THE LONGMYNDIAN SUPERGROUP : FACIES, STRATIGRAPHY AND STRUCTURE

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by

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ABSTRACT

THE LONGMYNDIAN SUPERGROUP: FACIES, STRATIGRAPHY AND STRUCTURE
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Aims of the investigation

The main aims of the investigation were as follows:

1) To determine the relationships between the Longmyndian and the Uriconian.
2) To describe and interpret the facies of the entire Longmyndian Supergroup.
3) To establish the relationship between the Stretton Group and the Wentnor Group.
4) To determine the provenance of the sediments.
5) To describe and interpret the small and large scale tectonic structures in the Longmyndian and to determine their relationships to regional structures, such as the Church Stretton fault system.

Results of the investigation

1) The Helmeth Grits were previously placed at the base of the Longmyndian Supergroup and were thought to unconformably overlie the Ragleth Tuffs of the Uriconian. This study groups the Ragleth Tuffs and Helmeth Grits together as the Ragleth Tuff Formation and this is included with the Longmyndian Supergroup. This formation has faulted contacts with the Uriconian Volcanic Complex and the Stretton Group of the Longmyndian. Conglomerates, which were previously included with the Uriconian are correlated with the Wentnor Group and they probably unconformably overlie the Uriconian. A new group of strata, the Linley beds, are recognised in the western part of the outcrop. The Longmyndian is interpreted to be faulted against the Western Uriconian, rather than to unconformably overlie it.

2) The Longmyndian Supergroup comprises basinal shales (the Stretton Shale Formation), turbidites (the lower parts of the Burway Formation), subaqueous delta deposits (at the top of the Burway Formation), distributary channel sandstones (the Cardingmill Grit), alluvial floodplain deposits with minor channel sandstones (the Synalds, Lightspout, Portway and Bridges Formations) and coarse braidplain deposits (the Bayston-Oakswood Formation and the Huckster Conglomerate). The alluvial floodplain deposits are comprised of an ENE to NE flowing river system and a W to WNW flowing sheetflood and sheetflow system which is probably the distal equivalent of a braidplain. All of the facies are extensively described and discussed.

3) The Wentnor Group is interpreted to conformably overlie the Stretton Group rather than to unconformably overlie it as previously thought.

4) The source for the sediments was mainly a magmatic arc of Uriconian type. Minor metamorphic detritus was probably derived from a sheared plutonic complex of Malvernian type.

5) The existence of a major syncline in the Longmyndian is confirmed and the parameters of this and other numerous, previously unrecognised folds are defined. Fault patterns indicate that a component of sinistral strike-slip was important during deformation and it is probable that both the Pontesford-Linley and Church Stretton fault systems are major strike-slip faults. Two maps of the Longmyndian are provided; one on a 1:10000 scale and one on a 1:25000 scale.

6) Small bedding plane markings of several types are described and their origins discussed. Some are possibly biogenic.

7) The plate-tectonic setting of the Longmyndian is discussed and a model of a forearc basin, which has been involved in a late Precambrian strike-slip orogeny, is proposed.

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PART 1

CHAPTER 1

INTRODUCTION AND HISTORY OF RESEARCH

1.1 INTRODUCTION

1.1.1 General geology

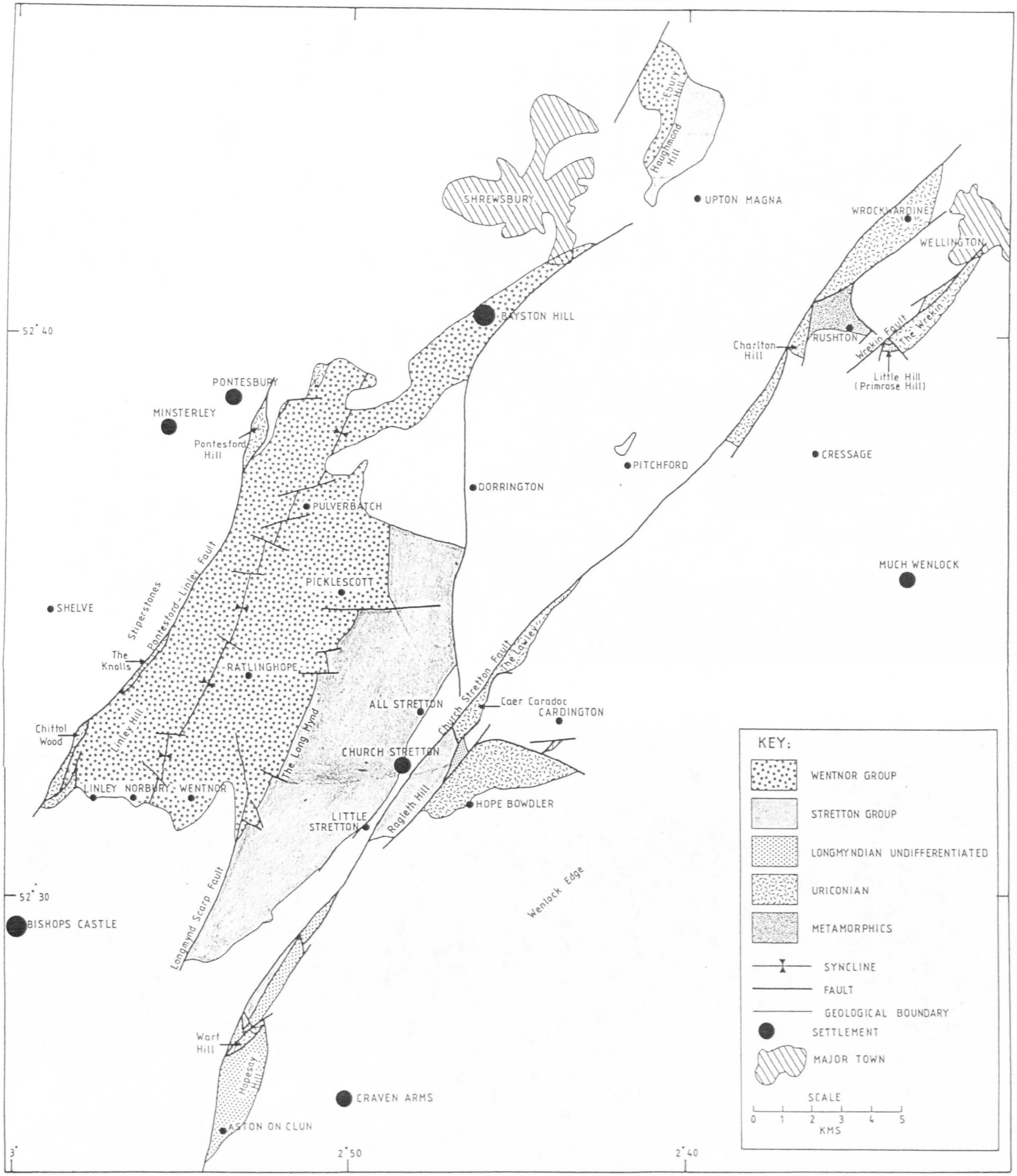
The Precambrian Longmyndian Supergroup is composed of approximately 6487m of volcaniclastics. It is folded into a large isoclinal fold whose axial trace trends NNE-SSW. The Longmyndian is bounded on both the west and east sides by major NNE-SSW trending faults: the Pontesford-Linley Fault in the west and the Church Stretton Fault in the east. However, some Longmyndian occurs within the fault zones and to the east of the main Church Stretton Fault (fig. 1). The Longmyndian is divisible into the lower Stretton Group, approximately 3464m thick and an upper Wentnor Group, approximately 3023m thick (fig. 1).

The second major group of Precambrian strata is the Uriconian Volcanic Complex which is composed of a calc-alkaline suite of rhyolites, andesites, dacites and minor granophyre, basalt, pyroclastics and intrusive dolerite. It is divisible by outcrop into the Western Uriconian, which outcrops between the Pontesford-Linley Fault and the Longmyndian and the Eastern Uriconian, which outcrops to the east of the main Church Stretton Fault (fig. 1).

Small, but important outcrops of Precambrian metamorphic rocks occur at Rushton: the Rushton Schists and directly south of the Wrekin, at Primrose or Little Hill: the Primrose Hill Gneiss and Schist (fig. 1).

The Precambrian is overlain unconformably by Palaeozoic strata of Cambrian to Carboniferous age or otherwise has faulted contacts. The Longmyndian is overlain unconformably in the south-west and south-east by the Llandovery and in the north-east by the

FIG. 1 DISTRIBUTION OF PRECAMBRIAN ROCKS IN SHROPSHIRE



Westphalian. The Eastern Uriconian is overlain unconformably by the Lower Cambrian, Wrekin Quartzite and in other places by the Harnage Shales of the Caradocian.

To the north-west, the Pontesford-Linley Fault separates the Longmyndian and Uriconian from the Ordovician of the Shelve inlier, which is complete from the Tremadoc up into the middle Caradoc Series. To the south-east, the Uriconian is mainly faulted against or overlain unconformably by the Caradoc Series, which is then unconformably overlain by the Silurian further to the east.

The main Precambrian outcrop extends from the Wrekin, near Wellington in the north-east, to Pontesford Hill, near Pontesbury in the north-west, and is then bounded by the Pontesford-Linley Fault as far as Linley in the south-west. The outcrop is limited by the Silurian to the south and extends to Hopesay Hill in the south-east. The main outcrop extends to the north-east to the Wrekin, largely bounded by the Church Stretton Fault to the east. Outcrops, largely of Uriconian occur within the Church Stretton fault system and to the east of the main Church Stretton Fault. A large inlier of Longmyndian outcrops north-east of Shrewsbury at Haughmond Hill (fig. 1). A small outcrop of the Precambrian occurs further to the north-east at Lilleshall.

1.1.2 Physiography

The Eastern Uriconian forms a discontinuous, NNE-SSW trending series of hills which are commonly lenticular in plan due to control by faults, which are abundant within the Church Stretton fault system. The high ground formed by the Eastern Uriconian of the Cardington and Hope Bowdler areas gradually slopes eastwards and south-eastwards beneath the unconformable Palaeozoic and has an abrupt northerly termination due to faulting. The Church Stretton Fault causes an abrupt scarp which separates the Eastern Uriconian hills from the low ground of the Church Stretton valley, which is occupied by Palaeozoic strata. The harder, nearly vertical strata of the Longmyndian rise relatively steeply from the Palaeozoic of the valley floor to form the western side of the Church Stretton valley, which then rises gradually to reach the plateau of the Long Mynd. The eastern side of the Long Mynd is dominated by numerous WNW-ESE trending deep valleys, which are largely fault controlled. The Long Mynd slopes northward to merge with the Shropshire plain and the Longmyndian is here covered with drift deposits of glacial origin. In the south, the Long Mynd is terminated by a steep scarp to the west, which follows the line of the Long Mynd Scarp Fault. This fault separates the hard, homogeneous sandstone of the Bayston-Oakswood Formation from the Silurian and the shalier Bridges Formation. In the north, the Long Mynd gradually slopes down toward the west to the River East Onny, reflecting the transition from the Bayston-Oakswood Formation to the Bridges Formation.

The low ground occupied by the River East Onny is underlain by the relatively soft Bridges Formation and further south by the Silurian. This rises westward to a low series of hills composed of sandstones of the Bayston-Oakswood Formation.

The Western Uriconian does not form such an imposing series of hills as the Eastern Uriconian. However, slightly elevated ground, occupied by the Western Uriconian, is recognisable in places; such as the Knolls and Chittol Wood. However, the Western Uriconian of Pontesford Hill near Pontesbury is of a similar elevation to the hills of the Eastern Uriconian. Its steep slopes and lenticular shape in plan are again controlled by faults.

The Pontesford-Linley Fault can be traced by a line of small valleys, but it does not form features of the scale of the Church Stretton Fault. The ground to the west of the Pontesford-Linley Fault rises quickly from the lower ground occupied by the Shineton Shales of Tremadoc age to the impressive ridge of the Stiperstones which is composed of quartzite of Arenig age.

Low hills, composed largely of Bayston-Oakswood Formation, rise from the Shropshire plain at Lyth Hill, Bayston Hill, Haughmond Hill and Ebury Hill. The position of these hills is largely controlled by faulting.

1.1.3 Aims and methods of study

The main aim of the study was to interpret and model the sedimentary facies of the Longmyndian. However, it became apparent during the course of the research that aspects of the stratigraphy and structure had to be investigated further in order that this aim could be satisfactorily accomplished. These aspects are as follows:

1. The relationship between the Longmyndian and the Uriconian.
2. The relationship between the Stretton Group and the Wentnor Group.
3. The structure of the Longmyndian, in order, primarily, that the palaeocurrent data could be properly analysed.

A study of the petrography was undertaken in order to determine the provenance of the sediments and to provide information on the relationships between the different formations. Some minor structures which have been attributed by some authors to organic activity and by others to inorganic origins were investigated in order to try and understand their origins.

The major part of the work involved the construction of sedimentological logs and the description of the physical characters of the sediments, in order that the environments of deposition could be established. Logs were constructed using the recommendations of Anderton (1985), so that sedimentary structures were symbolised as little as possible. Exposure in the Stretton Group is mainly provided by a number of parallel, WNW-ESE trending, deep valleys, which result in dip-sections, but elsewhere exposure is very poor. Since long and continuous sections are scarce, logs were constructed where the exposure was sufficiently continuous and where the representative facies could be related in vertical sequence. Most logs, because of these constraints, are only about 10m long. Exposure along strike is severely limited and consequently the analysis relies heavily on vertical sequences. Similarly, because of the steeply dipping nature of the beds, facies variations perpendicular to strike were unobservable, except for correlation of the Wentnor Group across the syncline axis. In the Stretton Group, numerous, small, isoclinal folds and small faults render it difficult to correlate individual horizons from one section to another.

Exposure within the Bayston-Oakswood Formation and Bridges Formation is very limited and vertical sections are commonly less than a few metres. Therefore, most of the area was investigated and numerous scattered exposures examined in order to build a catalogue of characteristic features. It was possible to construct only a few logs in these formations. Again, continuity along strike is very poor. Some information was gathered from large quarries in the Bayston-Oakswood Formation at Bayston Hill and at Haughmond Hill. Here, although faces are abundant, they mostly consist of joint and fault planes and the large scale architecture of the sandstones and the sedimentary structures are mostly obscure as a consequence. Much information was gathered from large blocks blasted from the quarry face and laid out on the quarry floor.

Because of the paucity of clear sedimentary structures observable in the field, numerous large blocks and smaller hand specimens were cut and polished and these provided much information.

In order to elucidate the structure, abundant bedding, cleavage, lineation and younging-directions were collected throughout the area. It was not intended to produce a map of the area, however, data are presented in Geological Maps 1 and 2 (in cover pocket), which of necessity draw heavily on previous maps for boundary positions. The detail on the maps was largely dictated by the detail of the sedimentological research.

1.2 History of research

1.2.1 Stratigraphy and structure

In the earliest accounts of the region, the Longmyndian was placed in the Cambrian and the Uriconian was considered to consist of intrusive "traps" of post-Silurian age. The succession in the Longmyndian was described as being from east to west, with a conformable transition into the Shineton Shales in the west (Murchison, 1839 and Ramsay, 1853). The Church Stretton fault system was considered to have acted as a conduit for the "traps" (Ramsay, 1853). These concepts were subscribed to by Salter (1856 and 1857), who divided the Longmyndian into nine numbered divisions, and by Murchison in 1867.

The true nature of the Uriconian was first demonstrated by Allport (1877) who concluded, from a petrographic study, that the Uriconian consisted of a regularly stratified series of extrusive rocks together with "agglomerates and ashes". He also recognised the unconformable relationship of the overlying "Caradoc strata" or Cambrian. He briefly described the Western Uriconian, which he considered to be comparable to the Eastern Uriconian. Almost simultaneously, Callaway (1877) declared the Uriconian of the Wrekin to consist of non-intrusive, bedded rocks.

Dr. C. Callaway pioneered much of the succeeding research. His papers were complemented by a number of articles on the petrography by Professor T. G. Bonney. Callaway firstly evolved a stratigraphy for the Cambrian of the Caradoc area from faunal evidence (Callaway, 1877). He then recognised that the Wrekin Quartzite is unconformable to the Eastern Uriconian (Callaway, 1878). The conclusion that the Eastern Uriconian is of Precambrian age was stated in 1879. In this paper, with the aid of Professor Bonney, the petrography of the Uriconian was thoroughly described, fragments of Uriconian were recognised in the Longmyndian (which was retained in the Lower Cambrian) and metamorphic rocks at Primrose Hill were described and assigned to the Precambrian "Malvernian System" and separated from the Eastern Uriconian by an inferred considerable unconformity. An argument, which later became contentious, involved the inclusion of a conglomerate at Charlton Hill and a feldspathic sandstone at Ragleth Hill with the Uriconian. Included granitic fragments were maintained to have been derived from the Ercall granophyre and similar granitic rocks from Primrose Hill, which were then necessarily older than the Uriconian.

In 1882, Callaway and Bonney assigned the Western Uriconian to the Precambrian because of its close similarities to the Eastern Uriconian and because of included fragments of Uriconian-type in the Longmyndian, which was still thought to be Lower Cambrian in age.

Callaway viewed the Uriconian as having been upthrust between faults and states: "The formation of the fault and the upthrust of the Archaens would therefore seem to be connected with the forces which, at the close of the Ordovician epoch threw the Ordovician and Cambrian deposits into folds." (Callaway, 1882, p. 123).

The term "Uriconian" was first used by Callaway in 1886. In the two papers of 1886 and 1887, Callaway developed conclusions based on the included fragments within the Longmyndian conglomerates and within the conglomerates which he included with the Uriconian. He identified a metamorphic terrain composed of gneiss, granite, quartzites, quartz schists, mica schists and "hypometamorphic grits", which supplied detritus to both the Uriconian and the Longmyndian and he reiterated that the Uriconian sourced the Longmyndian in part.

The term "Longmyndian" was first used by Callaway in 1887, who then considered that it might be Precambrian. On the discovery of a Lower Cambrian fauna in the Wrekin Quartzite and Comley Sandstone by Professor Lapworth in 1888, a Precambrian age for the Uriconian was confirmed and a Precambrian age for the Longmyndian was considered to be probable. The main reason for considering the Longmyndian to be Precambrian appears to have been that the length of time required for its deposition would not enable it to be placed stratigraphically beneath the Wrekin Quartzite and still be retained in the Cambrian.

In 1890, Professor J.F. Blake proposed radical departures from the accepted views. He proposed that an unconformity existed beneath the Wentnor Group and the Huckster Conglomerate (which he thought was an outlier of the Wentnor Group). He referred the Stretton Group to his Precambrian upper "Monian" and the Wentnor Group to the base of the Cambrian. Because there appeared to be no evidence for a Uriconian source for the Stretton Group, he therefore considered that the Eastern Uriconian volcanics "..... are younger than the slates and have been extruded from their midst." (Blake, 1890, p. 407). However, the Eastern Uriconian was considered to antedate the deposition of the Wentnor Group, as it had supplied detritus to it. The Western Uriconian was considered to be intrusive into the Wentnor Group, but was still thought to be Lower Cambrian in age. Supposed metamorphism at the margins of the Uriconian was cited as evidence for intrusion in both cases. The conglomerates and grits which outcrop east of the Church Stretton Fault and which were considered by Callaway to be part of the Uriconian were correlated with the Wentnor Group, assigned a basal Cambrian age and were considered to rest unconformably on the Uriconian and to antedate the Wrekin Quartzite. Blake agreed with Callaway that the Rushton Schists and Primrose Hill Gneiss and Schist were older than the other rock groups, but concluded that the Ercall granophyre intruded the Uriconian.

The conclusions of Blake (1890) were immediately criticised by Dr. Hicks in discussion of the paper, who pointed out the presence of Uriconian fragments in the Stretton Group and affirmed that the entire Longmyndian must therefore be younger than the Uriconian.

Callaway (1891) maintained his previous arguments and negated many of Blake's conclusions. He concluded that the Uriconian is not intrusive and from the evidence of included fragments he thought it to be older than the Longmyndian. The proposed unconformable outlier of the Wentnor Group, the Huckster Conglomerate (Blake, 1890) was shown to be a conformable part of the Stretton Group. A fault was favoured between the Longmyndian and the Uriconian, but an unconformable relationship was maintained to exist because of the discordances of strike between the Longmyndian and the Uriconian. He provided new evidence to demonstrate that the conglomerates of Charlton Hill are interbedded with the Uriconian and therefore part of it and not therefore of basal Cambrian age (Blake, 1890). Since fragments of Ercall granophyre and metamorphic rocks were found in the Uriconian conglomerates, these former were stated to be pre-Uriconian. Professor Blake's opinion that the granophyre intruded the Uriconian was then necessarily refuted and the junction was considered to be faulted. Callaway noted the difficulties in distinguishing between "a sedimentary deposit and a true lava-flow" in this area. This is obviously due to the volcaniclastic nature of the sediment and he concludes: ".... sedimentary material forms a larger part of the Uriconian rocks of the Church Stretton area than we had supposed." (Callaway, 1891, p. 120).

Lapworth and Watts (1894) introduced the terms "Western or Red Longmyndian Series" for the Wentnor Group and "Eastern or Grey Longmyndian Series" for the Stretton Group. Research during the early 1900's furnished a plethora of different interpretations of the stratigraphy which suffered from the lack of way-up criteria. In 1900, E.S. Cobbold presented many new ideas which were apparently influenced by discussions with Lapworth and Watts. Cobbold accepted a Precambrian age for the entire Longmyndian and that the Uriconian is older than the entire Longmyndian. He proposed that the Longmyndian passed down conformably into the Uriconian through the horizon of the Helmeth Grits and stated: ".... there are some indications that the volcanic rocks of Ragleth etc. are conformable with the base of the Lower Longmyndian and consequently may really be part of that great series." (Cobbold, 1900, p. 87). Cobbold placed an unconformity at the base of the "Ratlinghope Conglomerate" (now the Stanbatch Conglomerate) and suggested that the base of the

Huckster Conglomerate may be the base of a second unconformity. Cobbold also suggested a syncline within the Wentnor Group, the core of which was occupied by shaley beds and the limbs of which included the same "Ratlinghope Conglomerate". He agreed with Blake (1890) in assigning conglomerates and sandstones east of the Church Stretton Fault to the Wentnor Group. He recognised the pyroclastic nature of the Batch volcanics and concluded that the Longmyndian was composed of "..... fine volcanic ash deposited in water" and that ".... volcanic material was being ejected at no great distance during the deposition of the series." (Cobbold, 1900, p. 84). Cobbold introduced the terms: Upper Longmyndian, Lower Longmyndian, Huckster Conglomerate, Batch Volcanics, Carding Mill grits, Buxton Rock, Helmeth Grit, Stretton Shales and Habberley Line (for the Pontesford-Linley Fault), although he attributes some of these terms to Lapworth and Watts.

The true nature of the Western Uriconian was demonstrated by Boulton in 1904, who showed that Pontesford Hill is composed of a bedded series of rhyolites and pyroclastics. This disproved Blake's theory (1890) that the Western Uriconian was intrusive. In the discussion of Boulton's paper, Professor Lapworth reversed the stratigraphic positions of the Uriconian and Longmyndian. It appeared to him that the Longmyndian represented an increase in volcanic activity, which then resulted in the formation of the Uriconian volcanics which ".... rested transgressively upon the Lower Longmyndian and began with the so-called Helmeth Grits." (Lapworth, _in_ Boulton, 1904, p. 485). He proposed a threefold division of the Longmyndian, comprising a "Lower Division" (equated with the Charnian), an intermediate Bayston Group and an "Upper Division" (equated with the Torridonian).

Lapworth and Watts (1910) introduced the terms "Wentnor Series" and "Stretton Series" and they proposed that no unconformity existed between them. The five divisions of the Stretton Group proposed by Blake (1890) were retained and named in ascending order as the Stretton Shale "Group", Burway "Group", Synalds "Group", Lightspout "Group" and Portway "Group". These divisions have been retained essentially unchanged in the subsequent literature. The Wentnor Group was divided into a lower Bayston "Group" and an upper Ratlinghope "Group" and it appears that the syncline proposed by Cobbold in 1900 was not accepted. In agreement with Lapworth (in Boulton, 1904), the Western Uriconian was maintained to overlie the Wentnor Group and was called the "Linley (or Pontesford) Volcanic Series". However, the Eastern Uriconian was removed from its position above the Longmyndian (Lapworth, in Boulton, 1904) and put at the base and called the "Cardington Volcanic Series", so that now, the Eastern and Western Uriconian were separated in time by the Longmyndian.

Watts (1925) favoured the stratigraphical equivalence of the Western and Eastern Uriconian on lithological grounds. Watts quotes Cobbold as suggesting that the Eastern Longmyndian was inverted with the Helmeth Grits at the top. The Longmyndian might then have been folded into an anticline. This suggestion appears to have been formulated in order to retain the stratigraphical equivalence of the Western and Eastern Uriconian and to retain their position above the Longmyndian as suggested by Lapworth (in Boulton, 1904). The Ercall granophyre now appears to have been accepted as being intrusive into rhyolitic tuffs of the Uriconian rather than possessing a faulted contact, as proposed by Callaway (1891).

In 1925, Cobbold accepted that an unconformity may exist between the Stretton Group and the Wentnor Group, but noted that in the original proposal for an unconformity by Blake (1890), Blake had "not realised the effects of the great longitudinal and transverse faults that cut these Archaean rocks." (Cobbold, 1925, p. 364). From an excavation in the Cwms area, the Wrekin quartzite was found to be faulted against sandstones which were equated with the Wentnor Group. However, because of the large difference in strike, Cobbold concluded that the excavation ".... proved conclusively that the Cambrian rocks are unconformable to the Torridonian (?)." (Cobbold, 1927, p. 557).

In 1929, T.H. Whitehead of the Geological Survey added complexity to the interpretation of the Western Uriconian by recognising a rhyolite near Pontesford Hill, similar to the Western Uriconian, but of supposed Ordovician age, which rested unconformably on the Wentnor Group and was itself overlain unconformably by the Ordovician Harnage Shales. The Uriconian of Lyd Hole was shown to consist of bedded volcanics and therefore not to intrude the Wentnor Group.

In 1935, Cobbold and Whittard produced a key paper on the Eastern Uriconian/Longmyndian contact. They stated: ".... any discontinuity there may be between the two groups, if present at all, is of no great stratigraphical importance." (Cobbold and Whittard, 1935, p. 351). Fragments of Uriconian rocks were found in the Helmeth Grits, which appeared to pass into the Stretton Shale Formation by interbedding. The succession was therefore considered to be: Eastern Uriconian; Helmeth Grit; Stretton Shale Formation; in ascending order. The Helmeth Grit was described as consisting of impure lithic tuffs interbedded with shales. The former were taken to represent the dying stages of Uriconian volcanicity. The underlying Uriconian was described as consisting of tuffs interleaved with shales and it was noted that ".... the petrological identity of such tuffs in both groups of rocks made mapping extremely difficult." (Cobbold and Whittard, 1935, p. 354). The Stretton Group was considered to young in a westerly direction from the bedding/cleavage relationships and the direction of pitting of Arenicolites. The conclusion reached was that the Wentnor Group is younger than the Stretton Group. An unconformity was accepted between the two and additional arguments were put forward to support this. These were, the greater geographical distribution of the Wentnor Group and the absence of Cambrian and Ordovician on the Stretton Group; the implications being that the Wentnor Group was overstepping and that a mantle of unconformable Wentnor Group overlay the Stretton Group during the Cambrian and Ordovician. Cobbold and Whittard (1935) also proposed that the Wentnor Group was younger than the Western Uriconian, as the former contained fragments of the latter. In discussion, Whitehead subscribed to this view and proposed that a large stratigraphic break, as yet unrecognised, existed in the Longmyndian to explain the easterly-younging succession from the Western Uriconian to the Wentnor Group and the westerly-younging succession from the Eastern Uriconian to the Stretton Group. Mr. T.S. Westoll, in discussion of this paper, had come to the conclusion of a westerly-younging succession in the Stretton Group from the evidence of truncated cross-bedding.

The Geological Survey published a memoir of the Shrewsbury district in 1938 (Pocock et al., 1938) which accompanied a map of the area. The stratigraphy of the Stretton Group remained identical to that proposed by Lapworth and Watts (1910). However, the Wentnor

Group was subdivided into three divisions similar to those of Blake (1890). These divisions were called the Bayston "Group", Bridges "Group" and Oakswood "Group"; from east to west.

It was proposed that the entire "Western Longmyndian" younged in an easterly direction. This was based on: easterly-fining sequences in the associated conglomerates, the easterly transition from the Oakswood "Group" to the Bridges "Group", the contained fragments of Western Uriconian in the "Western Longmyndian" and the distribution of dips in the Bridges "Group", from which drag folding was inferred and which indicated an easterly-younging succession throughout the Bridges "Group". Evidence from Haughmond Hill was interpreted as showing an easterly-younging succession in the Wentnor Group and green beds on the eastern side of the hill, which were correlated with the Burway Formation, then appeared to follow stratigraphically. The memoir appears to have been written prior to the 1935 paper of Cobbold and Whittard. Consequently, an easterly-younging succession from Western Uriconian to Longmyndian to Eastern Uriconian was suggested although, in the stratigraphic tables, the entire Uriconian was placed beneath the Longmyndian. From petrographic considerations, the Primrose Hill Schist and Gneiss were thought to be due to the dynamic metamorphism of the Uriconian in conjunction with the intrusion of a granophyre. Similarly, the Rushton Schists were thought to have been derived from the "Eastern Longmyndian" by shearing and possible granophyre intrusion. The pre-Uriconian age of the metamorphic rocks was therefore thrown into doubt.

In 1948, Challinor presented way-up evidence from graded bedding for an easterly-younging succession in the Stretton Group of the Haughmond Hill inlier. This suggested to him that the survey's interpretation of an easterly-younging succession from Western

Uriconian to Longmyndian to Eastern Uriconian was correct. However Whitehead (1948) responded by confirming that, in the Long Mynd, a westerly-younging succession was apparent from the bedding/cleavage relationships, some sedimentary structures and from the upward transition from the Helmeth Grits into the Stretton Group. However, he confirmed that the Wentnor Group showed signs of inversion. A fault was proposed in the Haughmond Hill inlier to explain the juxtaposition of the probable Burway Formation with the Wentnor Group. Pocock and Whitehead (1948) proposed a fault between the Stretton Group and the Wentnor Group in the Long Mynd area, to explain the easterly-younging succession in the Wentnor Group and the westerly-younging succession in the Stretton Group. The possibilities of a syncline are not mentioned.

The beginning of the next decade was dominated by the researches of Whittard and James, the latter, under the direction of the former, produced a Ph.D. thesis on the structure and stratigraphy of the Longmyndian in 1952, together with a 2½" to the mile map of the central part of the Longmyndian outcrop. In 1952, James published an account of the Longmyndian/Western Uriconian contact. He included "green tuffs" and "green 'shaley' tuffs", previously thought to be Uriconian (Lapworth and Watts, 1910) with the sandstones and conglomerates of the Wentnor Group. An excavation near Chittol Wood proved that the Western Uriconian lay above the tuff beds which were shown to be inverted. Since no evidence for a fault was noted, an inverted unconformity was proposed.

Whittard (1952) largely pre-empted the conclusions of James's thesis. He proposed that the Wentnor Group rested unconformably on the Western Uriconian of the Linley area as proposed by James (1952a). In support of this unconformity, he proposed that the

"Ordovician" rhyolite near Pontesford Hill (Whitehead, 1929) was in fact Precambrian and probably dipped beneath the unconformable Wentnor Group. The Wentnor Group/Western Uriconian boundary in this area and in the area of Lyd Hole appeared to be stratigraphically transgressive and therefore an unconformity was proposed. From the opposed younging evidence and from the recognition of the lithological similarities between the Bayston and Oakswood "Groups", he proposed that, "Large scale and deep synclinal overfolding is required to explain the repetition of the Bayston Group in the form of the Oakswood Group." (Whittard, 1952a, p. 149). An unconformity within the Longmyndian was accepted, but he was uncertain as to its stratigraphic position. Mentioning that evidence was supplied by the mapping of James, he concluded that this unconformity ".... may occur at the base of the Huckster Conglomerate and the Portway Group would then be better classified with the Wentnor Series." (Whittard, 1952a, p. 145). Whittard first used the name "Pontesford-Linley line" for the Pontesford-Linley Fault and he proposed that this, together with F3 of the Church Stretton fault system, was a major tear fault. He also recognised other tear faults in both the Shelve inlier and the Longmyndian. The main period of folding and faulting in the Shropshire area was assigned to the Ordovician-Silurian boundary.

The thesis of James (1952b) reiterated the conclusions of Whittard (1952). Much evidence in support of the syncline was provided from way-up criteria. Two unconformities were proposed within the Longmyndian sequence. The first was placed beneath the Huckster Conglomerate where, although he recorded ".... failure to detect any marked angular discordance at the Huckster Conglomerate." (James, 1952b, p. 80), evidence was given from one section at

Deadman's Batch which was thought to show that the entire Lightspout Formation had been removed. Additional arguments in support of this unconformity were the slightly irregular base to the Huckster Conglomerate, the inclusion of fragments of shale, similar to the underlying shale, in the Huckster Conglomerate, the apparent absence of the Lightspout Formation in the Shrewsbury area and the apparent decrease in "tectonic grade" above the Huckster Conglomerate. A second unconformity was placed beneath the Bayston-Oakswood Formation, the main evidence for this being the thinning of the Portway Formation to the south. The absence of the Stretton Group in the western limb of the syncline was therefore attributed to the combined effects of the two unconformities, which resulted in the overstep of the Wentnor Group. James inferred that the Western Uriconian was the source for the volcanic fragments in the Longmyndian, from the inferred consistent tectonic subsidence in the east. James concluded that both the Uriconian and Longmyndian were folded together prior to the deposition of the Wrekin Quartzite and hence both the Longmyndian and the folding were of Precambrian age. James proposed the names "Strettonian" for the strata beneath the Portway Formation and "Wentnorian" for the Wentnor Group.

Whitehead (1955) accepted the syncline proposed by Whittard (1952) and James (1952b) and gave supportive evidence for way-up from sedimentary structures in the Shrewsbury area. The Wentnor Group of Haughmond Hill was now shown to young in a westerly direction, in contradiction to previous ideas (Pocock et al., 1938). The beds on the east side of the hill were attributed to the Burway Formation and because of the great thickness of this formation in this area, together with the easterly-younging evidence provided by Challinor (1948), folding was inferred to have occurred on a large scale. The apparent absence of the Lightspout, Synalds and Portway Formations at Haughmond Hill was now thought to be due to the unconformity at the base of the Wentnor Group rather than to the previously postulated fault (Whitehead, 1948).

James's thesis (1952b) was summarised in a paper published in 1956. Here, he called the Portway Formation the "Mintonian". The equivalent Bayston and Oakswood "Groups" were grouped into one and called the "Bayston-Oakswood Group". It was further argued that the Western Uriconian acted as the sole source for the Longmyndian, as the Eastern Uriconian was believed to have been covered by the conformable Stretton Shale Formation and it was noted that there was an apparent increase in the proportion of igneous material in the western part of the Bayston-Oakswood Formation, an increase in pebble size in the west and that there were included fragments in the Wentnor Group which were similar to those of the Western Uriconian of Lyd Hole and Pontesford Hill. In discussion, Whitehead thought that there could be a tectonic explanation for the thinning of the Portway Formation and he questioned the validity of including the "shaley tuffs" of the Chittol area with the Wentnor Group, rather than with the Uriconian. In discussion of James's paper, Professor Whittard presented new evidence for the unconformity beneath the Wentnor Group from the Brokenstones area. Here, the Wentnor Group was stated to rest on the Stretton Shale Formation and an unconformity was inferred. Thus, in this area, most of the Stretton Group appeared to have been removed beneath the unconformity.

Greig and Wright (1958) of the Geological Survey found no evidence for an unconformity beneath the Huckster Conglomerate. They correlated beds beneath the Huckster Conglomerate at Deadman's Batch with the Lightspout Formation rather than with the Portway Formation, as proposed by James (1952b and 1956) and the sequence here was therefore complete. They advised that the term "Mintonian" should be abandoned.

Dineley (1960) noted the similar orientations of transverse faults in the Long Mynd and Shelve areas and concluded that they might have been formed at the same time during the Taconian or Caledonian orogenies. The probability that the Church Stretton Fault and the Pontesford-Linley faults were tear faults was noted.

The unconformity between the Western Uriconian and the Wentnor Group at Lyd Hole, near Pontesford Hill (Whittard, 1952) was reinterpreted as a fault by Dean and Dineley (1961). In addition, the supposed Ordovician rhyolite of Whitehead (1929) south-east of Earl's Hill, which was subsequently reinterpreted by Whittard (1952) as Uriconian and used by him in arguments for an unconformity between the Western Uriconian and the Wentnor Group (Whittard, 1952), was now shown to consist of boulders and pebbles of rhyolite, which rested unconformably on the Precambrian. A Caradoc age was postulated for this conglomerate.

Dean (1964) described the Ordovician and adjacent strata in the area of Horderley and Brokenstones. It was concluded that sandstones and conglomerates, which were correlated with the Wentnor Group, rested unconformably on the Stretton Shale Formation in this area, as postulated by Whittard (<u>in</u> James, 1956). This unconformity lent support to the proposed unconformity beneath the Wentnor Group in the Long Mynd area. An unconformity was inferred from mapping in some cases, although the contacts were not exposed. The results of two excavations were detailed and both were interpreted as showing an unconformity (Dean, 1964, p. 270-271). Dean mapped and described the faults of the Church Stretton fault system in this area. Three main faults were recognised, as proposed by Cobbold (1927), from west to east: the Church Stretton Fault (F1), a normal fault of at least post-Silurian age; the Lawley Fault (F2), a low angle reverse fault and the Cwms-Hoar Edge Fault (F3), a tear fault with sinistral displacements. F2 and F3 were thought to be mainly Taconian, although F2 was considered to have probably been in existence as early as the Precambrian.

At the end of the decade, the Geological Survey produced a number of books and maps which refer to the area. A one inch to the mile map of the Church Stretton area (Sheet 166) was published in 1967 and this was accompanied by a memoir in 1968 (Greig et al., 1968). The syncline proposed by Whittard (1952) and James (1952b) was accepted with reservations, however the proposed unconformity beneath the Huckster Conglomerate was not, for reasons given by Greig and Wright (1958). The unconformity beneath the Bayston-Oakswood Formation, proposed by James (1952b and 1956), was accepted. However, the evidence for an unconformity between the Wentnor Group and Stretton Shale Formation in the Horderley area, proposed by Whittard (in James, 1956) and Dean (1964), was considered to be dubious and it was noted that ".... in every case, the junction is complicated by faulting." (Greig et al., 1968, p. 90). In the Haughmond Hill area, the Wentnor Group was interpreted to rest unconformably on beds which were tentatively correlated with the lower part of the Synalds Formation. An unconformity between the Longmyndian Helmeth Grit and the Uriconian of Caer Caradoc was proposed because, in this area, there was a marked divergence of strike between them and because the Helmeth Grit rested upon both the Ragleth Tuffs and Cwms Rhyolites of the Uriconian in a transgressive manner. Conglomerates which outcrop in the Hope Bowdler Hill area were interpreted as being interbedded with Uriconian volcanics, although the base of the conglomerates often appeared to be faulted or unconformable. From petrographic considerations, the Rushton Schists were assigned to a low grade regional metamorphic episode rather than to a local dynamic event associated with granophyre intrusion (Pocock et al., 1938). This metamorphism was considered to have occurred prior to the eruption of the Uriconian. The deposition of the Longmyndian was considered to be entirely post-orogenic and to have occurred in a subsiding trough, which was probably bounded by the Pontesford-Linley and Church Stretton fault systems and in which subsequent deformation was confined (Greig et al., 1968, p. 75). It was proposed that the Uriconian had suffered a pre-Longmyndian, "Charnoid" deformational episode, which had produced north-west trending fold axes and south-west dipping thrusts and shears that differed significantly from the north-east trending "Caledonoid" fold-trend in the Longmyndian. A pre-Caradocian age for the folding of the Longmyndian was necessitated by the interpretation of an unconformity between the Caradoc Series and the Wentnor Group in the area of Pontesford Hill (Dean and Dineley, 1961). The folding was concluded to have occurred in Precambrian times, as it was recognised that the Cambrian and lower Ordovician rocks of Shropshire did not have any indication of folding of the same magnitude as that in the Longmyndian. Minor folds were recognised in the Bridges Formation and it was proposed that parasitic folds existed at the base of the Burway Formation, in order to explain a zone of anomalous dips. Stratigraphical successions were formulated for parts of the Uriconian, but no overall sequence could be determined, due to the problem of correlation between the separate faulted areas. The terms Wentnor "Series" and Stretton "Series" were used by the survey.

The views of Greig et al. (1968) were reiterated by Wright (1968) and Hains (1969) of the Geological Survey in two booklets which accompanied 1:25000 maps of the area. The SO49 sheet was published in 1968 and the SO48 sheet was published in 1969.

Earp and Hains (1971) considered that the schist and gneiss of Primrose Hill was pre-Uriconian and equivalent to the Malvernian, rather than a local alteration of the Uriconian (Pocock et al., 1938). The "Charnoid" structural trend in the Eastern Uriconian was thought to be possibly due to volcanism and thus localised, although a widespread tectonic event was not dismissed. The Uriconian volcanism was stated to have occurred "along lines of crustal weakness trending broadly from north to south." (Earp and Hains, 1971, p. 6) and it appears that volcanism was thought to have been restricted to the pre-existing fault zones.

A geochemical analysis of the Uriconian was made by Thorpe (1972a) who concluded that the volcanics were a calc-alkaline, basalt-andesite-rhyolite association in which primary andesitic magma was evident. From the variability of K, Rb and Ba in the rhyolites, it was suggested that the lavas had suffered some alteration which had resulted in alkali migration. It was concluded that the Uriconian was similar to some island arcs (Thorpe, 1974). The Ercall granophyre was shown to be geochemically similar to the Uriconian and therefore probably part of it. However, granite and gneiss from Primrose Hill were shown to be significantly different from the Uriconian and a pre-Uriconian age was therefore considered probable (Thorpe, 1974). The latter conclusion contradicts that of Pocock et al. (1938) who thought that the Primrose Hill metamorphics consisted of altered Uriconian, but agrees with the conclusions of Earp and Hains (1971).

Dunning (1975) introduced the terms Wentnor Group and Stretton Group into the literature and substituted Formation for "Group" in the lower order divisions such as the Synalds "Group". Dunning briefly states that the overturned unconformity at Chittol (James 1952a), between the Wentnor Group and the Western Uriconian, could be interpreted as a fault.

A revised edition of the geological map 152 of the Shrewsbury area was published on a 1:50,000 scale in 1978. The stratigraphic hierarchy introduced by Dunning (1975) was used. An unconformable junction between the Wentnor Group and the Western Uriconian of the Lyd Hole near Pontesford Hill was retained, despite being previously interpreted as a fault by Dean and Dineley (1961). The Primrose Hill metamorphics were included with the Uriconian, in apparent agreement with Pocock et al. (1938). Davies (1978) reported on a previously undescribed outcrop of Western Uriconian near Bishops Castle. Here, the Ludlow Series is juxtaposed with some Western Uriconian tuffs along a probable tear fault, which is part of the Pontesford-Linley fault system. The structure within the Precambrian was found to consist of shallowly-dipping thrust faults. The Western Uriconian tuffs overlay "brecciated ashy turbidites" which were correlated with the "green shaly tuffs" of the Wentnor Group exposed at Chittol (James, 1952a).

A revised aeromagnetic map of the Church Stretton area was provided by Wilson (1980) who noted that the survey of Henson (1957) had been wrongly matched to the geology with a displacement of 1.2 km. Henson's interpretation, that the Uriconian consisted of SE dipping sheets, was thus thrown into doubt. Extensions of the Uriconian, beneath the Palaeozoic rocks, east and south-east of Hope Bowdler Hill and to the east of Stoneacton were noted, confirming

Henson's observations (1957). In addition, N-S trending anomalies were noted throughout the area. Many possessed no surface expressions and were attributed to concealed lines of weakness in the basement, which had not been reactivated.

Toghill and Chell (1984) produced a useful summary of the geology of Shropshire. The term Longmyndian Supergroup was introduced. This paper largely draws from the conclusions of previous authors. The Rushton Schists and Primrose Hill Gneiss and Schist were thought to be older than the Uriconian volcanics and to have formed in response to SE dipping subduction located in the Anglesey area. The Uriconian volcanics were postulated to be a late-orogenic igneous episode, probably extruded along the Pontesford-Linley and Church Stretton fault systems, which were initiated at this time. Deposition of the Longmyndian then followed in the fault-bounded basin. Following the eastward tilting and erosion of the Stretton Group, the Wentnor Group was deposited unconformably on the Stretton Group and overstepped onto the Western and Eastern Uriconian. The folding of the Longmyndian was maintained to have occurred prior to the deposition of the Lower Cambrian, Wrekin Quartzite.

Thorpe et al. (1984) concluded, from the geochemical data, that the Uriconian volcanics were of calc-alkaline or transitional alkaline character. Some Uriconian basalts possessed ambiguous characteristics suggestive of ocean floor or island arc tholeiites or active continental margins. Overall, the Uriconian basalts and minor intrusions appeared to be transitional between arc and within-plate lavas. The Uriconian Volcanic Complex was shown to be part of a late Precambrian to Palaeozoic magmatic and metamorphic episode which resulted from SE directed subduction. The basement of England and

Wales was thought to be the result of crustal growth by accretion of island arcs, associated accretionary prisms and forearc sediments within the Iapetus Ocean. Palaeomagnetic evidence suggested that the pre-Ordovician basement of England and Wales was a distinct microplate within Iapetus, emplaced in its present configuration as a result of subduction between Cambrian and Silurian times.

Woodcock (1984a) produced evidence which showed that strike-slip faulting might have been active in Wales, at least during the 50 Ma period from the Caradoc to the Silurian. Specifically, the Pontesford-Linley fault system, here referred to as the "Pontesford Lineament" with respect to its wider geographical distribution, was thought to show post-Caradoc and pre-late Llandovery strike-slip, probably dextral, on major NNE faults and their ENE splays and a probable post-Llandeilo dextral shift of 40 km. Additionally, the Church Stretton fault system was thought to show evidence for strike-slip movement; in its braided pattern, inclusions of basement slivers and more specifically, in the presence of strike-slip slickensides in the Old Radnor area on post-Wenlock faults and in the sinistral offset of the Carboniferous and possibly the Triassic north of Church Stretton.

Further evidence for strike-slip along the Pontesford Lineament was provided by Woodcock (1984b). The Pontesford-Linley Fault was shown to be part of a lineament, with various structural manifestations, which extends 250 km from Cheshire to Pembrokeshire, for which the term Pontesford Lineament was used. Woodcock proposed that the most important structural event on the Pontesford-Linley Fault was strike-slip faulting, probably dextral, in latest Ordovician and earliest Silurian times, probably juxtaposing unlike Ordovician terranes. However, he maintained that the

Pontesford-Linley Fault probably controlled the NW margin to a Longmyndian depositional trough in late Precambrian times. Similarly, the Pontesford Lineament recorded a major component of strike-slip elsewhere during late Ordovician to early Silurian time and intermittent activity was noted up to the Triassic. The Pontesford Lineament was shown to be one of an array of NE-SW lineaments in England and Wales, for which there was good evidence for a strike-slip component and which appeared to control the positions of sedimentary basins, particularly during late Ordovician and Silurian times.

In discussion of Woodcock's paper, Lynas (1985) stated that the structural evidence in the Shelve inlier supported the theory of dextral wrench faulting (Woodcock 1984b). Lynas put forward the possibility that the Precambrian 'basement' in Wales and perhaps the 'Midland Platform' is a collage of accreted tectonostratigraphic terranes, similar to those now recognised in California and Alaska and that fault patterns generated in the overlying Lower Palaeozoic (e.g. the Shelve inlier) could have been formed by the upward propagation and reactivation of the strike-slip faults that formed the boundaries to the Precambrian segments (e.g. the Pontesford-Linley Fault).

During 1983 and 1984, work has proceeded on the SO38 and SO39 areas by the BGS who have been remapping these areas. An unpublished field standard of the SO39 area by Dr. R. Langford has been consulted and open-file reports covering the Precambrian of this area (Langford and Lynas, 1985 and Cave et al., 1985) have been kindly provided by Dr. B. Lynas. Reference to this work has been made, where appropriate, in the following chapters.

1.2.2 Age-dating of the Longmyndian

The radiometric age-dates which have been obtained from the Precambrian rocks of Shropshire are summarised in Table 1.

The Wentnor Group was correlated with the Torridonian by Creer (1957), who showed that the mean axis of the normal direction of magnetisation was similar for both. A similar age was therefore inferred.

Harper (1966) reported a single K-Ar, whole-rock age of 470 Ma from Longmyndian "slates", from which he concluded that the Longmyndian had suffered some recrystallisation in early Ordovician times.

Greig et al. (1968) thought that the age of the Longmyndian was greater than 600 Ma as it was Precambrian, from correlation with the Torridonian (Creer, 1957), definitely younger than the Laxfordian orogeny (dated at 1600 Ma) and probably younger than 1160 Ma (Rb-Sr age of biotite from Lewisian Gneiss of the Loch Torridon area).

