gingerbread by bringing the entire party to her home for tea.

The final excursion of the season was led in October by Mr. P.H. Speed to examine the igneous rocks of the Wirksworth area. The outsize party of 63 members spent a most interesting autumn day visiting igneous outcrops at Wirksworth, Ible, Grange Mill and Bonsal.

The first indoor meeting of the winter season took place in November, when Dr. E.N.K. Clarkson, who had led a party to the Pentland Hills in July, came down from Edinburgh University to speak to the Society about trilobites. His lively description of the body structure and sense organs of these remote creatures, and his interpretation from trace fossils of their feeding habits and behaviour, together with his superb slides, brought them vividly to life.

In December, Mr. M. Pennington-George, an official engaged in the Selby project of the National Coal Board, gave a most informative lecture, in which he described the exploitation of the new Selby coalfield, and the new approach to the old problem of winning the coal as economically as possible.

The joint meeting held annually in January with the Matlock Field Club took place in Tawney House, Matlock. The speaker was Mr. J. Bennington, Inspector of Mines and Quarries, and his subject was the Mines and Quarries Inspectorate. His story began at the time of the first Factory Act, and he described the evolution of the Inspectorate from the early days of concern for employees to the present time, concluding with an account of present day practices.

Also in January, a meeting was held in Derby College of Technology. On this occasion, Dr. N.F.C. Hudson of the College's Department of Earth Sciences described certain areas of the Dalradian rocks of north-east Scotland, and compared and contrasted their chemical content and physical attributes.

The twelfth Annual Dinner was held early in February in the University of Nottingham Staff Club. As always, this was an informal and most enjoyable event.

The Presidential Address customarily takes place at the beginning of February to mark the anniversary of the founding of the Society in February 1964. For his first Presidential Address, the President, Dr. W.A. Cummins asked the rhetorical question, "Is plate tectonics the only answer?", and, addressing a very full house - some of whom half hoped that he might supply another answer - finally endorsed the plate theory as the most satisfactory at the present time.

1976 was an outstanding year of excellent meetings and splendid attendance both at indoor meetings and in the field. We are deeply grateful to our leaders, Dr. P.G. Baker, Dr. F.M. Taylor, Dr. R.B. Elliott, Mrs. M. Beaumont, and Mr. P.H. Speed, and to our speakers, Professor R.F. Peel, Dr. E.N.K. Clarkson, Mr. M. Pennington-George, Mr. J. Bennington, Dr. N.F.C. Hudson and our President.

Eleven monthly circulars were sent out during the year to give notice of meetings and other information of Society affairs. Postage is an expensive item and some of our members do the Society a great service by kindly delivering circulars by hand to addresses near to their own.

Membership of the Society increased by 22 during 1976, and at the end of the year the state of membership was as follows:

Honorary	Ordinary	Joint	Junior	Institutional	Total
2	269	126	23	108	528

With the 'Mercian Geologist', things went well in 1976. The Editor successfully published Vol. 6, no. 1 in October, and had Vol. 6, no. 2 ready for publication by the end of the year. The Editor is always ready to acknowledge the help he receives in the publication of the journal, but the Society is well aware of the moving spirit behind the scenes and the devoted and conscientious work carried out by the Editor himself.

The portable display unit designed by Mr. M.F. Stanley to publicise the Society completed a successful tour of Lincolnshire museums and W.E.A. centres during the year and was also on view at the annual conference of the Association of Teachers of Geology. Mr. Stanley has kindly assumed responsibility for the maintenance of the unit and we are indebted to Mr. D.N. Robinson for the supervision of its travels in Lincolnshire.

Once again the Society gratefully acknowledges the part played by Professor Lord Energlyn and the University of Nottingham in affording such a splendid venue for its meetings and the facilities of the Department of Geology for its general use.

D. M. Morrow

THE MERCIAN GEOLOGIST

Journal of the East Midlands Geological Society

The journal first appeared in December 1964 and since that time 24 parts, comprising 6 volumes have been issued; the last, vol. 6, no. 4, in April 1978. The Mercian Geologist published articles especially on the geology of the Midlands of England, but other articles have been published which relate to Midlands geology or are of current interest to geology generally. Contents include original papers, review articles, biography, bibliographies, excursion reports, book reviews and the Secretary's report on Society activities.

For Contributors:

Authors intending to submit manuscripts of papers for publication in the Mercian Geologist are asked to follow the format of papers included in a recent number of the journal, and if possible to provide two copies. As the journal is read by Members with a wide spectrum of geological interest and ability, authors are asked to ensure adequate introductions for their papers, particularly, if the subject has not been reviewed in the journal over the last few years. The paper should be complete in itself, without the need of the reader to refer to specialist journals not easily available to the average Member of this Society. It follows that the length of the paper may be greater than that published by some other journals but authors are asked to be as lucid and concise as possible and to avoid repetition.

Text-figs. normally occupy a full page of the journal, but part diagrams can be fitted into the typed page. Double page diagrams have been published with a single fold but each printed page has to be folded by hand. The standard reduction by our present printing process is approximately x 0.75. Thus the optimum size for the original diagram, including space for caption, index and explanation if required on the diagram, should be 285 x 190 mm. (285 x 380 mm with a single fold). Greater reduction is possible but care must be taken with the original to ensure that at the final reduced size (230 x 155 mm; or 230 x 310 mm) the smallest letters are no smaller than 1 mm and that there is a similar minimum spacing between letters and lines. Bar scales (metric) should be provided as the exact reduction cannot be guaranteed.

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If there are any points of difficulty, please do not hesitate to contact the editor during the production of the manuscript. The Editor's sole concern is to produce excellent quality papers to be enjoyed by all readers. Please send completed manuscripts to the editor.

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Ordinary Members	£4.00	Joint Members	£4.25
Institutional Members	£4.00		

Single Copies £2.25 (£2.00 Institutional Members; Other Members £1.50)
Complete volumes, 4 parts: £9.00 (£8.00 Institutions; Other Members £6.00).

Librarians in overseas libraries and geological institutions may take part in an exchange scheme organised by the Science Library of the University of Nottingham. About 200 institutions throughout the world receive the Mercian Geologist by sending in exchange, original geological periodicals.

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