Fitch et al. (1969) provided three K-Ar whole-rock determinations from the Uriconian as follows: Ercall rhyolite, 632 ± 32 Ma; basalt, Wrockwardine, 638 ± 81 Ma and rhyolite, Pontesford Hill, 677 ± 72 Ma. However, no comment was made on these ages.

Rb-Sr whole-rock analyses of Londmyndian sediments were reported by Bath (1974). These were: Burway Formation, 529 ± 6 Ma, Synalds Formation , 452 ± 31 Ma and Lightspout Formation, 529 ± 23 Ma. From considerations of the illite-crystallinity index, it was concluded that the sediments had suffered burial diagenesis only and had not recrystallised under metamorphic conditions. These ages were therefore thought to reflect dewatering, as a consequence of either

TABLE: 1 RADIOMETRIC AGE-DATING OF THE PRECAMBRIAN ROCKS OF SHROPSHIRE

HORIZON		DATE	METHOD	INTERPRETATION	AUTHOR
LONGMYNDIAN	Tuff	420±9	Fission-track	Uplift age.	Naeser *et al.* (1982)
	Slate	470	K-Ar whole rock	Recrystallisation in early Ord.	Harper (1966)
	Lightspout Fmn.	529±23	Rb-Sr whole rock	Dewatering due to burial or folding.	Bath (1974)
	Synalds Fmn.	452±31	" " "	" " "	" "
	Burway Fmn.	529±6	" " "	" " "	" "
	Stretton Shale Fmn. bentonites	526±18	Fission track	Uplift age.	Naeser *et al.* (1982)
	" " "	528±41	" " "	" "	" " " "
	Longmyndian	600	Initial Sr ratios	Maximum depositional age.	Bath (1974)
URICONIAN	Rhyolite,Ercall	632±32	K-Ar whole rock		Fitch *et al.* (1969)
	Basalt,Wrockwardine	638±81	" " "		" " " "
	Rhyolite,Pontesford Hill	677±72	" " "		" " " "
	Felsic tuff,Wrockwardine	558±16	Rb-Sr whole rock	Primary magmatic.	Patchett *et al.* (1980)
	Granophyre,Ercall	533±13	" " "	Primary intrusive.	" " " "
	" " "	533±12	" " "	" " "	Beckinsale *et al.* (1983)
	Wart Hill	325±7	Fission track	Uplift age	Naeser *et al.* (1982)
	Uriconian	<c.850	Initial Sr ratios	Differentiation from mantle source	Thorpe *et al.* (1984)
RUSHTON SCHIST	Rushton schist	536±8	Rb-Sr whole rock/ biotite	Cooling age	Patchett *et al.* (1980)
	" " "	661±20	Rb-Sr whole rock	Metamorphic age.	Beckinsale *et al.* (1983) and Thorpe *et al.* (1984)
	" " "	1600	Tmorb from Sm-Nd	Differentiation from source.	Beckinsale *et al.* (1983) and Thorpe *et al.* (1984)
	" " "	1300	Tchur from Sm-Nd	" " "	Thorpe *et al.* (1984)
	" " "	c.950	Initial Sr ratios	Pre-metamorphic age.	" " " "

burial or folding. From the initial strontium ratios, a maximum depositional age of 600 Ma was deduced and thus correlations of the Wentnor Group with the Torridonian (Creer, 1957) were dismissed.

Toghill (1975) recognised that if the maximum depositional age of 600 Ma (Bath, 1974) was accepted, then only 30 Ma was available in which to deposit the Longmyndian and to fold up the Uriconian prior to the start of the Cambrian (c. 570 Ma). Alternatively, he suggested a depositional age of 830-790 Ma, which would allow for the correlation of the Torridonian with the Wentnor Group, folding of the Longmyndian between 790 and 570 Ma and a late Cambrian metamorphic episode at around 529 $\pm$ 23 Ma which resulted in the dates of Bath (1974).

Dunning (1975) discussed correlations of the Precambrian of England and Wales. He concluded that since the Rushton Schist and Primrose Hill Gneiss and Schist are not as highly metamorphosed as the Laxfordian, an age of less than 1400 Ma was likely and since the metamorphics are older than the Uriconian, an age of greater than 620 Ma was probable. If correlations of the Wentnor Group with the Torridonian were correct, then the Uriconian was probably 950-1000 Ma old. However, if the Brand Group of the Charnian ($<$680 Ma, on fossil evidence) was correlated with the Stretton Group (600 Ma, Bath, 1974) and all or part of the Uriconian, then the latter was likely to be younger than 700 Ma. The 600 Ma depositional age for the Longmyndian of Bath (1974) was correlated with a widespread overprinting event in the Precambrian basement and 600 Ma was therefore thought to be a possible age for the Longmyndian folding. Also, if the correlation of the Wentnor Group with the Torridonian were correct, a depositional age of 900-950 Ma was possible.

Lomax and Briden (1977) found that the natural remanent magnetisation of both the Longmyndian and the Uriconian was imposed post-tectonically and consequently, the correlations between the Torridonian and the Longmyndian based on a depositional NRM for the latter (Creer, 1957) were dismissed. Since the palaeomagnetic poles for the dykes which cut the Longmyndian were different from the known Cambrian and Ordovician poles, a Precambrian age for the dykes was favoured. Both the Longmyndian deformation and dyke intrusion were thought to have occurred during the late Proterozoic period of rapid "apparent polar wander".

However, Piper (1978) noted that the NRM for the Longmyndian dykes was similar to that for the post-Ashgill and pre-Wenlock intrusions of the Shelve inlier to the west and a similar age was therefore implied. Choubert and Faure-Muret (1980) correlated the Longmyndian and Uriconian with other Precambrian strata in the "north peri-Atlantic and south Mediterranean mobile zones". It was concluded that the Uriconian lavas and Longmyndian gave rejuvenated ages as a result of the "Pan-Atlantic (= Pan African) Thermotectonic Episode", which was dated at 700 Ma to 500 Ma. From stratigraphic correlations, the Stretton Group was referred to the Middle Riphean (1400-1000 Ma), the unconformity dated as c. 1000 Ma and the Wentnor Group was referred to the Upper and latest Riphean (less than 1000 Ma). Supportive evidence was provided by Timofeyev _et al_. (1980) who reported finds of acritarchs in the Stretton Group. These included _Bavlinella faveolata_ Schep., which was thought to indicate a Middle Riphean age for the Stretton Group.

Patchett _et al_. (1980) obtained Rb-Sr whole-rock ages of 558 $\pm$ 16 Ma from a felsic tuff in the Wrockwardine area and 533 $\pm$ 13 Ma from the Ercall granophyre. These dates were interpreted as representing primary magmatic and intrusive dates, which indicated Uriconian volcanism in latest Precambrian to Lower Cambrian time. A Rb-Sr whole-rock/biotite age from the Rushton Schist of 536 ± 8 Ma was interpreted as a cooling age, associated with the end of the Uriconian igneous activity. Since the Lower Cambrian rested on the Ercall granophyre and the Rushton Schist, it was suggested that 533 Ma represented a maximum age for the base of the Cambrian in this area. The Longmyndian was therefore inferred to be possibly Cambrian. The K-Ar dates in excess of 600 Ma, obtained from the Uriconian (Fitch et al., 1969), were presumed to have suffered from excess argon.

Fission-track ages were reported by Naeser et al. (1982). These were: 325 ± 7 Ma (Uriconian -Wart Hill), 420 ± 9 Ma (Longmyndian tuffs) and 526 ± 18 Ma and 528 ± 41 Ma from bentonites in the Stretton Shale Formation. The two former samples came from west of the Church Stretton Fault (F1) and were taken to represent reset ages, reflecting deeper burial and higher temperatures than the two latter from the east side of F1. The dates from the Stretton Shale Formation, east of the Church Stretton Fault (F1) were thought to be due to deep burial, followed by rapid uplift during late Cambrian times.

Beckinsale et al. (1983) reported unpublished data which included a slightly revised Rb-Sr whole-rock date from the Ercall granophyre of 533 ± 12 Ma, a Rb-Sr whole-rock date of 670 ± 20 Ma from the Rushton Schists and Sm-Nd data for the Rushton Schists which yielded model ages (T morb) of about 1600 Ma. The latter date suggested that fragments of older continental crust may be present within the basement of England and Wales. It was maintained that the Ercall granophyre could not have been reset by a regional metamorphic event and therefore, the base of the Cambrian should be revised from 570-590 Ma to a value younger than 533 ± 12 Ma, at least in the Welsh Borderland.

The report of Bavlinella faveolata Shepeleva (Timofeyev et al. 1980) was questioned by Peat (1984b) who thought that this microfossil could not be reliably distinguished from framboidal pyrite and since Bavlinella faveolata was known to extend into the Lower Cambrian, a Middle Riphean age for the Stretton Group, based solely on this microfossil, was therefore put into doubt (Timofeyev et al., 1980).

Toghill and Chell (1984) suggest that if the fission-track dates of 526 ± 18 Ma and 528 ± 41 Ma from the Longmyndian (Naeser et al., 1982) were not reset, then they would fit in with the Rb-Sr dates of 558 ± 16 Ma for the Uriconian (Patchett et al., 1980) and would suggest either that the base of the Cambrian should be revised to be around 530 Ma or alternatively, that the Uriconian volcanics and Longmyndian are lowest Cambrian.

Thorpe et al. (1984) report a Rb-Sr, whole-rock isochron age of 667 ± 20 Ma for the Rushton Schists (similar to the unpublished result of Beckinsale et al., 1983). This age is taken to represent the true metamorphic age. The high initial Sr ratio indicated that the Rushton Schists might have been formed by the metamorphism of sedimentary rocks as old as c. 950 Ma. Maximum model ages from the Sm and Nd data were reported as Tchur, 1300 Ma and Tmorb, 1600 Ma. It was noted that the available Rb-Sr age dates were compatible with the metamorphic crystallisation of the Rushton Schists at 667 ± 20 Ma, eruption of the Uriconian volcanic rocks at 558 ± 16 Ma, followed by the emplacement of the Ercall granophyre, accompanied by limited recrystallisation of the Rushton Schists, at c. 533 Ma. From consideration of the Sr ratios it was considered probable that the igneous rocks could have left a mantle source not more than c. 300 Ma before the measured ages. These dates are in agreement with the proposed formation of the basement of England and Wales during the period 900 Ma to 400 Ma by accretion of island arcs, accretionary prisms and forearc basin sediments within the Iapetus Ocean. The relatively younger ages of the Uriconian volcanics were thought to reflect the north-westward migration of a SE dipping subduction zone, following the formation of the older Malvernian, Stanner-Hanter and Johnston Complexes, which all fall within the range 700-640 Ma.

Palaeomagnetic data were discussed by Thorpe et al. (1984) and Smith and Piper (1984). These indicated that the Longmyndian remanence was post-folding and linked to dewatering and uplift dated by both Rb-Sr studies (Bath,1974) and fission-track (Naeser et al., 1982) at c. 525-515 Ma. From the position of the palaeopoles, it was suggested that the Eastern Uriconian may be slightly older than the Western Uriconian.

Smith and Piper (1984) accepted that the Longmyndian is Lower Cambrian, as the geochronologic data appeared to constrain deposition of the Longmyndian to the period c. 540-530 Ma, followed immediately by rapid uplift. Gibbons (1984) remarked on the anomalous age of the Ercall granophyre and suggested that the granophyre may intrude the Lower Cambrian sediments (referring to a personal communication by Beckinsale), rather than be overlain unconformably by them (Beckinsale et al., 1983).

1.2.3 Sedimentology

In 1856, Salter postulated a shallow water environment of deposition for parts of the Stretton Group after finding ripple marks and very fine, branched lines of which he states: ".... mechanical markings produced by the minute drainage of the surfaces when the water retired; and hence that they afford proofs of quiet littoral action." (Salter, 1856, p. 251). Salter, in his subsequent paper of 1857, noted other shallow water phenomena such as: "The marks of tidal flows or currents." (Salter, 1857, p. 201) and fine ridges, which he attributed to ".... the quiet action of the surf on a level strand, or possibly the agitation of water by wind." (Salter, 1857, p. 202). He found evidence for subaerial exposure in mudcracks and rainprints. The latter were elongated and thought to be due to the impact of slanting rain.

The interpretation of littoral conditions accompanied by exposure was reiterated by Murchison in 1867 and Blake in 1890, who thought that the fine-banding in the Longmyndian was a criteria indicative of littoral conditions. Considering the Longmyndian succession as a whole, Blake noted the gradual coarsening-up of the sequence and interpreted this as being due to ".... the rising of the area of deposition with the denudation of new and less distant masses." (Blake, 1890, p. 390). This recognition of progradation put the view that the Stretton Shale Formation represented a basal deposit into doubt and Blake interpreted this formation as a product of tranquil conditions and noted that ".... it could not possibly be derived from the washing down of such volcanic rocks as are found on its eastern border." (Blake, 1890, p. 390). This obviously influenced his interpretation of the stratigraphic position of the Uriconian, which he thought post-dated the Stretton Group.

In 1900, Cobbold thought that the "ripple marks" of previous authors were due to a corrugation of the bedding, produced by the intersection of the cleavage and the bedding at a very acute angle. Cobbold recognised the pyroclastic nature of the Batch Volcanics and the volcaniclastic nature of the Longmyndian as a whole.

W.S. Boulton, in 1904, described the Western Uriconian of Pontesford Hill, in which he found some pyroclastic deposits. He proposed that they were deposited in shallow water, as they exhibited a fine and regular lamination, a "washed appearance" and the alteration of the glassy fragments to palagonite.

In 1935, Cobbold and Whittard described the Helmeth Grits as consisting of a number of beds of impure, lithic tuff interbedded with shale. The "grit" was envisaged as having formed by ".... agents of volcanicity from already consolidated Uriconian types." (Cobbold and Whittard, 1935, p. 354). This volcanicity was thought to represent the dying phases of the Uriconian volcanicity.

In 1938, Pocock et al. noted that the fragmental rocks of the Uriconian of Pontesford Hill were well bedded, often thinly laminated and had evidence for current action. It was concluded that the majority of these were deposited in water, in agreement with Boulton (1904). In agreement with Cobbold (1900), they proposed that the Longmyndian "ripple marks" were produced tectonically, although some were thought to be sedimentary. This was also the view of Pocock and Whitehead, 1948.

James (1956) demonstrated that the Western Uriconian had acted as the main source for the Longmyndian. He interpreted the features described as Arenicolites by Salter (1856 and 1857) as rain prints.

In 1958, Taylor reviewed the Precambrian sediments of England and Wales. Some description of the sedimentary structures of the Longmyndian was made, together with descriptions of the petrography. The Stretton Shale Formation was interpreted as representing a pre-orogenic phase of "quiet, muddy sedimentation". The Stretton Group was compared with the syn-orogenic flysch and greywacke sequences of the Alpine orogeny. Similarly, the Wentnor Group was compared with post-orogenic molasse.

Welded, vitric tuffs were described from the Western Uriconian of Earl's Hill by Dearnley (1966). Non-welded vitric and lithic tuffs were described from both the Western and Eastern Uriconian. Fragments of welded tuffs were found in many of the Uriconian tuffs and Longmyndian conglomerates and it was concluded that ".... ignimbrites are relatively widespread in the late Pre-Cambrian volcanic suites, indicating at least local terrestrial conditions." (Dearnley, 1966, p. 5).

Greig et al. (1968) described many sedimentary structures, mainly from the "Burway-Synalds Groups", such as rill markings, bubble impressions, pit and mound structures, mud cracks, rain prints and swash-like markings, from which they inferred that the sediments were deposited in a shallow water, probably tidal flat, environment. A shallow water depositional environment was also inferred from the grain size distributions and from the subgreywacke composition of the sediments, which was considered to be typical of a deltaic floodplain and the "closely associated marine environment." Since the Uriconian was interpreted as being largely terrestrial, the base at least of the Stretton Shale Formation was inferred to be of shallow water origin. It appears that a shallow water environment of deposition was thus inferred for the entire Longmyndian, which was then referred to as post-orogenic molasse. The geosynclinal, pre-orogenic sediments recognised by Taylor (1958) in the Stretton Group were stated to be of shallow water origin. The Longmyndian was thought to have been deposited in ".... a narrow crustal depression, which originated by subsidence of the underlying basement between major fault lines." (Greig et al., 1968). It was conjectured that the Pontesford-Linley and Church Stretton fault systems formed the margins of this trough. Some sedimentary structures were noted from the Ragleth Tuff Formation, which included graded bedding, current bedding and slump structures.

Baker (1973) suggested a deltaic origin for the Longmyndian. Since breccias were absent in the Longmyndian, it was suggested that the marginal faults (Greig et al., 1968) were largely synsedimentary.

Davies (1978) reported on a previously undescribed outcrop of Precambrian within the Pontesford-Linley fault system, near Bishops Castle. Here, Western Uriconian tuffs were found to overlie "brecciated ashy turbidites" in which "slump balls" were recognised. The latter were correlated with the "green shaley tuffs" of the Wentnor Group exposed at Chittol (James, 1952a).

An unpublished facies interpretation of the Longmyndian was made by G. Newall and E.J. Anderson who presented talks at the Geological Society of America and at B.S.R.G. The Stretton Shale Formation was interpreted as basinal shales, the Burway Formation as distal to proximal turbidites, the Cardingmill Grit as a shoreline, the Synalds Formation as a lower delta plain, the Lightspout Formation as a shoreline sequence, the Portway Formation as a meandering stream cycle sequence and the Bayston-Oakswood Formation as possible braided stream deposits (E.J. Anderson, personal communication).

Bland (1984) refers to the above unpublished facies interpretation and reiterates some of the above conclusions.

1.2.4 Palaeontology

Markings attributed to organic activity were first described by J.W. Salter in 1856. He recognised three forms:

1. Small depressions, regularly spaced, paired, elongated and parallel, which he attributed to the activities of marine worms and to which he gave the name Arenicola didyma (later renamed as Arenicolites didymus).

2. Small, undulating, shallow furrows which he considered to be worm trails.

3. A proposed trilobite cephalon which he named Palaeopyge ramsayi.

An associated structure, which he considered to be inorganic, consisted of numerous, elongated surficial hollows with numerous, raised and thread-like lines which were parallel to the elongation of the hollows. He remarked on the similarity of the former to rain prints, but favoured an origin due to ".... gaseous bubbles, or to the decomposition of small concretions, rather than that they indicate the marks of ancient showers." (Salter, 1856, p. 250). Similarly, he favoured an inorganic origin for the lines which he proposed were ".... lines of mineral structure." (Salter, 1856, p. 260).

In a subsequent paper (Salter, 1857) Salter proposed the new name Arenicolites for all worm burrows with double openings and as a consequence renamed Arenicola didyma as Arenicolites didymus, which, if it be organic, should be the type species of the genus Arenicolites. A new form was described, Arenicolites sparsus, which was divided into two forms, Junior or Juvenile and Major or Adult.

A. sparsus was described as being distinctly in pairs, although this does not appear to be true from his figured specimens (plate 5, figs. 1 to 4). A. sparsus Major was described as being larger, more scattered and with the edges of the holes a little raised. A. sparsus was differentiated from A. didymus ".... by the holes being remote, not close together, nor parallel to each other. They occur rather higher up in the series too." (Salter, 1857, p. 204). Salter found that his rain prints were "... not easy to distinguish from the larger annelide holes." (Salter, 1857, p. 203). The distribution of Arenicolites was found to be extensive and Salter states: "In a journey across the eastern portion of the Longmynd.... we found them at intervals all the way, until they ended with the sandstones of the Portway itself." (Salter, 1857, p. 201).

Salter's findings appear to have been accepted without further comment until a paper by Professor Blake in 1890. He was dubious that the depressions described by Salter were burrows made by annelids. Similarly, he considered the supposed trilobite cephalon, Palaeopyge ramsayi to be a mere artefact. However, he confirmed Salter's opinion of the existence of worm trails and states: "The surface of the slab is covered with very fine, discontinuous, curling, tube-like bodies which resemble exactly the castings of minute worms." (Blake, 1890, p. 391). He also reports finding impressions of Lingulae from near the "Church Stretton gasworks", but he does not describe or figure them and states that the specimens were broken during removal (Blake, 1890, p. 391). Despite the criticisms of Professor Blake, the review paper by Lapworth and Watts (1894) appears to accept Salter's interpretations for Arenicolites and Palaeopyge and does not question Blake's report of Lingulae.

E.J. Cobbold, in 1900, described many structures which he labelled as "quasi-fossils". With these he included Arenicolites, of which he noted that there is no trace of an organic structure and also that the depressions are scattered indiscriminately about, as well as occurring in pairs. In the Synalds Formation he noted linear markings which were almost invariably accompanied by Arenicolites. He also noted a form similar to Arenicolites, but more rectangular, which appears to have been found within the Ragleth Tuff Formation. Also from the Synalds Formation were noted paired spots ".... not like Arenicolites with occasionally wavy, radiating marks and concentric circular hollows." (Cobbold, 1900, p. 87). Other burrow-like markings which were found at the top of the Burway Formation and beneath the Cardingmill Grit were compared to Histioderma hibernicum from Bray Head. These markings were oval depressions which penetrated obliquely through the sediment up to 6" or more. Cobbold attributed this either ".... to the passage of some organism through the wet mud, or possibly, they might be due to the escape of bubbles of some gas." (Cobbold, 1900, p. 86). Other "quasi-fossils" noted from the Synalds Formation were small, flattened, white spheroids, which were infilled with a dark green material, and a vesicular looking rock with "tubules" of white carbonate, in both of which were found no traces of organic structure. From the Synalds Formation and also from some part of the Ragleth Tuff Formation were reported shallow depressions with oblique raised ridges which were compared to Cruziana but were rarely found in pairs. Palaeopyge was dismissed as an artefact due to structural corrugations of the surface. Nodules found within the Burway Formation were remarked on as being similar in form to lamellibranchs, but no trace of a shell structure was found.

In the review paper of Watts (1925), the find of *Palaeopyge*, which was now referred to as a pygidium, was considered to be dubiously organic. However, organic activity was accepted as resulting in "burrows and tracks of annelides." (Watts, 1925, p. 334). These were also the opinions of Lapworth and Watts in 1910 and of subsequent authors.

The paired nature of the holes of *Arenicolites* was considered to be "purely fortuitous" by James (1952b), who noted that individual depressions were more abundant. He states that "the bulk of the depressions are oval in shape and could be caused by rain striking at an angle to the surface." (James, 1952b, p. 26). However, he later notes that the oval shape could have been produced by tectonic deformation of a circular depression. He thought that the more circular depressions were certainly rain prints as they possessed a raised rim.

Greig *et al*. (1968) preferred an inorganic interpretation for *Arenicolites* and referred it to ".... inorganic bubble holes and pit and mound structures and in some instances to the counterparts of original mounds or small blister-like features." (Greig *et al*., 1968, p. 72). It was noted that the holes were mostly randomly distributed and not paired as stated by Salter (1856 and 1857) and no burrows were identified. *Palaeopyge ramsayi* was accepted as being inorganic as previously interpreted (Watts, 1925).

Timofeyev (1980) reported finds of acritarchs from Buxton Quarry, All Stretton. The acritarchs were assigned to *Protosphaeridium densum*, Timofeyev, *P. scabridum*, Timofeyev, *P. turberculiferum*, Timofeyev, *P. laccatum*, Timofeyev, *P. rigidulum*, Timofeyev, *Orygmatosphaeridium distributum*, Timofeyev, *Protosphaeridium* sp., *Trematosphaeridium* sp., *Synsphaeridium* sp., *Gloeocapsomorpha* sp. and *Bavlinella faveolata*, Shepeleva. The latter acritarch indicated a Middle Riphean age.

Bland (1984) referred arrays of straight to gently curving and occasionally diverging fine ridges and broad, shallow grooves, which were found on the top surfaces of fine sediment, to the genus *Arumberia*, Glaessner and Walter. This was tentatively interpreted as ".... the impression of a colony of flexible thin-walled tubular elements .... firmly adpressed or possibly fused into a dense sheaf, bundle or solid block, where they are in contact with the sediment surface." (Bland, 1984, p. 630-631). This structure was often found along with small, round or elliptical hollows and occasionally mounds. These were thought to be due to the collapse of flexible-walled, spheroidal bodies in the sediment, which were tentatively interpreted as a "dispersable resting stage" of *Arumberia* (Bland, 1984, p. 631). *Arenicolites*, Salter (1857) was interpreted as the impressions of *Arumberia* spheroids. *Arumberia* was found throughout the sequence above the Cardingmill Grit. Additionally, 5-50 mm diameter raised structures with central dimples from the Burway and Stretton Shale Formations were compared with *Cyclomedusa* and plain c. 2 cm diameter impressions in negative epirelief were compared with *Bronicella podolica*. The *Arumberia* spheroids were compared with *Chuaria*, Ford and Breed, although no organic material was found in the Longmyndian specimens.

Microfossils from the Longmyndian were described by Peat (1984a). A broad, nematomorph cryptarch, 155 µm wide, with marked transverse striae was reported from the Stretton Shale Formation. Curved ridges were noted on some of Salter's original specimens, which were of similar dimensions to this broad, nematomorph cryptarch and it was therefore postulated that the ridges could be the moulds of these. This cryptarch was tentatively interpreted as the organic lining of a metazoan dwelling-tube or burrow. Three forms were noted from the Lightspout Formation: broad nematomorph cryptarchs, 25 μm to 155 μm wide, thin nematomorph cryptarchs, 5 μm to 17 μm wide, which occurred singly or in dense mats on the bedding planes and sphaeromorph cryptarchs, 3 μm to 70 μm in diameter. The thin nematomorph cryptarchs were thought to represent photosynthetic microorganisms.

The report of Bavlinella faveolata Shepeleva (Timofeyev, 1980) was doubted by Peat (1984b) who thought that this microfossil could not be reliably distinguished from framboidal pyrite.

1.3 A Note on terminology

The term "volcaniclastic" was defined by Fisher and Schmincke (1984) as: ".... all clastic materials formed by any process of fragmentation, dispersed by any kind of transporting agent, deposited in any environment or mixed in any significant proportion with non volcanic fragments." (Fisher and Schmincke, 1984, p. 89). Suthren (1985) used this term in a similar way for "all clastic sediments and rocks, regardless of depositional process, whose particles are predominantly of volcanic origin." This thesis uses the term volcaniclastic, as defined by Fisher and Schmincke (1984) except for the added modifier that the particles are predominantly of volcanic origin, as used by Suthren (1985). In this sense, most of the Longmyndian Supergroup is composed of volcaniclastic sediment. Consequently, a distinction is required between deposits which result from volcanic action and those which result from a redistribution of these deposits by non-volcanic processes.

The term "pyroclastic" was defined by Fisher (1966) as: ".... an adjective applied to rocks produced by explosive or aerial ejection of material from a volcanic vent." (Fisher, 1966, p. 288) and was used similarly by Suthren (1985) for volcaniclastic sediments produced and emplaced by magmatic explosion. This thesis uses this term in a similar manner.

The term "epiclastic" was defined by Fisher (1966) as: "mechanically deposited sediments (gravel, sand, mud) consisting of weathered products of older rocks. Detrital material from pre-existent rocks." Suthren (1985) uses this term similarly for volcaniclastic sediments produced by sedimentary erosion, transport and deposition. Fisher (1961) considered it important to distinguish, if possible, fragments produced by explosion from those produced by weathering processes. Consequently, he proposed the classification of "reworked tuff" and defined this as pyroclastic material that ".... has been removed from its original place of deposition and redeposited before lithification, as for example "reworked tuff". Reworked tuff is to be distinguished from rocks formed by epiclastic processes." (Fisher, 1961, p. 1412). However, Fisher (1960) used the term epiclastic to "include volcaniclastic material transported by epigene geomorphic agents, including unconsolidated pyroclastic debris." (Fisher, 1961, p. 1413). Suthren (1985) differentiated between primary pyroclastic deposits and secondary (reworked) pyroclastic deposits but noted that secondary pyroclastic deposits were often considered to be essentially equivalent to epiclastic deposits by some authors. This thesis uses the term epiclastic s.s. as it was defined by Fisher (1966), where it was possible to distinguish weathered volcanic detritus from reworked pyroclastic detritus as defined by Fisher

(1961). When however, this distinction could not be made, the term epiclastic has been used in a broad sense (epiclastic s.l.), to indicate a fragmental rock which has been deposited by water, wind or gravity induced processes and has been derived from a previously deposited or emplaced rock, regardless of its origin and state of lithification, as used by Fisher (1960).

Since most of the Longmyndian Supergroup is composed of volcaniclastic rocks, care has to be taken to distinguish the process of deposition as sediment produced by mechanical processes from volcanic rocks may be lithologically similar to sediment produced by volcanic processes. For example, Fisher and Schmincke (1984) note that some pyroclastic deposits composed mainly of lithic pyroclasts may be identical to epiclastic s.s. deposits.

CHAPTER 2

STRATIGRAPHY

2.1 Introduction

The succession determined by this study is essentially similar to that proposed by recent authors (e.g. Toghill and Chell, 1984). However it differs in including the Ragleth Tuffs, here called the Ragleth Tuff Formation with the Longmyndian Supergroup rather than with the Uriconian. Additionally, an unconformity between the Wentnor Group and the Stretton Group, which has been accepted since the accounts by Greig *et al*. (1968), is not accepted here. The proposed succession is detailed in fig. 2 and a vertical column is included with map 2 (in cover pocket). The details of the main succession are well known (see Greig *et al*., 1968) and this account therefore addresses the changes that are proposed.

The main outcrop of the Longmyndian Supergroup occurs between the Pontesford-Linley and Church Stretton fault systems (fig. 1). The succession records a progradational basin infil approximately 6487m thick. The lower Stretton Group is composed of basin plain deposits, turbidites, shallow marine, alluvial floodplain and fluvial deposits in ascending order. The overlying Wentnor Group is mostly composed of alluvial braidplain deposits. It is later argued that, contrary to previous opinion, a major unconformity between the Wentnor Group and the Stretton Group is not present in this area.

Within the Pontesford-Linley and Church Stretton fault systems and to the east of the Church Stretton Fault, there are exposures of the Uriconian Volcanic Complex, which is generally held to be older than the Longmyndian Supergroup and to be either conformably overlain by it (e.g. James, 1952b) or possibly unconformably overlain by it

FIG 2 : THE STRATIGRAPHIC SUCCESSION OF THE PRECAMBRIAN OF SHROPSHIRE.

BRIDGES FORMATION (851m)

BAYSTON-OAKSWOOD FORMATION (1690m-2172m)

Western limb includes;	Eastern limb includes;
Lawn Hill Conglom. Mbr.(156m)	Stanbatch Conglom. Mbr.(122m)
Oakswood Conglom. Mbr. (101m)	Darnford Conglom. Mbr. (90m)
Radlith Conglom. Mbr. (83m)	Haughmond Conglom. Mbr.(106m)

-----------------(No Unconformity)-----------------------

PORTWAY FORMATION (742m)

Huckster Conglomerate Member (99m) at the base.

LIGHTSPOUT FORMATION (c.332m)

Red Lightspout Member (112m)

Green Lightspout Member (c.220m)

SYNALDS FORMATION (665m)

Red Synalds Member (112m) with Batch Volcanics near top.

Green Synalds Member (143m)

BURWAY FORMATION (709m)

Cardingmill Grit Member (52m) at top.

Buxton Rock Member (15m) at base.

STRETTON SHALE FORMATION (c.605m)

---------------------------------- CONTACT UNCERTAIN

? RAGLETH TUFF FORMATION (411m) STRATIGRAPHIC POSITION UNCERTAIN

Helmeth Grit Member (29m) at top.

------------------(Faulted contacts)---------------------

URICONIAN VOLCANIC COMPLEX

------------------(Faulted contacts)---------------------

RUSHTON SCHIST AND ? PRIMROSE HILL GNEISS AND SCHIST (e.g. Greig et al., 1968). However, the relationships between the Longmyndian Supergroup and the Uriconian Volcanic Complex are open to various interpretations and these are discussed in the following sections.

There are units, mostly exposed within the major fault systems, whose relationships with the surrounding rocks are equivocal because of faulting or poor exposure. These units have been informally named as the Linley beds, Knolls sandstone beds, Brokenstones sandstone beds and the Willstone Hill conglomerate beds. The three latter units may be correlated with the Wentnor Group and may have unconformable relationships with the Stretton Group or Uriconian Volcanic Complex. The stratigraphical positions of these units are discussed.

2.2 The relationships between the Stretton Group and the Wentnor Group

A conformable relationship between the Stretton Group and the Wentnor Group was accepted by Murchison (1839 and 1867), Ramsay (1853), Salter (1856 and 1857), Callaway (1891) and Lapworth and Watts (1910). Blake (1890) proposed an unconformity between the Wentnor Group and the Stretton Group and interpreted the Huckster Conglomerate as an outlier of the Wentnor Group. Callaway (1891) later showed the Huckster Conglomerate to be a conformable part of the Stretton Group. Cobbold (1900) placed an unconformity at the base of the "Ratlinghope Conglomerate", now the Stanbatch Conglomerate, and thought that the base of the Huckster Conglomerate might be a second unconformity. An unconformity at the base of the Wentnor Group was accepted by Cobbold (1925) and Cobbold and Whittard (1935). However, Whittard (1952) was uncertain of the position of the unconformity, which he thought might occur at the base of the Huckster Conglomerate. James (1952b and 1956) proposed two unconformities, one at the base of the Wentnor Group and the other at the base of the Huckster Conglomerate. However, Greig and Wright (1958) disproved the proposed unconformity at the base of the Huckster Conglomerate.

Dean (1964) accepted an unconformity between the Wentnor Group and the Stretton Group, for which he provided additional evidence from the Horderley and Brokenstones area. Greig *et al*. (1968) accepted an unconformity, but thought that the evidence of Dean (1964) was equivocal. An unconformity between the Stretton Group and the Wentnor Group was accepted by Wright (1968), Hains (1969) and Toghill and Chell (1984).

2.2.1 Arguments for an unconformity between the Stretton Group and the Wentnor Group

1) The apparent thinning of the Portway Formation to the south-west (James, 1952b).
2) The absence of the Stretton Group in the western limb of the syncline, the Stretton Group having been removed beneath the Wentnor Group unconformity, resulting in the overstep of the Wentnor Group onto the Western Uriconian (James,1952b).
3) The greater geographical distribution of the Wentnor Group (Cobbold and Whittard, 1935).

4) The absence of Cambrian and Ordovician strata on the Stretton Group. It was thus implied that a mantle of unconformable Wentnor Group covered the Stretton Group during the Cambrian and the Ordovician (Cobbold and Whittard, 1935).

5) The Wentnor Group rests directly on the Stretton Shale Formation in the Horderley and Brokenstones area (Dean, 1964).

6) The Wentnor Group rests unconformably on beds which are tentatively correlated with the lower part of the Synalds Formation in the Haughmond Hill area (Greig et al., 1968).

2.2.2 Arguments for conformity between the Stretton Group and the Wentnor Group

1) Tectonic rotation between the Stretton Group and the Wentnor Group.

The Portway Formation thins from approximately 1165m in the north-east, in the area of High Park Hollow (SO438972), to 643m in the south-west, in the area of Ashes Hollow (SO418937). This thinning is reflected in the anticlockwise rotation in strike of the base of the Wentnor Group with respect to the top of the Huckster Conglomerate. At Hawkham Hollow (SO433976), the base of the Wentnor Group and the top of the Portway Formation are exposed, but the actual contact is obscured. This is the only locality in the Long Mynd where the Portway Formation and the Wentnor Group relationships can be adequately determined as elsewhere, exposure is poor. At this locality, the Portway Formation and the Bayston-Oakswood Formation have similar dips and strikes and an angular unconformity is not apparent. Between the Huckster Conglomerate at High Park House

(SO443972) and the base of the Wentnor Group at Hawkham Hollow (SO433976), the strike is gradually rotated anticlockwise. A similar effect occurs at the head of Ashes Hollow (SO418937), where the strike appears to be rotated within the Portway Formation such that the base of the Wentnor Group appears to be conformable with the upper parts of the Portway Formation. The anticlockwise rotation of the base of the Wentnor Group from the top of the Huckster Conglomerate thus appears to be the result of an anticlockwise rotation of strike within the Portway Formation rather than an angular unconformity at the base of the Wentnor Group.

Fig. 12 shows that the average bedding plane, average cleavage plane and average minor fold, axial plane of the Stretton Group are rotated clockwise with respect to the average bedding and axial plane of the major syncline of the Wentnor Group. The rotation between the Stretton Group and the Wentnor Group thus appears to be tectonic. The rotation may be explained by movement along strike faults within the Portway Formation, which are not demonstrable because of poor exposure, or alternatively, the rotation is accomplished by shear and accommodation across cleavage planes throughout the Portway Formation. The rotation may be viewed as a decoupling between the more competent Wentnor Group and the less competent Stretton Group.

2) Faulted contacts between the Wentnor Group and the Western Uriconian

The Wentnor Group is faulted against the Western Uriconian at Lyd Hole, near Pontesford Hill (SJ415055). Elsewhere, the contact is not exposed or a thin strip of Linley beds separates the Wentnor Group from the Western Uriconian (from the Knolls, SO375978 to Linley, SO343936). The Linley beds have been included with the

Wentnor Group by James (1952a) and have been interpreted as showing an inverted and unconformable relationship with the Western Uriconian. However, the Linley beds are here considered to be a distinct unit. The contacts between the Western Uriconian and the Linley beds and the Linley beds and the Wentnor Group are interpreted as reverse faults. An unconformable relationship between the Wentnor Group and the Western Uriconian is thus refuted. Therefore the Wentnor Group cannot be shown to violently overstep the Stretton Group onto the Western Uriconian.

3) Faulted relationships between the Wentnor Group and the Stretton Shale Formation in the Brokenstones and Horderley area.

In the area between Horderley (SO408871) and Cwm Head (SO424887), strata correlated with the Wentnor Group and the Stretton Shale Formation are poorly exposed within the Church Stretton fault system. Some of the exposed contacts were interpreted as unconformities by Dean (1964) and Whittard (_in_ James, 1956). However, Greig _et al_. (1968) thought that the evidence for an unconformity was equivocal and stated: "... in every case, the junction is complicated by faulting." (Greig _et al_., 1968).

The sandstones in this area are geographically and structurally isolated from the Wentnor Group in the Long Mynd area, therefore correlation is based on lithological similarity. These sandstones are compositionally similar to the sandstones of the Wentnor Group (fig. 19). However, because of their geographical and structural isolation, the term Brokenstones sandstone beds is proposed for these strata exposed in the Church Stretton fault system as this does not assume direct correlation with the Wentnor Group.

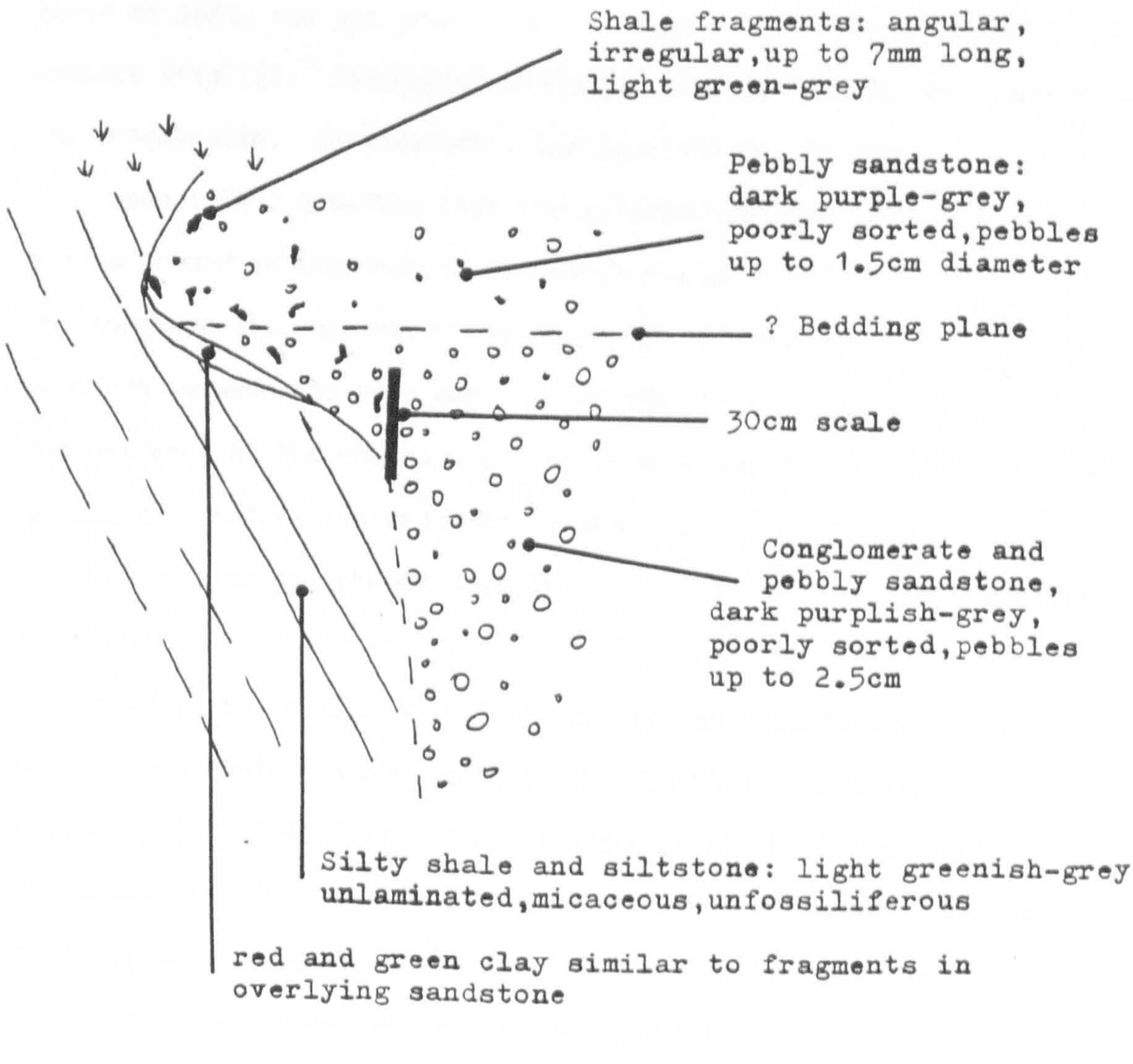
Dean (1964) describes three exposed contacts, one of which is a natural exposure in the Horderley area at SO40808649. Here, grey shales were seen, separated from purple sandstones by a band of maroon shales. The relationships were thought to be unclear. Maroon shales are commonly developed in fault planes elsewhere in the Longmyndian and are interpreted as fault gouge. The maroon shale at this locality might have formed in a similar manner. The remaining exposures were excavated. One excavation in the region of SO417880 is not visible. However, poor exposures of the Stretton Shale Formation and the Brokenstones sandstone beds are found in close proximity in this area, with highly divergent dips and strikes. Dean (1964, p.270) described an upward succession as follows:-

Maroon sandstone	3ft 3ins
Maroon conglomerate	5ft 2ins
Red-maroon shales	2ft 4 to 9ins
Fault breccia, etc.	0 to 4ins
Grey-green shales	1ft 8ins

This contact is here interpreted as a fault, since a fault breccia is developed at the change in lithology and the red-maroon shales might be interpreted again as a fault gouge.

One excavation is poorly exposed near the Round House at SO42008816. Here, a conglomerate and pebbly sandstone is seen in contact with greenish grey shales (fig. 3). The conglomerate and pebbly sandstone is lithologically and compositionally similar to the Brokenstones sandstone beds. However, the greenish grey shales differ significantly from the Stretton Shale Formation. The lithology is variably light grey, light greenish grey or light

FIG: 3 PHOTOGRAPH AND FIELD SKETCH OF THE JUNCTION BETWEEN THE BROKENSTONE SANDSTONE BEDS AND THE ? STRETTON SHALE FORMATION NEAR THE ROUND HOUSE, SO 42008816.



brownish grey, whereas the Stretton Shale Formation is exclusively light greenish grey. The lithology is typically silty and grades to siltstone. It is mostly unlaminated and contains significant amounts of mica. All these characteristics are not typical of the Stretton Shale Formation. This lithology is more closely comparable to a grey-green counterpart of the shales and siltstones which are interbedded with the sandstones of the Wentnor Group.

Bedding within the Brokenstones sandstone beds at this locality is obscure, but may be subhorizontal. Alternatively, the bedding could be subparallel to the contact and the contact could thus be interpreted as an erosional channel base. This is plausible if the silty shales are correlated with the Wentnor Group shales. A thin layer of soft, red and green clay is found in some parts of the contact (fig. 3). This clay is similar to fault gouges developed in the Longmyndian. Consequently, the junction may be faulted.

Dean (1964) observed that the conglomerate appeared to occupy a hollow eroded in the shales, as shales reappear on the other side of the conglomerate, approximately 6m to the north-west. However, the junction between the conglomerate and the shale is not exposed. The reappearance of the shales could be equally explained by faulting or by a normal succession from the conglomerate to the shale.

In conclusion, the evidence for an unconformity at this locality is equivocal.

Dean (1964, p.272) also inferred that an unconformity existed between the Wentnor Group and the Stretton Shale Formation near Pillocksgreen (SO423877). Here, two strips of the Brokenstones sandstone beds are separated by a narrow strip of Stretton shales. Faulting was considered to be unlikely, as two faults of enormous throw would have been required to juxtapose the two horizons and the rocks appeared to be uncrushed. However, the Geological Survey interpret these contacts as faults (sheet SO48). In the absence of exposures of the contact, it cannot be stated with certainty that these contacts are unconformable and since fault contacts within the Church Stretton fault system are common, fault contacts would appear to be more probable. The large, apparent throw on these faults may be the result of large strike-slip displacements as well as dip-slip displacements, if the Church Stretton fault system is considered to be a master wrench zone. The probability that the Church Stretton fault system is a master wrench zone is discussed in chapter 3. The absence of crushing of the rocks in this area does not preclude large strike-slip displacements, since a major strike fault between the Western Uriconian and the Wentnor Group at the Lyd Hole (SJ41500553, see section 2.7) is represented by only 40cm of fault gouge and the adjacent Longmyndian appears to be uncrushed.

In conclusion, the evidence for an unconformity in the area between Horderley (SO408871) and Cwm Head (SO424887) is dubious. In most cases, the junction could be explained by faulting. At only one, poorly exposed locality is the evidence equivocal (near the Round House, SO42008816) and here it appears possible that the shales may not be correlated with the Stretton Shale Formation.

4) Conformable base of the Wentnor Group in the Haughmond Hill area.

Pocock _et al_. (1938) proposed a conformable succession from west to east from the Haughmond Conglomerate through purple sandstones and shales to green sandstones and shales. The latter were correlated with the Burway Formation. However, Whitehead (1955) demonstrated a westerly-younging succession in the Wentnor Group at Haughmond Hill.

Since an easterly-younging succession was demonstrated in the Burway Formation by Challinor (1948), Whitehead (1955) proposed that the strata were folded. Whitehead (1955) noted the apparent absence of the Lightspout, Synalds and Portway Formations at Haughmond Hill and therefore proposed that there was an unconformity at the base of the Wentnor Group. However, Greig et al. (1968) proposed that the Wentnor Group rested unconformably on beds which were tentatively correlated with the lower part of the Synalds Formation.

Concordant dips and strikes across the boundary would appear to indicate a conformable relationship. In the vicinity of Douglas's Leap (SJ540136) the lithology is composed of numerous, fine-grained, sandstone beds between one and three metres thick, which fine upwards and indicate a westerly-younging succession. These sandstones are interbedded with thin siltstones and shales. This facies can be correlated with the Portway Formation and may be differentiated from the Synalds Formation, which is superficially similar, by the greater abundance of sandstone beds and the siltier nature of the associated fine-grained sediments. The Wentnor Group thus appears to rest conformably on the Portway Formation.

The Burway Formation is exposed in small quarries at Criftin (SJ55111381). A horizon close to the level of the Cardingmill Grit, which is composed of planar laminated sandstone, is exposed at SJ54581363. Between this locality and the ?Portway Formation in the vicinity of Douglas's Leap (SJ540136), the Synalds Formation and the Lightspout Formation (if present) would thus be of the order of 450m thick as opposed to the c. 1000m which is found in the Long Mynd. Geological Map 2 (in cover pocket) shows a complicated lineament pattern derived from aerial photos of Haughmond Hill. Somc strike-parallel lineaments could be interpreted as strike faults, one of which outcrops along the south-eastern side of Haughmond quarry (SJ54251455). These faults may explain the apparent thinning of the succession. Alternatively, a facies change between the Long Mynd and Haughmond Hill might have occurred such that the Synalds and Lightspout Formations are not represented or are thinned. Strike faulting is considered to be more likely. Since these faults are bedding-parallel, they are difficult to recognise. It is possible that the base of the Wentnor Group on Haughmond Hill is also a strike fault (see Geological Map 2, in cover pocket).

In conclusion, an unconformity is not apparent from the dip and strike geometries. Probable correlation of the underlying strata with the Portway Formation indicates a conformable succession. Strike faulting renders the interpretation of an unconformity, based on thinned sequences or absent formations, open to question in this area.

5) The greater geographical extent of the Wentnor Group

This observation has been used as an argument for an unconformity between the Wentnor Group and the Stretton Group by Cobbold and Whittard (1935) and by Greig *et al*. (1968, p.45). It was noted that the Stretton Group is confined to outcrops on the Long Mynd, on Haughmond Hill and within the Church Stretton fault system and that the Wentnor Group is found both in the Church Stretton area and in areas outside it: in the inliers of Pedwardine, Old Radnor and Huntley in Gloucestershire. This was taken to imply "a widespread and violent overstep of the Wentnor Series." (Greig *et al*., 1968).

No overstep of the Wentnor Group onto the Stretton Group can be demonstrated in the Church Stretton area. Outside the Church Stretton area the absence of the Stretton Group may be due to faulting and in the absence of other evidence, removal of the Stretton Group beneath a proposed unconformity at the base of the Wentnor Group cannot be inferred. Outcrops of the Longmyndian Supergroup appear to be controlled by major fault systems e.g. the Church Stretton fault system and the Pontesford-Linley fault system. These fault systems might have suffered strike-slip displacements as well as considerable vertical displacements (e.g. Woodcock, 1984a). Thus the outcrop pattern of the Longmyndian Supergroup may be the result of vertical and lateral dissection of the Longmyndian basin by faults and the geographical distribution cannot be the sole result of stratigraphical relationships.

6) The absence of Cambrian and Ordovician strata on the Stretton Group

This argument was used as evidence for an unconformity by Cobbold and Whittard (1935). This was taken to imply that a mantle of unconformable Wentnor Group covered the Stretton Group during the Cambrian and Ordovician.

Cambrian or Ordovician strata might have covered the Stretton Group directly and been removed subsequently beneath the Caradoc unconformity, which is manifest near Pontesford Hill and east of the Church Stretton Fault (F1), or beneath the Llandovery unconformity, which is manifest on the south-west and south-east margins of the Long Mynd. It is thus unnecessary to postulate an unconformable mantle of Wentnor Group, which is not now apparent. Alternatively, pre-Caradoc strata might not have been deposited in the Long Mynd area.

7) Lithological similarities between the Stretton Group and the Wentnor Group

Compositionally, the lithologies of the Wentnor Group are similar to the lithologies of the Stretton Group, in that both groups are composed of predominantly glassy volcanic rock fragments together with minor quartz and feldspar. When the modal percentages of these three components are plotted (fig. 20) they define similar compositional fields which are close to the field of magmatic arc provenances described by Dickinson and Suczek (1979).

Similar lithologies and facies to the Wentnor Group are developed within the Stretton Group. For example, the Huckster Conglomerate is of a similar grain size, has a similar composition and is composed of homogeneous, trough cross-bedded sandstone interpreted to have been deposited in a similar braided river or braidplain environment.

The Longmyndian Supergroup is interpreted as representing a progradational sequence from basin plain, to turbidite fan, deltaic and alluvial floodplain environments. The Bayston-Oakswood Formation is interpreted as a braidplain deposit, which appears to be the logical consequence of the progradational basin infil.

Although these features do not preclude an unconformable relationship between the Wentnor Group and the Stretton Group, they do suggest that the Wentnor Group is an intimate part of the same basin infil event and lend support to the conformable relationship derived from other considerations.

All the arguments for an unconformity between the Stretton Group and the Wentnor Group have been considered. In the main Longmyndian outcrop an unconformity does not appear to be likely and additional arguments have been considered which would indicate a conformable succession. However, in the Church Stretton fault system, the stratigraphic relationships between the probable Wentnor Group (Brokenstones sandstone beds) and the Stretton Group are uncertain. Since it is later argued that probable Wentnor Group, the Knolls sandstone beds and the Willstone Hill conglomerate beds, may rest unconformably on the Western and Eastern Uriconian volcanics respectively, then it appears possible that the Wentnor Group may overstep the Stretton Group onto the Uriconian Volcanic Complex at the basin margins. This relationship may now be manifest in the Brokenstones sandstone beds/Stretton Shale Formation relationship within the Church Stretton fault system, although unconformable relationships cannot be unequivocally demonstrated in the field.

2.3 The status of the Ragleth Tuff Formation

The Ragleth Tuff Formation has traditionally been included with the Uriconian Volcanic Complex. However, Callaway (1891) noted problems in distinguishing between "lava-flows" and sedimentary rocks in this area and concluded that "... sedimentary material forms a larger part of the Uriconian rocks of the Church Stretton area than we had supposed." (Callaway, 1891, p.120). Cobbold (1900) proposed three divisions for the succession in the Caer Caradoc area in response to this problem, these being volcanic, "volcanic sedimentary" and sedimentary in ascending order (Cobbold, 1900, p.97). The "volcanic sedimentary" division corresponds with the

Ragleth Tuff Formation and the overlying sedimentary succession begins with the Helmeth Grit. Cobbold also noted that the Ragleth Tuff Formation is "... conformable with the base of the lower Longmyndian and consequently may really be part of that great series." (Cobbold, 1900, p.87). Cobbold and Whittard (1935) accepted the inclusion of the Ragleth Tuff Formation with the Uriconian Volcanic Complex and the Helmeth Grit with the base of the Longmyndian. The Helmeth Grit was interpreted as an impure, lithic tuff and appeared to pass up into the Stretton Shale Formation by interbedding. The Helmeth Grit appeared to be similar to other "lithic tuffs" in the underlying Ragleth Tuffs and thus a conformable succession was proposed from the Ragleth Tuffs through the intermediate Helmeth Grit into the Stretton Shale Formation. However, Greig et al. (1968) thought that the base of the Helmeth Grit was an unconformity, as a divergence of strike was noted between it and the Uriconian of Caer Caradoc and the Helmeth Grit appeared to rest on different lithologies within the Uriconian of Caer Caradoc.

Several authors have noted features which are suggestive of a sedimentary origin for the Helmeth Grit and the Ragleth Tuff Formation, but these have not been commented on. Cobbold and Whittard (1935) noted the frequent presence of angular shale fragments up to 4 inches in length in the Helmeth Grit "grits", which were indistinguishable from the shales with which the "grits" alternated. Greig et al. (1968) noted sedimentary structures within the "grits" of the Ragleth Tuff Formation, these were traces of current bedding, slump structures and graded bedding. These "tuffs" were described as being interbedded with "... darker, greenish rocks of a more sedimentary aspect." (Greig et al., 1968, p.16).

2.3.1 The lithologies of the Ragleth Tuff Formation

Three main lithotypes are present:-

a) Very poorly sorted with fragments up to 4mm, dominantly composed of highly vesiculate and glassy fragments together with a high proportion of glomerocrystic plagioclase. This lithotype is interpreted to have been formed by pyroclastic activity, because of its poor sorting and high percentage of juvenile magmatic components, which appear to have suffered no reworking. This lithotype comprises less than 10% of the exposed Ragleth Tuff Formation.

b) Moderately to moderately well sorted, very fine to fine, with modal averages of quartz, 18%; feldspar, 22% and volcanic rock fragments, 60%. The volcanic rock fragments are predominantly glassy, but are distinct from those in lithotype A, in being non-vesiculate, not chloritised, commonly devitrified and with occasional hyalopilitic textures due to fine feldspar laths.

c) Siltstone, similar to the above in composition, together with shale.

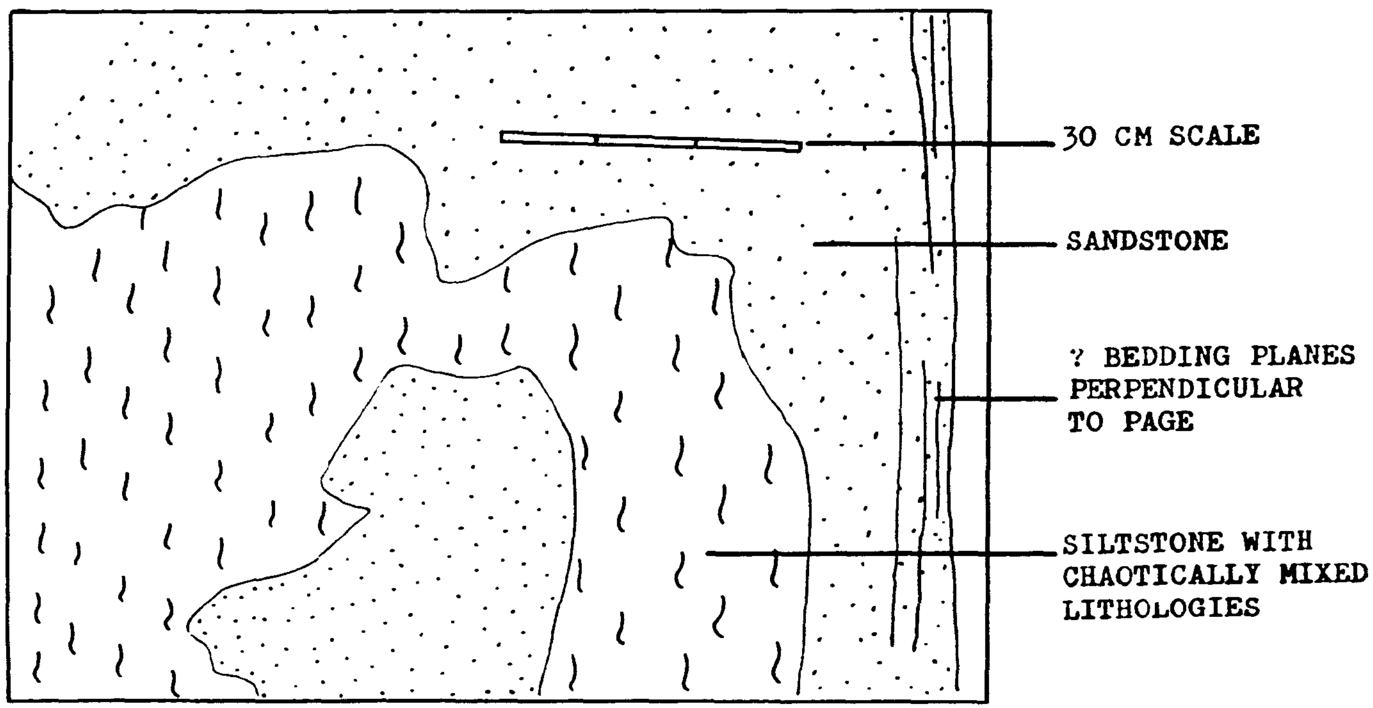
Lithotypes B and C comprise at least 90% of the exposed Ragleth Tuff Formation. Compositionally, lithotype B is identical to the sandstones of the Longmyndian Supergroup and plots of modal percentages fall within the Longmyndian compositional field (fig. 19).

In addition to these three lithotypes, there are occasional basic intrusions.

2.3.2 Sedimentary structures in the Ragleth Tuff Formation

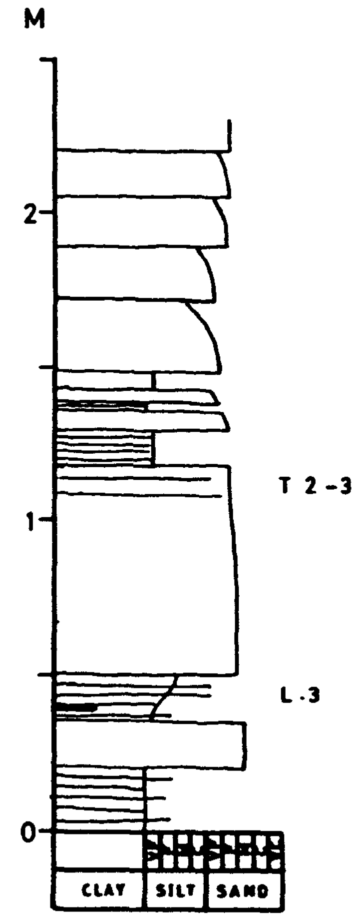
Lithotype B commonly occurs as graded units, 15 to 30 cm thick, with a sharp base. Occasionally, thicker units, up to 60 cm thick, occur. Most of the units appear to be massive, but occasional planar

FIG 4. SEDIMENTARY STRUCTURES IN THE RAGLETH TUFF FORMATION



LOG 53 RAGLETH TUFF FORMATION

RAGLETH HILL SO 45409178



lamination and ripple cross-lamination is visible. Lithotype C is commonly planar laminated on a fine scale. These features are illustrated by log 53, fig. 4. Rarely, soft-sediment deformation is visible (fig. 4). This might have been caused by extreme loading or slumping.

Lithotypes B and C are interpreted as epiclastic s.l., rather than pyroclastic, since they are well sorted and are composed of subrounded grains. They have similar compositional characteristics and sedimentary structures to some Longmyndian sandstones and are tentatively interpreted as the result of turbidite deposition (resulting in lithotype B) and deposition from suspension (resulting in lithotype C).

2.3.3 The lithologies of the Helmeth Grit

The Helmeth Grit is composed of two lithotypes:-

a) Fine to medium grained, moderately well sorted and subangular to subrounded. This lithotype is composed of quartz (33%), feldspar (33%) and non-vesiculate and glassy rock fragments (34%). These components are identical in nature to those of the Ragleth Tuff Formation sandstones and the Longmyndian sandstones, but the proportions of the components are different. A plot of the modal compositions of lithotype A (fig. 19) shows that although the lithotype falls outside of the Longmyndian compositional field, the composition reflects the trend of an increasing ratio of plutonic to volcanic components (trend 2, fig. 19). Lithotype A is similar in texture and in colour to the Ragleth Tuff Formation sandstones.

b) Shale. This is mostly finely laminated and is similar in character and colour to the shales of the Ragleth Tuff Formation.

2.3.4 Sedimentary structures in the Helmeth Grit

Exposures of the Helmeth Grit are poor and the thickness of the beds are difficult to determine. However, lithotype A forms beds which range from ½ cm thick to at least 15 cm thick. These beds are massive and lack sedimentary structures. Cobbold and Whittard (1935), with the availability of better exposure, report four or five beds of lithotype A, from 0.5 m to 4m thick, which appeared to be massive and contained angular shale fragments up to 4 inches in length, which were similar to the interbedded shales.

Lithotype B is finely laminated and constitutes the majority of the exposure.

The Helmeth Grit is considered to be essentially epiclastic s.l. since it is moderately well sorted, subangular to subrounded, is composed of fragments which are identical to those in the Longmyndian sandstones, plots on the trend of increasing plutonic to volcanic components of the Longmyndian Supergroup field (fig. 19) and possesses sedimentary structures which are tentatively interpreted as indicative of deposition from suspension (lithotype B) and possibly subaqueous gravity flows (lithotype A), although the evidence for the mode of deposition of the latter is equivocal.

2.3.5 Conformable base to the Helmeth Grit

The lithologies of the Helmeth Grit and the Ragleth Tuffs are similar, which suggests that there is no unconformity between them. Cobbold and Whittard (1935) thought that the Helmeth Grit is conformable with the Ragleth Tuff Formation. However, Greig et al. (1968) postulated that the base of the Helmeth Grit is unconformable on the Uriconian of the Caer Caradoc area, with which the Ragleth Tuffs were included, such that the Helmeth Grit is generally accepted to form the unconformable base of the Longmyndian Supergroup. The reason for placing an unconformity at the base of the Helmeth Grit is based on two lines of evidence. Firstly, the Helmeth Grit appears to cut across the strike of the Ragleth Tuffs on the south-west slopes of Caer Caradoc (SO 469 946). Secondly, the Helmeth Grit apparently rests on different lithologies on the south-west slopes of Caer Caradoc, these being the Ragleth Tuffs and the Cwms Rhyolites (at SO 469 948).

The Helmeth Grit does not outcrop in the area where the unconformity is proposed. Fragments of sandstone which occur in the soil zone in this area might have been derived from the Ragleth Tuff sandstones, which are lithologically similar. Cobbold and Whittard (1935) did not show the Helmeth Grit to extend to the area of Caer Caradoc. Magnetic evidence detailed by Greig et al. (1968, p.316-317) cannot be used to infer the presence of the Helmeth Grit here, since the anomaly found was not shown to be a unique characteristic of the Helmeth Grit. A faulted contact is inferred, with a NNW-SSE trend similar to the fault trend at the base of the Ragleth Tuff Formation. This fault would explain the juxtaposition of the Stretton Shale Formation against the anomalously easterly-younging Ragleth Tuff Formation in this area (geological map 2, in cover pocket), the absence of exposures of the Helmeth Grit, and would obviate the anomalous north-west swing in strike of the proposed Helmeth Grit in this area. Secondly, the Cwms Rhyolites at SO469948 appear to be a thrust sheet which rests upon the Ragleth Tuffs. Therefore, even if the Helmeth Grit were present, it cannot be demonstrated to be in sedimentary contact with the Cwms Rhyolites.

Elsewhere, the Helmeth Grit displays dips and strikes which are concordant with those in the Ragleth Tuff Formation. Minor discrepancies may be explained by minor folding and rotation of the strike by faults, which appear to be numerous in this area.

It is therefore concluded that the Helmeth Grit is conformable with the Ragleth Tuffs.

2.3.6 The relationship between the Helmeth Grit and the Stretton Shale Formation

The Helmeth Grit was thought to pass conformably into the Stretton Shale Formation by Cobbold and Whittard (1935). This was based on the proposed lithological identity of the shales in the Helmeth Grit with those of the Stretton Shale Formation and it was thought that the "grits" of the Helmeth Grit represented waning volcanic activity which then became absent in the Stretton Shale Formation.

The shales of the Helmeth Grit are distinct from the shales of the Stretton Shale Formation. Although the shales within the Helmeth Grit are similarly thinly laminated, they are medium greenish grey and purplish grey, whereas the Stretton Shale Formation is exclusively light greenish grey. This is significant, because colours of shale in the Long Mynd appear to be stratigraphically related. There is a complete absence of sandstone in the Stretton Shale Formation and the lithologies of the Helmeth Grit are therefore distinct from those of the Stretton Shale Formation. No exposure has been found which juxtaposes the Stretton Shale Formation and the Helmeth Grit when these units are distinguished using the above criteria. Consequently, a conformable relationship cannot be presumed and a faulted contact may be equally likely. Where exposures allow, the Stretton Shale Formation always shows steep dips of 77° to 88° directly west of the junction, whereas the Helmeth Grit mostly shows shallow dips of between 38° and 54° on the west slopes of Ragleth Hill (SO 44859205). At the junction between the Helmeth Grit and the Stretton Shale Formation, igneous rocks were found which are similar to the Uriconian and which are extensively fractured. The western extremity of this exposure appeared to be a fault. The occurrence of igneous rocks at this locality appears to be anomalous and suggests either that intrusion has occurred along a fault zone, or that fragments of the Uriconian Volcanic Complex were brought up along a fault.

It is concluded that the relationships between the Helmeth Grit and the Stretton Shale Formation are obscure. A conformable junction cannot be presumed and a faulted junction appears probable.

2.3.7 The relationship between the Ragleth Tuff Formation and the Uriconian Volcanic Complex.

Greig et al. (1968, p.15) stated: "The Ragleth Tuffs appear to follow the Cwms Rhyolites conformably...". Consequently, the Ragleth Tuffs are accepted to be part of the Uriconian Volcanic Complex and to lie conformably at the top of the succession.

However, the Ragleth Tuff Formation displays faulted contacts with the Uriconian Volcanic Complex in most of the outcrop. Only at one locality has the succession been regarded as conformable (by Greig et al., 1968), this is on the SE facing slopes of Caer Caradoc at SO 474948. Here, the Ragleth Tuff Formation is isolated from the Uriconian Volcanic Complex to the north-east and the main outcrop of the Ragleth Tuff Formation to the south-west by NW-SE trending faults. The Ragleth Tuff Formation has been intruded by dolerite such that only remnants of it remain. These define a north-west plunging syncline. The contact between the Ragleth Tuff Formation and the Cwms Rhyolites is largely obscured by the same dolerite intrusion and elsewhere the contact is poorly exposed. Since the contact is obscured or poorly exposed, no certain conclusion can be reached about its nature and it is concluded that a conformable relationship cannot be presumed from this area.

The junction between the rhyolite and the Ragleth Tuffs is mapped as a line which continues as the junction between the dolerite intrusion and the rhyolite. This would appear to be improbable if the dolerite is a later intrusion into the Ragleth Tuff Formation. Two alternatives are possible: either that the dolerite/rhyolite junction represents the original Ragleth Tuff Formation/rhyolite junction and the dolerite has preferentially intruded the Ragleth Tuff Formation or the junctions are expressions of the same fault.

If the junction did represent the junction of the Ragleth Tuff Formation with the Cwms Rhyolites, then this junction is discordant to the syncline defined by the Ragleth Tuff Formation. The junction appears to follow closely the contours of the hill, in which case it is either vertical and not therefore conformable with the Ragleth Tuff Formation or it is a gently sloping, planar surface which is not concordant with the Ragleth Tuff Formation syncline.

These exposures may be better interpreted as showing a gently sloping sheet of dolerite which, together with the Ragleth Tuff Formation, have been thrust over the Cwms Rhyolites. Similar thrusts are apparent beneath the Caer Caradoc Rhyolites and beneath the Cwms Rhyolites and Caer Caradoc Andesites in the immediate vicinity.

It has been shown that the Ragleth Tuff Formation is composed of epiclastic s.l. deposits which are tentatively interpreted as turbidites. The Uriconian Volcanic Complex is composed of rhyolites, andesites, basalts and welded tuffs which were probably formed subaerially. Consequently, a conformable succession between the Ragleth Tuff Formation and Uriconian Volcanic Complex is unlikely.

2.3.8 Conclusions regarding the stratigraphic position of the Ragleth Tuff Formation and the Helmeth Grit

The Helmeth Grit has been shown to consist of lithologies which are similar to the Ragleth Tuff Formation, but distinct from the Stretton Shale Formation. The Helmeth Grit has been shown to rest conformably on the Ragleth Tuff Formation, but the contacts with the Stretton Shale Formation are not certain and may be faulted. It is therefore proposed that the Helmeth Grit should be included with the Ragleth Tuff Formation rather than with the Stretton Shale Formation and should be classified as a member in that formation.

Both the Helmeth Grit and the Ragleth Tuff Formation in general, have been shown to comprise epiclastic s.l. deposits, which are lithologically distinct from the Uriconian Volcanic Complex, but are similar in character to the Longmyndian Supergroup. It is therefore proposed that the Ragleth Tuff Formation (including the Helmeth Grit Member) should be included with the Longmyndian Supergroup rather than with the Uriconian Volcanic Complex.

The eastern boundary of the Ragleth Tuff Formation has been shown to be faulted against the Uriconian Volcanic Complex. The western boundary of the Ragleth Tuff Formation (the Helmeth Grit Member/Stretton Shale Formation boundary) has been shown to be poorly exposed and possibly faulted and it was concluded that a conformable relationship could not be presumed. The Ragleth Tuff Formation cannot be related stratigraphically to the adjoining horizons on the evidence of its contacts. Other criteria may be employed to determine the stratigraphic position of the Ragleth Tuff Formation.

The Ragleth Tuff Formation is tentatively interpreted as being comprised of deep water deposits, which are turbidites interbedded with shales. Deposits of a similar nature occur in the Stretton Shale Formation and the Burway Formation in the Long Mynd area. Consequently, the Ragleth Tuff Formation is tentatively placed within the Stretton Group of the Longmyndian Supergroup. However, two factors conflict with this. Firstly, both the Helmeth Grit and the Ragleth Tuff Formation in general have modal QFL compositions which plot outside of the field of the majority of the Stretton Group (fig. 19). The modal values are more similar to some values which have been obtained from the Wentnor Group and which have greater amounts of plutonic as opposed to volcanic detritus. It is possible that some stratigraphically higher levels have greater proportions of plutonic detritus due to dissection of the magmatic are source with time. Consequently, the Ragleth Tuff Formation may be stratigraphically higher than the Burway and Stretton Shale Formations. Secondly, it is later argued that the colour of the sediment reflects the environment of deposition, such that reddening is associated with subaerial deposition. Since purplish grey colours are commonly developed in the Ragleth Tuff Formation and only very rarely in the Burway and Stretton Shale Formations, the Ragleth Tuff Formation may come from a stratigraphically higher level.

Although the position of the Ragleth Tuff Formation is uncertain, it is tentatively retained with the Stretton Group, following the accepted practice of placing the Helmeth Grit at the base of the Stretton Group, but the possibility that it is a down-faulted segment of a higher stratigraphic level cannot be discounted.

2.4 The Brokenstones sandstone beds

The term, Brokenstones sandstone beds, is proposed as a name for all predominantly sandstone lithologies which outcrop within the Church Stretton fault system and which are thought to belong to the Wentnor Group. Previous authors have referred these beds to the Wentnor Group. However, since these beds are geographically and structurally isolated from the Wentnor Group in the Long Mynd area, the term Brokenstones sandstone beds is proposed as a better term as this does not assume direct correlation with the Wentnor Group in the Long Mynd area.

The relationships between the Brokenstones sandstone beds and the Stretton Shale Formation in the area between Horderley (SO 408871) and Cwm Head (SO 424887) have been discussed in section 2.2.2. It was concluded that all the contacts are faulted, excepting one locality, where the contact is uncertain (at the Round House, SO 42008816). This differs from the views of previous authors (e.g. Dean, 1964) who thought that the Brokenstones sandstone beds rested unconformably on the Stretton Shale Formation. Other contacts with the Uriconian and with the Ragleth Tuff Formation which occur outside of the Horderley-Cwm Head area are faulted. Consequently, all the contacts with the Uriconian Volcanic Complex and the Longmyndian Supergroup are faulted.

The Brokenstones sandstone beds are thought to be overlain unconformably by the Hoar Edge Grit of Caradoc age (Dean, 1964). Greig et al. (1968) maintained that the Lower Cambrian Wrekin Quartzite overlay the Brokenstones sandstone beds unconformably in the Cwms area (SO 475942). However, the contact was only revealed in one excavation by Cobbold in 1927 and Cobbold (1927, p.557) states: "... the actual junction was found, at the spot chosen, to be faulted". Cobbold inferred an unconformity because of the differences in strike between the two horizons. Alternatively, rotation might have occurred during or after juxtaposition by faulting. Consequently, since the junction was found to be faulted, no unconformity can be inferred. The only direct evidence for the age of the Brokenstones sandstone beds therefore indicates a pre-Caradoc age only. The Brokenstones sandstone beds are similar lithologically, compositionally and in facies to the Wentnor Group. The modal compositions plot in similar positions to the Wentnor Group on a QFL diagram (fig. 19). Therefore, correlation of the Brokenstones sandstone beds with the Wentnor Group is probable.

2.5 The Willstone Hill conglomerate beds

This name is proposed for conglomerates which outcrop in the area of Willstone Hill and between Hope Bowdler Hill and the Cwms at SO 485944, SO 479942 and SO 476938. These conglomerates have been included with the Uriconian Volcanic Complex and have been placed stratigraphically between rhyolites and andesites (Greig et al., 1968, p.18-19). However, it is proposed that these conglomerates are not part of the Uriconian Volcanic Complex. They show similarities with the Wentnor Group, but because of their structural and geographical isolation from the Longmyndian of the Long Mynd area and because of their stratigraphically uncertain position, the term Willstone Hill conglomerate beds is proposed as a preferred term to Wentnor Group or Uriconian conglomerates.

Greig et al. (1968, p.19 and p.265-266) maintained that the Willstone Hill conglomerate beds are outcrops of one horizon which was folded about two fold axes. One anticlinal fold axis trended NE to SW between the outcrops at SO 485944 and SO 479942. The other synclinal fold axis trended NW to SE, bisecting the curved outcrop in the area of SO 485944 (see geological map 2). However, mapping by the author indicates that the disposition of the dips and the younging-directions in the SO 485944 outcrop area does not lend support to either of the proposed folds. The north-eastern edge of the outcrop at Willstone Hill (SO 485944) is interpreted as a fault. The north-western and western limits of this outcrop are defined by a dolerite intrusion and the south-eastern limit is defined, in part, by the Willstone Hill Thrust. All other contacts of the Willstone Hill conglomerate beds with the Uriconian Volcanic Complex are of uncertain type. If stratum contours are drawn at the contact of the conglomerate with the andesite in the area of Willstone Hill (SO 485944) a south-easterly dipping plane is defined which bears no relationship with the strike of the overlying conglomerate. This plane may therefore be a fault. An unconformity is unlikely because of the high disparity of dip between the conglomerate and the plane. At least three thrusts occur in the immediate vicinity, these being the Willstone Hill Thrust, the Sharpstones Thrust and the Hill End Thrust. The Willstone Hill Thrust forms the south-eastern limit to part of the Willstone Hill conglomerate beds outcrop. It is plausible therefore, that the conglomerate in the area of Willstone Hill (SO 485944) may be a thrust sheet. Unfortunately, younging-directions determined from the Willstone Hill conglomerate beds in this area are not certain and therefore these observations are tentative. Greig et al. (1968) noted disparities in dip between the rhyolites of the Battle Stones (SO 486944) and the conglomerates of Willstone Hill (SO 485944). They therefore proposed local folding or a local unconformity. However this contact is better interpreted as a NNW-SSE trending fault. Near the Gaer Stone (SO 474935) either a faulted or an unconformable junction of the conglomerate with the underlying rhyolite was noted by Greig et al. (1968).

It is concluded that the contacts between the Willstone Hill conglomerate beds and the Uriconian Volcanic Complex are open to interpretation. Disparities of dip and strike between the Uriconian Volcanic Complex and the Willstone Hill conglomerate beds suggest that the Willstone Hill conglomerate beds are either faulted against or unconformable with the Uriconian Volcanic Complex in some areas.

The conglomerates are similar in texture and grain size to the conglomerates of the Bayston-Oakswood Formation. Compositionally they are similar to the Wentnor Group conglomerates. The majority of the plots of the modal compositions of the Willstone Hill conglomerate beds show a paucity of feldspar (fig. 19). However, the analyses are mostly based on the composition of large clasts. The feldspar is therefore contained within the plutonic and volcanic lithic fragments. The average modal composition of the Willstone Hill conglomerate beds is more quartzose than the average modal composition of the Wentnor Group (fig. 18). This difference may reflect a difference in stratigraphic level, with the enrichment of plutonic rock fragments, quartzites and quartz schists, following the greater exposure of the metamorphic and plutonic basement during arc dissection. If the conglomerates were interbedded within the

Uriconian Volcanic Complex, it would be unlikely that they would have rounded clasts and textural, grain size and compositional similarities with the Wentnor Group. For these reasons it is therefore concluded that the Willstone Hill conglomerate beds may be correlated with the Wentnor Group and probably unconformably overlie the Uriconian Volcanic Complex.

2.6 The Linley beds

This name is proposed for a group of greenish grey coloured shales, siltstones and mostly very fine grained sandstones which outcrop in a narrow strip along the western margin of the Wentnor Group and along the eastern margin of the Western Uriconian, from the area of the Knolls (SO 376978) to Linley (SO 343936). These beds have been included with the Western Uriconian (Lapworth and Watts, 1910), with the Wentnor Group (James, 1952a, 1952b and 1956) and more recently with the Portway Formation (Cave *et al*., 1985 and Langford and Lynas, 1985). The name Linley beds is preferred since the stratigraphic position and relationship of these beds are considered to be uncertain. The name is derived from the area of best exposure in the vicinity of Linley (SO 343936).

2.6.1 The lithologies of the Linley beds

Log 54 (fig. 108) illustrates the coarser lithology of the Linley beds, this being sandstone, which is mostly very fine to fine grained, planar laminated and greenish grey in colour. These sandstones form thick homogeneous sequences, with occasional normal grading visible. The finer grained lithologies are shale and siltstone which are either massive or very thinly planar laminated. In the shale and siltstone there are occasional beds from 10cm to 2m thick of very fine to fine grained sandstone, which occasionally display fining-upward trends. Shale and siltstone is the predominant lithology. All the lithologies are greenish grey. However, occasional reddish tints are developed in the siltstone and shale. At Chittol (SO 34969495) is an exposure of highly altered, lapilli tuff with lapilli of altered vesiculate glass. This lithology is unique and is not developed elsewhere in the Linley beds. This appears to be associated with laminated siltstone and shale which may be correlated with the Linley beds. However, these lithologies may consist of very fine pyroclastic material at this locality.

These lithologies are very distinct from the adjacent Bayston-Oakswood Formation, which is invariably composed of homogeneous, red, fine to medium grained, cross-bedded sandstone. Therefore the Linley beds should not be included with the Bayston-Oakswood Formation. There are distinct differences between the Portway Formation and the Linley beds. The Portway Formation is mostly purplish red in colour and only rarely greenish grey. Colour of the sediment is later argued to be a primary feature, related to the environment of deposition. Reddish colouration appears to be associated with a subaerial environment. It is unlikely therefore that the Linley beds have been deposited under the same conditions as the Portway Formation. Fine to medium grained, cross-laminated and cross-bedded sandstones with rip-up clasts and erosional bases are common in the Portway Formation and ripple cross-lamination is commonly developed in the finer lithologies. None of these features are developed in the Linley beds. Consequently, correlation of the Linley beds with the Portway Formation is not favoured. The lithology and facies of the Linley beds show similarities to the upper parts of the Burway Formation and the lower parts of the Synalds Formation and it is therefore possible that the Linley beds may be correlated with either of these.

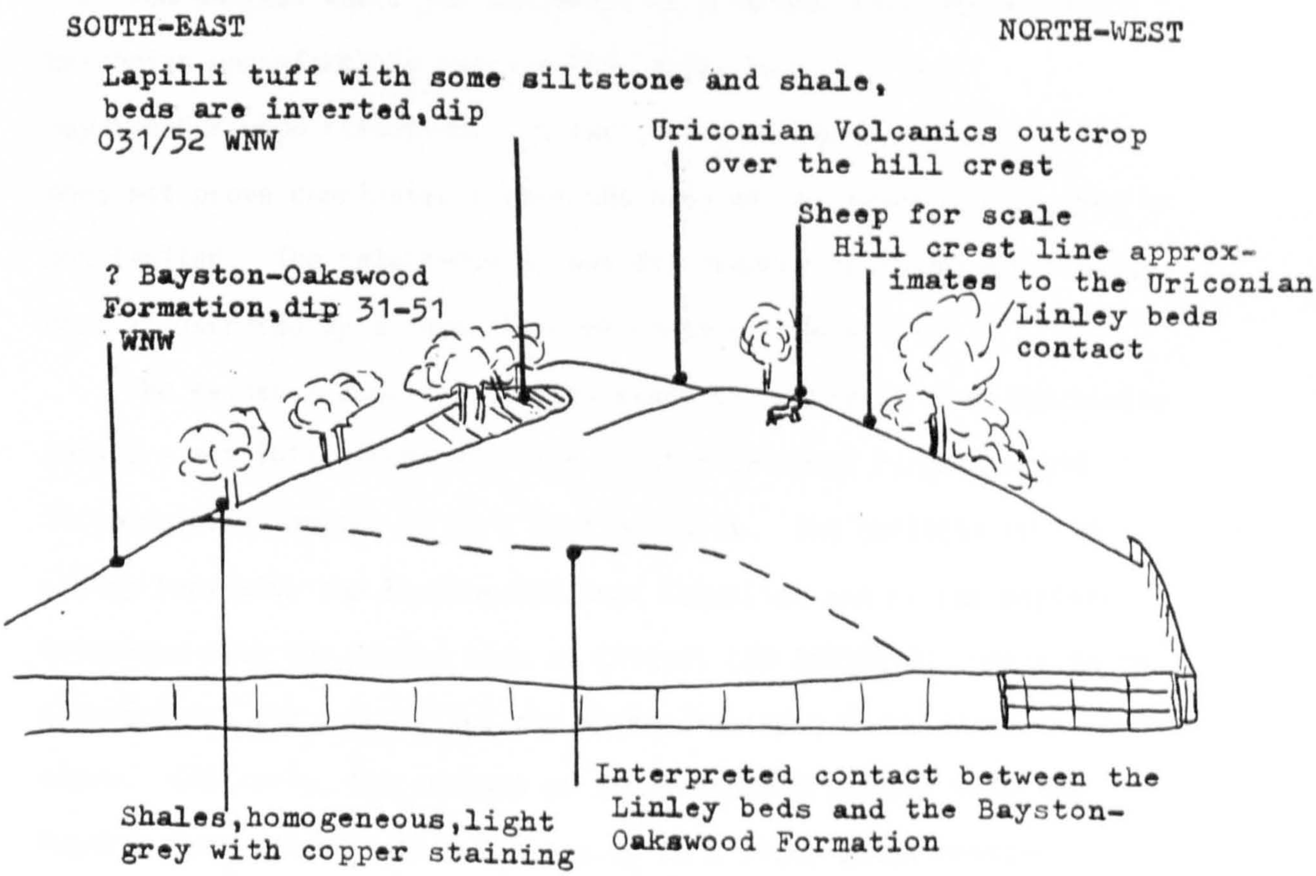
Compositionally, the Linley beds are identical to the Longmyndian Supergroup. Plots of the modal percentages of quartz, lithics and feldspar fall mostly within the Stretton Group field (fig. 19). However, one plot falls close to the field of the Wentnor Group.

2.6.2 The contacts of the Linley beds with the Western Uriconian

James (1952a) proposed that an inverted and unconformable relationship existed between the Linley beds at Chittol (SO 34969495) and the Western Uriconian and since he correlated the Linley beds with the Wentnor Group this was used as a key argument for an unconformable relationship between the Stretton Group and the Wentnor Group.

The Western Uriconian forms the isolated knoll of Chittol (SO 34939477) and is composed of andesite and rhyolite which are thought to be extrusive. The underlying beds are stratigraphically inverted as determined from grain size changes. These beds are composed of lapilli tuffs and light greenish grey siltstone and shale which is either massive or thinly, planar laminated and may be of pyroclastic origin in part. These lithologies are distinct from the Wentnor Group and therefore their inclusion with it, as proposed by James (1952a), is not accepted. Since the siltstone and shale lithologies are similar to those of the Linley beds, the Chittol beds are tentatively correlated with the Linley beds. However,

FIG. 5 THE RELATIONSHIPS BETWEEN THE WESTERN URICONIAN, THE ? LINLEY BEDS AND THE BAYSTON -OAKSWOOD FORMATION AT CHITTOL KNOLL (SO 34939477)-PHOTOGRAPH AND FIELD SKETCH.



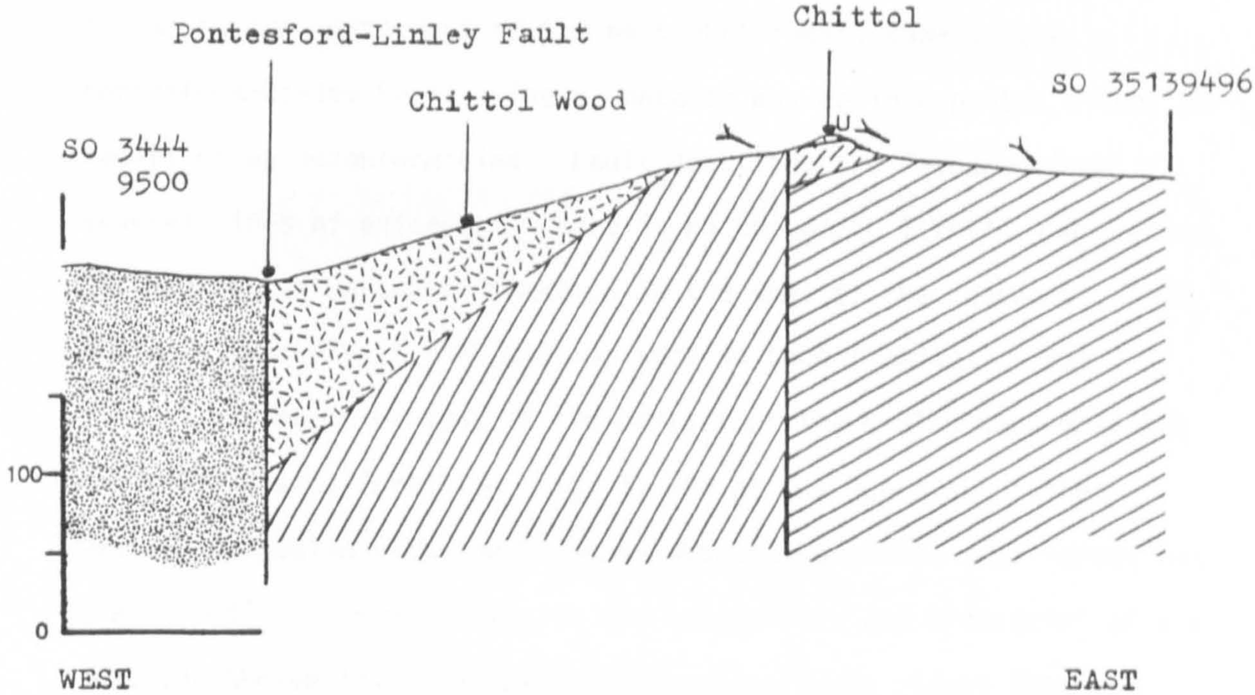
pyroclastic deposits are not known elsewhere in the Linley beds. Alternatively, these beds may be correlated with some pyroclastic units of the Uriconian, such as the Earl's Hill tuff beds.

The Bayston-Oakswood Formation is not exposed at this locality, but is exposed nearby at SO 34839480. The Bayston-Oakswood Formation is interpreted to form the low ground below the break in slope of the Chittol knoll. Unfortunately, the contacts between the Linley beds and the Western Uriconian and between the Linley beds and the Bayston-Oakswood Formation are not exposed. The field relationships are illustrated by fig. 5 and are shown on map 2 (in cover pocket).

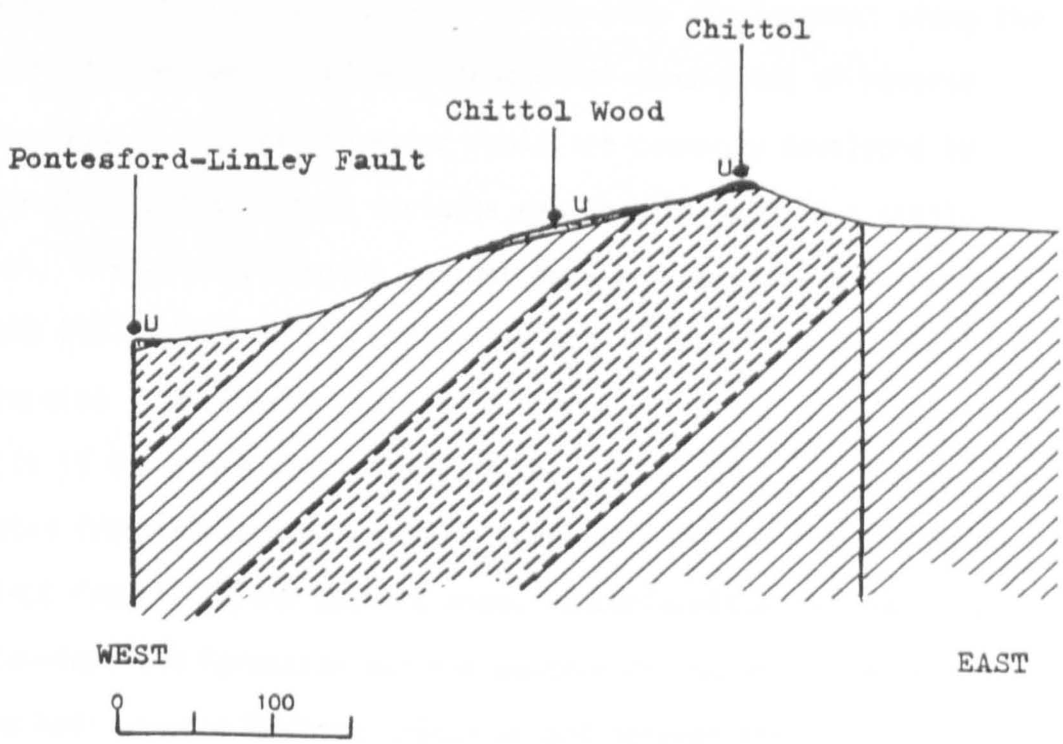
The contacts of the Western Uriconian with the Linley beds are obscured. However, James (1952a) thought that the Linley beds were interbedded with the Bayston-Oakswood Formation and from combined evidence from the areas of Chittol Wood (in the area of SO 343949) and from Chittol knoll (SO 34939477) he proposed that the Western Uriconian unconformably overlay the Linley beds and the Bayston-Oakswood Formation. In fact, the evidence of James (1952a) does not prove conclusively that the base of the Western Uriconian is not faulted. The relationships and differences in interpretation are best illustrated by a comparison of cross-sections (fig. 6).

The evidence acquired by this study demonstrates that the Linley beds are not interbedded with the Bayston-Oakswood Formation and lithologically appear to be a distinct unit. The contacts of the Linley beds with the Bayston-Oakswood Formation and of the Western Uriconian with the Linley beds at Chittol (SO 34939477) appear to be planes which are subparallel and which cross-cut the bedding in both cases. Similarly, the contact of the Western Uriconian with the Bayston-Oakswood Formation appears to be a plane which cross-cuts the bedding of the underlying Bayston-Oakswood Formation in the area of

FIG. 6 CROSS-SECTIONS THROUGH THE URICONIAN AND BAYSTON-OAKSWOOD FORMATION CONTACT IN THE CHITTOL AREA.



SECTION 1 SCALE 1:5000,1cm=50m,no vertical exaggeration



SECTION 2 SCALE 1:5069,1cm=51m,no vertical exaggeration
(After James,1952b,fig. 1)

KEY:

Bayston-Oakswood Formation

Linley beds = "green shaly tuffs" of James 1952b

Younging direction

Western Uriconian (U)

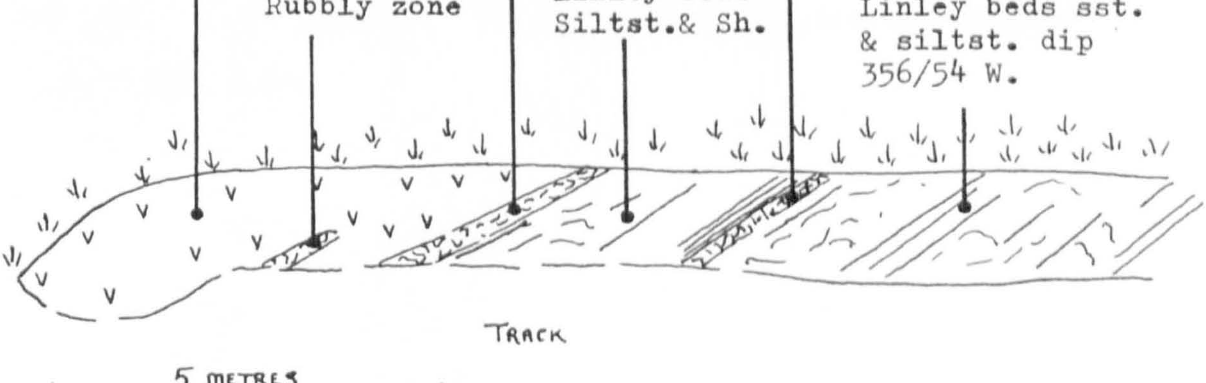
Habberley Shale Formation (Ordovician)

Chittol Wood (SO 343949). All the contacts dip at low angles of up to 35° to the north-west or the west-north-west, towards the Pontesford-Linley Fault. These contacts may be interpreted either as faults or as unconformities. Fault junctions are favoured from several lines of evidence. There is no change in lithology or facies of the Bayston-Oakswood Formation in the area of the contacts. This would be unlikely if the junction were an unconformity. Copper mineralisation is evident at the contact of the Linley beds with the Bayston-Oakswood Formation (fig. 5) and at the junction of the Western Uriconian with the Bayston-Oakswood Formation (the latter has been mined). Mineralisation in the Longmyndian and Ordovician of the adjacent Shelve inlier usually occurs along fault planes (Dines, 1958). The contacts plot as flat planes which are subparallel. Woodcock (1984b) has demonstrated strike-slip displacement along the Pontesford Lineament. "Flower structures" consisting of reverse faults dipping towards the major fault are commonly developed by strike-slip faults. These contacts may be explained in a similar manner. A similarly dipping contact between the Western Uriconian and the Linley beds in the Linley Big Wood area (at SO 34219357) is interpreted as a reverse fault.

It is therefore concluded that the Linley beds are structurally isolated from the Bayston-Oakswood Formation and are lithologically distinct from it in the Chittol area. Unconformities between the Bayston-Oakswood Formation and the Western Uriconian, between the Linley beds and the Western Uriconian and between the Bayston-Oakswood Formation and the Linley beds cannot be adequately demonstrated. Alternatively, all of these contacts may be explained by faulting, for which there is some evidence. Consequently, the conclusions of James (1952a) that at Chittol the Wentnor Group

FIG: 7 THE CONTACT BETWEEN THE WESTERN URICONIAN AND THE LINLEY BEDS NEAR LINLEY BIG WOOD AT SO 34219357. PHOTOGRAPH AND FIELD SKETCH.

PHOTOGRAPH shows the Uriconian/Linley beds contact from the section below. Scale bar is 30cm long. Massive rhyolites in the top left corner of the photo overlie a 30cm thick rubbly zone.



unconformably overlies the Western Uriconian, together with the implicit removal of the Stretton Group beneath a Wentnor Group unconformity are not accepted.

The contact between the Linley beds and the Western Uriconian is exposed along the banks of a track in the Linley Big Wood area at SO 34219357. Rhyolites are exposed to the west of the contact and typical lithologies of the Linley beds to the east. These are sandstones which are very fine to fine grained, with minor siltstone and shale. All the lithologies are greenish grey and thinly to very thinly planar laminated. The contact between the volcanics and the Linley beds consists of a rubbly zone, 30cm thick, which is subparallel to the bedding of the Linley beds. Similar, subparallel, rubbly zones approximately 30cm thick are developed within the Linley beds and within the Western Uriconian (fig. 7). These rubbly zones are interpreted as fault breccias.

An exposed contact between the Western Uriconian and the Linley beds has not been seen elsewhere. However, Cave et al. (1985) report an excavated junction between pale mudstones and the Uriconian in the Linley Big Wood area at SO 34309393. This junction is interpreted as a fault, with a series of anastomosing faults being developed, with crush zones up to 160mm wide in the sediments. The Western Uriconian/Linley beds junction was excavated by James in the area of the Knolls at SO 36589679. The following succession was noted by James (1956, p.326):-

Uriconian	2ft 10 inches
Soft green shale, containing rounded Uriconian fragments	0ft 6 inches
Hard green shale	0ft 6 inches
Green shale with cherty tuffaceous bands	0ft 9 inches

From this exposure, an inverted unconformity appears to have been inferred. However, no evidence of way-up could be found in the area of the excavation. The rounded pebble of Uriconian was found within the shale three inches from the contact. This occurrence is anomalous, in that unique sedimentation conditions would have been required to isolate a rounded fragment of Uriconian within shale. The occurrence of soft shale with Uriconian fragments is strongly suggestive of a fault gouge. James reports the occurrence of a faulted Uriconian/Linley beds contact in the vicinity of the above excavation "... two thousand feet to the west of Squilver Farm." James (1952b, p. 63). Dines (1958) reports a faulted contact in mines west of Squilver Farm, in the Knolls area, which may be the same contact noted by James (1952b) and he states: "At the Knolls, an adit with portal 600 yd west of Squilver Farm was driven 48 yards WNW in 1943... . About half way along its course, the adit passes through a well-marked fault plane with a westerly hade. At the mouth of the adit, the country rock is Longmyndian conglomerate, but beyond the fault, the strata appear to be older and may be of Uriconian age." Dines (1958, p.40). No other exposures of the Linley beds/Western Uriconian contact are known.

It is concluded that an unconformity between the Linley beds and the Western Uriconian cannot be demonstrated. The available evidence indicates that the junction is faulted. All of the fault planes dip to the west or the north-west and are therefore reverse faults and thrusts. Inversion of the Western Uriconian has been inferred from inversion of the Linley beds (James, 1952a, 1952b and 1956). However the inversion of the Western Uriconian in these areas has not been specifically demonstrated and the Western Uriconian may not in fact be inverted.

2.6.3 The contacts of the Linley beds with the Bayston-Oakswood Formation

The Linley beds are in contact with the Bayston-Oakswood Formation from the area of the Knolls (SO 376978) to Linley (SO 343936) and except for the area of Chittol Wood (SO 343949), the Linley beds separate the Bayston-Oakswood Formation from the Western Uriconian in this area. James (1952a, 1952b, 1956) proposed that the Linley beds are interbedded with and are a conformable part of the Bayston-Oakswood Formation. However, it has been shown that the Linley beds are a distinct lithological unit which cannot be demonstrated to be interbedded with the Bayston-Oakswood Formation. The area of the contact is poorly exposed. In the majority of cases, the junctions appear to be subparallel to the bedding and would thus appear to be conformable. Consequently, the Linley beds have been placed stratigraphically beneath the Bayston-Oakswood Formation and were correlated with the Portway Formation by Cave *et al*. (1985) and Langford and Lynas (1985). However, it has been argued that the Linley beds may not be correlated with the Portway Formation, on lithological and facies grounds,and alternatively may be correlated with the top of the Burway Formation, or with the similar base of the Synalds Formation. The presence of a conformable junction is therefore questioned.

It has been shown that the contact of the Bayston-Oakswood Formation with the Linley beds in the area of Chittol knoll (SO 34939477) cross-cuts the bedding of both horizons and appears to be a mineralised fault with a dip of 22° to the north-west (fig. 6 and section 2.6.2). Faulted contacts which are subparallel to bedding and which dip in a westerly to north-westerly direction can be demonstrated in this area (the Uriconian/Linley beds contact in the Linley Big Wood area, SO 34219357). There is no change in lithology or of facies of either the Linley beds or of the Bayston-Oakswood Formation towards the contact. Cave et al. (1985) propose that there is a transition from the Linley beds to the Bayston-Oakswood Formation in the Linley Big Wood area (SO 345942). However, exposures indicate a rapid alternation of the distinct lithologies along strike, which may be best explained by a number of small ENE-WSW trending dip faults, since no interbedding of or gradation between the lithologies was noted. Near Cold Hill Farm (SO 36139627) a faulted junction is exposed between the Linley beds and the conglomerates of the Bayston-Oakswood Formation. However, this fault junction is oblique to the normal Linley beds/Bayston-Oakswood Formation contact and thus appears to be a cross-fault.

It is concluded that the junctions between the Linley beds and the Bayston-Oakswood Formation are of uncertain type, but may be faults, for which there is some evidence. Similar orientations of these faults with those at the base of the Western Uriconian suggest reverse faulting or thrusting, with planes dipping towards the Pontesford-Linley fault, reminiscent of "flower-structures" developed by strike-slip faults. If the contacts are reverse faults, then this may lend support to the correlation of the Linley beds with a stratigraphically lower horizon, such as the Burway Formation, or the lower parts of the Synalds Formation.

2.7 The contacts of the Bayston-Oakswood Formation with the Western Uriconian

The Linley beds isolate the Bayston-Oakswood Formation from the Western Uriconian, from the area of the Knolls (SO 376978) to Linley (SO 343936). However, the Bayston-Oakswood Formation is found in contact with the Western Uriconian in the area of Chittol Wood (SO 343949) and in the area of the Lyd Hole at SJ 41500553. The former contact has already been discussed in section 2.6.2. A faulted contact is proposed as the favourable interpretation and the unconformable contact proposed by James (1952a, 1952b and 1956) cannot be demonstrated and appears to be unlikely.

At the Lyd Hole (SJ 41500553), the Western Uriconian/Bayston-Oakswood Formation boundary has been interpreted as an unconformity (Whittard, 1952) or a fault (Dean and Dineley, 1961). Pocock et al. (1938) noted some crushed rhyolite at the contact, but thought that movement had occurred within the rhyolites and that "...there may not be any serious displacement between these and the Longmyndian" (Pocock et al., 1938, p.35). This view was supported by the finding of uncrushed rocks on either side of the contact, within a few inches of one another, at one point. They noted the presence of volcanic material in the Bayston-Oakswood Formation, which was almost identical with the Uriconian rocks of the Lyd Hole. Although it was not stated, it appears therefore that they favoured an unconformable junction. Blake (1890) proposed that the Uriconian Volcanics of the Lyd Hole were intrusive. However, the lithologies of the Uriconian do not indicate that they are intrusive (Pocock et al., 1938).

Several lines of evidence suggest that the Uriconian/Bayston-Oakswood Formation contact is faulted rather than unconformable. There is no change in the lithological character or in the facies of the Bayston-Oakswood Formation as the contact is approached. This would be unlikely if the contact were an unconformity. Similarities between the volcanic fragments, of which the Bayston-Oakswood Formation is composed at this locality, and the volcanics of the Lyd Hole is not a criterion by which an unconformity can be implied, since the Western Uriconian is lithologically similar to the Eastern Uriconian, from which the fragments could have been derived, and derivation from the Western Uriconian does not preclude a later faulted junction. The junction is vertical, straight and planar and parallel to bedding in the adjacent Bayston-Oakswood Formation. Directly at the contact, 40cm of brecciated siltstone and silty shale, which appears to be homogeneous, is developed. This is interpreted as a fault gouge. The Uriconian of the Lyd Hole is highly faulted, but none of the faults were observed to pass through the Bayston-Oakswood Formation/Uriconian contact. The contact is subparallel to the Pontesford-Linley fault and appears similar to other subvertical strike faults with associated fault gouges developed elsewhere in the Longmyndian (e.g. at Haughmond quarry, SJ 54251455).

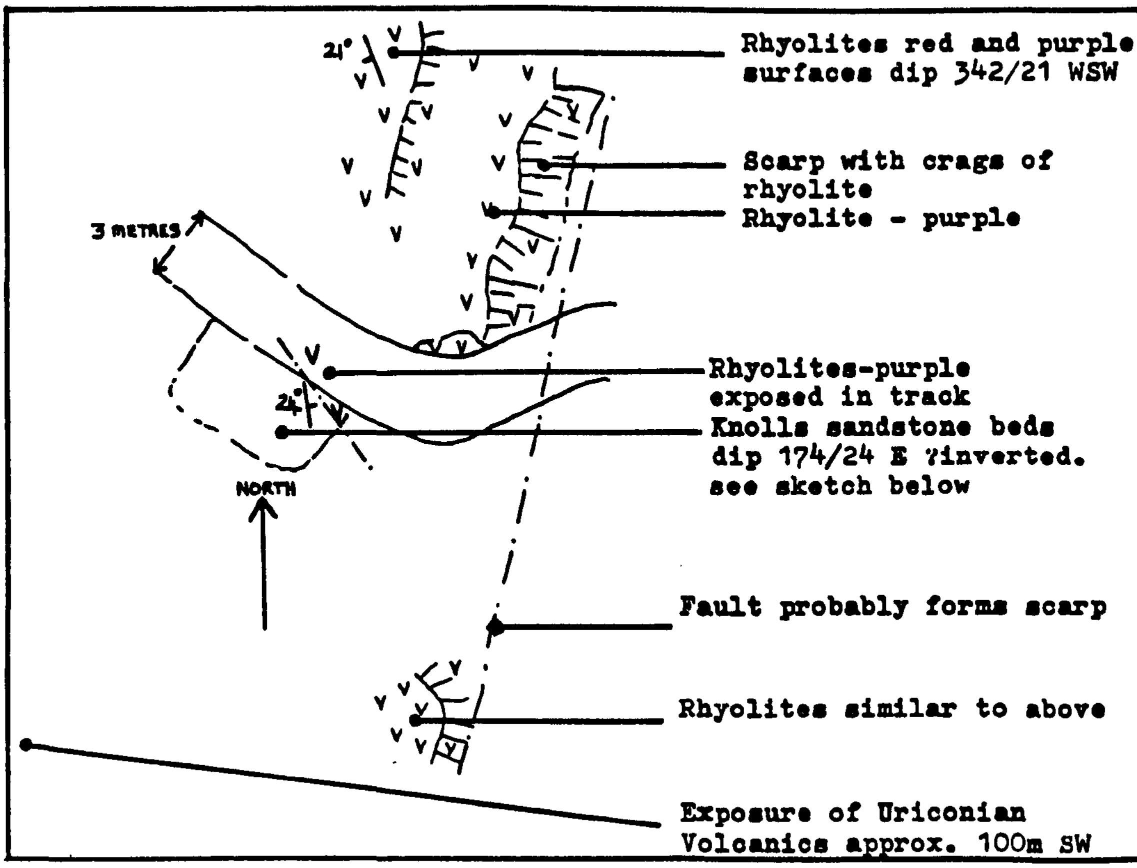
It is therefore concluded that the junction of the Bayston-Oakswood Formation with the Western Uriconian at Lyd Hole is faulted. Since the contact at Chittol Wood (SO 343949) is interpreted as a fault, there are no known exposures where an unconformable relationship between the Bayston-Oakswood Formation and the Western Uriconian can be demonstrated and therefore the proposed removal of the Stretton Group due to overstepping of an unconformable Wentnor Group onto the Western Uriconian (James, 1952a, 1952b, 1956) is not accepted.

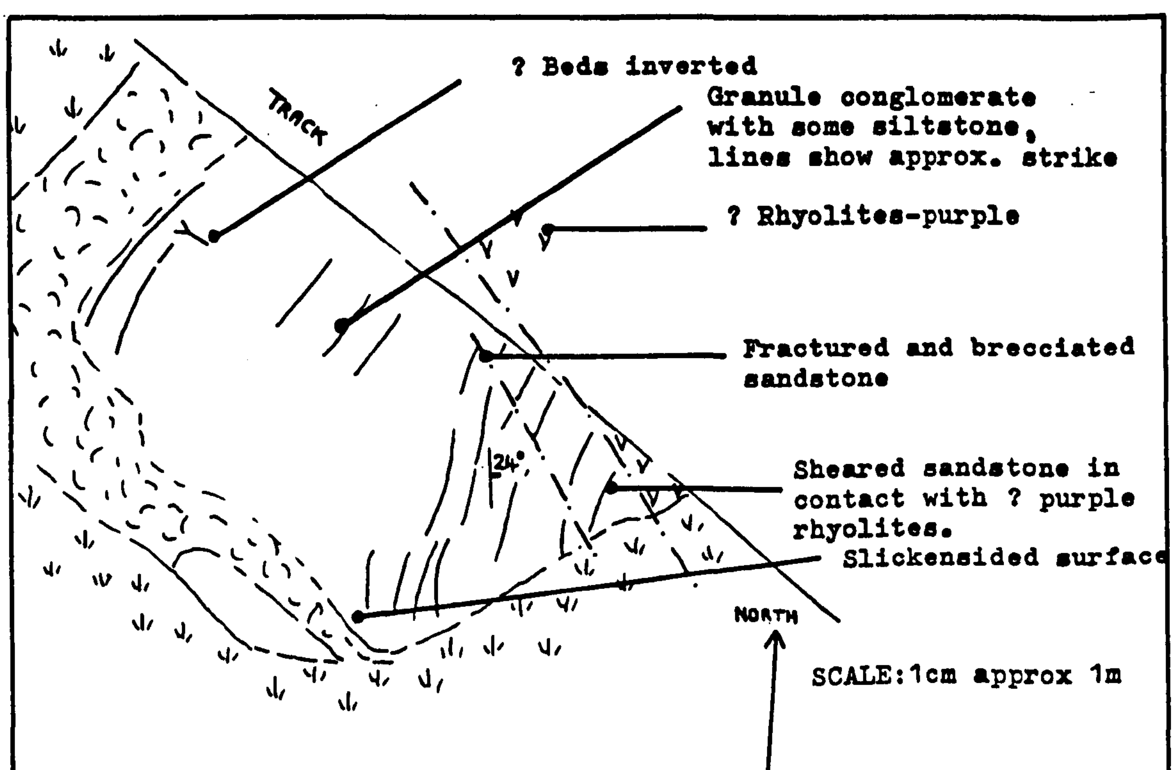
2.8 The Knolls sandstone beds

This name is proposed for sandstones which outcrop within the outcrop of the Western Uriconian near Linley. The name is derived from the area where the exposures occur, near the Knolls, directly south of Linley Big Wood. The best exposure occurs in a track at SO 34489352 which was first noted by R. Langford of the BGS (Cave, Lynas and Langford, 1985). These sediments have a relatively high magnetic susceptibility, unlike the Longmyndian Supergroup and were thus correlated with the Western Uriconian (Cave, Lynas and Langford, 1985). Similar magnetic sediments were found at SO 33779335 and SO 34009360. The latter showed signs of silicification and strain (Cave et al., 1985). The exposure at SO 34489352 was excavated by Langford and the BGS and this revealed a contact between the sediments and a basalt. The contact appeared to be transgressive and the basalt was therefore considered to be probably intrusive (Cave et al., 1985).

The field relationships are illustrated by fig. 8. At the time of examination, the previous trench appeared to have been infilled. The sediments are surrounded by the Uriconian volcanics on the north-east, south-west and south-east sides. No exposures were found to the north-west of the exposed sediments. One contact was exposed at the north-east side of the outcrop, between the sediment and the Uriconian volcanics. Here, the sediments displayed evidence of

FIG: 8 RELATIONSHIPS BETWEEN THE KNOLLS SANDSTONE BEDS AND THE WESTERN URICONIAN AT SO 34489352-FIELD SKETCHES.





shearing and brecciation and a faulted contact is therefore inferred. Slickensided surfaces were noted elsewhere in the sediments, which suggests some dislocation. The sediments are compositionally similar to the Longmyndian Supergroup, in that they are composed of predominantly glassy volcanic rock fragments with minor quartz and feldspar. Plots of these components on a QFL diagram fall within the field of the Longmyndian Supergroup and the compositions are not dissimilar to some of the more lithic sandstones of the Wentnor Group (fig. 19). However, some differences were noted between these sediments and those of the Longmyndian Supergroup, these being: the poorly sorted nature of the sediment, the rapid variations in grain size, from granule conglomerate to siltstone,and the immature nature of some of the lithic components. These latter include large fragments of reticulate, vesiculate glass and vesiculate glass with "pipe-vesicles". Since the composition of the sandstones is similar to the Longmyndian Supergroup, the cause of the anomalously high magnetic susceptibility (Cave et al., 1985) is uncertain. Patches of copper mineralisation were noted in places.

It is concluded that because of the poorly sorted and immature nature of the sediment, the Knolls sandstone beds might have been deposited in close proximity to the Uriconian volcanics, possibly within or on them. The lack of plutonic rock fragments suggests that these sediments were probably deposited prior to the dissection of the volcanic complex and therefore are probably not post-orogenic. These sediments may be viewed as representatives of the Longmyndian Supergroup, deposited in a more proximal setting with respect to the volcanic source. However, the actual junctions of the Knolls sandstone beds with the Western Uriconian are uncertain and are faulted in part.

2.9 The relationships between the Longmyndian Supergroup and the Uriconian Volcanic Complex

James (1952b) argued for a conformable base to the Stretton Shale Formation on the Eastern Uriconian, together with an unconformable base to the Wentnor Group on the Western Uriconian. Implicitly, the Western Uriconian remained as a source for the Longmyndian Supergroup sediments. The Uriconian Volcanic Complex was envisaged as being incorporated with the major syncline, into which the Longmyndian Supergroup was folded.

It has been proposed that the Ragleth Tuff Formation is part of the Longmyndian Supergroup and has faulted contacts with the Eastern Uriconian. All the other contacts of the Stretton Group with the Eastern Uriconian appear to be faulted. Similarly, all the contacts of the Longmyndian Supergroup, specifically the Linley beds and the Bayston-Oakswood Formation, with the Western Uriconian are interpreted as faults. Consequently, the relationship between the main outcrop of the Longmyndian Supergroup and the Uriconian Volcanic Complex cannot be demonstrated by superposition. Probable Wentnor Group, the Knolls sandstone beds and the Willstone Hill conglomerate beds, may rest unconformably on the Western and Eastern Uriconian volcanics respectively. It is therefore possible that the Wentnor Group oversteps onto the Uriconian Volcanic Complex at the basin margins.

The provenance of the Longmyndian Supergroup is interpreted as a magmatic arc with lithologies similar to the Uriconian Volcanic Complex. Consequently, it is concluded that the Uriconian Volcanic Complex acted as a source for the Longmyndian Supergroup. Since the lithological components of the sediments suggest that the magmatic arc provenance was largely undissected and since pyroclastic deposits are common in the Longmyndian Supergroup it is concluded that deposition of the Longmyndian Supergroup was essentially synchronous with the formation of the Uriconian Volcanic Complex. The lack of facies which are indicative of proximity to a volcanic source within the Longmyndian Supergroup and the fact that the palaeocurrent data does not indicate immediate derivation from the Uriconian Volcanic Complex in its present geographical positions suggests that the Longmyndian Supergroup has been tectonically juxtaposed with the Uriconian Volcanic Complex.

2.10 The stratigraphic position of the Rushton Schist and Primrose Hill Gneiss and Schist

Both the Rushton Schists and the Primrose Hill metamorphic rocks are unconformably overlain by the Wrekin Quartzite (Pocock *et al.*, 1938). A Precambrian age is therefore demonstrated. Pocock *et al.* (1938) concluded that the Rushton Schist may be "... Eastern Longmyndian rocks in a sheared and hornfelsed condition." (Pocock *et al.*, 1938, p.31). Similarly, the Primrose Hill rocks were concluded to have been "... produced by the injection of granophyre into a group of tuffs and lavas, in conjunction with moderately intense dynamic metamorphism." (Pocock *et al.*, 1938, p.24). Greig *et al.* (1968) concluded that the Rushton Schists are low-grade regional metamorphic rocks and noted that a metamorphic suite of fragments in the Longmyndian are similar to the Rushton Schists. Thorpe (1974) demonstrated that there are chemical dissimilarities between the Uriconian volcanics and the Primrose Hill granite and gneiss. The latter appeared similar to a hybrid igneous rock in composition. It was concluded that the Primrose Hill rocks are probably pre-Uriconian in age.

The Rushton Schists are varied in character. In one example, a quartz-muscovite-biotite-chlorite-garnet schist shows poikiloblastic and idioblastic garnets. A secondary crenulation cleavage, together with alteration of biotite to chlorite, is developed. In the Malvern igneous complex, garnetiferous rocks are found in a small area of less than 0.1 percent of the complex. These are biotite-muscovite-garnet schists carrying garnets retrogressed to chlorite. Other garnetiferous rocks, including biotite-plagioclase-quartz-garnet schist are found as inclusions within diorite (Lambert and Holland, 1971). They concluded that these specimens "... suggest the existence of a former greywacke-type metamorphic basement into which the Malvernian diorites were intruded." (Lambert and Holland, 1971, p.345). Garnet is found as an accessory mineral throughout the Longmyndian Supergroup, together with schistose quartzites, quartz-muscovite schists and quartz-muscovite-chlorite schists. Thorpe _et al_. (1984) reported a Rb-Sr whole-rock isochron age of 677$\pm$20 Ma for the Rushton Schists and suggested that the Sr ratios indicated derivation from sedimentary rocks as old as c.950 m.y. The Rushton Schists are intruded by a felsite dyke which is considered to be of late Uriconian age. The derived fragments in the Longmyndian Supergroup demonstrate the existence of a metamorphic terrain of which the Rushton Schists and the Primrose Hill rocks are probably a selvage.

The Primrose Hill rocks have characters which are analogous to the metamorphosed igneous rocks of the Malverns, which have been produced by shearing and hydrolysis (Lambert and Holland, 1971). One example of the varied rock suite is a plagioclase-quartz-biotite-chlorite-muscovite-garnet gneiss which is of a similar metamorphic grade to the Rushton Schists and which shows evidence for a secondary schistosity together with retrograde metamorphism similar to the Rushton Schists. The Primrose Hill rocks include what appear to be intrusive granophyres, which possess cataclastic textures only and which include xenoliths of metamorphic rock, which contain abundant garnet and chlorite (from Pocock <u>et al</u>., 1938, p.23). These acidic rocks may be analogous to the acidic intrusions in the Rushton Schist and the late-Uriconian Ercall granophyre. These Primrose Hill metamorphic rocks therefore appear to be analogous to the Rushton Schists.

The metamorphic rocks of Shropshire are of a similar grade to the metamorphic rocks of the Malvernian igneous complex and they occur in a similar structural setting. It is interesting to note that both metamorphic terrains are associated with shearing and occur at the southern ends of the volcanic complexes.

2.11 The absolute ages of the Precambrian rocks of Shropshire

The metamorphic rocks are bracketed in age by the dates of 667±20 Ma, for metamorphism of the Rushton Schists, and maximum model ages of Tchur, 1300 Ma and Tmorb, 1600 Ma for the Rushton Schists (Thorpe et al., 1984). The magmatic arc is dated as 700 to 643 Ma (Thorpe et al., 1984). These dates bracket the age of formation of the Stanner-Hanter Complex, the Malvernian Complex and the Johnston Complex. Only one Rb-Sr age date of 558±16 Ma has been obtained from the Uriconian Volcanic Complex by Patchett et al. (1980). However, this age was determined from "felsic tuffs" which were thought to have been "... subjected to a variable amount of sorting by water at the time of deposition." (Patchett et al., 1980, p.650). These "tuffs" therefore might not be part of the Uriconian Volcanic Complex, but alternatively might be representatives of the Ragleth

Tuff Formation of the Longmyndian Supergroup. This age date therefore might not be representative of the Uriconian Volcanic Complex.

The deposition of the Longmyndian Supergroup is considered to be essentially synchronous with the evolution of the magmatic arc. Consequently, the age of the Longmyndian Supergroup is considered to be in the region of 700 Ma to 643 Ma and is probably closer to the latter date. The oldest dates obtained from the Longmyndian are within a narrow range of 526±18 Ma to 529±23 Ma derived by Rb-Sr whole-rock analyses (Bath, 1974) and fission-track analyses (Naeser _et al_., 1982). Both Bath (1974) and Naeser _et al_. (1982) considered these dates to be reset ages reflecting dewatering due to burial or folding (Bath, 1974) or uplift (Naeser _et al_., 1982). Younger dates are reported by both authors of 325±7 Ma for the Uriconian (Naeser _et al_., 1982), 420±9 Ma for the Longmyndian Supergroup (Naeser _et al_., 1982) and 452±31 Ma for the Longmyndian Supergroup (Bath, 1974). Harper (1966) also reported a single K-Ar whole-rock age of 470 Ma from the Longmyndian Supergroup.

The Ercall granophyre has been dated as 533±12 Ma by Patchett _et al_. (1980). This date is very similar to the reset ages of the Longmyndian Supergroup of c.526 Ma and consequently may be a reset date. However, the age of the Ercall granophyre is similar to magmatic ages of 540±57 for the Charnwood south diorites (Cribb, 1975) and 549±19 for the Sarn complex (Beckinsale _et al_., 1984). The contact of the Ercall granophyre with the Wrekin Quartzite is uncertain and may be faulted or intrusive (cf. Thorpe _et al_., 1984). Consequently, the Ercall granophyre may or may not be pre-Late Tommotian.

The absence of an Ediacarian fauna despite the wide variety of marine and marginal marine environments suggests that the Longmyndian Supergroup is older than the Ediacarian period, the base of which is dated as approximately 650-660 Ma. (Glaessner, 1984a). Glaessner (1984b) concluded that the Ediacarian fauna required no special conditions of environment or preservation. Consequently, their absence in the Longmyndian Supergroup may be a reflection of its age rather than unfavourable conditions. This conclusion is supported by the finding of an Ediacarian fauna in the Charnian and in the Carmarthen area (Cope, 1977). These strata are similarly volcaniclastic and a similar setting to the Longmyndian is envisaged for the Charnian. Moseley and Ford (1985) conclude that the Charnian was deposited in a basin adjacent to a contemporaneous, explosive volcanic arc on its north-west margin. Such a setting is analogous to the relationship between the Longmyndian Supergroup and the Uriconian Volcanic Complex. Ford (1980) suggested that the Charnian fauna represents the period from 552 to 684 Ma. Glaessner (1984b) suggests that the Charnian sediments may be between 620 and 580 Ma, which would agree with the dating of other occurrences of Charnia and Charniodiscus. All of these dates agree with the hypothesis that the Longmyndian sediments were deposited in the region of 700 to 643 Ma.

The geological history, derived from radiometric age dating, may be summarised as follows: A magmatic arc developed at c.700 Ma. Shortly after, there was some metamorphism of sediments and the early volcanics (e.g. Rushton Schists, parts of the Primrose Hill complex and parts of the Malvernian igneous complex). Both may be related to initiation of a SE dipping subduction zone (e.g. Thorpe et al., 1984). Volcanic activity of this magmatic arc continued to c.643 Ma, during which time the Longmyndian Supergroup was deposited. Volcanic activity may have continued up to ?c.580 Ma in some areas, together with sedimentation in adjacent basins e.g. Charnian. The Longmyndian Supergroup was uplifted at c.526 Ma and this was accompanied either by late acidic intrusions (e.g. Ercall granophyre) or by resetting of some of the age dates of the Uriconian Volcanic Complex. Further deformation of the Longmyndian occurred at c.450 Ma. Using the geochronostratic scale of the Precambrian-Cambrian transition provided by Glaessner (1984a, p.140), the Uriconian Volcanic Complex and the Longmyndian Supergroup are probably Vendian and the Longmyndian Supergroup is probably Varangerian to early Ediacarian.

2.12 Subdivision of the Synalds Formation

The Synalds Formation has been subdivided into a lower Green Synalds Member, 143m thick, and an upper Red Synalds Member, 522m thick. These members are mappable units and the boundaries between them appear to be concordant with the bedding.

2.12.1 The Green Synalds Member

The base of the member is in the transition from the Cardingmill Grit Member of the Burway Formation and is defined as the change from the predominantly thick and homogeneous sandstone of the Cardingmill Grit to the predominantly siltstone succession of the Green Synalds Member. The top of the member appears to be well defined as the rapid transition from the predominantly green siltstones of the Green Synalds Member to the predominantly red siltstones of the Red Synalds Member. The type section is exposed in Ashes Hollow between SO 43259305 and SO 43179319. The lithology of the member mainly consists of thinly laminated siltstones and medium to thick-bedded, very fine grained sandstones. Towards the base, very thick-bedded, very fine to fine grained sandstones are developed which are up to 5m thick. The interbedded siltstones are commonly purplish red towards the base. Wavy, non-parallel laminated siltstones are common at the top of the member. Very thinly laminated, medium to dark grey siltstones are occasionally developed. The transition to the Red Synalds Member is usually accompanied by colour mottling.

2.12.2 The Red Synalds Member

The base of the member is defined above. The top of the member is defined as the change from the predominantly red siltstones of the Red Synalds Member to the predominantly green siltstones of the basal Lightspout Formation, the Green Lightspout Member. The type section is exposed in Ashes Hollow between SO 43179319 and SO 42759387. The lithology of the member mainly consists of laminated, purplish red siltstones with minor, medium-bedded, very fine grained sandstones. Rarely, very thick-bedded sandstones are developed up to 3m thick. Occasional, pale green colour-mottling is developed. The upward transition to the Lightspout Formation is usually accompanied by a slight increase in grain size and folding occurs directly at and above the transition.

2.13 Subdivision of the Lightspout Formation

The Lightspout Formation has been subdivided into a lower Green Lightspout Member, 220m thick and an upper Red Lightspout Member, 112m thick. These members are mappable units and the boundaries between them appear to be concordant with the bedding.

2.13.1 The Green Lightspout Member

The base of the member is defined in section 2.12.2 as the top of the Red Synalds Member. The top of the member is defined as the rapid transition from green siltstones and sandstones of the Green Lightspout Member to the red siltstones and sandstones of the Red Lightspout Member. The type section is exposed in the Cardingmill Valley and in Lightspout Hollow between SO 43709502 and SO 42859519. The lithology typically consists of laminated, pale green siltstones. Very thick-bedded, fine grained sandstones are common, from approximately 1m to 8m thick. There are occasional, thin to thick-bedded, very fine grained sandstones and very thinly laminated, medium to dark grey siltstones. This member is characterised by the development of numerous, isoclinal folds of short wavelength. The thickness estimate of 220m for this member is an approximation which tries to account for the folding. The development of red siltstones at some horizons may be due to the exposure of the underlying Red Synalds Member in the hinge zones of anticlines or is more probably due to a change in the lithology of the Green Lightspout Member.

2.13.2 The Red Lightspout Member

The base of the member is defined above, as the transition from the Green Lightspout Member. The top of the member is subjectively defined as the transition from the interbedded red siltstones and sandstones of the Red Lightspout Member to the thick, homogeneous sandstone of the Huckster Conglomerate Member of the Portway Formation. The type section is exposed near Motts Road between SO 43299547 and SO 43209554. The lithology mainly consists of red, laminated siltstones, with thin to thick-bedded, very fine grained sandstones. Towards the top of the member, fine to medium grained and thick to very thick-bedded sandstones with common siltstone rip-up clasts are developed, which mark the transition to the Huckster Conglomerate.

CHAPTER 3

STRUCTURE

3.1 Introduction

Measurements of the orientations of cleavage and bedding planes and bedding/cleavage intersection lineations were made throughout the Longmyndian Supergroup. These data were plotted on stereographs and on geological maps (geological maps 1 and 2, in cover pocket). Structural parameters, such as the orientations of fold axes and fold axial planes were derived from these data. It is apparent from these parameters that there are consistent discrepancies in orientation between the Stretton Group and the Wentnor Group structures and this is discussed.

From considerations of the fault orientations and the strain ellipsoid orientations, it is argued that the Longmyndian has been involved in strike-slip deformation. The probability that both the Church Stretton and the Pontesford-Linley fault systems are major strike-slip faults is discussed.

Some minor tectonic structures which simulate sedimentary structures, such as pseudo-ripples and tectonic lineations, are discussed and the evidence for their mode of origin is presented.

3.2 The nature of the main syncline

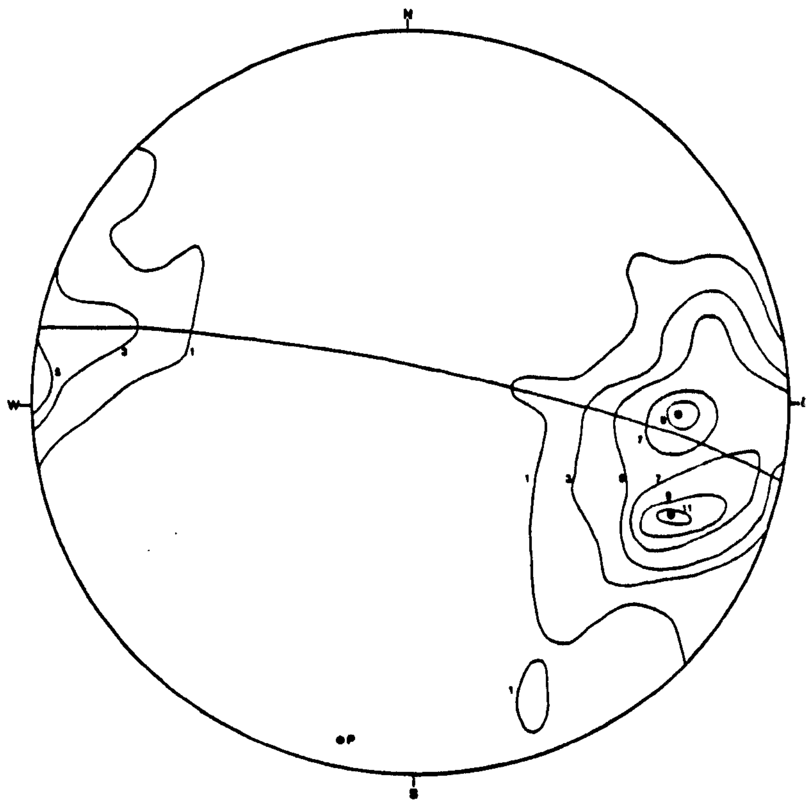
Abundant observations of way-up from sedimentological criteria have been made. These include the gross grain size trends visible at outcrop, together with observations of cross-bedding, cross-lamination and grain size gradations at outcrop and in cut and polished hand specimens. These data extend the limited observations of Greig et al. (1968) and James (1952b) and confirm the existence of a large syncline.

The Bridges Formation occupies the core of the syncline and the western and eastern limbs of the syncline are composed of the Bayston-Oakswood Formation and the Bayston-Oakswood Formation and Stretton Group respectively. The axial trace trends NNE to SSW from Longden (SJ 443065) in the NNE to Norbury (SO 364928) in the SSW. The fold is isoclinal and mostly overfolded. However, there is a systematic variation in the nature of the fold from tight, with an upright to slightly inclined axial plane in the NNE, to an isoclinal overfold with an inclined axial plane, which dips approximately 60° WNW, in the SSW. In the SSW, the axial surface appears to be curviplanar and concave upward, this being manifest in the steeper dips in the ENE and shallower dips in the WNW.

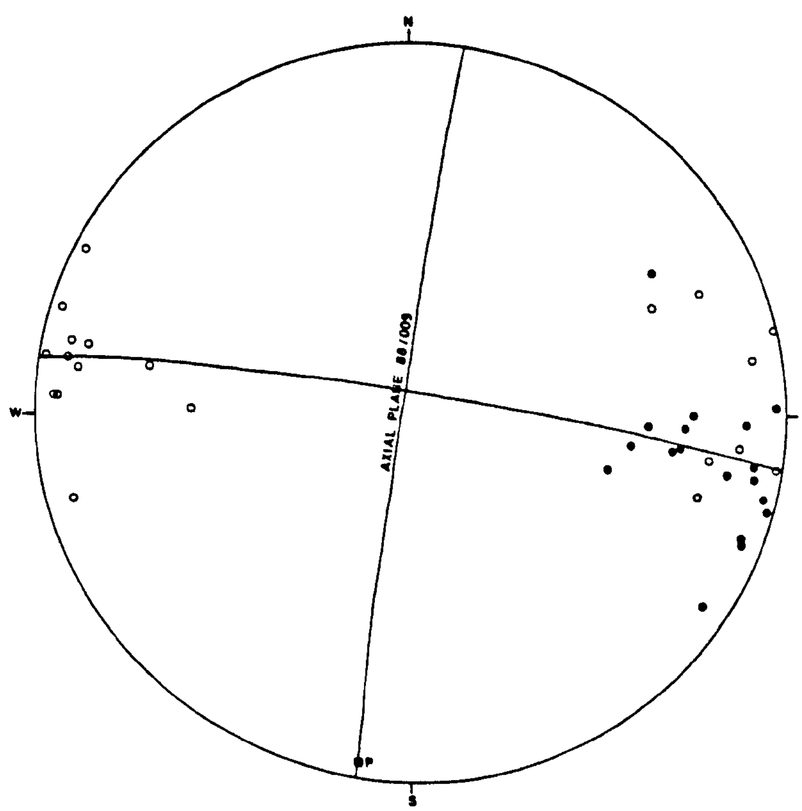
A stereographic plot of poles to bedding for the Wentnor Group (fig. 9, stereographic plot 1) defines an approximate fold axis with a plunge of 8° towards 192° (SSW). This value is an approximation, since the fold is mostly isoclinal and the girdle is consequently poorly defined. The definition of a girdle is also complicated by WNW-ESE trending folding or non-cylindricity of the fold hinge, which causes scatter of the bedding poles away from the main syncline bedding pole girdle (fig. 9, stereographic plot 1). A stereographic plot of poles to bedding from the area of Cothercott (SJ 415015), where the fold is tight rather than isoclinal and where the axial plane is approximately upright, defines a fold axis which plunges 5° in the direction 189° (SSW) (fig. 9, stereographic plot 2). James (1952b) defined a fold axis with a plunge of 10° in the direction 215°. However, this figure was determined from the bedding/cleavage

FIG: 9 STEREOGRAPHIC PLOTS 1, PERCENT POLES TO BEDDING FOR THE WENTNOR GROUP AND 2, POLES TO BEDDING FOR THE WENTNOR GROUP IN THE COTHERCOTT AREA (SJ 415015)

STEREOGRAPHIC PLOT 1. PERCENT POLES TO BEDDING
WENTNOR GROUP N=256
P=8/192



STEREOGRAPHIC PLOT 2. POLES TO BEDDING
WENTNOR GROUP, SUB AREA 5.
○ E YOUNGING
● W YOUNGING
P 5/189



intersection lineations in the Stretton Group. There is a clear discrepancy between the structural orientations in the Stretton Group and the structural orientations in the Wentnor Group (see following sections). Consequently, the structures of the Stretton Group cannot be used to define the main syncline axis which lies in the Wentnor Group.

The outcrop of the Bridges Formation thins from SSW to NNE, from the area of Bridges (SO 393964) to Longden (SJ 443065) and this indicates a low angle plunge to the SSW, in agreement with the calculated values. The anomalous thickening of the Bridges Formation south of Bridges may be due to faulting. Since cleavage is very poorly developed in the Wentnor Group, the fold axis and variations in dip of the fold axis cannot be adequately determined from the bedding/cleavage intersection lineations. Consequently, a fold axis plunge of 8° in the direction 192° (SSW) has been used throughout the Wentnor Group during rotations of the palaeocurrent data.

3.3 Minor folds in the Wentnor Group

From the way-up criteria, it is evident that minor folds exist in the Bridges Formation (Geological Map 2 - in cover pocket). These were previously noted by Greig et al. (1968) from the area of Bridges (SO 393964). From this study, the existence of minor folds is demonstrated elsewhere in the Bridges Formation and they may be more common than are depicted on Map 2 (in cover pocket), since exposure is mostly poor and way-up criteria are often lacking. These minor folds appear to be co-axial and co-planar with the main syncline. No minor folds are apparent in the Bayston-Oakswood Formation, which may be due to its greater competence and homogeneity.

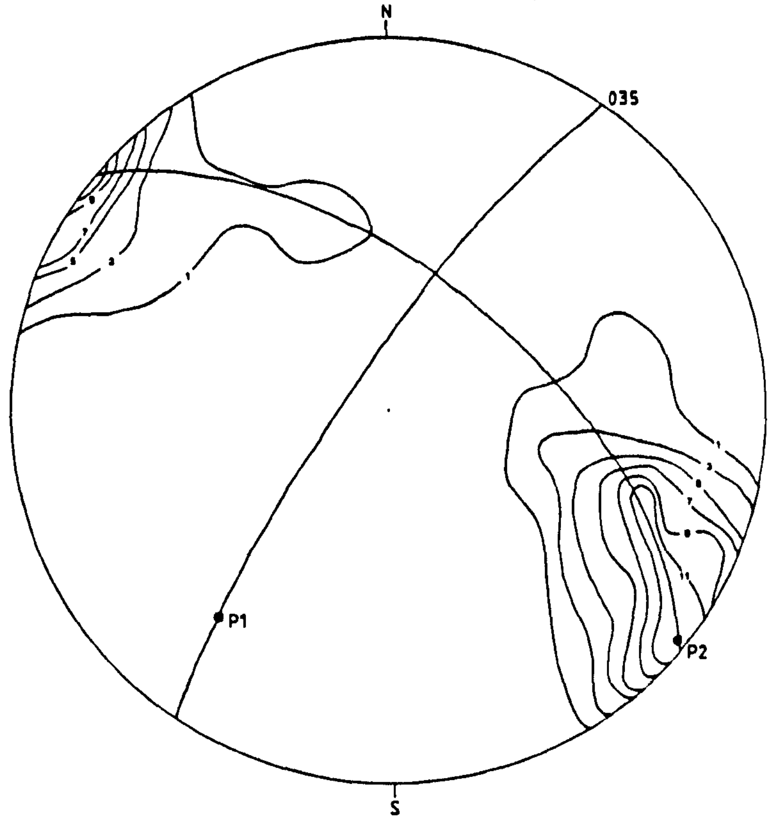
3.4 Minor folds in the Stretton Group

Abundant evidence for way-up has been collected from outcrops and from cut and polished hand specimens. In conjunction with observations on the bedding/cleavage relationships, the way-up evidence has enabled the recognition of numerous, minor folds in the Stretton Group, the extent of which has not been previously recognised. These folds have upright axial planes and are isoclinal to tight with large amplitudes and wavelengths commonly of the order of 90m. The hinges of the folds are not commonly exposed. Faults which are subparallel to the fold axial planes are commonly developed. The folds are commonly developed at certain horizons where there is either an internal variability in competence, as in the Lightspout Formation, or a contrast in competence with the adjacent strata, such as at the base of the Burway Formation. The distribution of folding is depicted on Geological Maps 1 and 2 (in cover pocket). However, not all the fold axes or folded zones are depicted on these maps, since an exhaustive delineation of these structures was not made. Consequently, the degree of minor folding is greater than that depicted on these maps (except for the sections through the Stretton Group in Ashes Hollow; the Lightspout Formation in the Cardingmill Valley, Lightspout Hollow and Mott's Road areas and the Lightspout Formation in the Long Batch and Jonathon's Hollow areas, where a thorough analysis of structure was made in order to determine the extent of the minor folding in these areas). A stereographic plot of poles to bedding in the Stretton Group defines a clear girdle which indicates a fold axis which plunges 30° in the direction 220° (SW). The axial traces of these folds strike

FIG: 10 STEREOGRAPHIC PLOTS 3, PERCENT POLES TO BEDDING, STRETTON GROUP AND 4, MINOR FOLD AXES, STRETTON GROUP.

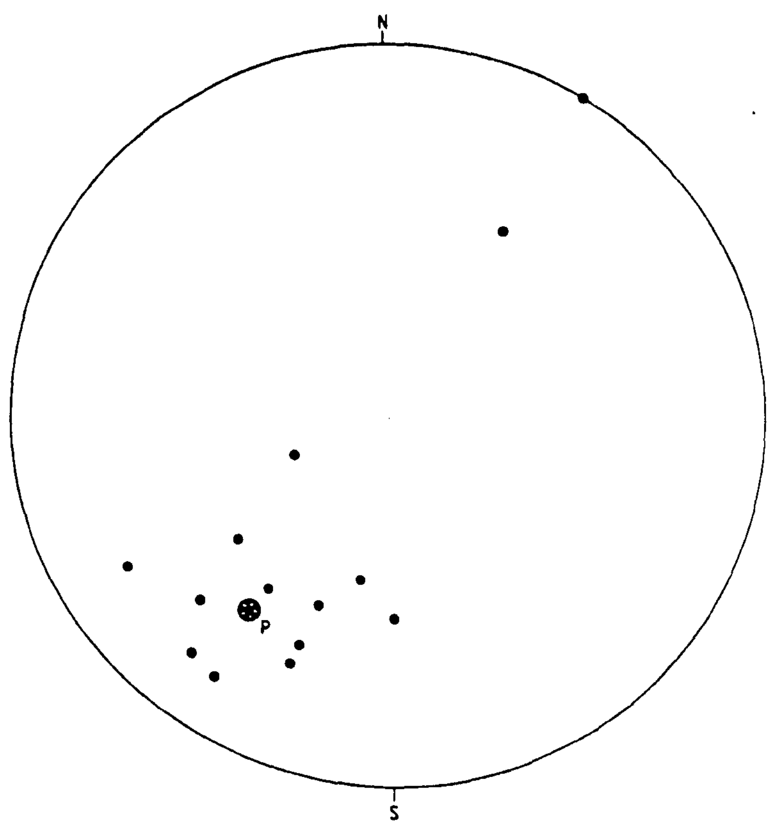
STEREOGRAPHIC PLOT 3, PERCENT POLES TO BEDDING
STRETTON GROUP

N = 236
P1 = 30/220 (fold axis)
P2 = 3/129 (point max.)
Axial plane = 035/81



STEREOGRAPHIC PLOT 4, MINOR FOLD AXES,
STRETTON GROUP.

N = 15
P = 36/216 (average)



approximately NNE-SSW (035°-215°) and the axial planes, from stereographic constructions, dip 035/81 (fig. 10, stereographic plot 3).

The fold axes of the minor folds are not coaxial with the fold axis of the major syncline. The Stretton Group fold axes have been rotated clockwise with respect to the major syncline and the plunges of the fold axes are steeper by 8° to 30°. The fold axis rotation is a reflection of the clockwise tectonic rotation of the Stretton Group as a whole, with respect to the Wentnor Group.

In addition to the macroscopic folds described above are occasional mesoscopic folds, which are asymmetric, parasitic folds which are commonly "Z" shaped when viewed from the SSW. Fold wavelengths vary from several centimetres to a few metres. The orientations of these fold axes can be measured directly in the field and yield an average plunge of 36° in the direction 216° (fig. 10, stereographic plot 4). The plunge of these mesoscopic folds is therefore approximately coaxial with the plunge of the macroscopic folds.

In addition to the above folds, chevron folds are occasionally developed which have wavelengths of several centimetres. These are most abundant in the Stretton Shale Formation.

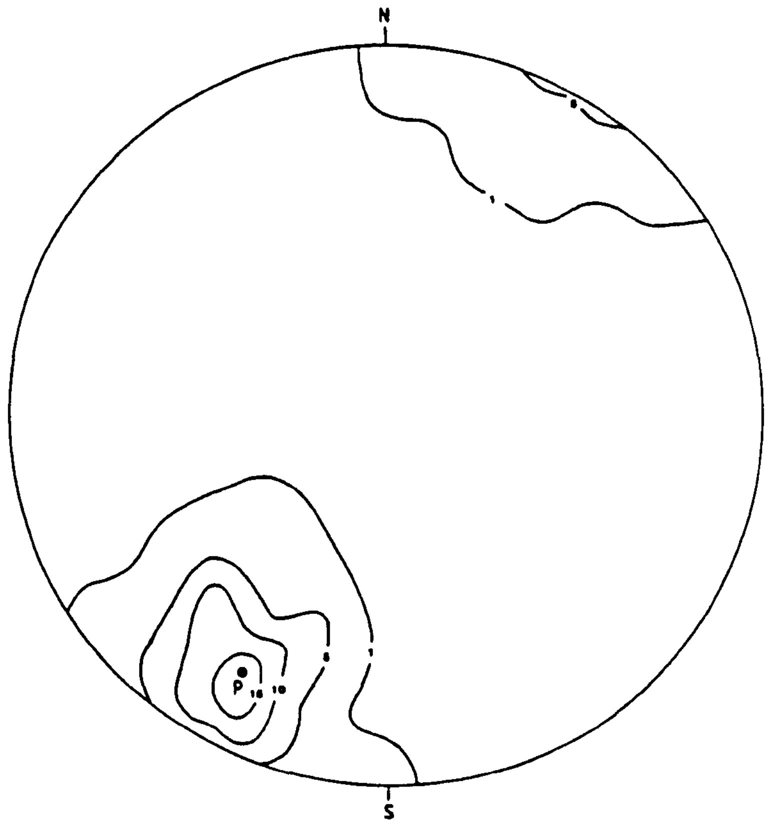
3.5 Cleavage in the Stretton Group

A cleavage is commonly developed in the finer grained lithologies of the Stretton Group. The main horizons in which cleavage is extensively developed are the Synalds Formation and the Portway Formation. A stereographic plot of the bedding/cleavage intersection lineations (fig. 11, stereographic plot 5) defines a point maximum reading which plunges 22° in the direction 210°.

FIG:11 STEREOGRAPHIC PLOTS 5, PERCENT BEDDING/CLEAVAGE INTERSECTION LINEATIONS, STRETTON GROUP AND 6, PERCENT POLES TO CLEAVAGE, STRETTON GROUP.

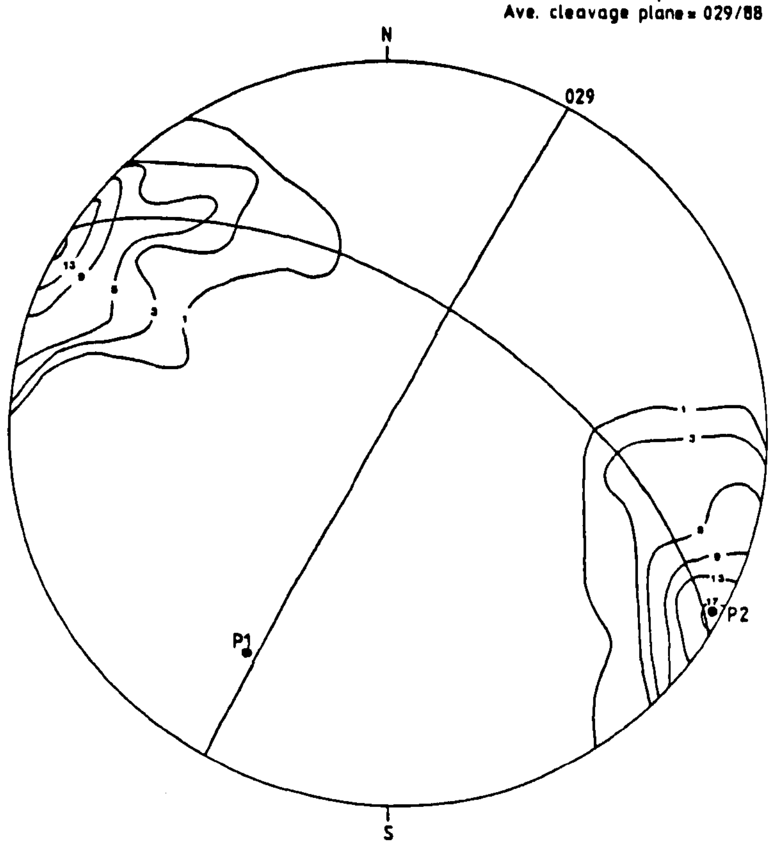
STEREOGRAPHIC PLOT 5, PERCENT BEDDING / CLEAVAGE INTERSECTION LINEATIONS, STRETTON GROUP.

N = 170
P = 22 / 210 (point max.)



STEREOGRAPHIC PLOT 6, PERCENT POLES TO CLEAVAGE, STRETTON GROUP

N = 134
P1 = 30/213 (fan axis)
P2 = 2/119 (point max.)
Ave. cleavage plane = 029/88



However, the fold axis of the main syncline has been shown to plunge 8° in the direction 192°. Since the bedding/cleavage intersection lineation would be expected to be coaxial with the main fold axis, there appears to be an anomalous clockwise rotation of the bedding/cleavage intersection lineation. A similar rotation of the macroscopic and mesoscopic folds was noted in the previous section. The bedding/cleavage intersection lineation is approximately coaxial with the fold axes of the macroscopic and mesoscopic folds.

A stereographic plot of poles to cleavage (fig. 11, stereographic plot 6) defines a clear girdle with an axis which plunges 30° in the direction 213°. This axis is approximately coaxial with the macroscopic and mesoscopic fold axes and may be due to cleavage fanning around these axes. If this hypothesis is correct, then this would imply that the cleavage and the macroscopic and mesoscopic folds are a product of the same stress field and are probably synchronous. The average cleavage plane is approximately coplanar with the axial planes of the macroscopic folds and dips 029/88.

3.6 The relationship between the Wentnor Group and the Stretton Group

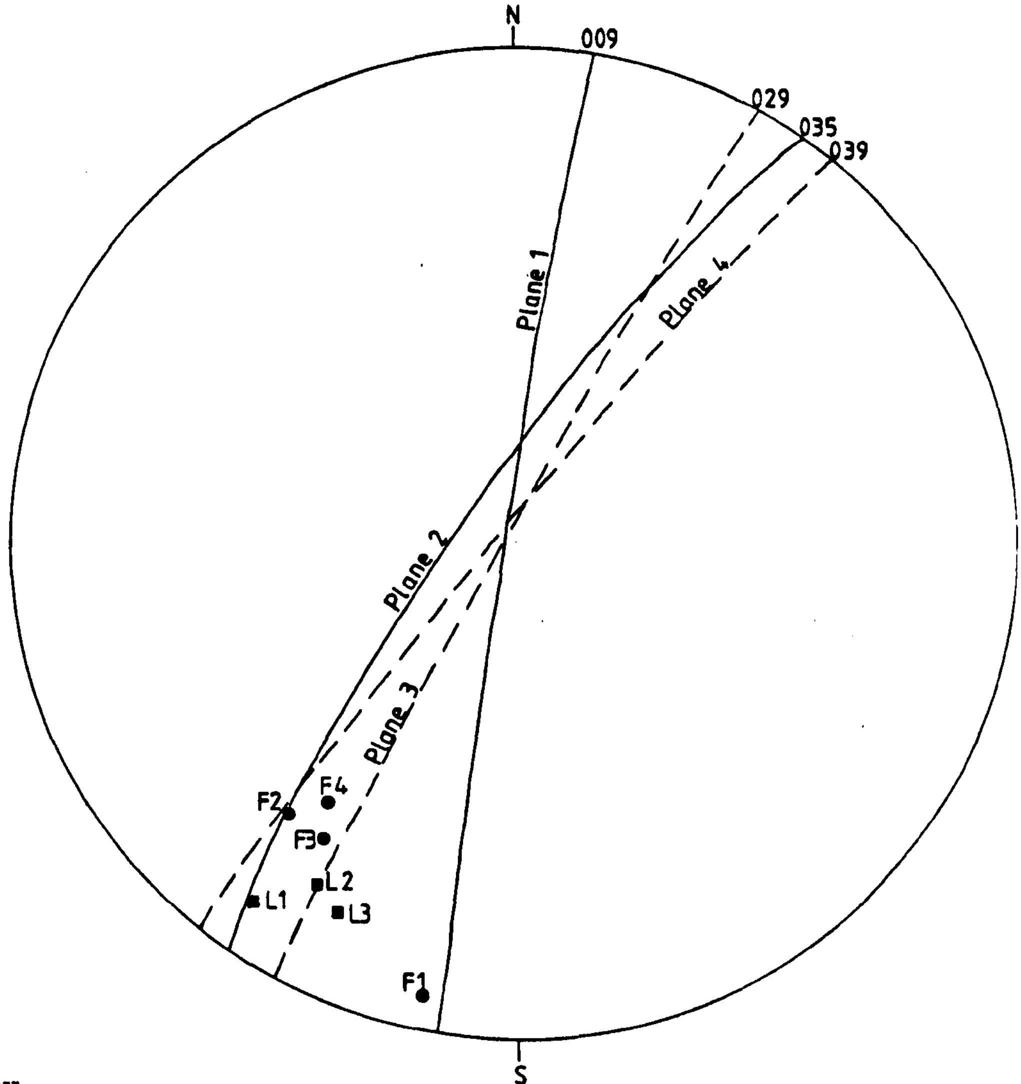
It has been noted in the preceding sections that the fold axes and the axial planes of the macroscopic and mesoscopic folds in the Stretton Group and the cleavage and bedding/cleavage intersection lineations in the Stretton Group have been rotated clockwise, with respect to the axial plane and fold axis of the main syncline. This clockwise rotation is also manifest in the clockwise rotation of the bedding strike in the Stretton Group, with respect to the bedding strike in the Wentnor Group. These rotations are of the order of 22°

(see fig. 12, stereographic plot 7). The angular discordance of strike between the Wentnor Group and the Stretton Group has been interpreted as an unconformity by Greig _et al_. (1968) and James (1952b). However, it is apparent that the angular discordance between the Wentnor Group and the Stretton Group is a manifestation of a tectonic rotation of all the structural elements.

The angular discordance of strike between the Wentnor Group and the Stretton Group appears to be the result of a gradual rotation of strike within the Portway Formation rather than an angular discordance at the base of the Wentnor Group. At Hawkham Hollow (SO 433976), the Portway Formation and the Bayston-Oakswood Formation have similar dips and strikes and an angular discordance is not apparent. Between the Huckster Conglomerate at High Park House (SO 443972) and the base of the Wentnor Group at Hawkham Hollow (SO 433976), the strike of the bedding is gradually rotated anticlockwise. A similar effect occurs at the head of Ashes Hollow (SO 418937), where the strike of the bedding appears to be rotated within the Portway Formation, such that the base of the Wentnor Group appears to be conformable with the upper parts of the Portway Formation.

The rotation of strike within the Portway Formation may be explained by movement along strike faults within the Portway Formation, which are not demonstrable because of poor exposure, or alternatively, the rotation is accomplished by shear and accommodation across cleavage planes throughout the Portway Formation. The rotation may be viewed as a decoupling between the more competent Wentnor Group and the less competent Stretton Group.

FIG:12 STEREOGRAPHIC PLOT 7, STRUCTURAL SYNTHESIS OF THE LONGMYNDIAN SUPERGROUP



KEY:

PLANE 1 = Axial plane of major syncline in the Wentnor Group in the Cothercott area (SJ 415015),009/88.

PLANE 2 = Axial plane of macroscopic folds in the Stretton Group, 035/81.

PLANE 3 = Average cleavage plane for the Stretton Group,029/88.

PLANE 4 = Average bed plane for the Stretton Group,039/87.

F1 = Fold axis of the major syncline in the entire Wentnor Group, 8/192.

F2 = Fold axis of the macroscopic folds in the Stretton Group,30/220

F3 = Axis of cleavage fanning in the Stretton Group,30/213.

F4 = Average fold axis of mesoscopic folds in the Stretton Group, 36/216.

L1 = Average bedding/cleavage intersection lineation for the apparently unfolded areas in the Stretton Group,12/216.

L2 = Average bedding/cleavage intersection lineation for the entire Stretton Group,22/210.

L3 = Average bedding/cleavage intersection lineation for the apparently folded areas in the Stretton Group,19/206.

The south-westward thinning of the Portway Formation, which is the result of the rotation of the Wentnor Group with respect to the Stretton Group, thus appears to be a tectonic phenomena rather than the result of an unconformity at the base of the Wentnor Group.

3.7 The treatment of palaeocurrent data for the Stretton Group

The bedding/cleavage intersection lineations from apparently unfolded areas in the Stretton Group plunge at significantly lower angles than the fold axes of the macroscopic and mesoscopic folds (fig. 12, stereographic plot 7). However, the exact relationships between the bedding/cleavage intersection lineations and the Stretton Group fold axes are not certain, since the bedding/cleavage intersection lineations from the folded areas of the Stretton Group and the average bedding/cleavage intersection lineation for the whole of the Stretton Group are not exactly coincidental with the macroscopic and mesoscopic fold axes of the Stretton Group (fig. 12, stereographic plot 7). It appears that there might have been a slight anticlockwise rotation of the cleavage and the bedding/cleavage intersection lineation, with respect to the fold axes and the axial planes of the macroscopic folds of the Stretton Group (fig. 12, stereographic plot 7).

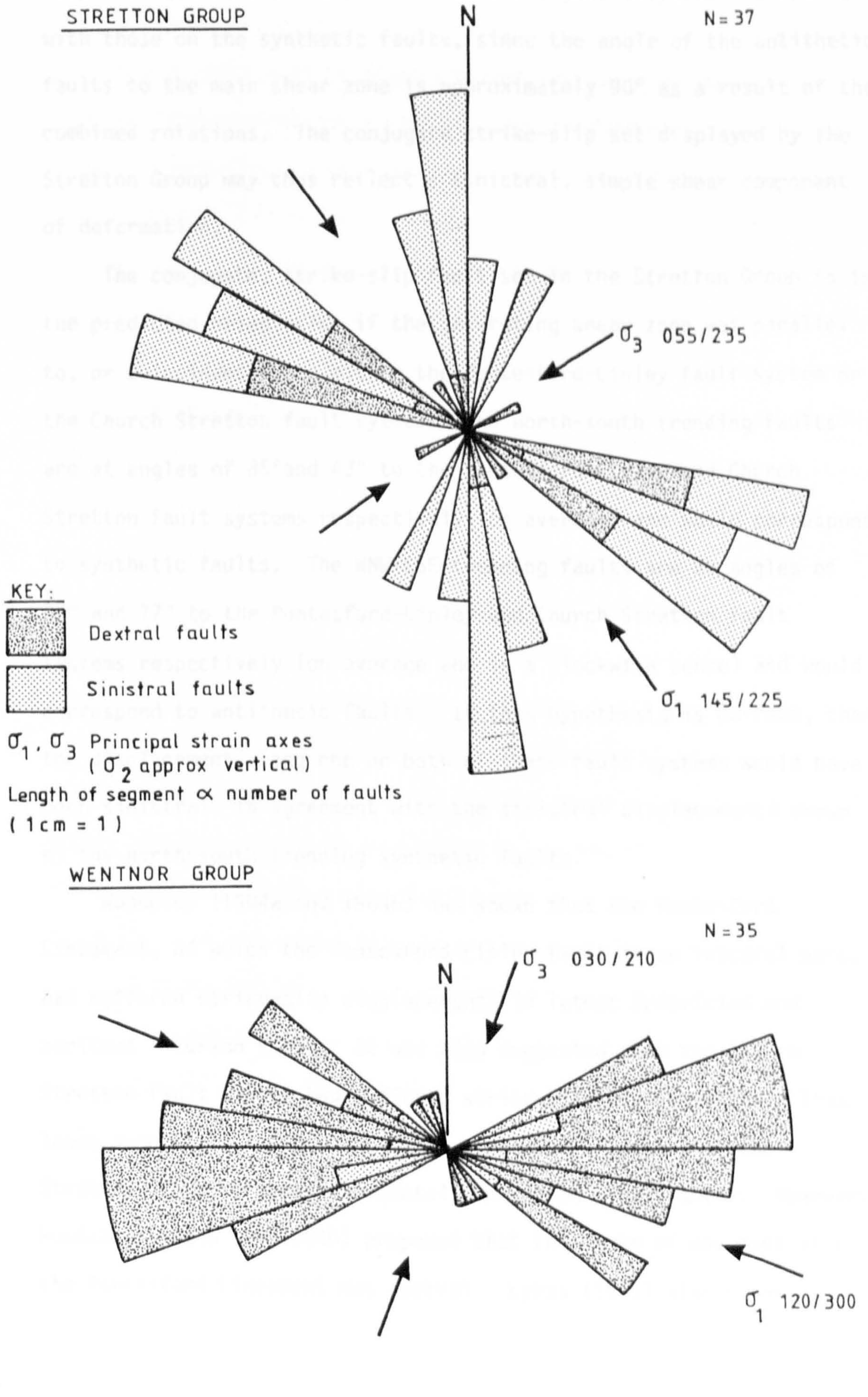
Since the bedding/cleavage intersection lineations approximate to the plunge of the macroscopic and mesoscopic folds and since these folds are either impersistent or are unable to be recognised in places, the plunge of the local bedding/cleavage intersection lineation was used for each palaeocurrent rotation.

3.8 Faulting in the Longmyndian

3.8.1 Fault patterns in the Stretton Group

A clear fault pattern is evident in the Stretton Group (fig. 13 and geological map 2). Two distinct fault sets are present: an approximately north-south set and an approximately WNW-ESE set. The strike of these faults is not affected by the topography and the faults from both sets are apparently predominantly vertical. The faults also appear to have a predominant strike-slip component, otherwise very large throws would be required to produce the observed horizontal displacements of the steeply dipping beds. There is a significant difference between the fault sets, in the preferred amount and sense of displacement. The north-south faults are almost exclusively sinistral and commonly show large sinistral displacements (e.g. the Ashes Hollow Fault and the Yewtree Bank Fault). Conversely, the WNW-ESE trending faults have predominantly small displacements which are both dextral and sinistral. The number of sinistral faults in the WNW-ESE set is approximately equal to the number of dextral faults and the displacements on both types are similar. These two fault sets are interpreted as a strike-slip conjugate set. A mechanism by which the amount of displacement on a conjugate strike-slip set remains highly inequal is detailed by Wilcox _et al_. (1973). During simple shear, synthetic faults, which are developed at angles of 10° to 30°to the shear zone, tend to be rotated away from the shear zone by the imposed external rotation, however, the internal rotation counteracts this and little rotation of the synthetic faults results. Therefore, these faults remain in a favourable position for shear displacements. Conversely, antithetic faults, which are developed at high angles to the shear zone, are rotated away from the shear zone both by the imposed external

FIG: 13 FAULT ORIENTATIONS IN THE LONGMYNDIAN



rotation and the internal rotation. Consequently, lateral displacements on the antithetic faults are generally small compared with those on the synthetic faults, since the angle of the antithetic faults to the main shear zone is approximately 90° as a result of the combined rotations. The conjugate strike-slip set displayed by the Stretton Group may thus reflect a sinistral, simple shear component of deformation.

The conjugate, strike-slip fault set in the Stretton Group is in the predicted orientation if the generating shear zone was parallel to, or coincident with, either the Pontesford-Linley fault system or the Church Stretton fault system. The north-south trending faults are at angles of 35°and 43° to the Pontesford-Linley and Church Stretton fault systems respectively (on average) and would correspond to synthetic faults. The WNW-ESE trending faults are at angles of 85° and 77° to the Pontesford-Linley and Church Stretton fault systems respectively (on average and in a clockwise sense) and would correspond to antithetic faults. If this hypothesis is correct, then the displacement along one or both of these fault systems would have been sinistral, in agreement with the sinistral displacements shown by the north-south trending synthetic faults.

Woodcock (1984a and 1984b) has shown that the Pontesford Lineament, of which the Pontesford-Linley fault is an integral part, has suffered strike-slip displacements in latest Ordovician and earliest Silurian times. It was also suggested that the Church Stretton fault system had suffered strike-slip displacements. This lends support to the hypothesis that the fault pattern in the Stretton Group reflects horizontally directed simple shear. However, Woodcock (1984a and 1984b) proposed that the sense of movement along the Pontesford Lineament was dextral. Lynas (1985) also noted dextral strike-slip faulting in the Ordovician of the Shelve inlier, which is adjacent to the Longmyndian inlier and on its north-west margin. However, both these authors detail evidence from Lower Palaeozoic rocks. It is plausible that the sense of movement has changed from predominantly sinistral to predominantly dextral with time. Alternatively, adjacent areas may show different senses of displacement. For example, Murphy (1985) describes a reversal of strike-slip sense and associated structures over only a few kilometres, due to the influence of an isolated rigid terrane, in a Caledonian sinistral transpressive deformation, in eastern Ireland. Therefore the evidence of Woodcock (1984a and 1984b) and Lynas (1985) does not preclude the hypothesis of a sinistral, simple shear deformation in the Stretton Group.

3.8.2 Fault patterns in the Wentnor Group

Overall, the Wentnor Group fault trend is dissimilar to the fault trend in the Stretton Group. The main fault trend is WSW-ENE and the faults are mainly dextral, with moderate to small displacements (in comparison with the N-S fault set in the Stretton Group) and they appear to be similar to the WNW-ESE set of the Stretton Group. There is no clearly defined conjugate fault system as in the Stretton Group (fig. 13 and geological map 2). The Stretton Group dextral fault trend averages approximately 305°, whereas the Wentnor Group dextral fault trend averages approximately 285°. The difference between the two trends is approximately 20°, with the Stretton Group trend rotated clockwise with respect to the Wentnor Group. A similar relationship has been demonstrated for other structural elements, such as the bedding, cleavage, fold axes, fold axial planes and bedding/cleavage intersection lineations, which were previously discussed in section 3.6, and this difference is a further manifestation of the tectonic rotation between the Wentnor Group and the Stretton Group, which was previously estimated to be of the order of 22° (similar to the fault trend rotation), from the combined structural criteria. The Wentnor Group faults appear to post-date the folding, as the axial trace of the major syncline is displaced dextrally by these faults.

The anomalous thickening of the Bridges Formation, south of Bridges, may be due to NNW-SSE faults, but this cannot be proved because of poor exposure. These faults would be subparallel to the fault which forms the eastern margin of the Bridges Formation in this area.

3.8.3 Strike faults

It is difficult to assess the degree and magnitude of strike faulting in the Longmyndian since these faults are difficult to recognise. However, it is thought that strike faults may be an important element in the structure of the Longmyndian and may occur at the boundaries of some stratigraphic units and in apparently thinned units in the Longmyndian.

It is proposed that the Church Stretton fault system and the Pontesford-Linley fault system have a component of strike-slip movement. Since the strike of the bedding in the Longmyndian is subparallel to these fault systems, it is probable that strike-slip movement has occurred along the bedding planes or along faults subparallel to the bedding planes, in response to the same simple shear strain.

Strike faults, which are parallel to bedding, outcrop along the south-eastern side of Haughmond quarry (SJ 54251455). The fault planes are characterised by a fine-grained fault gouge, several centimetres thick. In some cases these faults appear to define facies boundaries. If this is correct, then considerable displacement along these faults might have occurred. A lineament pattern is present on Haughmond Hill, which is apparent on aerial photographs. A number of lineaments occur parallel to bedding. These may be interpreted as strike faults, similar to the faults exposed in Haughmond quarry (SJ 54251455) since this could account for the apparent thinning of the combined Synalds and Lightspout Formations from c.1000m in the Long Mynd area to c.450m in the Haughmond Hill area. The north-western margin of Haughmond Hill is faulted against the Carboniferous along a major strike fault.

The Longmynd Scarp Fault approximates to a bedding-parallel strike on the south-west slopes of the Long Mynd. This fault is easily recognisable because of the juxtaposition of the Llandovery with the Longmyndian.

Strike faults are evident in Ashes Hollow at SO 43659288. Here, the limbs of a minor syncline have been faulted. The north-western margin of the minor syncline passes abruptly across a linear valley into strata which young in the opposite direction. Consequently, a strike fault appears to have removed the complementary anticline of the fold pair. A similar fault is present on the south-eastern margin of this syncline.

The tectonic rotation of the Wentnor Group with respect to the Stretton Group (section 3.6) is a late feature. The rotation appears to have occurred in the Portway Formation, which is very poorly exposed in the Long Mynd area. It is considered to be highly likely that this rotation has been accomplished by movement along a strike fault in the Portway Formation, which could be a north-easterly continuation of the Longmynd Scarp Fault. This fault has suffered post-Llandovery movement and this late movement could have produced the observed tectonic rotation of the Longmyndian structural elements.

3.8.4 The Church Stretton and Pontesford-Linley fault systems

Both of these fault systems have been postulated to show strike-slip displacements (Woodcock,1984a, 1984b and Lynas, 1985).It has been shown that the fault pattern in the Stretton Group may be explained in terms of a horizontally directed simple shear with a sinistral sense of rotation. Additionally, the fault orientations are as predicted if the generating shear zone was parallel to, or coincident with, either the Pontesford-Linley fault system or the Church Stretton fault system (section 3.8.1). Additional evidence for a horizontally directed simple shear component is provided by observations on the bulk strain: the X axis of the strain ellipse is subhorizontal and subparallel to the Pontesford-Linley and Church Stretton fault systems (section 3.9). Both the Pontesford-Linley and Church Stretton fault systems show an extremely complex, braided pattern of faulting, in which strata of Precambrian, Cambrian, Ordovician and Silurian age have been involved and in which the majority of the contacts are faulted. Some of the main faults show a change in the sense of vertical displacement along strike (e.g. F3 and F2-F3 of the Church Stretton area, see geological map 2). Both braiding of faults and changes in the sense of vertical displacement along fault strike are considered to be typical of master wrench zones (Wilcox *et al.*, 1973). The Church Stretton fault system displays two main fault trends: a NW-SE minor fault trend, with dominantly vertical displacements and a braided NNE-SSW major fault trend, with dominantly strike-slip displacements. Additionally, reverse faults are developed subparallel to the trend of the fault systems in both the Church Stretton and Pontesford-Linley fault systems. These reverse faults show a dominantly south-easterly directed movement. These fault trends are similar to the pattern developed in the adjacent Stretton Group of the Longmyndian and may be similarly explained in terms of a horizontally directed simple shear component.

Greig et al. (1968) proposed that a north-west to south-east minor fold trend in the Uriconian Volcanic Complex, within the Church Stretton fault system, is similar to the trend of the folding in the Charnian and since these structures are not manifest in the Longmyndian Supergroup and since the Longmyndian Supergroup is supposed to cover the Uriconian Volcanic Complex unconformably, this "Charnoid" fold trend in the Uriconian Volcanic Complex was considered to be pre-Longmyndian in age. However, it is apparent that the deformation in the Longmyndian Supergroup and the Church Stretton fault system are of the same type. No superimposed trends are apparent which need to be explained in terms of different orogenic episodes. The proposed "Charnoid" trending folds may be better explained by NW-SE trending tensional faults, associated with dolerite intrusions, which have produced a NW-SE trending syncline in the Caer Caradoc area (SO 474950) and NE directed movements across the Sharpstones and Hill End reverse faults, which have produced a WNW-ESE trending syncline in the Woodgate Tuff beds (SO 480932). These mechanisms and geometries can be explained by the same deformation which affected the Longmyndian and produced the Church Stretton fault system and it is not necessary to postulate an earlier "Charnoid" orogeny.

Determination of the sense of displacement of the Church Stretton and Pontesford-Linley fault systems is hampered by the fact that the sense of displacement along a master wrench zone may change with time and that minor faults within the master wrench zone may show different senses of displacement. Greig et al. (1968) proposed that sinistral movement had occurred along the Cwms-Hoar Edge Fault (F3) accompanied by north-easterly directed overthrusting of the Uriconian Volcanic Complex over the Ordovician along the Sharpstones Thrust. Dean (1964) concluded that F3 is probably a tear fault with a sinistral displacement of the Alternata Limestone in the south-western part of the Church Stretton fault system. Woodcock (1984a) proposed that the Carboniferous and possibly the Triassic had been offset sinistrally along the Church Stretton fault system. This evidence for sinistral movement along and within the Church Stretton fault system lends support to the hypothesis of a sinistral simple shear deformation of the adjacent Stretton Group from a consideration of the contained minor structures.

Conversely, dextral movements have been proposed along the Pontesford-Linley fault system (Woodcock, 1984b) and Lynas notes evidence for dextral strike-slip in the adjacent Ordovician Shelve inlier (Lynas, 1985). The opposite senses of movement along the fault systems may be compared to a similar situation in a Caledonian sinistral transpressive orogeny in eastern Ireland (Murphy, 1985). Here, local dextral movement in an overall sinistral régime is found on the margin of a rigid terrane. Sinistral movements are found on a strike-slip fault (the Lowther Lodge Fault) approximately 4km away from dextral movements on the margin of the rigid terrane (the Bellewstown terrane). The adjacent strata display minor structures which accord with the local strike-slip movement (Murphy, 1985). The variation in the sense of strike-slip movement for the Longmyndian Supergroup may be modelled on this example. If the overall régime were dextral, as shown by the Pontesford Lineament (Woodcock, 1984b), then local sinistral strike-slip might have been produced adjacent to the Eastern Uriconian Volcanic Complex rigid terrane, as shown by the Church Stretton fault system and adjacent Stretton Group. The scale of the Pontesford-Linley/Church Stretton system is of the same order as the Lowther Lodge/Bellewstown system, with the distance between the Church Stretton and Pontesford-Linley fault systems being approximately 9km. Implicit to this argument is the requirement that the Eastern Uriconian forms a large rigid terrane whereas the Western Uriconian does not. The Eastern Uriconian is in fact shown to extend south-eastwards beneath the Palaeozoic cover on magnetic evidence (Wilson, 1980) whereas the Western Uriconian occurs in small, isolated lenticles.

The contacts of the Longmyndian Supergroup with the Uriconian Volcanic Complex (including the Western Uriconian and the Eastern Uriconian) have been shown to be faulted. The Uriconian Volcanic Complex is considered to have developed approximately synchronously with the Longmyndian Supergroup. However, the lack of facies which are indicative of proximity to a volcanic source within the Longmyndian Supergroup and the fact that the palaeocurrent data do not indicate immediate derivation from the Uriconian Volcanic Complex, in its present geographical positions, suggests that the Longmyndian Supergroup has been tectonically juxtaposed with the Uriconian Volcanic Complex along the Church Stretton and

Pontesford-Linley fault systems. The recognition of strike-slip movements along these fault systems suggests that the discrepancies noted above and the tectonic juxtaposition could be explained by large horizontal displacements. Consequently, large geographical displacements of the Longmyndian Supergroup, with respect to the Uriconian Volcanic Complex, have probably occurred. However, the amount of these displacements cannot, as yet, be quantified.

3.9 Strain and the Longmyndian structure

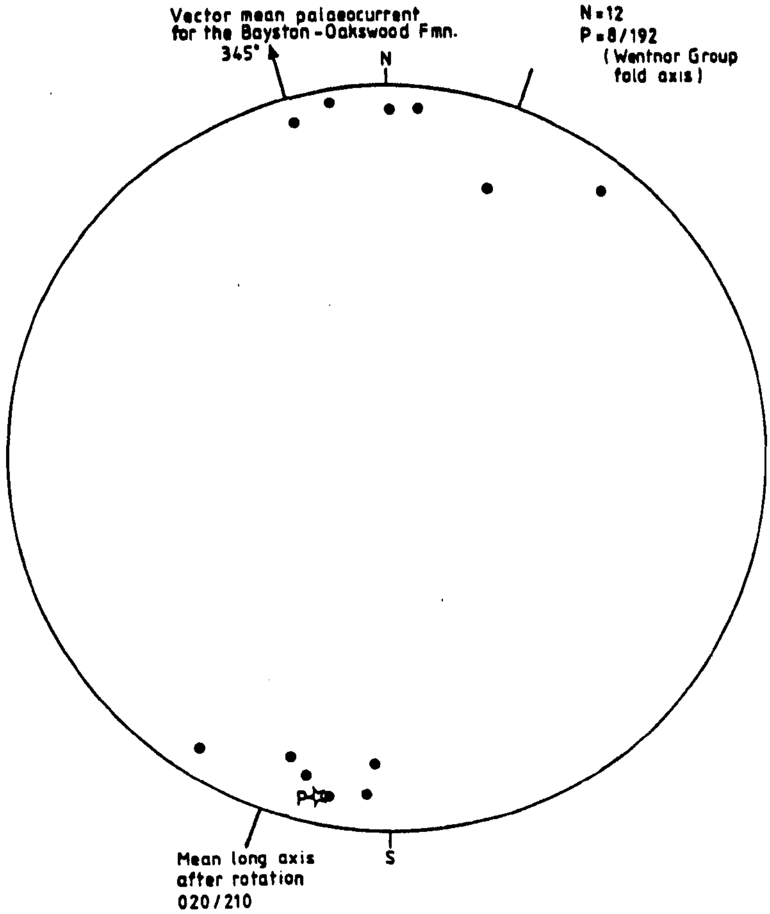
No quantitative measurements of strain were made, however qualitative observations were made on the amount and direction of strain. Two features enable the strain to be determined: deformed, elongate pebbles from the Wentnor Group and deformed pits from the Stretton Group, which have been variably interpreted as rain prints (James, 1952b), Arenicolites burrows (Salter, 1856 and 1857), inorganic bubble holes and small, blister-like features (Greig et al., 1968) and biogenic spheroids of Arumberia (Bland, 1984). The origin of these structures is discussed in chapter 5.

3.9.1 Strain in the Wentnor Group

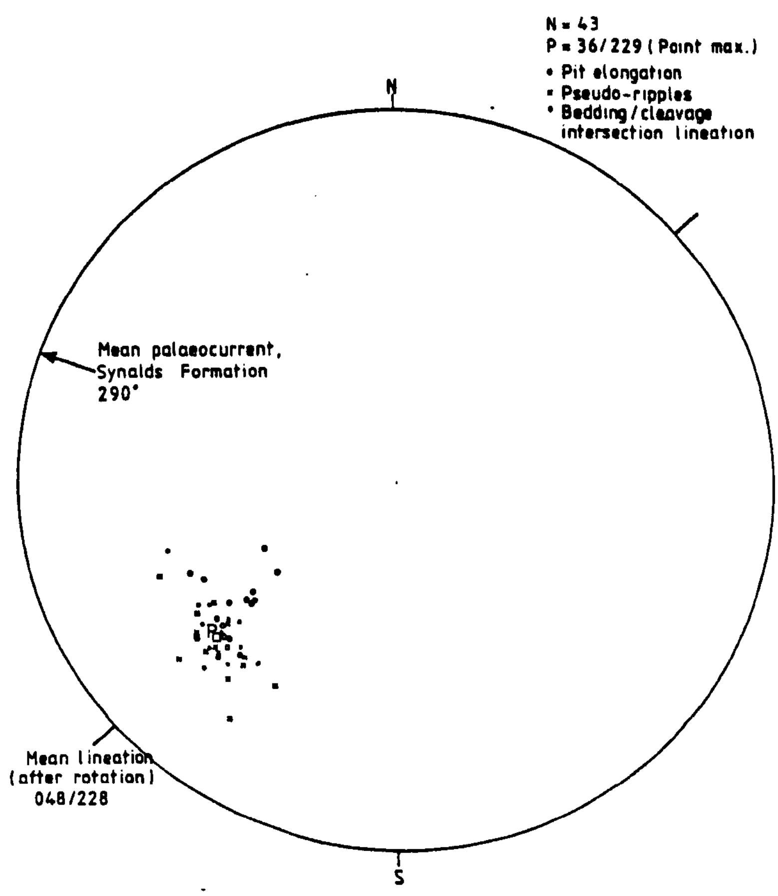
A poorly developed pebble-elongation was noted in the conglomerates of the Bayston-Oakswood Formation. A visual estimation of the average pebble elongation direction was made in the field and a single reading at each outcrop was taken. Measurements were made on both sides of the fold axis in approximately equal proportions. The orientations of the average pebble long axes were plotted on a stereograph (fig. 14, stereographic plot 8) and these define a lineation which is approximately coaxial with the major syncline fold axis. If the pebble long axis fabric was sedimentary in origin, then

FIG: 14 STEREOGRAPHIC PLOTS 8, PEBBLE LONG AXES, WENTNOR GROUP AND 9, LINEATIONS IN THE SYNALDS FMN. ASHES HOLLOW.

STEREOGRAPHIC PLOT 8. PEBBLE LONG AXES. WENTNOR GROUP



STEREOGRAPHIC PLOT 9. LINEATIONS IN THE SYNALDS FORMATION OF ASHES HOLLOW (SO 430937 area)



an approximately normal orientation to the mean palaeocurrent direction would be expected (Rust, 1972). However, the Wentnor Group pebble long axis fabric (after rotation) differs from the vector mean palaeocurrent for the Bayston-Oakswood Formation by 36° (fig. 14, stereographic plot 8). This indicates that the pebble long axis lineation is not sedimentary in origin. The analysis of deformed pebble fabrics is complicated by the presence of a primary, sedimentary pebble orientation and non-spherical, initial pebble shapes (Park, 1983). Since pebble lineation readings were taken on both limbs of the major syncline, then any primary fabric will be effectively removed by averaging of the data. It is therefore concluded that the average pebble long axis lineation fabric is a tectonic feature. The pebble long axis linear fabric is thus interpreted to represent the X axis of the strain ellipsoid and is approximately coaxial with the fold axis of the major syncline, which plunges 8/192. This value agrees with the orientations of the principal strain axes which were derived from considerations of the fault orientations: $\sigma 3$ is subhorizontal and strikes 030/210, $\sigma 1$ is subhorizontal and strikes 120/300 and $\sigma 2$ is subvertical (fig. 13).

3.9.2 Strain in the Stretton Group

Small projections on the base of laminae and their counterparts consisting of pits on the tops of the laminae are abundant in the Synalds Formation. The shape of these features varies from circular (fig. 15, photo A) to strongly elliptical (fig. 15, photo B). The origin of these structures is discussed in chapter 5. It is proposed that they were originally spherical and that deformation into elliptical shapes, in varying degrees was due to strain. Observations throughout the Long Mynd area indicate that the long

FIG: 15 STRAIN MARKERS IN THE SYNALDS FORMATION

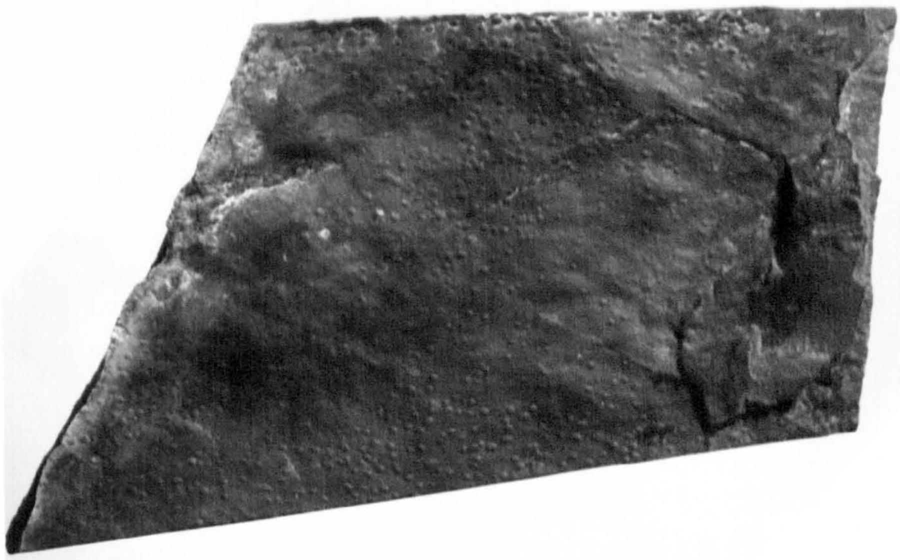


PHOTO A CIRCULAR PROJECTIONS FROM THE BASE OF A BED (Scale bar = 2cm).

PHOTO B ELLIPTICAL PROJECTIONS FROM THE BASE OF A BED PRODUCED BY STRAIN FROM THE CIRCULAR PROJECTIONS SHOWN ABOVE (Scale bar = 2cm).

axes of these features are approximately coaxial with the bedding/cleavage intersection lineations. A stereographic plot of the pit long axes was made in the area of SO 430937 in Ashes Hollow (fig. 14, stereographic plot 9). This plot shows that the pit long axes are coaxial with the bedding/cleavage intersection lineation in this area. Therefore, the X axis of the strain ellipsoid is approximately coaxial with the bedding/cleavage intersection lineation of the Stretton Group which plunges 22/210. Embleton (1976 - Univ. Liverpool B.Sc. dissertation, unpublished,p.28) came to a similar conclusion and calculated the X axis of the strain ellipsoid to plunge 27/214. Since the cleavage contains the X and Y axes of the strain ellipsoid, then the Y axis plunges 68/024. Since the Z axis is normal to the cleavage, then the Z axis plunges 2/119.

The principal strain axes can be derived from considerations of the fault orientations (fig. 13): σ_1 is approximately horizontal and strikes 145/225, σ_2 is approximately vertical and σ_3 is approximately horizontal and strikes 055/235. These values are similar to, but not coincidental with, the values derived from considerations of the other strain markers. This discrepancy may be the result of the approximations undertaken or it may reflect a slight anticlockwise rotation of the cleavage with respect to the faults. A similar anticlockwise rotation of the cleavage with respect to the bedding and of the bedding/cleavage intersection lineation with respect to the minor fold axes of the Stretton Group was previously noted.

3.9.3 The nature of the Longmyndian deformation from considerations of strain

The anticlockwise rotation of the X axis of the Wentnor Group with respect to the Stretton Group is another example of the tectonic rotation previously noted (section 3.6). It was concluded that this rotation is probably due to faulting in the Portway Formation.

Modelling of the Longmyndian deformation in terms of pure shear with a subhorizontal σ_1 axis would require a large normal deviatoric, extensional component in the horizontal plane to produce the observed subhorizontal X axes. Murphy (1985) modelled the production of a subhorizontal X axis, in a compressional régime in the Caledonides of eastern Ireland, in terms of initial horizontal shortening, with production of a vertical X axis, followed by transcurrent simple shear, which produced a subhorizontal X axis. This finite strain may also be considered to be the product of these two superimposed infinitesmal strains acting synchronously during transpression. The observed strains in the Longmyndian may be modelled in a similar manner, but it is not known whether the deformation was transpressional or compressional followed by transcurrent movement.

Murphy (1985) has modelled a clockwise transection of the cleavage with respect to the fold axes as a result of sinistral transpression. Conversely, an anticlockwise rotation of the cleavage with respect to the fold axes was considered to be the result of dextral transpression. It has been noted that the cleavage and the bedding/cleavage intersection lineation in the Stretton Group appears to have suffered a slight anticlockwise rotation with respect to the bedding, fold axial plane and fold axes (fig. 12, stereographic plot

7). This may be the result of transpression, however a dextral sense would be indicated whereas a sinistral sense was interpreted from the fault patterns (section 3.8.1).

3.10 The age of the deformation

3.10.1 The age of the folding of the Longmyndian

The Pontesford Shales, which are correlated with the Harnage Shales of the Caradoc Series, overlie the Uriconian and the Bayston-Oakswood Formation unconformably (Dean and Dineley, 1961). The Longmyndian folding therefore appears to be pre-Caradoc in age. Similarly, the Harnage Shales overlie the folded Uriconian in the Hope Bowdler area. The unconformity at the base of the Lower Cambrian, Wrekin Quartzite, which rests on faulted Uriconian, suggests that folding in the Longmyndian might have been of Precambrian age. However, this unconformity might have been due to uplift and faulting within the Church Stretton fault system, which might not have been synchronous with the main folding event of the Longmyndian.

The oldest radiometric age dates obtained from the Longmyndian are within a narrow range of 526±18 Ma to 529±23 Ma derived by Rb-Sr whole-rock analyses (Bath, 1974) and fission-track analyses (Naeser *et al*., 1982). Naeser *et al*. (1982) interpreted the fission-track ages to be the result of rapid uplift during Cambrian times. Bath (1974) interpreted the Rb-Sr whole-rock ages as a dewatering episode due to burial or folding. Consequently, it appears probable that the folding of the Longmyndian occurred before c.530 Ma (from the evidence of Naeser *et al*., 1982) or during c.530 Ma (from the evidence of Bath, 1974). From the stratigraphic evidence discussed above, folding probably occurred prior to the base of the Lower

Cambrian, Wrekin Quartzite and therefore the interpretations of Naeser et al. (1982) are more likely, that is, that the dates of c.530 Ma are due to uplift of the Longmyndian.

3.10.2 The age of faulting in the Pontesford-Linley and Church Stretton fault systems

It is generally considered that deposition of the Longmyndian Supergroup occurred in a basin which was bounded by the Church Stretton and Pontesford-Linley fault systems (e.g. Toghill and Chell, 1984 and Greig et al., 1968). However, the sedimentary facies and palaeocurrents of the Longmyndian Supergroup do not lend support to this hypothesis and it has been previously argued that the Longmyndian Supergroup has been tectonically juxtaposed with the Uriconian Volcanic Complex and that considerable strike-slip displacements might have been involved. Woodcock (1984b) recognises Caradoc to Llandovery deformation along the Pontesford Lineament, probably comprising dextral strike-slip and associated folding, followed by reactivation during later periods. Prior to this time, there is evidence for a mismatch of sequences across the Pontesford Lineament (Woodcock, 1984b). These mismatches may be explained by later strike-slip faulting. Mapping of the Longmyndian indicates that the Pontesford-Linley fault obliquely cuts across the strike of the Longmyndian Supergroup (e.g. around SJ 393006, Geological Map 2, in cover pocket). Therefore, the Pontesford-Linley fault appears to post-date the folding of the Longmyndian. Consequently, the Pontesford-Linley fault system may be wholly lower Palaeozoic in age and there is no evidence for its existence during the Precambrian from the Longmyndian.

There is some evidence to indicate that there were Precambrian movements along the Church Stretton fault system. The generally accepted unconformity at the base of the Lower Cambrian, Wrekin Quartzite suggests that some movement occurred prior to this. There is much evidence for intermittent movement of the Church Stretton fault system up to at least Carboniferous and probably Triassic times (Woodcock, 1984a and Greig et al., 1968). The facies and palaeocurrents of the Longmyndian Supergroup do not indicate that the Church Stretton fault system was in its present geographical position with respect to the Longmyndian Supergroup. However, this does not preclude the existence of the Church Stretton fault system during deposition of the Longmyndian Supergroup followed by later strike-slip faulting oblique to the Longmyndian/Uriconian basin margin. Consequently, the Church Stretton fault system might have been active during the Precambrian and possibly during the deposition of the Longmyndian Supergroup (though not in the present geographical positions).

3.11 Minor tectonic structures which simulate sedimentary structures

3.11.1 Pseudo-ripples

Many authors have noted ripple marks in the Stretton Group. Symmetrical ripple marks have been noted by Greig et al. (1968). The presence of ripple marks has been used by some authors to imply a shallow water origin for parts of the Stretton Group (e.g. Salter, 1856 and 1857 and Greig et al., 1968). However, other authors have proposed that some "ripple marks" were due to structural effects, such as corrugation of the bedding by the cleavage (Cobbold, 1900, Pocock et al., 1938 and Pocock and Whitehead, 1948).

Numerous bedding plane corrugations were noted in the laminated siltstones of the Synalds Formation and occasionally in similar lithologies in the Lightspout Formation and the Portway Formation. The corrugations in the Synalds Formation have an average height (crest to trough) of 1.2cm and an average wavelength of 5cm (fig. 16, photo A). The crests and troughs of these corrugations have rounded, subsymmetrical profiles and are discontinuous longitudinally, giving a lensoid appearance on the bedding planes. These corrugations are almost identical to similar markings in the Lightspout Formation and the Portway Formation and are probably the "ripples" and "wave-ripples" of previous authors.

It was generally noted throughout the Long Mynd area that the crests and troughs of these corrugations are subparallel to the bedding/cleavage intersection lineations. A number of readings of the corrugation orientations were taken from the Synalds Formation in Ashes Hollow (area around SO 430937) and were plotted on a stereograph (fig. 14, stereographic plot 9). This plot showed that the corrugations, in this area, are identical in orientation to the pit long axes (X axis of the strain ellipsoid) and the bedding/cleavage intersection lineations and significantly different from the normal to the mean palaeocurrent direction. Since the orientations of the corrugations are similar in the other formations, despite wide variations in the palaeocurrent direction, it is concluded that these corrugations are mostly a tectonic lineation and appear to be a type of mullion structure.

FIG: 16 MINOR TECTONIC STRUCTURES IN THE STRETTON GROUP WHICH SIMULATE SEDIMENTARY STRUCTURES.

PHOTO A PSEUDO-RIPPLES IN THE SYNALDS FORMATION,TOP OF BED (Lens cap diameter 5cm).

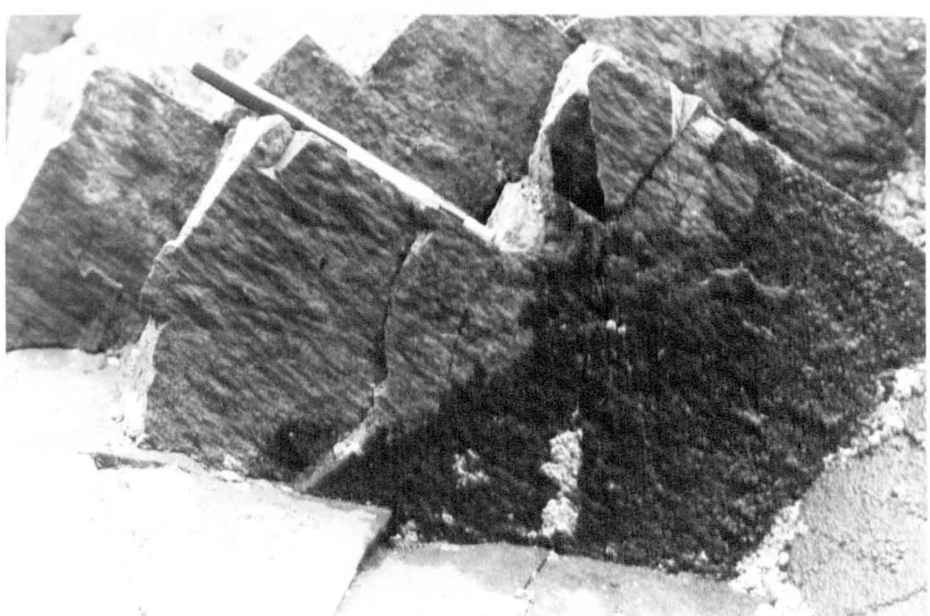
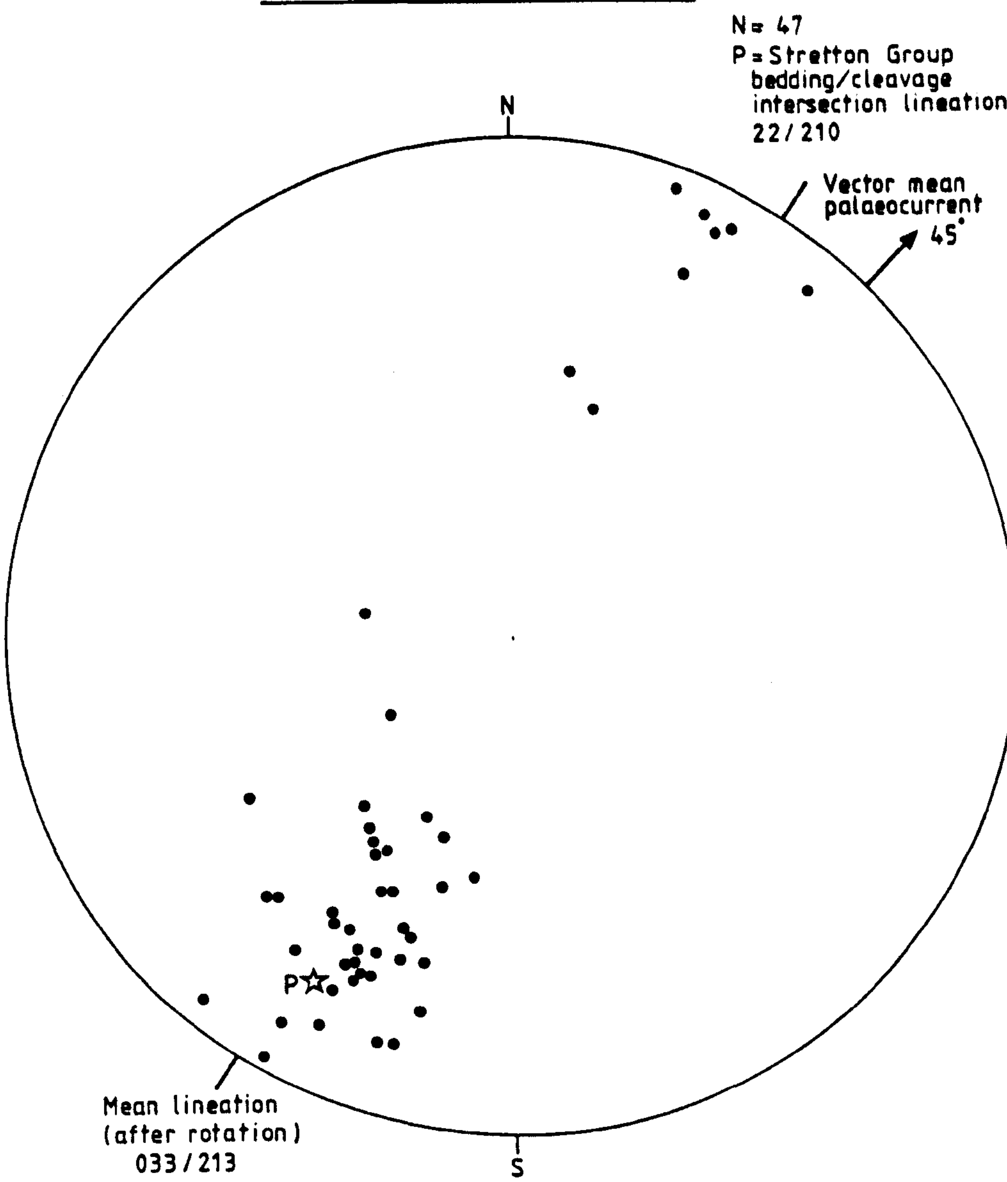


PHOTO B RIDGES AND GROOVES IN THE TURBIDITES OF THE BURWAY FORMATION (INTERPRETED AS BEDDING/CLEAVAGE INTERSECTION LINEATIONS),Pencil length c. 14cm, BASE OF BED.

3.11.2 Lineations in the turbidite facies of the Burway Formation

Lineations on bedding planes are common in the turbidite facies of the Burway Formation. They are commonly seen on the bases of sandstone beds and on the tops of the underlying siltstone units. Many of the lineations appear similar to the longitudinal ridges and grooves, which are due to current action, figured by Dżułyński and Walton (1965). Other forms of these lineations appear very similar to the bedding/cleavage intersection lineations seen elsewhere in the Stretton Group and are clearly of tectonic origin. The lineations consist of ridges projecting from the base of a bed, grooves in the base of a bed or both ridges and grooves, together with their opposing counterparts on the tops of beds. The lineations vary from straight and parallel to slightly sinuous and from continuous to very discontinuous. The wavelength of the lineations is commonly a few to several millimetres and the height commonly fractions of a millimetre (fig. 16, photo B). The orientations of the lineations were plotted on a stereograph (fig. 17, stereographic plot 10). The lineations form a cluster, whose average orientation is approximately coincidental with the average bedding/cleavage intersection lineation of the Stretton Group. The lineations cannot be separated into a clearly tectonic or possibly sedimentary grouping either in the field or on the stereographic plot. If the lineations are rotated to remove fold plunge and bedding dip, then the direction of the lineations closely corresponds with the vector mean palaeocurrent direction for the overlying subaqueous delta facies of the Burway Formation (no palaeocurrent directions were obtainable for the turbidite facies of the Burway Formation). It would thus be possible for some of the lineations to be sedimentary in origin. However,

FIG: 17 STEREOGRAPHIC PLOT 10, BEDDING PLANE LINEATIONS, BURWAY FORMATION TURBIDITES,



since all of the lineations form a discrete cluster and since this approximates to the average bedding/cleavage intersection lineation of the Stretton Group, then a tectonic origin is favoured.

CHAPTER 4

PETROGRAPHY AND PROVENANCE

4.1 Introduction

The Longmyndian Supergroup is composed of sediments which are predominantly volcaniclastic and which appear to have been derived from a magmatic arc. This magmatic arc is represented by the Uriconian Volcanic Complex and by similar late Precambrian igneous complexes represented by the Stanner-Hanter, Malvernian and Johnston igneous complexes. The petrography of the Longmyndian Supergroup can be used to delineate the evolution, nature and degree of dissection of this magmatic arc.

In addition to volcaniclastic detritus, metamorphic sources are represented by minor percentages of metamorphic clasts in the Longmyndian Supergroup. The petrography of these metamorphic clasts, their provenance and the implications concerning the nature of the late Precambrian magmatic arc are discussed.

During the analysis of the petrography, observations were made on the type of alteration of the sediment of the Longmyndian Supergroup. A very low metamorphic grade of probable prehnite-pumpellyite facies is indicated. The nature of the alteration is discussed in the following sections.

4.2 The Uriconian Volcanic Complex - general characteristics

The Uriconian Volcanic Complex is generally thought to have been the source for the volcaniclastic detritus in the Longmyndian Supergroup (e.g. Greig *et al*., 1968, James, 1952b and Toghill and Chell, 1984). The petrography of the Uriconian Volcanic Complex is summarised by Thorpe (1982) and Sabine and Sutherland (1982) and a more detailed description is provided by Greig *et al*. (1968). The

Uriconian Volcanic Complex is composed of basalt, andesite, dacite and rhyolite, associated tuffs and intrusions of dolerite and acid rocks (granophyre and rhyolite), (Thorpe, 1982 and Sabine and Sutherland, 1982). Measured sections indicate that intermediate and acid rocks predominate (Thorpe, 1982). Thorpe (1972a, 1974, 1979, 1982 and 1984) has discussed the geochemical characteristics and the plate-tectonic setting of the Uriconian Volcanic Complex. Thorpe et al. (1984) concluded that the Uriconian Volcanic Complex is calc-alkaline or transitional alkaline in character. Thorpe (1982, 1979, 1974 and 1972a) interpreted the plate-tectonic setting of the Uriconian Volcanic Complex to be a volcanic arc in a continental margin setting. Thorpe (1979) envisaged the formation of the Uriconian Volcanic Complex by partial melting of oceanic crust and mantle material followed by some contamination with crustal Sr or by melting of sialic material formed by such a system not more than 800 Ma ago. Thorpe et al. (1984) interpret the Uriconian Volcanic Complex to be part of the late Precambrian igneous basement of England and Wales, which is also represented by the Stanner-Hanter, Malvernian and Johnston igneous complexes, which have similar, though not identical, volcanic arc characteristics. The formation of the volcanic arc is generally related to late Precambrian, south-easterly dipping subduction (Thorpe et al., 1984 and Toghill and Chell, 1984).

4.3 "Tuffs" in the Uriconian Volcanic Complex

The nature of the Ragleth Tuffs has been discussed in section 2.3 and it was proposed that these "tuffs" were dominantly epiclastic (s.l.) in origin. Consequently, the Ragleth Tuffs, which were previously included with the Uriconian Volcanic Complex (e.g. James, 1952b and Greig et al., 1968) are included with the Longmyndian Supergroup and are called the Ragleth Tuff Formation.

There are a number of other "tuff" horizons which are traditionally included with the Uriconian Volcanic Complex and which have been examined at outcrop and petrographically by the author. These horizons are the Woodgate tuff beds and the Earl's Hill tuff beds. Several features suggest that some lithologies in these beds may be reworked pyroclastics or epiclastic deposits (s.s.) as opposed to primary pyroclastics (terminology after Fisher, 1961 and Suthren, 1985). The characteristics of these lithologies are discussed below.

4.3.1 The Woodgate tuff beds

The Woodgate tuff beds outcrop in the area of SO 480932. Outcrops of a very coarse, pale brown, fragmental rock were found within the mapped area of the Woodgate tuff beds (Geological Survey, sheet SO49) at SO 4813 9337 and at SO 4835 9331. These outcrops are poor and the stratigraphic relationships are therefore uncertain. This lithology is composed predominantly of plutonic rock fragments (65%) which are composed of medium crystalline quartz and subhedral plagioclase, together with isolated crystals of quartz (15%) and plagioclase (10%), which are similar to those within the plutonic rock fragments. Aphanitic and holohyaline volcanic rock fragments are present in small amounts (c. 10%). This lithology is interpreted as epiclastic (s.s.). Lithologically and compositionally it is very similar to the Helmeth Grit Member of the Ragleth Tuff Formation. The preponderence of plutonic detritus in this sandstone, as opposed to the mainly volcanic detritus present in the Longmyndian Supergroup, suggests that it might have been deposited during a late dissection of the Uriconian arc, during which time the plutonic roots of the magmatic arc were being eroded. Considering the position of this sandstone outcrop within the main outcrop of the Uriconian Volcanic Complex, it appears possible that this sandstone may be part of an unconformable mantle to the Uriconian Volcanic Complex. Alternatively, it might have been derived from a locally exposed plutonic source during the evolution of the Uriconian Volcanic Complex.

A fine to medium grained, medium purplish red, moderately well sorted and subangular fragmental rock occurs at SO 4851 9308. This rock is composed of a high percentage of quartz (50%), together with perthitic alkali feldspar (10%), plagioclase feldspar (5%) and 30% lithic fragments, which are mainly polycrystalline quartz with sutured contacts and strained extinction with minor aphanitic and vitric, volcanic rock fragments. This composition is atypical of the Longmyndian Supergroup in that the percentage of quartzose fragments is very high (fig. 18, plot WT1). This lithology is interpreted as epiclastic (s.s.) and has probably been derived from a largely plutonic and metamorphic source.

In an adjacent exposure (at SO 4860 9308, near Woodgate Cottage) occurs a medium grained, planar bedded, fragmental rock. This has been referred to as a crystal-vitric tuff by Greig et al. (1968, p. 24). Laminae which are predominantly composed of lithic grains and crystals alternate with other laminae rich in glass shards. Compositionally and texturally, the lithic and crystal laminae are similar to some sandstones in the Longmyndian Supergroup and are composed of quartz (15%), plagioclase (25%) and aphanitic and vitric volcanic rock fragments (55%). However, the composition is significantly different to that of any similar facies in the main

Longmyndian Supergroup, in the presence of angular, cuspate glass shards (5%) and vesiculate volcanic rock fragments. In the more vitric laminae, non-welded and angular glass shards comprise 60 to 90% of the total components. The origin of this rock is uncertain. The compositionally differentiated lamination might have been produced by processes operating during pyroclastic ash fall (Fisher and Schminke, 1984). Alternatively, compositional layering might have been produced by current reworking of a primary pyroclastic deposit. However, if current reworking has occurred, then this must have been only minor, since the angularity of the glass shards is preserved.

In conclusion, the Woodgate tuff beds appear to have two distinct lithologies: a sandstone composed of plutonic and metamorphic rock fragments and their derived crystals and possible pyroclastic deposits. The relationships between these two rock types and their overall stratigraphic position is conjectural.

4.3.2 The Earl's Hill tuff beds

The Earl's Hill tuff beds outcrop in the area of SJ 407048. They have been described by Boulton (1904) and Pocock et al. (1938). Boulton stated that "the tuffs of Pontesford were deposited in water is abundantly clear from the fine and regular lamination of some of the tuffs and hälleflintas and the pronounced bedding of some of the volcanic grits." (Boulton, 1904, p. 474). Similarly, Pocock et al. (1938) postulated ".... some denudation and resorting of previously existing volcanic material." (Pocock et al., 1938, p. 36). Two distinct lithologies were found in the Earl's Hill tuff beds and these are described in the following account.

Exposures at SJ 40650464 on the western flanks of Earl's Hill reveal a pale green lapilli tuff. This tuff is composed of 65% glass shards and glassy fragments, which are highly vesiculate, unwelded and which have cuspate, angular edges. The lapilli are composed of isotropic and chloritised, brown glass and they occasionally contain small plagioclase phenocrysts. They comprise the remaining 35% of the rock.

The second lithological type is exposed between Earl's Hill and Pontesford Hill at SO 4093 0507. This is a well sorted, very fine to fine grained fragmental rock with subrounded grains. The rock is thinly planar laminated and in places cross-bedded with set thicknesses of 12 cm. The rock is composed predominantly of aphanitic and glassy volcanic rock fragments (90%), with plagioclase (10%) and only traces of quartz. The volcanic rock fragments include hyalopilitic types with small plagioclase laths, which are occasionally devitrified, flow-banded rhyolites and chloritised glass. On a QFL plot, the composition is similar to some of the sandstones of the Longmyndian Supergroup (fig. 18). However, it is slightly different in containing a small percentage of vesiculate volcanic rock fragments, which are generally absent in the Longmyndian Supergroup. This lithology is interpreted as epiclastic (s.s.) from the lithology, composition, texture and sedimentary structures. This confirms the conclusions of Boulton (1904) and Pocock et al. (1938).

The relationships between these two lithologies and their overall stratigraphic position is conjectural.

4.3.3 Conclusions regarding the nature of "tuffs" in the Uriconian Volcanic Complex

It has been argued that some of the "tuff" beds commonly included with the Uriconian Volcanic Complex by previous authors (e.g. Greig et al., 1968), specifically the Ragleth Tuff Formation, the Woodgate tuff beds and the Earl's Hill tuff beds, contain distinct epiclastic (s.s.) deposits as well as pyroclastic deposits. In the case of the Ragleth Tuff Formation, it was argued that there was justification for including it with the Longmyndian Supergroup rather than with the Uriconian Volcanic Complex, since it contained predominantly epiclastic (s.s.) deposits (section 2.3). It is debatable whether the Woodgate tuff beds and the Earl's Hill tuff beds should also be included with the Longmyndian Supergroup since they contain appreciable volumes of epiclastic (s.s.) deposits.

4.4 The Longmyndian Supergroup - general characteristics

The Longmyndian Supergroup is composed of clastic sediments. Some authors have reported calcareous nodules in the Stretton Shale Formation (Cobbold, 1900, James, 1952b and Whittard et al., 1953). Reade and Holland (1908) analysed one nodule and reported a calcium carbonate content of over 50%. Most authors refer to a cone-in-cone structure as characteristic of these nodules. No calcareous nodules were found by the author. However, at SO 4405 9200, a thin layer (7 cm thick) of a highly weathered and earthy textured rock was found in the Stretton Shale Formation. Voids suggest the former presence of sparry calcite. This lithology is similar to that reported by Cobbold (1900, p. 81). These occurrences suggest that the carbonate is of diagenetic and not of sedimentary origin, since the carbonate is present only as nodules of sparry calcite exhibiting cone-in-cone structure. Calcite cement is not uncommon in the sandstones of the Longmyndian Supergroup, which attests to the presence of carbonate rich pore fluids during diagenesis.

The clastic sediments of the Longmyndian Supergroup vary from shales to conglomerates. All lithologies contain the same essential clast types. The sediments are predominantly epiclastic. Primary pyroclastic deposits, including ash flow deposits, are present and reworked pyroclastic deposits can occasionally be recognised (terminology after Suthren, 1985 and Fisher, 1961). All of the sandstones of the Longmyndian Supergroup may be classified as either litharenite or arkosic arenite (terminology after Pettijohn et al., 1972). Litharenite is the main type of sandstone, comprising at least, approximately 90% of the sediment. Arkosic arenites are present in the Wentnor Group, the Helmeth Grit, the Brokenstones sandstone beds and the Willstone Hill conglomerate beds (fig. 19). The lithic fragments in the sandstones are predominantly of volcanic origin and therefore the sandstones may be classified as volcaniclastic. However, some of the sandstones, in particular the arkosic arenites, may be classified as plutoniclastic (terminology after Ingersoll, 1983), in that they are composed mainly of fragments of plutonic rocks. Lithic fragments of metamorphic origin are a common minor component throughout the Longmyndian Supergroup. Lithic fragments of sedimentary origin are not common except shale rip-up clasts in the sandstones of the Wentnor Group, which are abundant.

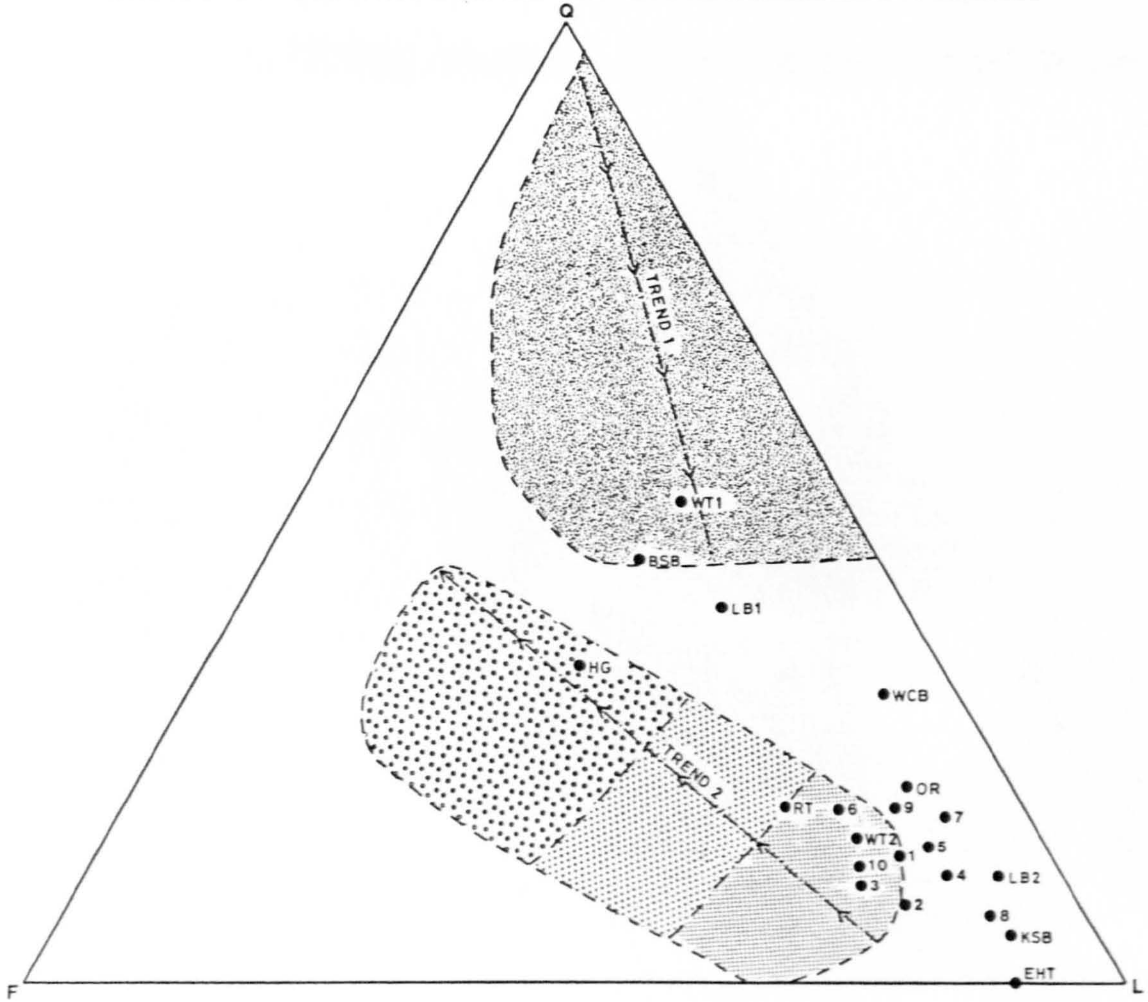
The majority of the epiclastic deposits are well sorted to moderately well sorted, based on a visual estimation. Greig et al. (1968) report an average coefficient of sorting of 1.24 and refer to the degree of sorting as good, with only a slight variation throughout the Longmyndian. Poorly sorted sediment is present as conglomerate and pebbly sandstone in the Bayston-Oakswood Formation and in the Huckster Conglomerate. The grains are generally subrounded to subangular, the greater degree of angularity being shown by feldspar crystals and feldspathic volcanic rock fragments. The release of feldspar phenocrysts from feldspathic, volcanic rock fragments can maintain the angularity of these grains during transport (e.g. Davies et al., 1978). Consequently, the sediment may be better referred to as being subrounded. Well rounded pebbles are present in the Huckster Conglomerate and the conglomerates of the Bayston-Oakswood Formation.

4.5 Compositional analysis of the Longmyndian Supergroup

Visual estimates of the percentages of the components of the sandstones were made throughout the Longmyndian Supergroup on a total of 83 thin sections. These components were classified and divided into stable quartzose grains, including monocrystalline and polycrystalline quartz (Q), monocrystalline feldspar grains including plagioclase and alkali feldspar (F) and unstable polycrystalline lithic fragments including volcanic, metavolcanic, sedimentary and metasedimentary fragments (L), (after Dickinson and Suczek, 1979). These components were recalculated such that Q + F + L equalled 100% and the percentages were then plotted on triangular QFL plots after Dickinson and Suczek (1979), (figs. 18, 19 and 20).

It is estimated that these visual estimates of percentages have a precision of ± 5%. All percentages were initially estimated to the nearest 5%. However, no point counts of the Longmyndian sediments were made, so the absolute accuracy of these data cannot be ascertained. Greig et al. (1968) provide 6 approximate modal analyses of Longmyndian sandstones (table 2, p. 63). The average

FIG.18 AVERAGE MODAL COMPOSITIONS - ENTIRE LONGMYNDIAN



KEY:

Field of recycled orogen provenances

Field of magmatic arc provenances-dissected

" " " " -transitional

" " " " -undissected

Trend 1 = increasing ratio oceanic/continental components
Trend 2 = increasing ratio plutonic/volcanic components
Q = total quartzose grains
F = total feldspar grains
L = total lithic fragments
(After Dickinson and Suczek,1979,fig.1,p2171)

KEY TO HORIZONS. (numbers in parentheses refer to number of samples)

1 Burway Formation-turbidites (7)
2 Burway Formation-subaqueous delta (5)
3 Cardingmill Grit (2)
4 Synalds Formation-green member (4)
5 Synalds Formation-red member (6)
6 Lightspout Formation (11)
7 Huckster Conglomerate (2)
8 Portway Formation (5)
9 Bayston-Oakswood Formation (13)
10 Bridges Formation (4)
WT1 Woodgate tuff beds (1)
WT2 Woodgate tuff beds (2)
BSB Brokenstones sandstone beds (3)
LB1 Linley beds (1)
LB2 Linley beds (2)
HG Helmeth Grit (6)
WCB Willstone Hill conglomerate beds (6)
OR Old Radnor (1)
RT Ragleth Tuff Formation (4)
KSB Knolls sandstone beds (2)
EHT Earls Hill tuff beds (1)

FIG. 19 MODAL COMPOSITIONS OF STRATA EXPOSED IN FAULT SYSTEMS

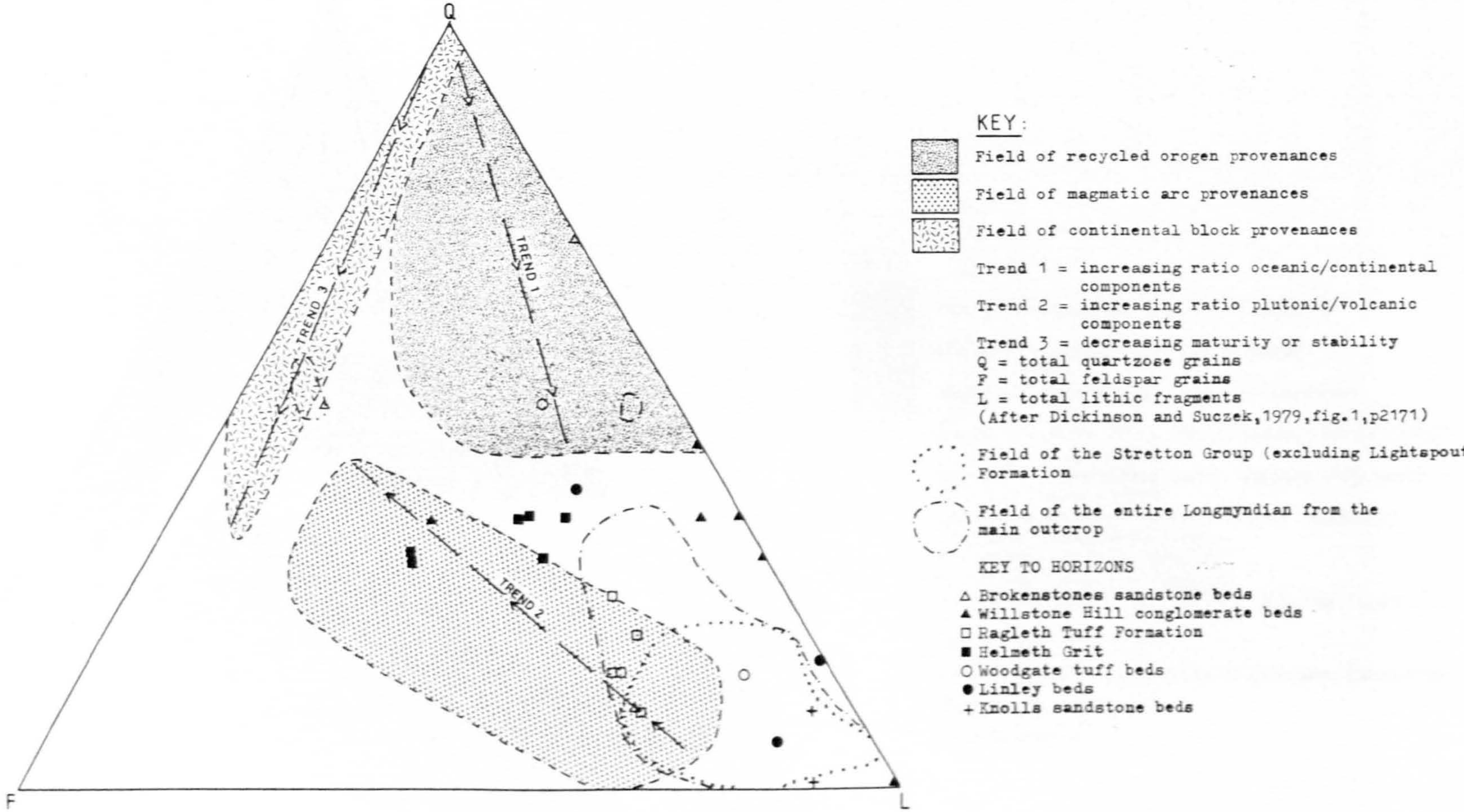
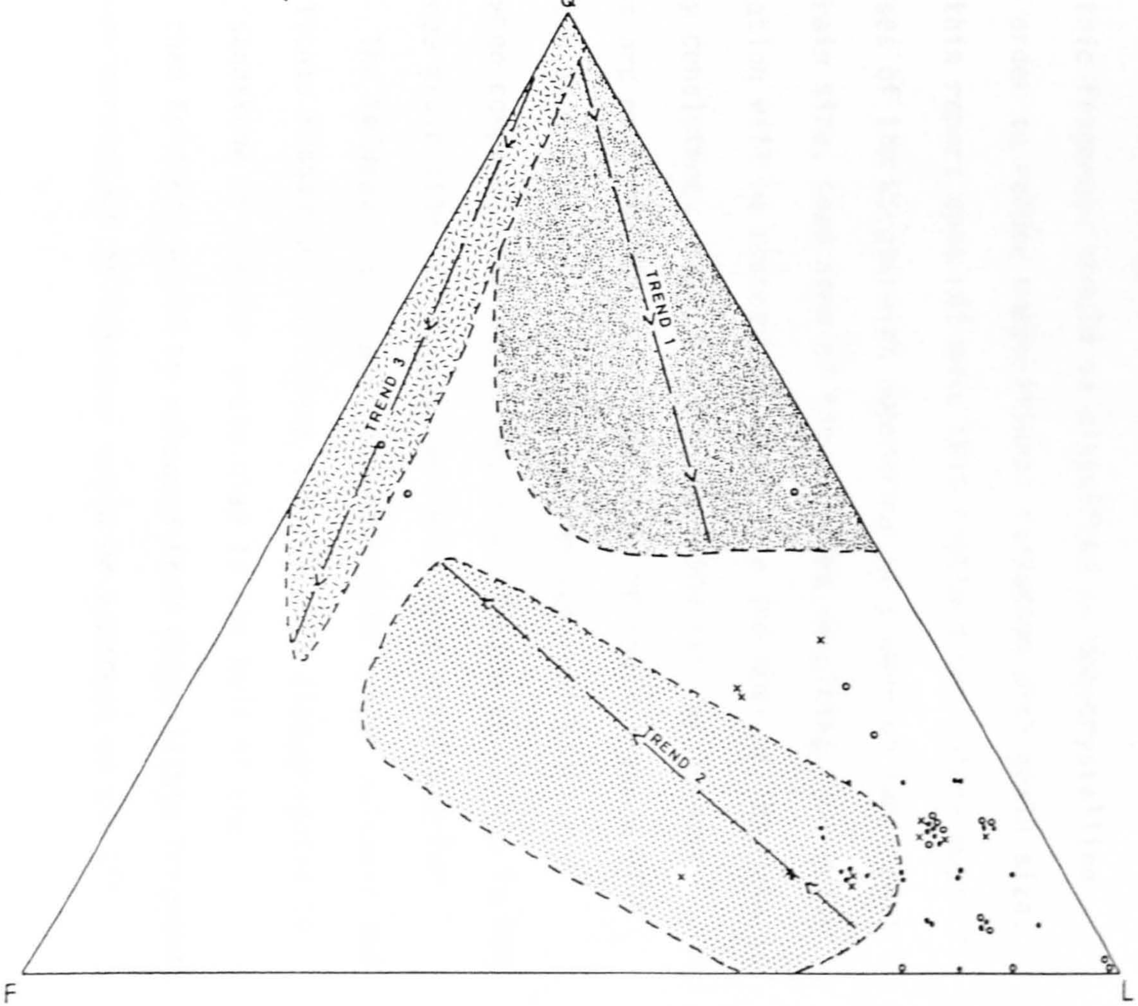


FIG. 20 MODAL COMPOSITIONS OF THE MAIN LONGMYNDIAN OUTCROP



KEY:

- Field of recycled orogen provenances
- Field of magmatic arc provenances
- Field of continental block provenances

Trend 1 = increasing ratio oceanic/continental components
Trend 2 = increasing ratio plutonic/volcanic components
Trend 3 = decreasing maturity or stability
Q = total quartzose grains
F = total feldspar grains
L = total lithic fragments
(After Dickinson and Suczek, 1979, fig.1, p2171)

KEY TO HORIZONS

- • Stretton Group (excluding Lightspout Formation)
- × Lightspout Formation
- ○ Wentnor Group

Longmyndian sandstone composition from this data is Q 35%, F 8% and L 57%. This composition differs from the data collected by this author in the higher quartz content of the former which differs by 20% on average.

Greig et al. (1968) might have included glassy and cryptocrystalline, silicic volcanic clasts with the Q component rather than with the L component or alternatively, there exists a real error in visual estimation. The former explanation is considered to be the most likely. The inclusion of rhyolitic fragments with the L component rather than with the Q component is favoured, since this reflects the volcanic source on the QFL plots.

Ingersoll (1983) proposed that crystals, larger than silt size, within lithic fragments should be classified as monocrystalline grains in order to reduce compositional variation with grain size. However, this report does not make this distinction. Since most of the analyses of the Longmyndian Supergroup were made on sandstones of similar grain size, then some of the problems resulting from grain size variation will be reduced in importance and most of the data is internally consistent. The problems resulting from grain size variations are manifest mainly in the plots of the Willstone Hill conglomerate beds (fig. 19). These show an apparent paucity of feldspar when compared with the Longmyndian sandstones. This is due to the large grain size of the samples which were available for analysis. The feldspar is mainly present within larger volcanic and plutonic lithic fragments. If these samples were disaggregated to produce a sandstone of similar grain size to the bulk of the analyses, then feldspar would be released from these lithic fragments and a higher percentage of feldspar would be apparent on the QFL plot. Bearing in mind the limitations of the data, useful conclusions can be drawn from the QFL plots in conjunction with petrographic analysis.

4.6 Clast types in the Longmyndian Supergroup

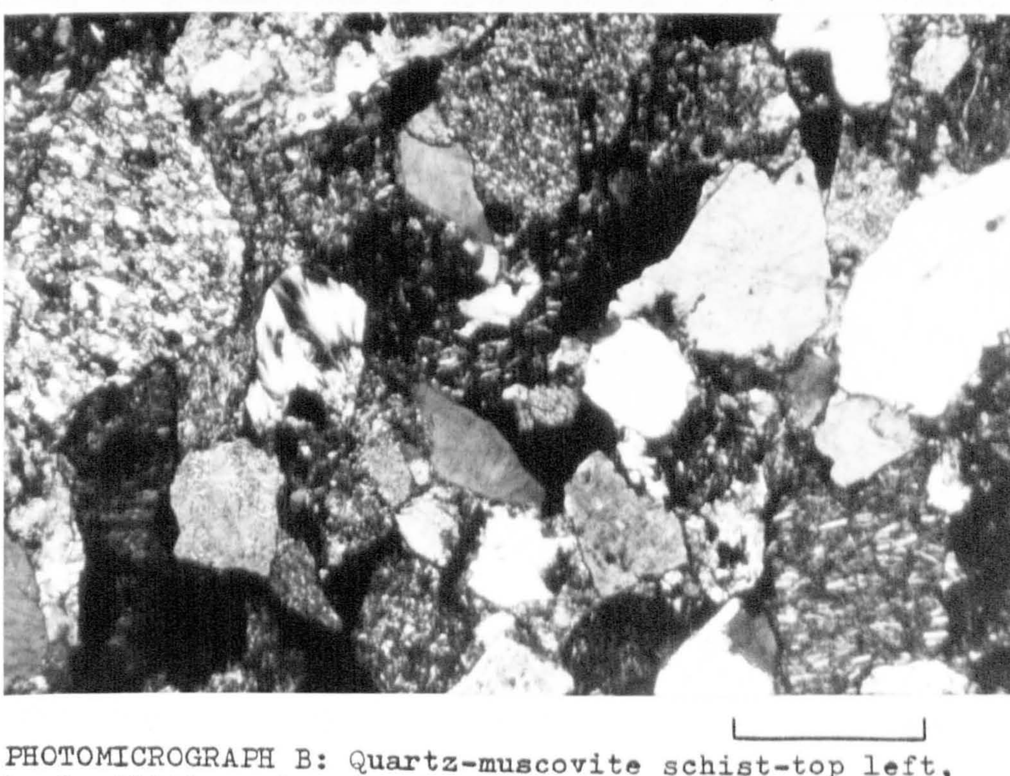
Clasts and the associated monomineralogic grains are of three main types: volcanic, plutonic and metamorphic in decreasing order of abundance. The volcanic and plutonic clast types can be directly matched with rock types present in the Uriconian Volcanic Complex. The origin of the metamorphic clasts is more problematic. A metamorphic source similar to the Rushton Schists and the Mona Complex was envisaged by Greig *et al*. (1968). The type of clasts present in the Longmyndian Supergroup are described in the following sections.

4.6.1 Plutonic rock fragments

Plutonic rock fragments are mostly represented by microgranodiorite and quartz monzodiorite. This is fine to medium crystalline, equigranular and anhedral, with common consertal textures and with occasional subhedral crystals of plagioclase (fig. 21, photo A). Microcline and muscovite are occasionally present in small percentages. Devitrified and cryptocrystalline, pale light brown glass may occur. The quartz commonly displays undulose extinction and occasionally is broken down into a fine, polygonal mosaic with dentate grain boundaries. Micrognanite occurs in subordinate amounts to the microgranodiorite. The microgranite is fine to medium crystalline and anhedral. Micrographic textures are common and occasionally granophyric. The alkali feldspar is commonly heavily sericitised and commonly displays microperthitic textures.

FIG: 21 PHOTOMICROGRAPHS ILLUSTRATING CLAST TYPES IN THE LONGMYNDIAN SUPERGROUP.

PHOTOMICROGRAPH A: Quartzite-top centre, granitic clast left margin, quartz-muscovite schist centre, feldspar-phyric volcanic clast-right centre. Lawn Hill Conglomerate. Scale bar 0.5mm.



PHOTOMICROGRAPH B: Quartz-muscovite schist-top left, hyalopilitic volcanic clast-bottom right, aphyric and cryptocrystalline volcanic clast-top centre, together with quartz-clear and feldspar crystals-turbid and sericitised. Oakswood conglomerate.
Scale bar 0.5mm.

Plagioclase occasionally occurs as phenocrysts. Mafic minerals in all these granitic rocks are represented solely by muscovite and occasional chlorite with abundant opaques, possibly after biotite.

4.6.2 Volcanic rock fragments

The volcanic rock fragments are predominantly glassy, of intermediate to acidic type and probably of mainly dacitic to rhyolitic composition (fig. 21, photos A and B). The most abundant fragment is a pale, light brown, cryptocrystalline rock. This is occasionally devitrified into an anhedral, polygonal, very finely crystalline and quartzo-feldspathic mosaic, which may vary in grain size within a fragment. Rarely, radiate or spherulitic textures may be developed. Occasionally, strained extinction is evident in the devitrified grains. Dusty hematite is widespread and is often concentrated in patches. Very fine mica is often scattered throughout and granular epidote is occasionally present. Chlorite commonly occurs in blebs and patches. Phenocrysts of anhedral and embayed quartz and opaques are rarely present. Phenocrysts of feldspar occur in some fragments and these are most commonly subhedral plagioclase. Smaller feldspar laths are common and are often flow aligned. Flow-banding can also be occasionally recognised in the glassy matrix.

In some instances, a distinct porphyritic rock can be recognised. This has predominantly subhedral plagioclase phenocrysts, which may be glomerocrystic, in a light brown, glassy matrix similar to that described previously. Some fragments of a more basic composition, which are possibly andesitic, are composed of subhedral and often flow-aligned plagioclase, with traces of anhedral quartz and anhedral, intergranular opaques with insertal, chloritised or hematised glass. Vesiculate textures are occasionally developed.

Pyroclastic rock fragments are rarely present and are mostly vitric tuffs, with small percentages of feldspar crystals, and rare lithic fragments. Varieties include welded and non-welded shard tuffs and pumice tuffs.

4.6.3 Metamorphic rock fragments

The metamorphic rock fragments are represented by variably schistose quartzites, quartz-muscovite schists and quartz-muscovite-chlorite schists (fig. 21, photos A and B). Grain sizes can be highly variable, from very fine to very coarse, both within a fragment and between fragments. Some of the large quartz grains are broken down into polygonal subgrains. The quartz usually exhibits highly undulose extinction and intergranular textures are usually consertal and dentate. A schistosity is defined by elongation of the quartz grains and this is variably developed. Muscovite is usually present in small quantities, but can comprise up to 25% of the fragment. The muscovite is usually lepidoblastic. Chlorite may occur in traces and is usually lepidoblastic, but it may also occur in irregular patches. Anhedral to subhedral opaques may occur in traces and they can also show a dimensionally preferred orientation. Other accessory minerals are epidote, which is finely crystalline and polygonal, and rare biotite, which is usually yellowish brown to green in colour.

Occasional igneous textures can be recognised, such as relict flow-banding and perlitic textures. One fragment displays a micrographic intergrowth of quartz and feldspar and has a schistosity defined by mica, elongate quartz and some opaques. Some fragments have patches of pale, light brown, devitrified glass which is often chloritised. Clusters of plagioclase are rarely present. The plagioclase is anhedral to subhedral and polygonal with a fine mica alteration. These textures and the occasional development of a schistosity in the plutonic rock fragments suggest that some of the metamorphic fragments were originally igneous rocks.

Garnet is occasionally found as individual grains, throughout the Longmyndian Supergroup, but it has never been found in any of the lithic fragments.

4.6.4 Sedimentary rock fragments

Sedimentary rock fragments are rare in the Longmyndian Supergroup and they have only been found in the Willstone Hill conglomerate beds and the Lawn Hill Conglomerate by the author. These fragments are sandstones which are similar to those in the Longmyndian Supergroup. They are principally composed of cryptocrystalline or feldspar-phyric volcanic rock fragments, with subordinate quartz and minor plagioclase. The sandstones show the same type of alteration as the Longmyndian Supergroup. Greig *et al.* (1968) also found fragments of sandstone of Longmyndian type in the Stanbatch, Darnford, Lawn Hill and Oakswood Conglomerates.

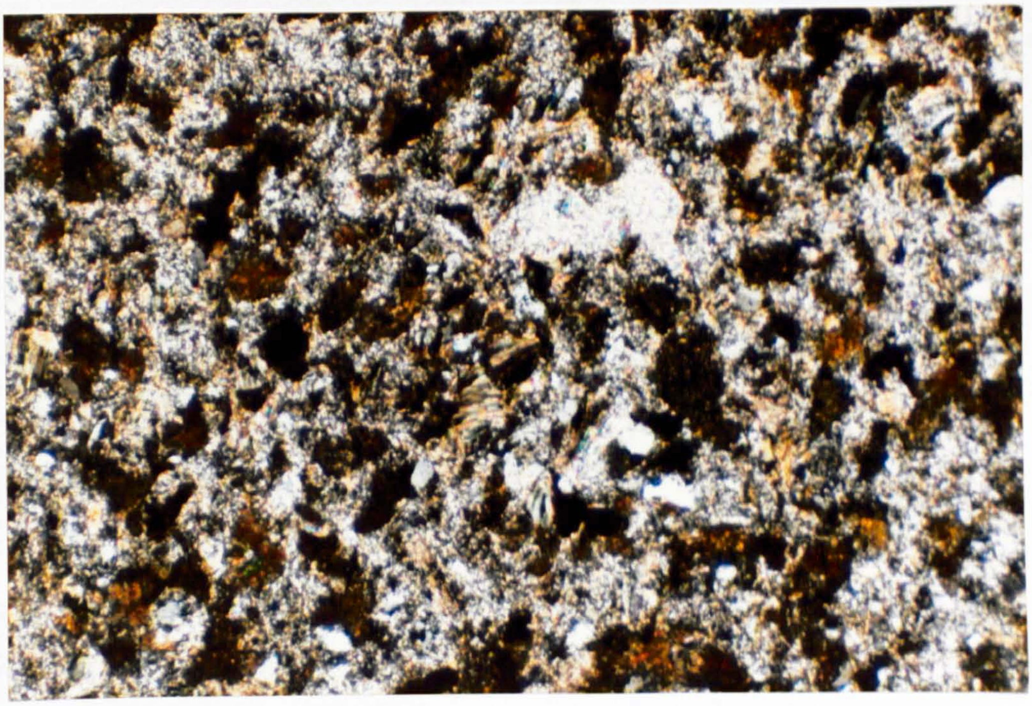
4.7 Alteration of the Longmyndian Supergroup

Alteration products of the sediments of the Longmyndian Supergroup include epidote, chlorite, mica, calcite, hematite and sphene. These minerals are also commonly found within veins.

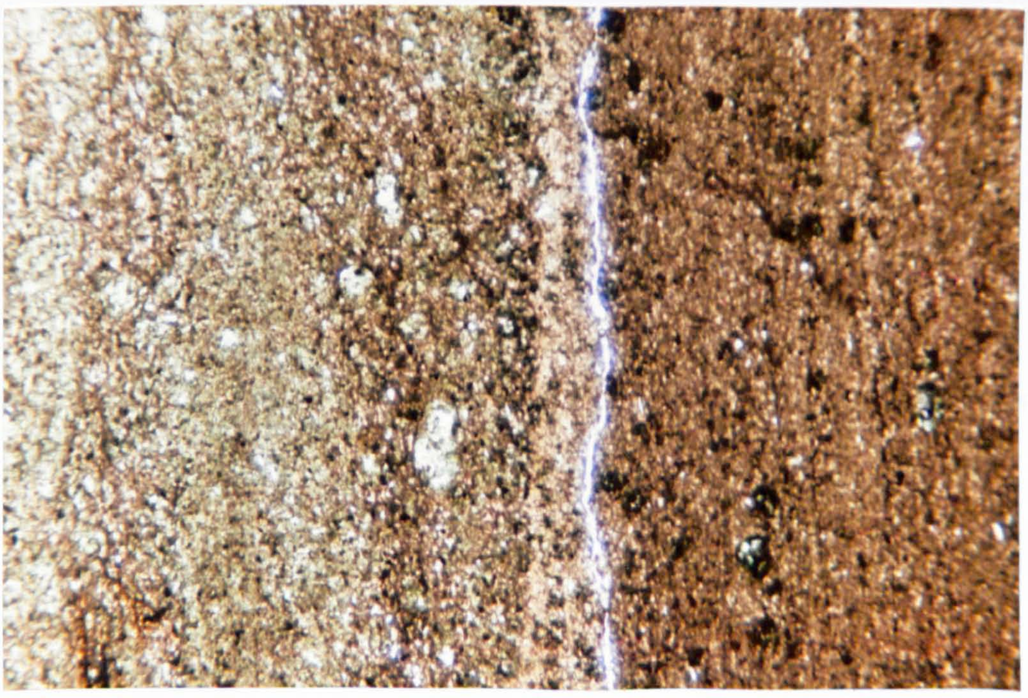
Feldspars are commonly partially altered, with the development of chlorite patches and scattered very fine mica. Epidote is the most widespread alteration product and is found replacing the glassy volcanic fragments as well as within the pore spaces (fig. 22, photo A). Epidote occurs as dusty granules, as clusters of fine, granular crystals and occasionally as isolated, subhedral, fine crystals. Epidote may comprise up to 50% of a sample and, in the cases where it is abundant, it is commonly found concentrated in laminae which may reflect an original compositional variation. Chlorite is common and may be found in patches, replacing the volcanic rock fragments, or in the pore spaces (fig. 22, photo B). Chlorite may also occur in large flakes, often as stacks with minor mica (fig. 22, photo A). Some chlorites contain lamellae rich in opaques and occasionally, lamellae of biotite, which suggest that some of the chlorites might have been derived by the alteration of biotite, which is present in small percentages in some samples. Calcite is common and occurs as irregular, replacive patches and rarely in small concretions. Fine mica commonly replaces the glassy volcanic fragments together with chlorite. Hematite occurs as very fine, dusty particles which are found within pore spaces, but it may also be replacive. Detrital opaques are commonly replaced by sphene and occasionally by hematite. Some sandstones in the Bayston-Oakswood Formation are silicified, with only remanent patches of sandstone, within a very fine quartzose matrix. This silicification is commonly accompanied by barytes mineralisation.

Some of the alteration products may occur in subspherical bodies of coarse to very coarse grain size, usually within altered siltstones (fig. 22, photo B). In places, these subspherical bodies become highly irregular and grade into the more usual, patchily

FIG: 22 PHOTOMICROGRAPHS ILLUSTRATING THE TYPES OF ALTERATION IN THE LONGMYNDIAN SUPERGROUP.



PHOTOMICROGRAPH A: Highly altered siltstone with muscovite/ chlorite stacks-centre,anhedral epidote-brownish colours and groundmass rich in fine mica and chlorite.Relict cross lamination is visible in hand specimen.Lightspout Formation.Scale bar 0.25mm.



PHOTOMICROGRAPH B: Siltstone,highly chloritised with patches of chlorite.Patches on the right side have thin rims of dusty hematite and ?epidote.Synalds Formation. Scale bar 0.5mm.

developed alteration. Thin veins may connect some of these subspherical bodies. In some instances, remanents of the original lithology may be found within the bodies and the margins of the bodies digitate with the matrix. These features suggest that these bodies have originated by growth within the matrix rather than by the replacement of an originally spherical object. The subspherical bodies usually consist of a discrete centre and a thin rim. The centre is usually occupied by radiate chlorite and occasionally by radiate calcite. The rims are usually composed of epidote, occasionally with hematite.

The sediments in the Longmyndian Supergroup are predominantly greenish grey or purplish red. These colour differences reflect the nature of the alteration rather than the composition of the clasts. The greenish grey sediments have abundant chlorite and common fine mica which occur in chloritised and altered glass and in the pore spaces, whereas hematite only occurs in traces. The purplish red sediments, in contrast, have abundant dusty hematite and less chlorite and fine mica than the greenish grey sediments.

Greig et al. (1968) found prehnite in vesicles and veins in the Uriconian Volcanic Complex. This is associated with quartz, epidote and chlorite. Possible pumpellyite was found in some samples of the Uriconian Volcanic Complex by the author, but microprobe analysis would be necessary to confirm its presence. The mineral phases which are found throughout the Longmyndian Supergroup and the Uriconian Volcanic Complex indicate a low grade metamorphism of probable prehnite-pumpellyite facies.

4.8 Provenance of the Longmyndian Supergroup

The volcanic and plutonic clasts are very similar in type to the lithologies of the Uriconian Volcanic Complex. The Uriconian Volcanic Complex or another very similar late Precambrian volcanic complex was therefore the main source for the Longmyndian sediments. Granitic rocks are not widespread in the Uriconian Volcanic Complex but are represented by the Ercall Granophyre.

The metamorphic clasts are of less obvious derivation. A schistosity was noted in some of the plutonic rock fragments and therefore, some of the metamorphic clasts might have been derived from variably metamorphosed plutonic rocks. However, many of the metamorphic clasts are quartzites which are not obviously derived from plutonic rocks because of their highly quartzose composition. Additionally, many of the quartzites are more schistose than any of the schistose plutonic clasts. There are no similar schistose quartzites present in the Uriconian Volcanic Complex and a schistosity is not seen in any of the lithologies of the Uriconian Volcanic Complex.

Lambert and Holland (1971) interpret the Malvern igneous complex to consist mostly of diorites and tonalites which have suffered intense shearing and hydrolysis. This shearing resulted in the alteration of hornblende to chlorite and epidote, the chloritisation of biotite and the alteration of plagioclase to muscovite. The metamorphic phases present in the Malvernian igneous complex are very similar to those present in the metamorphic clasts of the Longmyndian Supergroup. Lambert and Holland (1971) note the presence of quartz aggregates in the leucocratic granites of the Malvernian igneous complex. This texture was attributed to solid-state recrystallisation of the primary feldspar and a transition to metamorphic textures.

This texture is apparent in some of the clasts in the Longmyndian Supergroup. Lambert and Holland (1971) also note the presence of fracture zones, which contain schistose rocks which consist of quartz, oxides and hydrolised silicates and which lack feldspar and biotite. These represent the last stage of crystallisation of the complex which was accompanied by brittle deformation. Both of these mechanisms might have provided the quartzitic metamorphic clasts in the Longmyndian Supergroup. Therefore, it is considered that the Malvernian igneous complex, or an igneous complex of similar type, could have contributed both metamorphic and plutonic detritus to the Longmyndian basin.

Garnet is found in traces throughout the main Longmyndian Supergroup. However, no lithic clasts containing garnet have been found. The Rushton Schists are varied in character. One example is a quartz-muscovite-biotite-chlorite-garnet schist with poikiloblastic and idioblastic garnets. The Primrose Hill "gneiss" is of a similar grade to the Rushton Schists and also contains garnets. In the Malvernian igneous complex, garnetiferous rocks are found in small areas which constitute less than 0.1 percent of the complex. These are biotite-muscovite-garnet schists which carry garnets retrogressed to chlorite. Other garnetiferous rocks, including biotite-plagioclase -quartz-garnet schist, are found as inclusions within diorite (Lambert and Holland, 1971). It is thus likely that the garnet in the Longmyndian Supergroup has been derived from metamorphic rocks which are similar to the Rushton Schists, Primrose Hill gneiss and the Malvernian garnetiferous rocks.

Since all of the fragments in the Longmyndian Supergroup could have been derived from volcanic, metamorphosed igneous and metamorphosed sedimentary rocks, which are now represented by the

Uriconian Volcanic Complex, the Malvernian igneous complex and the Rushton Schists, Primrose Hill gneiss and Malvernian garnetiferous rocks in southern Britain, it is not necessary to postulate that the Mona Complex acted as a source for the Longmyndian Supergroup (as postulated by Greig et al., 1968).

The geochemical composition of sediments may be related to the plate tectonic settings of their sedimentary basins (Bhatia, 1983). Bhatia (1983) was able to distinguish between oceanic island arc, continental island arc, active continental margin and passive continental margin basin settings by using the overall chemical composition of the sediments and by the use of some discriminating parameters. From the geochemical data provided by Reade and Holland (1908) and by Greig et al. (1968, table 5, p. 67), it is possible to compare the composition of the Longmyndian Supergroup with the data of Bhatia (1983), (table 2). The data from the Longmyndian Supergroup compares best with the data from Bhatia's continental island arc setting. Thorpe (1982, 1979, 1974 and 1972a) argued that the Uriconian Volcanic Complex was a continental margin island arc, using the geochemical data from the volcanics.

Dickinson and Suczek (1979) established distinctions in sandstone compositions with respect to type of provenance by means of QFL plots. QFL plots of the Longmyndian Supergroup (figs. 18, 19 and 20) show that the Longmyndian Supergroup plots close to the field of magmatic arc provenances defined by Dickinson and Suczek (1979). In addition, Valloni and Maynard (1981) differentiated forearc from back-arc sediments on the basis of the preponderence of vitric volcanic clasts and the higher lithic component, which could be differentiated on a QFL plot (Valloni and Maynard, 1981, fig. 6, p. 80). The majority of the Longmyndian Supergroup plots close to the

TABLE 2 GEOCHEMICAL COMPOSITION OF THE LONGMYNDIAN SUPERGROUP AND PLATE TECTONIC SETTING.

	AVERAGE LONGMYNDIAN SUPERGROUP	CONTINENTAL ISLAND ARC	OCEANIC ISLAND ARC
SiO_2	68.62	70.69	58.83
TiO_2	0.64	0.64	1.06
Al_2O_3	15.89	14.04	17.11
Fe_2O_3	4.82	1.43	1.95
FeO	1.48	3.05	5.52
MnO	0.12	0.10	0.15
MgO	2.08	1.97	3.65
CaO	2.01	2.68	5.83
Na_2O	2.58	3.12	4.10
K_2O	1.70	1.89	1.60
P_2O_5	0.07	0.16	0.26
Fe_2O_3*+MgO	8.54 (6-10)	6.79 (5-8)	11.73 (8-14)
Al_2O_3/SiO_2	0.23 (0.1-0.3)	0.20 (0.15-0.2)	0.29 (0.24-0.33)
K_2O/Na_2O	0.66 (0.4-1.3)	0.61 (0.4-0.8)	0.39 (0.2-0.4)
Al_2O_3/Na_2O+CaO	3.46 (2.4-5.9)	2.42 (0.5-2.5)	1.72 (1-2)

All Longmyndian Supergroup analyses recalculated to 100% on a volatile free basis. Averages and ranges from 6 analyses given by Greig et al. (1968, table 5, p67).
Comparative analyses of continental and oceanic island arc sediments from Bhatia (1983).
Fe_2O_3*=total iron as Fe_2O_3.
Numbers in parentheses are ranges of the analyses.

field of forearc basin sediments, as defined by Valloni and Maynard (1981). In addition, the preponderance of cryptocrystalline, glassy volcanic clasts in the Longmyndian Supergroup, as opposed to feldspar rich volcanic clasts or other lithic fragments agrees with the characteristics of forearc basin sediments (Valloni and Maynard, 1981 and Maynard et al., 1982). Dickinson and Suczek (1979) define a trend of increasing plutonic to volcanic components during dissection of a magmatic arc by erosion (trend 2, figs. 18, 19 and 20) and were able to distinguish undissected, transitional and dissected magmatic arc provenances. The majority of the Longmyndian Supergroup plots close to the field of undissected magmatic arc provenances defined by Dickinson and Suczek (1979), (fig. 18).

The QFL modal compositions of the entire Longmyndian Supergroup show a variation in composition which subparallels the trend of increasing plutonic to volcanic detritus (fig. 18). This variation in the Longmyndian Supergroup can be interpreted in terms of the variability of plutonic detritus as opposed to volcanic detritus. In the main Longmyndian Supergroup there appears to be no systematic change in the average composition with stratigraphic level (fig. 18, plots 1 to 10). However, plots of the QFL modal compositions of the main Longmyndian Supergroup (fig. 19) show that the Wentnor Group, including the Brokenstones sandstone beds and the Willstone Hill conglomerate beds, possesses a great variety of compositions from highly lithic to highly quartzo-feldspathic, whereas the Stretton Group, excepting the anomalous Lightspout Formation, is restricted to mainly lithic compositions. This variability could be a result of the dissection of the magmatic arc source, resulting in the exposure of plutonic as well as volcanic rocks during the deposition of the Wentnor Group. Some spatial variability in source is recognisable in the Wentnor Group. The Haughmond Conglomerate is unusually rich in andesitic detritus. The anomalous composition of the Haughmond Conglomerate and its thinning and eventual absence to the south, suggest a local andesitic and probably uplifted source. However, the close correspondence of the average composition of the Bayston-Oakswood Formation with that of a sample of the lithologically similar Old Radnor sandstone (fig. 18) demonstrates the general lack of overall spatial variation in the type of source over wide areas.

The most notable feature of the QFL plots is the relatively quartzo-feldspathic composition of the Helmeth Grit (fig. 19). As a consequence, it has previously been argued (section 2.3.8) that the Helmeth Grit may come from a stratigraphically higher level than the Stretton Group. The sandstones of the Ragleth Tuff Formation appear to be less quartzo-feldspathic than the Helmeth Grit, but on average more quartzo-feldspathic than the average Stretton Group and consequently a similar argument may apply.

The composition of the Knolls sandstone beds is in general more lithic than the compositions of the Longmyndian Supergroup (fig. 18). The petrographic and lithological characters of the Knolls sandstone beds indicate that the sediment is relatively immature. For these reasons, it was previously argued that the Knolls sandstone beds may be more proximal representatives of the Longmyndian Supergroup, with respect to the volcanic source (section 2.8).

The modal compositions of the Brokenstones sandstone beds and Willstone Hill conglomerate beds show a wide variation and they are more quartzo-feldspathic than the Stretton Group (fig. 19). For these reasons and because of the similarities in facies and lithology to the sandstones of the Wentnor Group, it has previously been argued (chapter 2) that the Brokenstones sandstone beds and the Willstone Hill conglomerate beds should be correlated with the Wentnor Group. The majority of the plots of the Willstone Hill conglomerate beds show a paucity of feldspar. However, the analyses are mostly based on the composition of large clasts and quartzo-feldspathic plutonic clasts have been counted as lithic fragments. One analysis of a sandstone gave a more representative modal composition of quartz (35%), plagioclase (35%), glassy volcanic rock fragments (30%) and matrix (10%). The average modal composition of the Willstone Hill conglomerate beds is more quartzose than the average modal composition of the Wentnor Group (fig. 18). This difference may reflect a difference in stratigraphic level, with the enrichment of plutonic rock fragments, quartzites and quartz schists following the greater exposure of the metamorphic and plutonic basement during arc dissection. This observation agrees with the conclusion that the Willstone Hill conglomerate beds may overlie the Uriconian Volcanic Complex unconformably (section 2.5), the inference being that the Willstone Hill conglomerate beds may represent a late stratigraphic equivalent of the Wentnor Group which has overstepped onto the Uriconian Volcanic Complex.

The highly quartzose sandstone present in the Woodgate tuff beds (fig. 18, plot WT1) and a sandstone rich in plutonic rock fragments, both of which were previously discussed in section 4.3.1, may represent part of an unconformable mantle to the Uriconian Volcanic Complex, which is postulated to be represented by the Willstone Hill conglomerate beds, on the basis of their quartzose and plutonic character.

The Brokenstones sandstone beds also occasionally possess lithologies which are rich in quartz (fig. 19). However, because of the wide variations in modal composition and because of the small number of analyses, no valid distinctions can be made from the Wentnor Group.

CHAPTER 5

PALAEONTOLOGY, PROBLEMATICA AND PSEUDOFOSSILS

5.1 Introduction

Numerous, small, ring-shaped, blister and dimple-like bedding plane markings are present beneath the Cardingmill Grit, at the top of the Burway Formation. In addition, larger circular structures, which are raised in positive epirelief with central pits, occur at the same horizon. In the Synalds Formation, small pits occur on the bedding planes in abundance. In addition, fine, subparallel, raised ridges are common and the two may be found together. Occasionally, these features can be found in the finer grained lithologies of the Portway Formation and the Lightspout Formation. Small, sinuous and curved ridges (c. 1mm across) can rarely be found in the Lightspout Formation. One sample from the Green Synalds Member of the Synalds Formation shows small (fractions of a mm across), curved and straight lineations in positive and negative hyporelief which are associated with small, circular blisters.

Most of the above have been commented on by previous authors (Salter, 1856 and 1857; Blake, 1890; Cobbold, 1900; Watts, 1925; James, 1952b; Greig et al., 1968; Bland, 1984 and Peat, 1984a). The interpretations of some of these features have ranged from biogenic to inorganic. In many cases, no definite conclusions have been reached about the origins of these features in the present study. Some of the biogenic interpretations are questioned (e.g. Arumberia of Bland, 1984). The following therefore provides a descriptive account of these features and discusses the various interpretations without, in many cases, reaching a definite conclusion about their origins.

5.2 Large circular features

These are present in the thinly laminated siltstones at the top of the Burway Formation. Several examples were found in Ashes Hollow at SO 43399297. These examples all occur on one bedding plane with a spacing of 2.5 to 4cm. The feature is rare and has been found at approximately the same horizon in Callow Hollow and rarely, lower down in the Burway Formation of Ashes Hollow.

The features consist of a circular depression, 12 to 16mm in diameter and approximately 1.5mm deep, on the top surfaces of bedding planes. They commonly have a diffuse, raised outer margin approximately 1.5mm high. Occasionally, narrow concentric rings can be recognised at the transition from the raised outer margin to the central depression. Fig. 23, photo A shows three of these features on the top surface of a bedding plane. Similar features are shown by Greig *et al*. (1968, plate 5, fig. A). Bland (1984) refers to similar raised structures 5-50mm in diameter (p. 629). Greig *et al*. (1968) record domes up to 3cm in diameter and up to 3mm high (p. 70).

Greig *et al*. (1968) interpret these structures as "pit and mound structure ('mud volcano')" (p.xi) and note the similarities between them and air-heave structures, which are caused by rising air bubbles, and sand domes, which are found on present day beaches. Bland (1984, p.629) states that they are "..., possibly related to *Cyclomedusa*?" and appears therefore to favour a biogenic interpretation. Clues to the origin of these structures are provided by cross-sections (fig. 23, photo B). The surface morphology is reflected in the similar morphology of the underlying silt laminae, the disturbance of which may extend downwards for at least 7mm (Greig *et al*., 1968, p.70) and Bland (1984, p.629) notes a stacking of the feature in columns through 0.2 to 10cm of finely laminated sediment.

FIG:23 LARGE,CIRCULAR FEATURES FROM THE TOP OF THE BURWAY FORMATION,ASHES HOLLOW,SO 4339 9297.

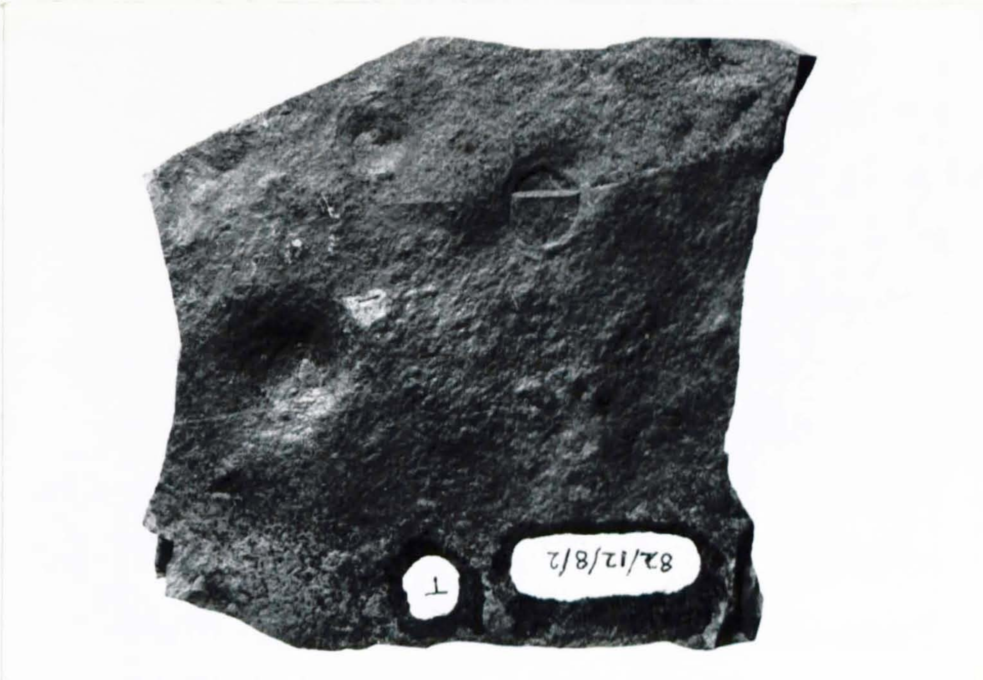


PHOTO A: Three large,circular features consisting of low, raised,marginal rims with a central depression.Top of bed. Scale bar = 2cm.

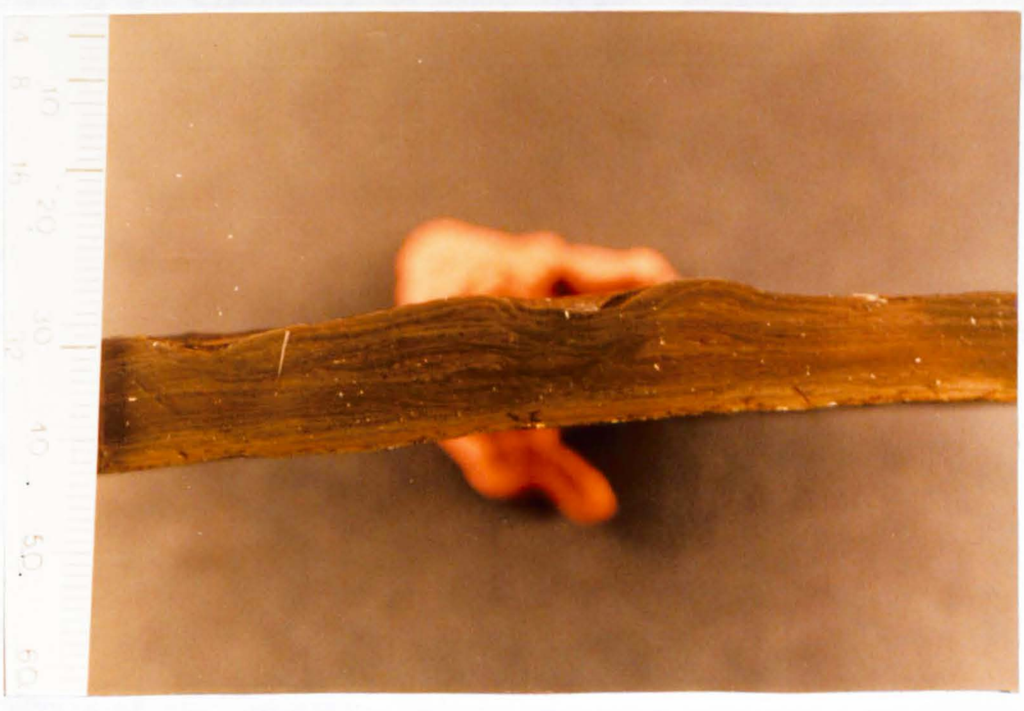


PHOTO B: Cross-section through two,large,circular features similar to those shown in photo A.Note the continuation of the surface feature downward through the underlying,thinly laminated silt,the absence of a central piercement zone and the deformation of the laminae both beneath the rim and the central depression.Scale bar is graduated in mms.

Greig et al. (1968, p.70) note that the central surface depression is underlain by broken and deformed lamination which partly truncates the underlying bedding planes. The disturbed nature of the lamination, which underlies the central depression, is apparent in fig. 23, photo B. However, the lamination does not appear to have been pierced by a highly disturbed central zone, nor by sediment of significantly different grain size. Oldershaw (1960) interprets very similar structures of Lower Proterozoic age as sand volcanoes, which formed during the dewatering of a turbidite bed. These sand volcanoes are 6 to 100mm in diameter and occur in laminated siltstones. However, a direct comparison cannot be made, since the central depression in these sand volcanoes is underlain by a narrow pipe of coarser material which pierces the lamination and this is not present in the Longmyndian structures. Cloud (1960) describes a mechanism by which a raised ring and central depression may be produced by gas ejection or by water escape. Plummer (1980) describes similar features (figs. 1b and 1c), 6-20mm in diameter and with a relief of up to 2mm in the Late Precambrian, Moorillah Formation of the Flinders Ranges, S. Australia. The proposed mechanism of formation is by gas or air escape. Twenhofel (1921) proposed that the discharge of upward currents during the settling of large quantities of suspended sediment can produce small mounds 3-10mm in diameter with small, crater-like depressions on their summit.

If the feature were the impression of a medusoid, as is suggested by Bland (1984), then it would be unlikely that the disturbance of the lamination would extend far below the surface. However, the disturbance can extend for at least 7mm through the sediment. The favoured mechanism of formation is by updoming of the lamination, followed by collapse of the centre after removal of the cause of the upwelling. This would explain the lack of any piercement structures. The upwelling might have been caused by the localised accumulation of trapped gas, air or water. The accumulation of gas is not favoured, since this would have required a significant proportion of organic material. Although Peat (1984a) has noted the presence of microfossils, they only appear to occur in large numbers in some beds of the Synalds and Lightspout Formations. It is unlikely that sufficient gas could therefore have been generated. The accumulation and trapping of air in sediment can occur on floodplains and in intertidal environments (Twenhofel, 1921). Plummer (1980) postulated that trapped air or gas could migrate to the surface during low tide forming craters 6-20mm in diameter and with a relief of up to 2mm. However, the environment of deposition in which the Longmyndian features are found is interpreted as a prodelta and consequently the trapping and accumulation of air in the sediment is unlikely. Consequently, these features are thought to be due to the accumulation of water, accompanied by soft sediment deformation, producing a domed structure, followed by dissipation, probably by upward migration through tensional cracks developed in the lamination as a consequence of the upwelling, and subsequent collapse of the central part of the dome into the area from which the water has been expelled. This mechanism is similar to those proposed by Oldershaw (1960), Cloud (1960) and Plummer (1980) for the production of similar structures.

5.3 Small, ring-like features

These are present in the thinly laminated siltstones at the top of the Burway Formation, directly beneath the Cardingmill Grit. These siltstones are interbedded with thick to very thick-bedded, fine grained sandstones. Numerous examples have been found by the author in Ashes Hollow at SO 43399302, where they are found in abundance on certain bedding planes. They are present at the same horizon throughout the outcropping area of the Longmyndian Supergroup.

The features most commonly consist of circular rings (occasionally elliptical due to strain) 4mm or less in diameter (fig. 24, photo A). The circumference of the ring is raised in positive hyporelief and is also found as impressions in negative epirelief. Thin sections through these rings reveal that the interiors of the ring are composed of silt and clay which is identical to that of the surrounding sediment (fig. 25, photos A and B). Fig. 25, photo B shows that the interiors of the rings can be very thinly laminated. In this case, the lamination is oblique and truncated, suggesting considerable soft sediment deformation in the lamina in which the ring structures are now found. Most commonly, the ring features occur in the finer grained laminae. Occasionally, ring structures can be found in positive epirelief (fig. 26, photo A) and in corresponding impressions in negative hyporelief (fig. 24, photo B). A cross-section through the specimen shown in fig. 26, photo A, shows that these positive epirelief structures occur on top of a sandstone bed and are in continuity with and compositionally indistinct from it (fig. 26, photo B). They are therefore significantly different from the positive hyporelief structures noted previously, both in the lithology in which they occur and of which they are composed and in

FIG: 24 PHOTOGRAPHS OF SMALL RING-LIKE FEATURES FROM THE TOP OF THE BURWAY FORMATION.

PHOTO A: Small ring-like features projecting from the base of thinly laminated siltstone (positive hyporelief), Ashes Hollow. Scale bar graduated in cms and mms.

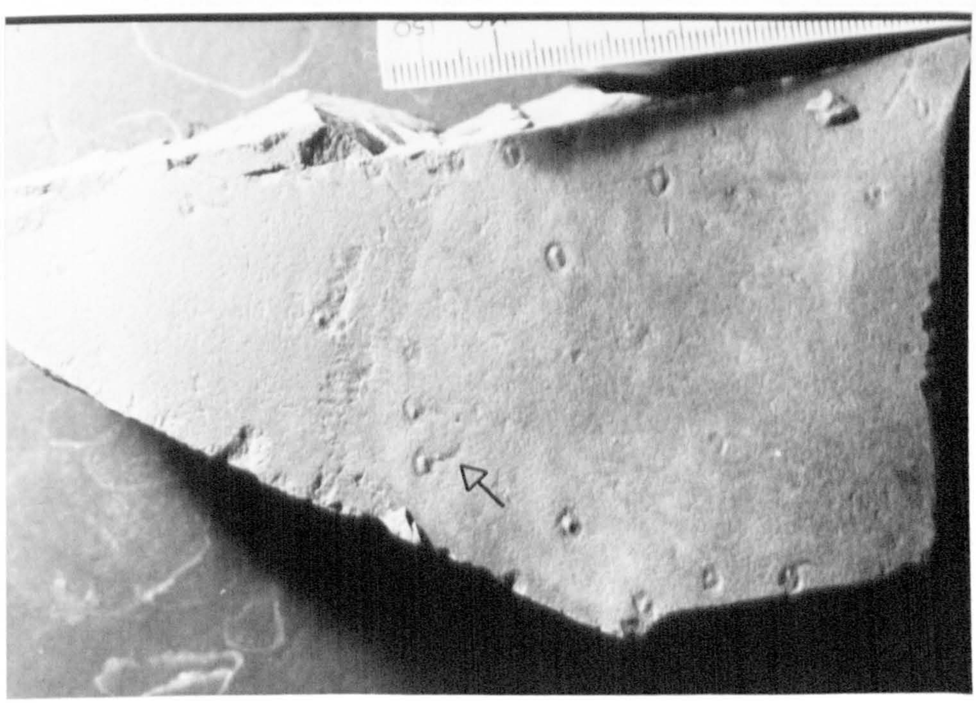


PHOTO B: Small, ring-like features impressed in the base of a bed (negative hyporelief), Haughmond Hill. Note the small, curvilinear depressions which extend from some of the rings. Scale bar graduated in mms.

FIG: 25 PHOTOMICROGRAPHS OF THIN-SECTIONS THROUGH SMALL, RING-LIKE FEATURES FROM THE TOP OF THE BURWAY FORMATION.

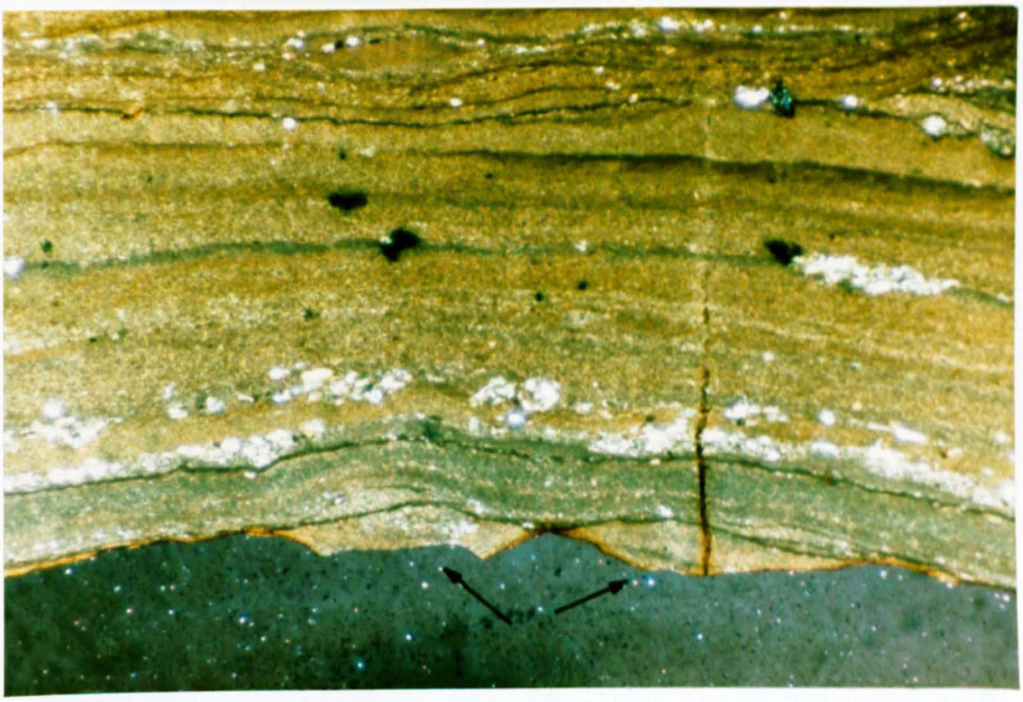


PHOTO A: Thin-section, normal to bedding, through a ring-like feature (bottom centre) projecting from the base of the bed (positive hyporelief). Note the similarity of the lithology within and outside the ring and the pinch and swell of the lamination in the upper part of the photo. Left side of the feature has been slighly eroded during grinding. Scale bar=1mm.

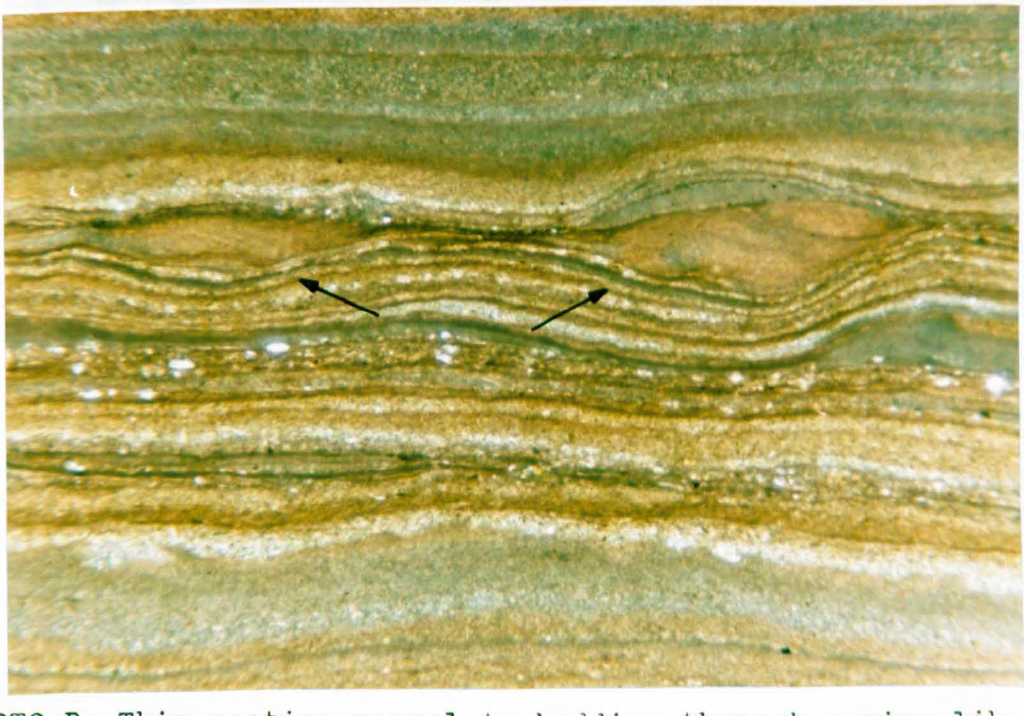


PHOTO B: Thin-section normal to bedding through a ring-like feature (top centre). Note the pinch and swell of the lamina in which the feature occurs and the obliquity and truncation of the lamination within the feature. Scale bar=1mm.

FIG:26 PHOTOGRAPHS OF SMALL RING-LIKE FEATURES FROM THE TOP OF THE BURWAY FORMATION.

PHOTO A: Raised rings on the top of a sandstone bed (positive epirelief). Scale bar graduated in cms and mms.

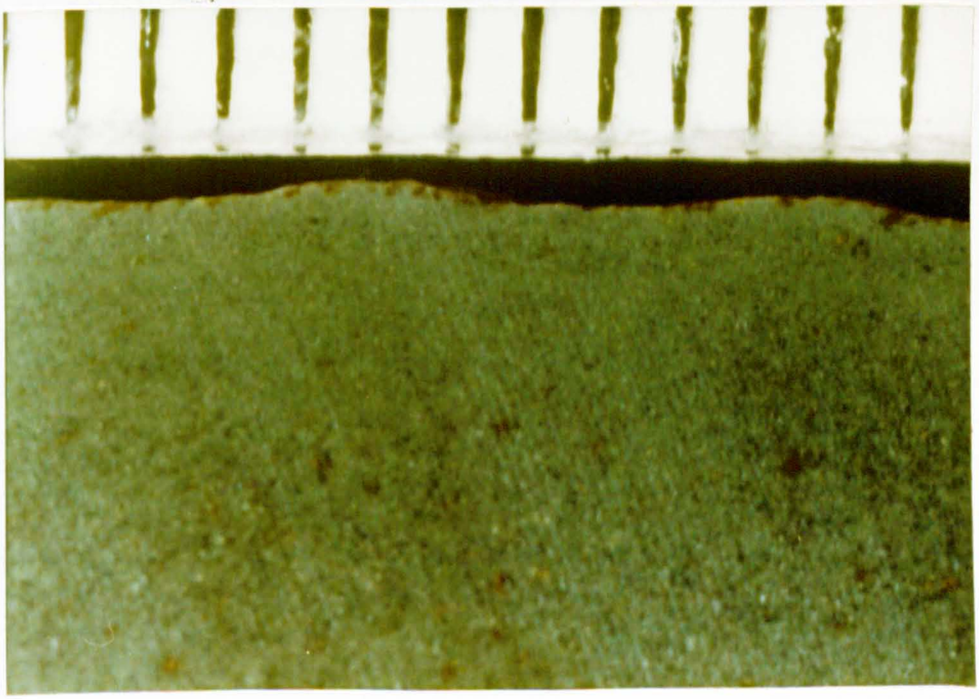


PHOTO B: Cross-section, normal to bedding, through the rings shown in photo A. Note that the rings are lithologically similar to the underlying sandstone and in continuity with it. Lines from top left to bottom right are saw marks. Scale bar graduated in mms.

FIG: 27 SMALL RING-LIKE FEATURES FROM THE PRECAMBRIAN INLIER OF CARMARTHEN.

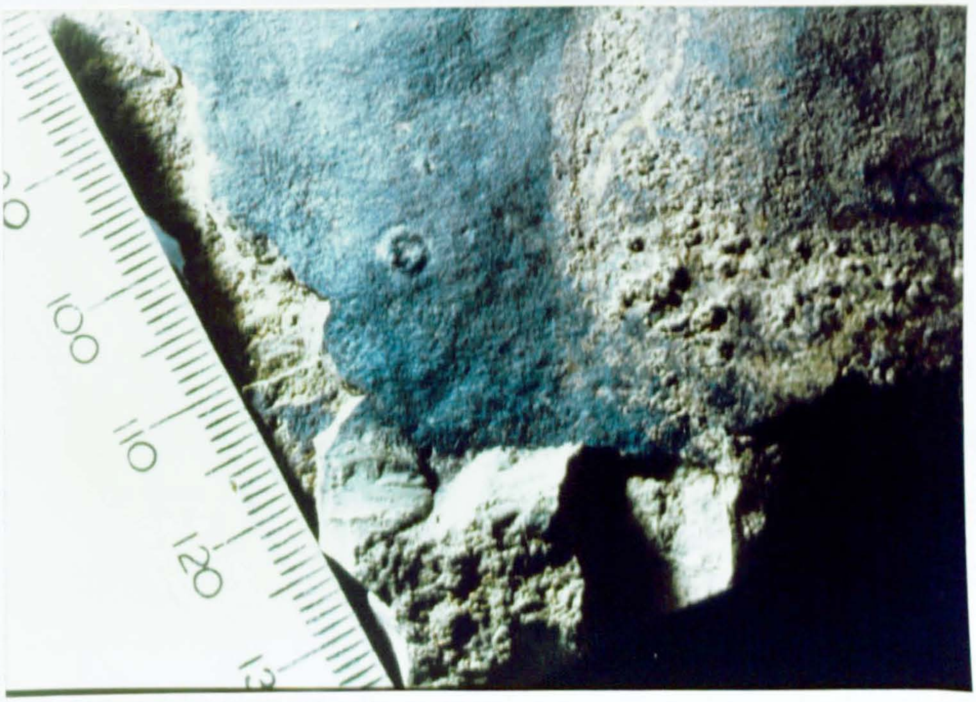


PHOTO A: Isolated ring. Scale bar in cms and mms.

PHOTO B: Small rings with possible trail (bottom left). Scale bar in mms.

the direction in which the ring is raised. Rarely, discontinuous linear features, 1mm in width, extend from the rim of the ring structures (fig. 24, photo B). These have been found in negative hyporelief together with ring features which are in negative hyporelief.

Similar structures are figured by Greig et al. (1968, plate 4, fig. A). Greig et al. (1968) refer to these as "pit-and-mound structure" and envisage a mechanism of formation by either loading or by the movement of entrapped air bubbles, followed by slumping in the wake of the rising bubbles. Similar, previously undescribed, small ring-like features have been found by the author in the Precambrian inlier of Carmarthen (fig. 27, photos A and B). These are associated with a well developed Ediacara-type fauna (Cope, 1977, 1979 and 1982). The lithology in which the Carmarthen examples are found consists of finely laminated siltstones which are volcaniclastic. This is very similar to the lithology in which the Longmyndian ring features are found. Similar, small ring-like features are also figured by Crimes and Germs (1982, plate 1, fig. 1). These were found in the late Precambrian, Kuibis Subgroup of the Nama Group of Namibia in south-west Africa. They were considered to be biogenic and were referred to the genus Bergaueria. However, these specimens differ slightly from the majority of the Longmyndian ring-like features in consisting of a mound with only a small central depression. They are therefore more similar to the positive epirelief rings shown in fig. 26, photo A than the positive hyporelief rings shown in fig. 24, photo A. In all of these occurrences, that is, in the Carmarthen inlier, in the Kuibis Subgroup and in the Longmyndian, the environment of deposition is considered to be shallow marine.

These ring-like features are not similar to raindrop impressions, since the latter generally consist of a pit with a raised rim on the top surface of the bed, whereas the ring-like features mostly consist of a moat-like depression in the top surface of the bed. The positive epirelief rings shown in fig. 26, photo A are also dissimilar from raindrop impressions in consisting essentially of a raised mound with a smaller central depression on the top surface of the bed. The environment of deposition of the strata in which these ring features are found is interpreted as a subaqueous delta. Consequently, raindrop impressions would not be expected.

The Longmyndian ring-like features are very similar to those figured by Cloud (1960, plate 1, figs. B to E). The latter are interpreted as gas blisters and were found in the Devonian of Pennsylvania. Cloud (1960, p.37) envisages a process of formation by development of a simple mound or blister followed by collapse, leaving a doughnut-shaped rimmed depression or dimpled blister. Such a mechanism of formation could explain the morphology of the ring-like structures in the Longmyndian, but would require the direction of gas blistering to be either predominantly downward, to produce the positive hyporelief rings (fig. 24, photo A) or predominantly upward, to produce the positive epirelief rings (fig. 26, photo A). It was previously argued (section 5.2), that features which were formed by either gas generation or by the trapping and subsequent escape of air are unlikely at this horizon. It is plausible that they might have been caused by dewatering. Such a dewatering mechanism was postulated for the production of the large circular features described in section 5.2, which are similar in form to the positive epirelief rings (fig. 26, photo A), but larger.

Twenhofel (1921, p.369) suggested that small mounds, 3 to 10mm in diameter, with a small crater-like depression on the summit, were formed by the rapid settling of clay, accompanied by small upward currents of water. This mechanism could explain the positive epirelief mounds with central depressions (fig. 26, photo A). However, these features are composed of sand and not mud. It is difficult to envisage that the positive hyporelief structures were produced by such a process, since they are in the reverse orientation to that expected.

In conclusion, these features might have been formed by inorganic phenomena such as gas, air or water escape. The pinch and swell of the laminae in which the ring features are found and the obliquity and truncation of the lamination within one ring feature (fig. 25, photo B), strongly suggests an origin which has involved soft-sediment deformation. Such deformation might have been caused by degassing or by dewatering. It is considered that dewatering is the most likely cause of these features. However, since late Precambrian medusoid impressions commonly consist of circular and ring-like features and since there are similarities between Bergaueria (Crimes and Germs, 1982) and the Longmyndian ring-like features, the possibilities that some of these features may have had a biogenic origin should not be dismissed.

5.4 Blister-like features from the Burway Formation

These occur in the thinly laminated siltstones at the top of the Burway Formation, directly beneath the Cardingmill Grit. These siltstones are interbedded with thick to very thick-bedded, fine grained sandstones. Numerous examples have been found by the author in Ashes Hollow at SO 43399302, where they can be found in abundance on certain bedding planes.

The features are subcircular, dome-shaped and project downwards from the base of the bed (positive hyporelief). Corresponding depressions are found in the top of the bed (negative epirelief). They are 2 to 3mm in diameter and commonly have a small, central, pimple-like projection (fig. 28, photo A). Occasionally, the margins of the feature are indented, resulting in a radial, lobate morphology (fig. 28, photo B). Cross-sections normal to bedding show that the blister is infilled with coarse silt and fine sand which is similar in composition to and in continuity with the adjacent laminae. A thin tube, less than one millimetre in width, of coarse silt and fine sand extends upwards from the blister and cuts through the overlying thinly laminated silt and clay (fig. 29, photo A). This tube may connect several blister-like features over several millimetres (fig. 29, photo B). It is the cause of the pimple-like projection seen in many of the features. The variability of the morphology in plan view displayed by these features (fig. 28, photo A) is the result of the variability in the level at which any one lamination will cut through the blisters and tubes. These blister-like features occur at the same horizon as the small ring-like features. However, the two structures have not been found together on the same sample. Although it is plausible that the blister-like features could be causally related to the small ring-like features, this relationship could not be demonstrated from the material collected.

These features were noted by Cobbold (1900), who attributed them either "... to the passage of some organism through the wet mud, or possibly, they might be due to the escape of bubbles of some gas."

FIG:28 <u>BLISTER-LIKE FEATURES FROM THE TOP OF THE BURWAY FORMATION.</u>

PHOTO A: Blister-like projections from the base of the bed (positive hyporelief).Note the small pimple-like projections which occur individually and which are also superimposed centrally on the blisters.Scale graduated in cms and mms.

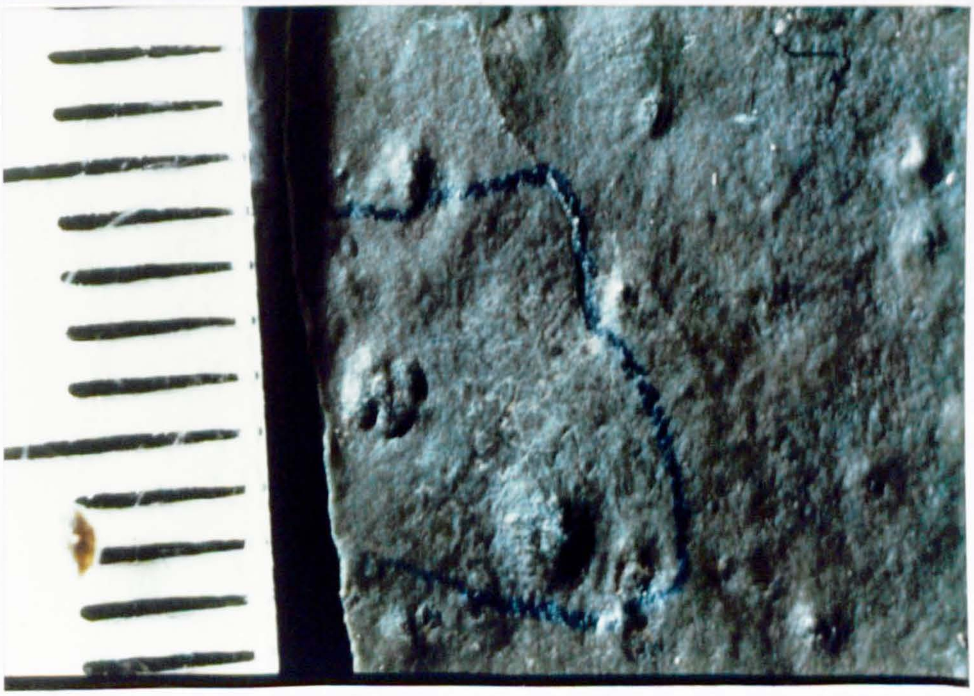


PHOTO B: Blister-like projection from the base of the bed (positive hyporelief).Note the radial,lobate form of the blister and the central,pimple-like projection.Scale graduated in mms.

FIG:29 CROSS-SECTIONS, NORMAL TO BEDDING, THROUGH BLISTER-LIKE FEATURES FROM THE TOP OF THE BURWAY FORMATION.

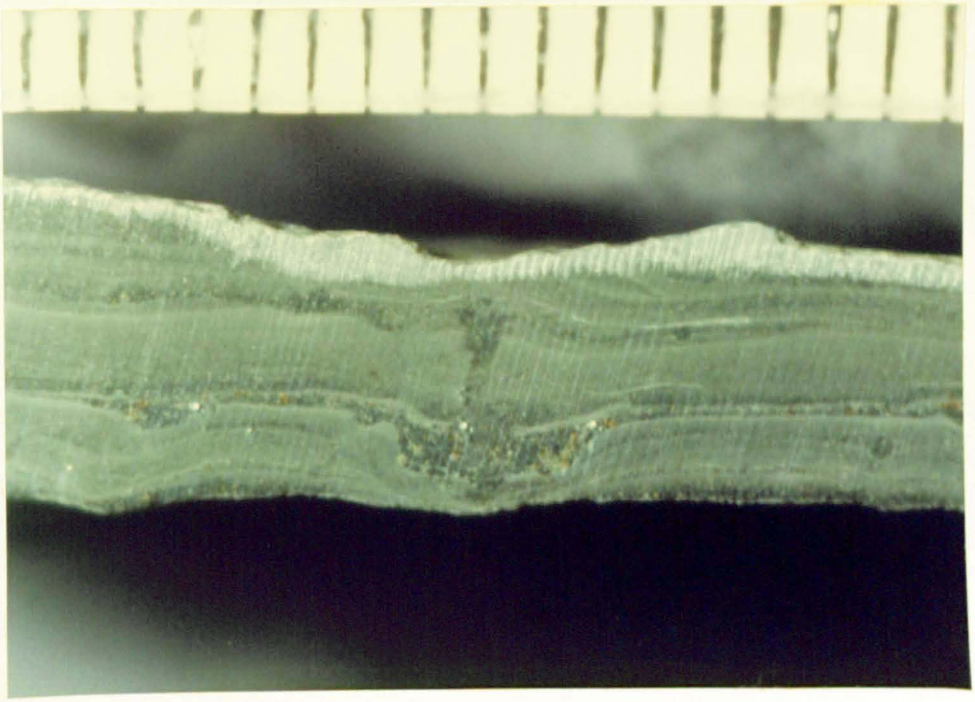


PHOTO A: Cross-section, normal to bedding, through a blister-like feature (slightly left of centre). Note the infilling of very fine sand, similar to and in continuity with the adjacent lamination, the sand filled tube connecting with the overlying sand lamina and the deformation of the underlying silt and clay laminae. Scale graduated in mms.

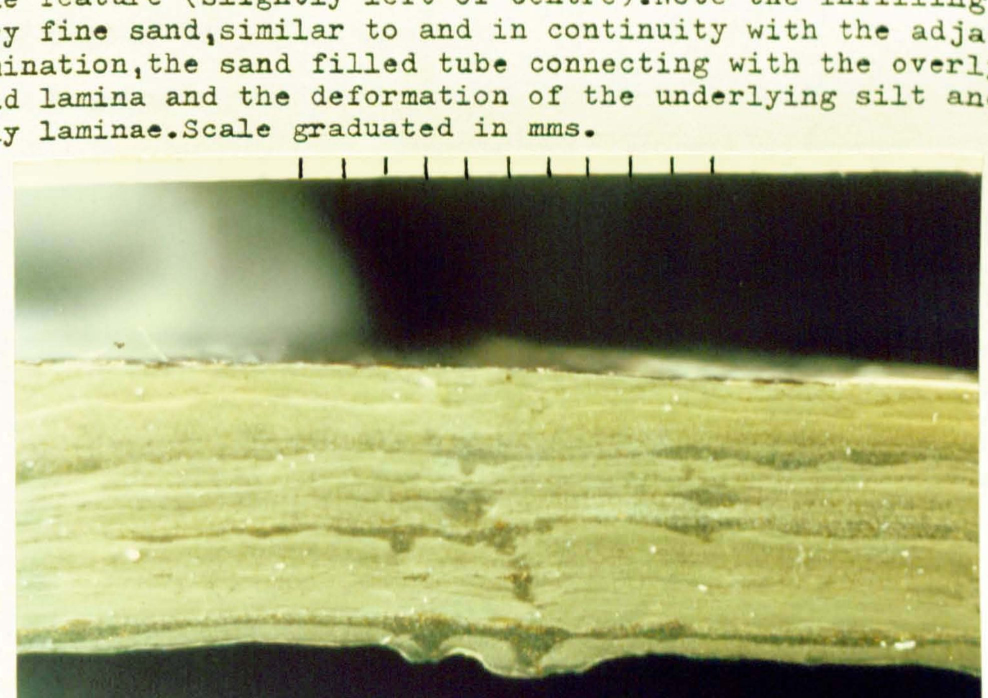


PHOTO B: Cross-section, normal to bedding, through several blister-like features. Note the pinch and swell of the silt and sand laminae, the sand-filled tube connecting several of the blisters and the sandy infillings, similar to and in continuity with the adjacent laminae. Scale graduated in mms.

(Cobbold, 1900, p.86). Cobbold compared them to Histioderma hibernicum from Bray Head. A similar structure is figured by Greig et al. (1968, plate 5, fig. D), which is referred to as "pit-and-mound structure". Their specimen is the syntype of Arenicolites sparsus, Salter, figured by Salter (1857, plate v, fig. 2) (Greig et al., 1968, p.69). Greig et al. (1968) also record that similar "... downward protruberances of coarser sand cores into finely banded mudstone occur in the shales of the Burway Group. The bedding surface is studded with 1 to 3mm diameter pits and mounds corresponding to the sandstone core infillings" (Greig et al., 1968, p.70). These features are referred to as "pit-and-mound structure" and are attributed either to loading or to the movement of entrapped air bubbles, followed by slumping in the wake of the rising bubbles.

Very similar structures, both in size and form, are present in the Vendian of the White Sea region of the USSR, where they are abundant and are associated with the moulds of large medusoids and trace fossils. These structures have not been previously described. The host rock is siltstone which accumulated in the shallow water environment (M.A. Fedonkin, pers. comm.). Fedonkin (pers. comm.) is uncertain whether these structures from the White Sea region of the USSR are biogenic or inorganic.

Crimes and Germs (1982) figure similar blister-like features from the Schwarzrand Subgroup of the Nama Group (Precambrian-Cambrian) of Namibia (south-west Africa), (plate 1, figs. 2 and 3). These are only known from a plaster cast of a lost original. Although they are 5-12mm in diameter and therefore larger than the Longmyndian examples they are similar in showing a radiating lobate structure and they commonly have a central, raised, pimple-like projection. In this respect, they are very similar to the

Longmyndian lobate forms shown in fig. 28, photo B. Crimes and Germs (1982) refer to the Nama Group features as Brooksella sp., Walcott 1896 and favour an origin by the activities of a small worm-like animal. Cloud (1973) interprets Brooksella canyonensis as a deformed sedimentary blister, whose petallate margin is the product of small-scale jointing. However, it is not inferred that all of the specimens of Brooksella sp. are inorganic.

If the blister-like feature were a resting trace of biogenic origin (as suggested by Crimes and Germs (1982) as one mechanism for the production of Brooksella sp.), then some mechanism is required to explain the coarse silt and sand filled tube. It is possible that such a tube is an escape burrow. This would require the ability of the organism to drastically change shape. The irregularity of the feature, however, does not suggest a resting trace. Alternatively, and plausibly, these features may have been produced by the burrowing and feeding mechanisms of a worm-like creature. These mechanisms could be compared with those of Gyrophyllites, which consists of a vertical dwelling shaft punctuated by horizontal rosettes of feeding lobes (Fürsich and Kennedy, 1975). However, these feeding lobes are usually found in the finer grained lithologies, whereas the blister-like features of the Longmyndian occur in the coarser lithologies.

Fenton and Fenton (1937) describe and figure similar structures, up to 3mm in diameter, consisting of pits with central pimples, from the Appekunny Formation of the Precambrian Belt Supergroup (plate 5, figs. 5 and 6). These occur in finely laminated, siliceous, green argillite, which, from their descriptions, appears to be similar to the lithology in which the Longmyndian structures occur. These are interpreted as sleet pits.

However, the environment of deposition in which the Longmyndian features are found is interpreted as a subaqueous delta. Therefore, sleet pits would be unlikely to occur. The Longmyndian structures are not similar to those produced by degassing since the latter usually consist of a raised blister with a central depression on the top of the bed (Cloud, 1960).

Fig. 29, photo B demonstrates that the coarser lithologies show considerable pinch and swell of the laminae. Beneath some of the blister-like projections, the lamination within the finer grained lithologies can be seen to be passively deformed. Both of these features show that soft-sediment deformation has occurred. Soft-sediment deformation was also shown to have occurred within the associated sediment which contains the small ring-like features (section 5.3). These features may be explained by a loading and also probably dewatering mechanism. Compaction may have resulted in unequal pore pressure between the coarser laminae which equilibrated by forcibly opening conduits through the intervening mudstone, resulting in the coarse silt and sand filled tube. This could have been accompanied by loading and spreading of the coarse silt and sand, resulting in the blister-like projection. The lobes of this projection may be explained by the rise of the underlying mud, as incipient flames, during the radial spreading of the coarse silt and sand.

It is concluded that these features might have been formed by inorganic phenomena, such as dewatering, accompanied by soft-sediment deformation and loading. Alternatively, they could be interpreted as biogenic structures resulting from the activities of a worm-like organism, possibly similar in habit to Gyrophyllites.

5.5 Small pit and blister-like features from the Synalds Formation

These are abundant in the laminated siltstones of the Synalds Formation and are occasionally present in similar lithologies in the Portway, Lightspout and Bridges Formations. They are most abundant in the Red Synalds Member of the Synalds Formation and can be found at this horizon throughout the outcropping area of the Longmyndian Supergroup.

They consist of small, circular to elliptical, shallow pits on the top surfaces of the beds (negative epirelief) and corresponding small projections on the bases of the beds (positive hyporelief). Their diameter ranges from less than 1mm (fig. 30, photo A), up to 4mm (fig. 30, photo B) and is most commonly 1 to 2mm. Their depth is usually less than 1mm. It was previously argued (section 3.9.2) that the commonly elliptical form of these features (fig. 15, photo B) is due to tectonic deformation of the circular forms (fig. 15, photo A), since the long axes of the pits are coaxial with the bedding/cleavage intersection lineations. The features are usually found crowded together on certain bedding planes. They are sometimes associated with narrow, subparallel, linear features (discussed in section 5.8), but this is due to random imposition and no causal relationship can be inferred (discussed in section 5.9). In some instances, small, linear and occasionally sinuous, discontinuous and irregularly orientated, narrow ridges and grooves are associated (fig. 31, photo A). These are discussed in section 5.6.

These features are figured by Greig et al. (1968, plate 4, fig. G and plate 5, figs. C, D, E, F and G), who describe them variously as bubble impressions, raised mounds and small blister-like mounds. The specimen shown in plate 5, fig. G is the counterpart of a syntype of Arenicolites sparsus, Salter. This specimen is very similar to

FIG:30 SMALL,BLISTER-LIKE FEATURES FROM THE SYNALDS FORMATION.

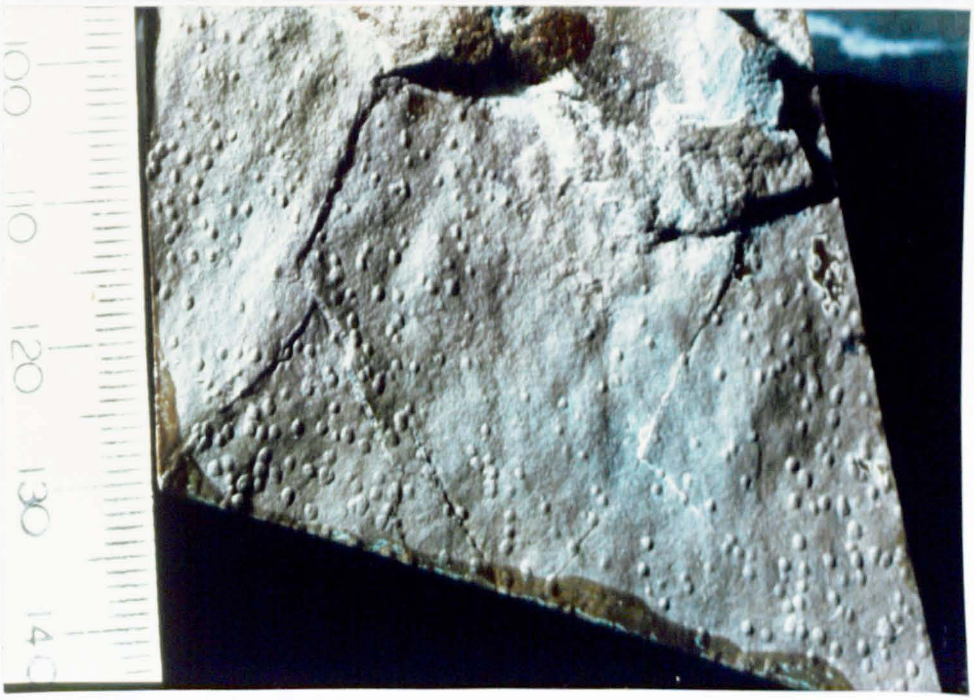


PHOTO A: Small,blister-like features projecting from the base of thinly laminated siltstone (positive hyporelief) Green Synalds Member.Scale graduated in cms and mms.

PHOTO B: Blister-like projections from the base of a sandstone bed (positive hyporelief).Red Synalds Member.Scale graduated in cms and mms.

FIG: 31 SMALL, BLISTER-LIKE AND IRREGULAR LINEAR FEATURES FROM THE SYNALDS FORMATION AND IRREGULAR LINEAR FEATURES FROM THE LIGHTSPOUT FORMATION.

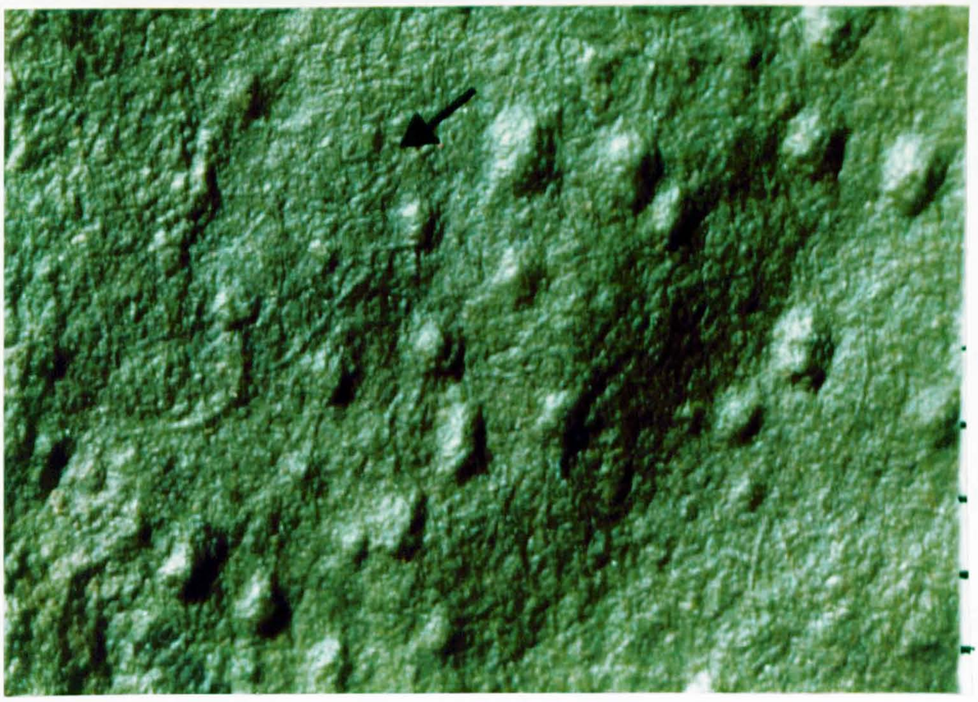


PHOTO A: Small, blister-like features projecting from the base of thinly laminated siltstone (positive hyporelief). Note the presence of very thin linear features with positive and negative relief. Note one sinuous and curved form (above left of centre). Scale graduated in mms.

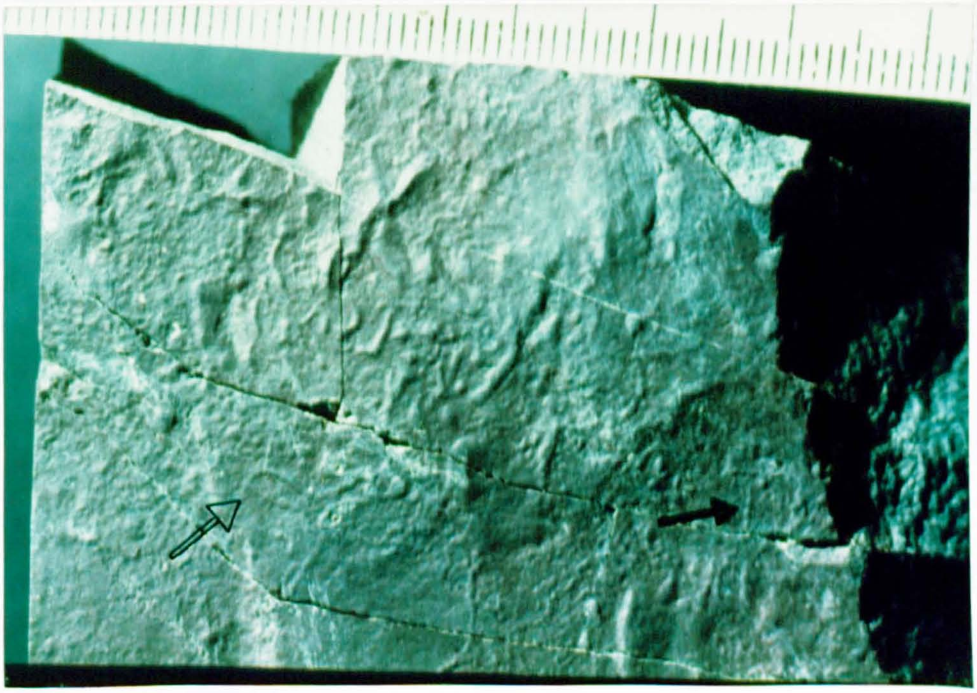


PHOTO B: Irregular, linear features from the Lightspout Formation. Note the presence of irregular nodes, some of which approach rows of small, blister-like protruberances e.g. lower right . Note also the recurved and sinuous form of some of the features e.g. lower left of centre. Scale graduated in mms.

that shown in fig. 31, photo A. The small depressions were interpreted as bubble impressions and the impressions of wave foam. An arrangement of the impressions in orientated lines or streams was interpreted as being due to current activity and was compared to the impressions of bubbles produced by wave activity in shallow water. It was also noted that some of the structures occurred as raised mounds or blisters on the tops of the beds (positive epirelief). However, such raised features have only been seen in very few samples by the author (fig. 32, photo A). Greig et al. (1968) interpret these as mud bubbles or blisters, probably similar in origin to pit and mound structures and argue that some of the specimens of Arenicolites sparsus, Salter in fact consist of these small raised blister-like features and their counterparts. However, the majority of the author's samples appear to be positive hyporelief structures and their counterparts.

Bland (1984) figures similar features (figs. 1a and b and figs. 2a, b and c). He interprets them as impressions, caused by the collapse of flexible-walled spheroidal bodies of biogenic origin. These spheroidal bodies are interpreted as a dispersable "resting stage" of "Arumberia", the latter consisting of fine ridges and grooves (discussed in sections 5.8 and 5.9) and Bland therefore favours a genetic link between the linear features and the small pit and blister-like features (discussed in section 5.9). Bland (1984) argues that spherical to elliptical bodies up to 6mm in diameter from the Lightspout Formation (his fig. 1c) are similar, chlorite filled spheroids. However, close examination of his figure does not suggest that these are related to the pit and blister-like forms from the Synalds Formation. Their interior appears to be composed of sediment, very similar to the surrounding sediment. If this

FIG: 32 SMALL BLISTER-LIKE FEATURES FROM THE SYNALDS FORMATION AND ACCRETIONARY LAPILLI FROM THE LIGHTSPOUT FORMATION.

PHOTO A: Small blister-like features from the Red Synalds Member, projecting from the base of the bed (positive hyporelief) Note the slightly larger depressions (negative hyporelief). Scale bar= 1cm.

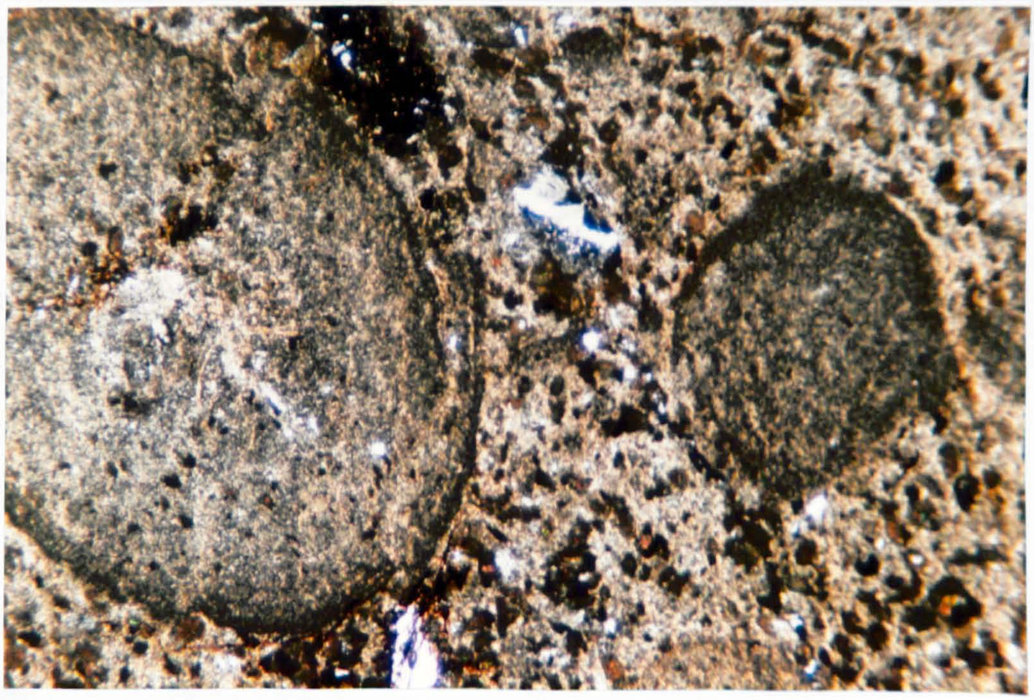


PHOTO B: Accretionary lapilli in highly altered dust-tuff from the Lightspout Formation. The groundmass is highly altered to fine mica and patchy epidote(brownish colours). Phenoclasts are occasionally present (top centre). Scale bar = 1mm.

structure were organic, then this would require removal of the interior (?by decay) of a solid object or puncturing of the exterior, if the object were hollow, and subsequent infilling by the surrounding sediment. Either mechanism would not favour the preservation of the subspherical exterior "wall". These structures have greatly different sizes, ranging from less than 1mm up to 6mm, whereas the pit and blister-like features from the Synalds Formation are commonly of a similar size in any one sample. A very similar specimen was collected from the Lightspout Formation in situ by the author. This sample is similar both in appearance and texture to the specimen figured by Bland. This sample is a coarse, grey coloured, poorly sorted and highly altered siltstone. The sample is unlaminated and is interpreted as a dust tuff with plagioclase phenoclasts and feldspar-phyric lithic phenoclasts. The subspherical objects are interpreted as accretionary lapilli (fig. 32, photo B). These occasionally possess cores of crystal or lithic fragments. These accretionary lapilli are similar in size to the objects figured by Bland (1984, fig. 1c). Bland (1984) also interprets a chlorite filled subspherical object, 2.4mm in diameter, from the Lightspout Formation (his fig. 1d) as a mineralised Arumberia spheroid. However, it was previously argued (section 4.7) that alteration products of the Longmyndian sediments commonly consist of subspherical bodies, composed of radiate chlorite and occasionally calcite, with thin rims of epidote, occasionally with hematite, and that these subspherical bodies have originated by growth within the matrix, rather than by the replacement of an originally spherical object. The object figured by Bland (1984, fig. 1d) appears very similar to these diagenetic features. Bland (1984, p.627) argues that an organic wall decayed leaving a "microporous region that scatters light". However, in thin section, similar features examined by the author show no trace of any organic material. The "opaque cream layer" noted by Bland (1984, p.627) is the common appearance of very finely granular epidote in reflected light and thin sections confirm that this is the composition of the "wall". Bland (1984, p.626) argues that the "position of the spheroid wall... had fallen away from one side of the cavity before mineralization". However, this feature could be easily explained by the removal by weathering or during cutting of part of an irregular margin. Bland (1984, p.627) also argues that chlorite-filled spheroids have been found together with "Arumberia" arrays in situ. The only sample of such a feature found by the author was not associated with the narrow linear features. In this specimen, it was clear that chlorite mineralisation had occurred along a narrow bedding plane fracture and this had resulted in the small pits as well as the surrounding sediment surface being coated in chlorite. During weathering and removal of the sample, it is plausible that the chlorite would be preferentially concentrated in the small pits giving the appearance of chlorite-filled spheroids. In conclusion, the hypothesis that these small pits and blister-like features are the result of mineralised, originally organic spherical objects is questioned.

Similar features are figured by Salter (1856 and 1857), who interpreted them as worm burrows and named varieties of the features as Arenicolites didymus and Arenicolites sparsus. Both of these were considered to occur distinctly in pairs. However, this does not appear to be true from Salter's figured specimens of Arenicolites sparsus (Salter, 1857, plate 5, figs. 1-4) and Greig et al. (1968) note that the type specimen of Arenicolites didymus shows a random arrangement of pits, except on portions of the specimen. The samples collected by the author do not show a distinct arrangement in pairs and therefore this feature is considered to consist of distinctly separate pits. Since Salter used the term "Arenicolites" for "all worm burrows with double openings" (1857, p.204), then the term "Arenicolites" should not be used for these features. Confusingly, Salter (1856, fig. 4) shows a very similar feature to "Arenicolites", which he interprets as being due to gaseous bubbles, or to the decomposition of small concretions and he also notes their resemblance to rain prints.

Banks (1970) figured very similar, blister-like projections from the Innerelv Member of the Stappogiedde Formation of the late Precambrian to Lower Cambrian Vestertana Group of Finnmark, Norway (plate 1a). These blister-like projections are circular protruberances, 1-2mm in diameter on the bases of siltstone laminae in interlaminated siltstone and claystone. Their distribution was noted to be mostly random and occasionally they occurred in pairs. These projections were interpreted as "undoubted trace fossils" and were considered to be the passive infillings of vertical burrows (Banks, 1970, p.26).

Similar, simple, subcircular, blister-like features are figured by Urbanek and Rozanov (1983, plates XLIX, 5; LII, 4; LIII, 3; LXVIII, 1 and LXVIII, 2). However, these features are generally larger than those in the Longmyndian, the former being mostly 2-6mm in diameter. They are interpreted as simple burrows (Urbanek and Rozanov, 1983). Associated with the small burrowing traces shown by Urbanek and Rozanov (1983, plate LIII, 3) is a sinuous crawling trail (Urbanek and Rozanov, 1983, p.88). This association may be similar to the association of the small blister-like features from the

Longmyndian with small, linear and occasionally sinuous, discontinuous, narrow ridges and grooves (fig. 31, photo A). These are discussed in section 5.6.

James (1952b) considered the paired nature of the holes of Arenicolites to be "purely fortuitous" and noted that individual depressions were more abundant. He states that, "the bulk of the depressions are oval in shape and could be caused by rain striking at an angle to the surface" (James, 1952b, p.26). However, he later notes that the oval shape could have been produced by tectonic deformation of a circular depression. He thought that the more circular depressions were certainly rain prints as they possessed a raised rim.

Discussion

The small, blister-like features generally do not resemble raindrop imprints. The latter often have slightly elevated rims and vary in size from about "... one-half an inch to that of a pin-head or even less" (Twenhofel, 1921), whereas the blister-like features in the Synalds Formation are mostly of a uniform and similar size throughout the formation. In addition, Twenhofel (1921) notes that after only a few minutes of rain, a mud surface becomes thoroughly sculptured through the presence of a multiplicity of coalescing pits, whereas these features often show similar densities of discrete pits throughout the Synalds Formation. This would have required consistent rain showers of similar duration and of similar strength.

Moussa (1974, p.1118) notes that "... most, if not all, so called rain-drop impressions are, as Desor correctly interpreted them in 1850, the result of air bubbles rising through sediments". Such air bubbles may originate from the entrapment of air during the flooding of a river floodplain (Twenhofel, 1921 and Moussa, 1974). Since the environment of deposition of the Synalds Formation is interpreted as an alluvial floodplain, then this mechanism of formation for the blister features in the Longmyndian is plausible. Greig et al. (1968) interpret some of these features as bubble impressions due to floating wave-foam bubbles. However, although foam impressions consist of hemispherical pits without raised rims and are therefore, in this manner, similar to the Longmyndian pits, foam impressions generally occur as clusters of adhering pits with a wide range of sizes on the same surface (Reineck and Singh, 1975, p.52), whereas the pits in the Longmyndian are discrete and of a similar size on one surface. Some of the impressions are similar to the surface markings produced by foam blown across a sediment surface (Reineck and Singh, 1975, fig. 71). However, the pit elongation and the linearity so produced in the pits of the Longmyndian is coaxial with the bedding/cleavage intersection lineation and is therefore a tectonic feature rather than a sedimentary one.

The evidence for the production of these pits and small blister-like features by mineralised organic spheroids of "Arumberia" (Bland, 1984) has been examined and previously discussed and is considered to be inadequate. Cross-sections through one sample of these small, blister-like features and their counterparts consisting of pits reveal a lack of mineralisation, a lack of significant grain size difference at the pit/blister boundary and a lack of any pierced lamination suggestive of deep burrowing. The lack of significant grain size difference at the pit/blister boundary suggests that these features are not the product of loading.

Conclusions

Two mechanisms of production appear to be the most likely from the previous discussion. The pits and blisters could have been produced by the escape of entrapped air during the flooding of an alluvial floodplain. Alternatively, these features might have been produced by biological activity (e.g. Banks, 1970, p.26). The association of pits and blisters with small, possible trails (section 5.6), the uniformity of size of the pits and, in some instances, the apparent arrangement of the features in irregular linear and curving rows, suggests that a biogenic origin is plausible. This apparent linear arrangement is noticeable on fig. 30, photo A and fig. 32, photo A and may be comparable with the "trails of progressive motion with periodic burrowing into the sediment" shown by Urbanek and Rozanov (1983, plate LII, 4 and p.151). This apparent linear arrangement would require the application of statistical tests to confirm its presence. These have not been carried out.

5.6 Small, irregular linear features from the Synalds Formation

These can be found occasionally in the laminated siltstones from both the Green Synalds Member and the Red Synalds Member of the Synalds Formation. They consist of very thin (fractions of a mm wide), discontinuous, irregularly orientated linear markings. They may be straight, curved or sinuous and they occur in positive or negative relief with very shallow depths. They are often associated with small (1mm or less in diameter), blister-like mounds (positive hyporelief) and corresponding pits (negative epirelief). These latter features are discussed in section 5.5. One example of these linear features is shown in fig. 31, photo A.

Similar structures are figured by Greig et al. (1968, plate 5G). This specimen is the counterpart of part of a syntype of Arenicolites sparsus, Salter. These were interpreted as grooves, which represented "... the impressions made by drifting objects, floating bubbles or seaweed-like material" (Greig et al., 1968, p.71). Blake (1890) noted that one of the slabs with Arenicolites sparsus (Salter, 1857, plate 5, fig. 1) was "... covered with very fine discontinuous curling tube-like bodies, which resemble exactly the castings of minute worms" (Blake, 1890, p.391).

Peat (1984a) also commented on the presence of "... numerous, unbranched, curved and sometimes recurved ridges" on some of Salter's original material (Peat,1984a, p.18). These ridges were thought to be biogenic and either the moulds of broad nematomorph cryptarchs or the casts of burrows. The latter explanation was considered to be more probable.

These linear features are very similar to the simple crawling traces figured by Urbanek and Rozanov (1983, plates LIII, figs. 2 and 3 and LIV, fig. 6). Associated with one sinuous crawling trace figured by these authors are burrowing traces (Urbanek and Rozanov, 1983, plate LIII, fig. 3). These burrowing traces are very similar to the small, blister-like features from the Synalds Formation (section 5.5). Occasionally, a sinuosity is developed in the Longmyndian linear markings which is similar to Cochlichnus sp. figured by Urbanek and Rozanov (1983, plate LIV, fig. 6).

The simple crawling traces figured by Urbanek and Rozanov (1983, plates LIII, figs. 2 and 3 and LIV, fig. 6) are generally larger than the linear features from the Longmyndian. Similarly, the small, blister-like features from the Longmyndian are smaller than the similar burrowing traces shown by Urbanek and Rozanov (1983) (discussed in section 5.5). Consequently, the proportions between the linear features and the blister-like projections from the Longmyndian are comparable with the similar crawling traces and burrowing traces figured by Urbanek and Rozanov (1983). Possibly related features are "discoidal moulds with an extending worm-like body." (Urbanek and Rozanov, 1983, p.61 and plate LI, fig. 1). Occasionally, the thin linear features from the Longmyndian can be seen to extend from and terminate at the blister-like projections (e.g. fig. 31, photo A). Again, the features from the Longmyndian are smaller, but in similar proportion.

Conclusions

It is considered unlikely that these small, linear structures are impressions made by bubbles, as suggested by Greig *et al*. (1968). The linear features are all similar in size. This would require the unlikely condition that all the generating bubbles were of similar size. The variability in direction of the linear features would require that the generating currents were very variable in direction. The very small size would require unusually small bubbles. The linear features occur in negative and positive relief and are consequently unlike groove marks.

Peat (1984a) suggested that these structures are the moulds of tubes, similar to a broad nematomorph cryptarch from the Stretton Shale Formation. This he tentatively interpreted as the organic lining of a metazoan dwelling tube or burrow.

It appears plausible that these small, linear features are biogenic. There is insufficient evidence to suggest that they are genetically related to the small, blister-like projections. However, their close association, the consistent size differential between possible examples of comparable features from Russia and the Longmyndian and the interpretation of "discoidal moulds with an extending worm-like body" (Urbanek and Rozanov, 1983, p. 61 and plate LI, fig. 1) merit further investigation.

5.7 Irregular, linear features from the Lightspout Formation

These are rare and only one example has been found by the author at SO 3882 8806. The lithology comprises thinly laminated, reddish purple siltstones, with some thick to very thick-bedded, very fine grained sandstone beds. They occur on a thin lamina of siltstone. The outcrop area is highly faulted and these beds are tentatively correlated with the Red Lightspout Member of the Lightspout Formation. The environment of deposition is interpreted as an alluvial floodplain. The lithology and facies are similar to those of the Red Synalds Member of the Synalds Formation.

The sample on which these features were found was not *in situ*. However, the sample came from a newly opened excavation, implying derivation from an immediately adjacent exposure. The features occur in positive relief on a thin lamina of very fine grained siltstone and consist of very irregular, discontinuous, curved and occasionally sinuous, linear ridges up to 1mm wide. Occasionally, the linear form is composed of irregular nodes and swellings, some of which approach the form of rows of small blister-like protruberances (fig. 31, photo B).

These features are similar to the small, irregular linear features from the Synalds Formation. However, the latter are commonly smaller (fig. 31, photo A). A similar, though slightly larger structure is figured by Crimes and Germs (1982, plate 1, fig. 4) and is interpreted as ?*Chondrites* sp. These traces are irregular, discontinuous and occasionally slightly sinuous. However, these traces differ from the Longmyndian features in possessing both ridges and furrows and in showing a slightly dendritic, diverging pattern (Crimes and Germs, 1982, p.895).

Banks (1970) figured very similar structures, both in size and form, from the Precambrian/Cambrian, Manndraperelv Member of the Stappogiedde Formation of Finnmark, Norway (his plate 1, fig. b). These are interpreted as hypichnial and exichnial casts. They show a variety of burrowing patterns which are remarkably similar to those of the Longmyndian and include discontinuous, nodal linear traces and sinuous and curved traces. In contrast to the Longmyndian, the traces occur in turbidite facies (Banks, 1970).

Some of the biogenic traces from the Late Precambrian to Cambrian of Russia are similar to the Longmyndian examples (Urbanek and Rozanov, 1983, plates LIII, 2 and LIV, 6) and the latter occasionally show a sinuosity similar in form to Cochlichnus sp. (Urbanek and Rozanov, 1983, plate LIV, 6).

Discussion

These features are unlikely to be groove casts because of their irregular, curved and occasionally sinuous shapes. They do appear similar to some wrinkle marks (Reineck and Singh, 1975, fig. 80, p.56), which are also 0.5 to 1mm thick and a few millimetres long and although generally parallel, they can be curved. Wrinkle marks can be produced by the action of wind on wet sediment (Reineck and Singh, 1975, p.56). Since the environment of deposition of the Red Lightspout Member is considered to be an alluvial floodplain, then these conditions may have been met. Alternatively, wrinkle marks may be produced by aseismic soft-sediment deformation (Allen, 1985).

However, the wrinkle marks illustrated by Allen (1985) are mostly larger than the Longmyndian features and consist of both depressions and ridges.

Conclusions

These structures may be interpreted as wrinkle marks. However, because of their similarities to some biogenic traces (e.g. Banks, 1970 and Urbanek and Rozanov, 1983), both in size and in the variety of forms, and because of their greater degree of organisation, compared to wrinkle marks, then a biogenic origin is favoured.

5.8 Subparallel, linear features from the Synalds Formation

These are most abundant in the Red Synalds Member of the Synalds Formation. They also occur in similar lithologies in the Lightspout Formation and in the Portway Formation. Similar features occur in some parts of the Burway Formation.

These features consist of subparallel, raised linear ridges on the top surfaces of thinly laminated siltstone (positive epirelief) (fig. 33, photo A). The ridges are narrow, often cuspate and are separated by shallow furrows which have a width of 0.5 to 3mm. The furrows are flat to concave in cross-section. The relief (from ridge top to furrow bottom) is usually less than 0.5mm. As the ridge spacing increases, their relief tends to increase (fig. 33, photo B). The ridges are usually parallel, but they can diverge or converge (e.g. fig. 33, photo B). This characteristic is manifest where the surface on which the feature occurs is curviplanar. Rarely, the ridges can bifurcate during divergence. The features are

FIG:33 SUBPARALLEL,LINEAR FEATURES FROM THE SYNALDS FORMATION.

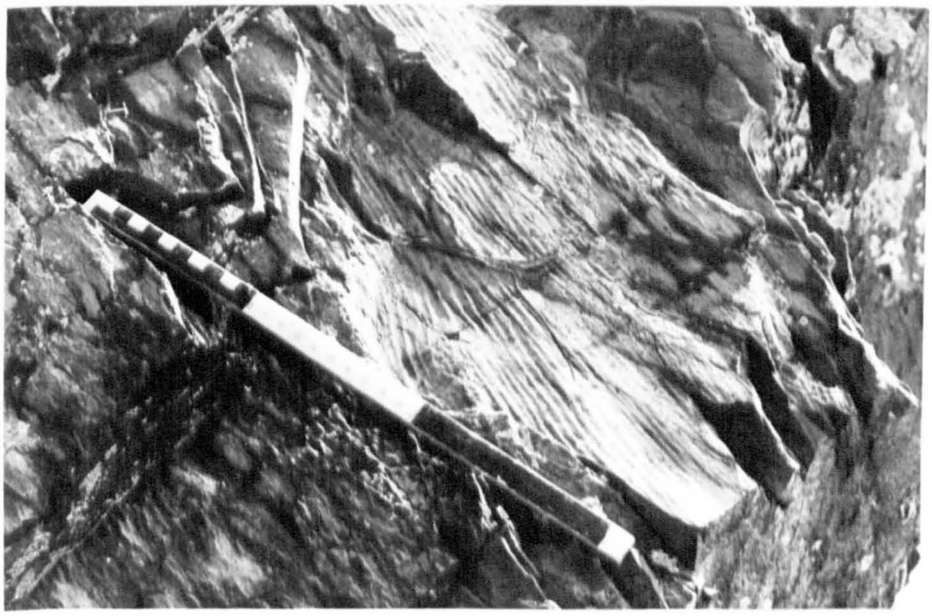


PHOTO A: Subparallel,linear features from the Synalds Formation.Top surface of bed.Note the presence of elongated small pits on the same surface (right of centre) and the penetrative cleavage (right corner).Scale is marked in decimetres and centimetres.

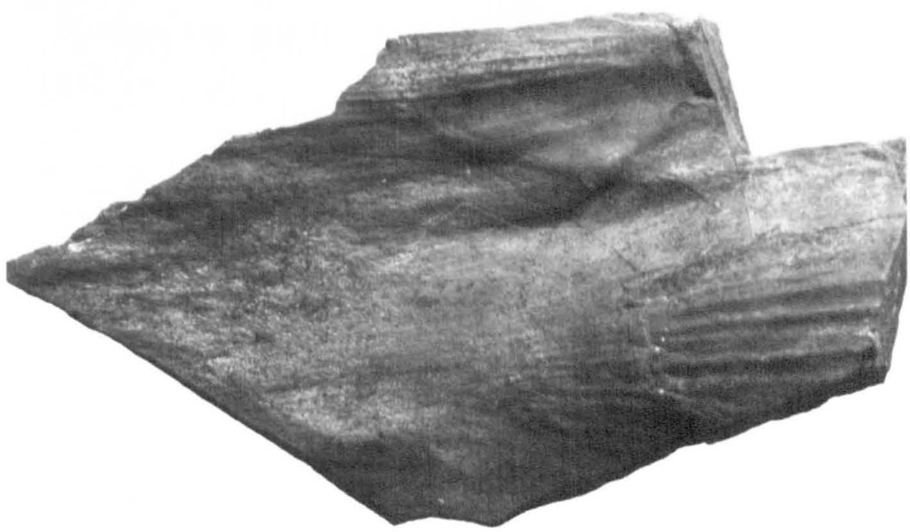


PHOTO B: Linear features from the Synalds Formation.Top surface of bed.Note the non parallel,divergent nature of the lines over the raised bedding surface undulation (pseudoripple).Scale bar = 2cm.

occasionally associated with small blister-like and pit features (discussed in section 5.5), this is due to random imposition and no causal relationship can be inferred (discussed in section 5.9).

Cross-sections through the ridges and grooves occasionally show an asymmetry. In these cases, the inclination of the ridges is subparallel to small fracture planes in the underlying rock. In some cases, these fracture planes can be identified as cleavage planes.

Salter (1856) figured similar markings from the Longmyndian (plate IV, fig. 4) and concluded that they were inorganic and were lines of "... mineral structure". However, Greig et al. (1968, p.71) interpreted the raised ridges on the top of Salter's specimen as groove casts. However, their figured specimen (Greig et al., 1968, plate 5F) shows pits which normally occur on the top surfaces of the bed. Consequently, these linear features are better interpreted as raised ridges on the top surface of the bed (positive epirelief) rather than as groove casts (positive hyporelief). This specimen would then agree with the characteristics normally observed by the author, that is, of pits associated with raised ridges. This was Salter's original interpretation (Salter, 1856, p.250).

Bland (1984) figures very similar linear structures from the Longmyndian (figs. 1a and 1b) and from the Late Precambrian of Newfoundland and France (figs. 2a and 2b). He interprets these as Arumberia, Glaessner and Walter and favours a mechanism of formation by "... the impression of a colony of flexible, thin-walled tubular elements... firmly adpressed or possibly fused into a dense sheaf, bundle or solid block, where they are in contact with the sediment surface" (Bland, 1984, p.630-631). However, the diagnosis of Arumberia as upright cups by Glaessner and Walter (1975) is difficult to reconcile with the absence in the vast majority of cases of any such structure in the Longmyndian. The Longmyndian linear features, in contrast, are invariably subparallel and straight over wide areas of the sediment surface. Thus, even if these linear features are organic, they should not be referred to the genus Arumberia, Glaessner and Walter.

The linear features in the Longmyndian are similar to longitudinal ridges and furrows produced by horizontal current vortices (Dżułyński and Walton, 1965). However, Dżułyński and Walton (1965) state that the modal width of ridge separation is 5-10mm with a range of 3mm to 5cm, whereas, the range of separation of the Longmyndian linear features is 0.5 to 3mm. Additionally, the direction of these lineations, which is fairly consistent throughout the Synalds Formation, is not consistent with the palaeocurrent orientations. Whereas these lineations, if current produced, would correspond to a NNE-SSW trending current direction, the vector mean palaeocurrent direction for the Synalds Formation is WNW. Therefore, these lineations could not have been produced by the observed palaeocurrents. Bland (1984) gives the impression that Arumberia may be current orientated and states: "Ridges may diverge from a sunken apical area, or stream from an oblique linear edge, or parabolic crescentic edge, bordering the upstream side of a shallow depression in the sediment" (p.625). However, it is apparent in the Longmyndian that the linear features are orientated almost at right angles to the palaeocurrent.

Discussion

These linear features are very similar to lineations in the Burway Formation. These latter were interpreted as tectonic, being coincidental with the average bedding/cleavage intersection lineation of the Stretton Group (section 3.11.2). Observations throughout the Synalds Formation confirm that these linear features are parallel to a tectonic lineation defined by the elongation of small pits, the bedding/cleavage intersection lineation and pseudo-ripple long axis orientations. A stereographic plot from one area shows that there is a close correspondence between the pit elongation direction, the pseudo-ripple long axis orientation and the lineation direction (fig. 14, stereographic plot 9). General observations show that many of these linear features are demonstrably the product of the intersection of bedding and cleavage. Cross-sections through some samples show an asymmetry. In these cases, the inclination of the ridges is subparallel to the small fracture planes in the underlying rock and the surface linear features are a product of the intersection of these planes with the bedding. It would thus appear that these linear features are a tectonic phenomenon. However, in some cases these linear features are non-parallel and may converge at high angles (e.g. Bland, 1984, fig. 1a; Greig et al., 1968, plate 5E; fig. 33, photo B this thesis; and Salter, 1857, pl. V, fig. 8). Salter (1857) interpreted these as wave-ripples or "surf-lines". Greig et al. (1968) interpreted them as grooves, similar to rills, noted their occurrence with ridges, irregular mounds and depressions and suggested that they were all the result of intertidal current action. However, these linear features are associated with small blisters. These latter project in positive relief. Consequently, this specimen may be better interpreted as showing grooves in negative hyporelief along with blister-like projections in positive hyporelief. This specimen would then agree with the characteristics normally observed by the author, that is, of pits associated with raised ridges on the tops of the beds. This orientation does not allow for these linear features to be interpreted as rill-marks since they are in negative hyporelief rather than in negative epirelief. Bland (1984) interprets the same specimen as Arumberia, showing casts of ridges and spheroid impressions.

Observations by the author demonstrate that non-parallel linear features are developed only where the surface is curviplanar. This suggests that there is a causal relationship between the bedding surface morphology and the non-parallel nature of the linear features. Close examination of Bland's figure 1a (1984, p.626) reveals that despite the non-parallelism of the lines in this instance, the blister elongation is always subparallel to the linear features. This subparallel relationship is always maintained throughout the Synalds Formation. It was previously concluded that blister and pit elongation is due to strain of an originally circular feature. Consequently, the variability of the direction of the linear features appears to be related to the variability of the strain in the rock and is related to the curviplanar bedding surface. Cross-sections through samples showing curviplanar bedding surfaces reveal an internal soft-sediment deformation with swellings of the coarser lithologies. The surface morphology is therefore a reflection of the irregular swellings in the underlying lamination and is not a current formed feature (as suggested by Greig et al., 1968). The variability of strain is therefore a reflection of the variability in density along the bedding plane. Consequently, the non-parallelism of these linear features in these cases may be explained by the variability in the strain orientation and a tectonic origin, as previously concluded, cannot be dismissed on the evidence of non-parallelism.

Conclusions

There is no evidence to suggest that these linear features were produced by a biogenic mechanism. The absence of any organic material associated with these features, despite the finding of organic material in cryptarchs in unrelated samples (Peat,1984a), seriously questions the interpretations of Bland (1984, p.630) that a "... dense sheaf, bundle or solid block" of organic material was the causative agent. If the structure is biogenic, then the term Arumberia should not be applied, since the form in the Longmyndian does not agree with the diagnosis of Arumberia, Glaessner and Walter as upright cups.

Sedimentary causes of these features, such as rill-marks and longitudinal ridge and groove marks of current origin, have been discussed and have not been found to be adequate.

It is concluded that these linear features are tectonic in origin, for which evidence has been presented.

5.9 Discussion of Arumberia Bland (1984)

Bland (1984, p.625) assigned "... arrays of straight to gently curving, parallel to subparallel fine ridges and broad shallow grooves.." from the Longmyndian to the genus Arumberia, Glaessner and Walter (1975). The nature of these linear features has been discussed (section 5.8) and it was concluded that they are tectonic in origin. The interpretation of these linear features as an organic "... carpet like growth with no distinct apices" (Bland, 1984, p.631) does not compare with the diagnosis of Arumberia in terms of upright cups (Glaessner and Walter, 1975). Therefore, even if these linear features were biogenic, then the name Arumberia, Glaessner and Walter (1975) should not be applied.

Bland (1984) maintained that small hollows and mounds were often associated with Arumberia and that these were formed ".. by the collapse of flexible-walled spheroidal bodies in the sediment" (p.627). However, these features have been discussed (section 5.5) and it was concluded that the evidence for the presence of mineralised, organic spheroids in the Longmyndian was insufficient and that alternatively, these features may be interpreted as air escape features or possibly as small burrows.

The observations by the author that there is a random superposition of lines on pits in the Longmyndian is not agreed with by Bland (1984, p.627), who states that "Arrays of ridges are superimposed on patches of spheroid impressions... far more often than would be expected by chance". This is not the opinion of the present author. It is maintained that statistical tests would be required before any conclusion that the pits are genetically associated with the linear features and also not solely the result of favourable lithological conditions can be asserted. Bland (1984) notes that Arumberia, Glaessner and Walter (1975) does not have pits or blisters associated with it as part of the diagnosis. Therefore, if the linear features were assignable to Arumberia, Glaessner and Walter (1975) then the association of pits and blisters is atypical.

Bland (1984, p.625) maintains that the pits and blisters are "... mostly clustered in patches 10-50cm across,the same shape as Arumberia arrays (Fig. 2c)", however this has not been observed by the author and Bland's illustrative fig. 2c does not show the pits associated with arrays of ridges of Arumberia as he suggests. Bland (1984) also implies a genetic association between the pits and blisters and the linear features of "Arumberia" by stating that "... in some cases it can be seen that they are arranged in straight to gently curving rows parallel to the Arumberia ridges". It has been observed by the author that pit elongation and alignment is subparallel to the linear features. This phenomenon can be explained by the strain of originally spherical impressions into ellipsoids whose long axes are coincident with the bedding/cleavage intersection lineation, which is represented by the linear features. The observation by Bland (1984, p.627) that "within an array, elliptical spheroid impressions are orientated parallel to the ridges and grooves that pass closest to them, even when the ridges in different parts of the array point in quite different directions", has been discussed in section 5.8 and it was concluded that the non-parallelism of the linear features, together with the coincident pit elongation, could be explained by a variability in strain orientation caused by lithological inhomogeneity. The author has observed that generally, the linear features can be traced through the pit and blister features without any change in character. This is observable on Bland's (1984) figs. 1b and 2b. It appears that the lines have been superimposed on the pit and blister features. This relationship agrees with the interpretation of the blister-like and pit features as early sedimentary or biogenic features, which have later undergone strain into ellipsoids with the superimposition of a tectonic lineation.

In conclusion, Arumberia should not be applied as a term to the linear features of the Longmyndian. The linear features and the blister-like and pit features in the Longmyndian are genetically unrelated phenomena. There is insufficient evidence to suggest that the pit and blister-like features in the Longmyndian are due to the collapse of flexible walled, spheroidal bodies of organic origin. In the opinion of the author, Arumberia from the Longmyndian, as described by Bland (1984), is a pseudofossil composed of a tectonic lineation which is randomly superimposed on genetically unrelated pit and blister-like features. However, these latter may have a biogenic origin. This possibility has been discussed in section 5.5.

5.10 Microfossils

Peat (1984a) described cryptarchs from the Longmyndian. Four forms were described: a broad, transversely striated nematomorph from the Stretton Shale Formation, 155μm wide; broad nematomorph cryptarchs, 25 μm to 155 μm wide, from the Lightspout Formation; narrow nematomorph cryptarchs, 5 μm to 17 μm wide, from the Lightspout Formation and sphaeromorph cryptarchs, 3 μm to 70 μm wide, also from the Lightspout Formation.

The facies in which the specimens from the Lightspout Formation were found consisted of "... finely laminated dark shales and siltstones.." (Peat, 1984a, p.17). Similar facies were found by the author at four different localities: SO 43429508 (Green Lightspout Member), SO 42869514 (Green Lightspout Member), SO 41939367 (Portway Formation) and SO 43189311 (Green Synalds Member). The facies is composed of planar, parallel and very thinly laminated mudstone. The grain size is predominantly very fine siltstone and the colour is medium grey overall, the individual laminae being grey where organic rich and greenish grey where poor in organic material. The laminae are commonly normally graded. This facies forms beds 10cm to 1m thick and commonly 50cm thick and is always associated with greenish grey, laminated siltstones and very thick beds of very fine grained sandstones with fining upward profiles. The environment of deposition is interpreted as an alluvial floodplain and the very thinly laminated mudstone was probably deposited in

FIG:34 PHOTOMICROGRAPHS ILLUSTRATING SOME MICROFOSSILS FROM THE LONGMYNDIAN SUPERGROUP.

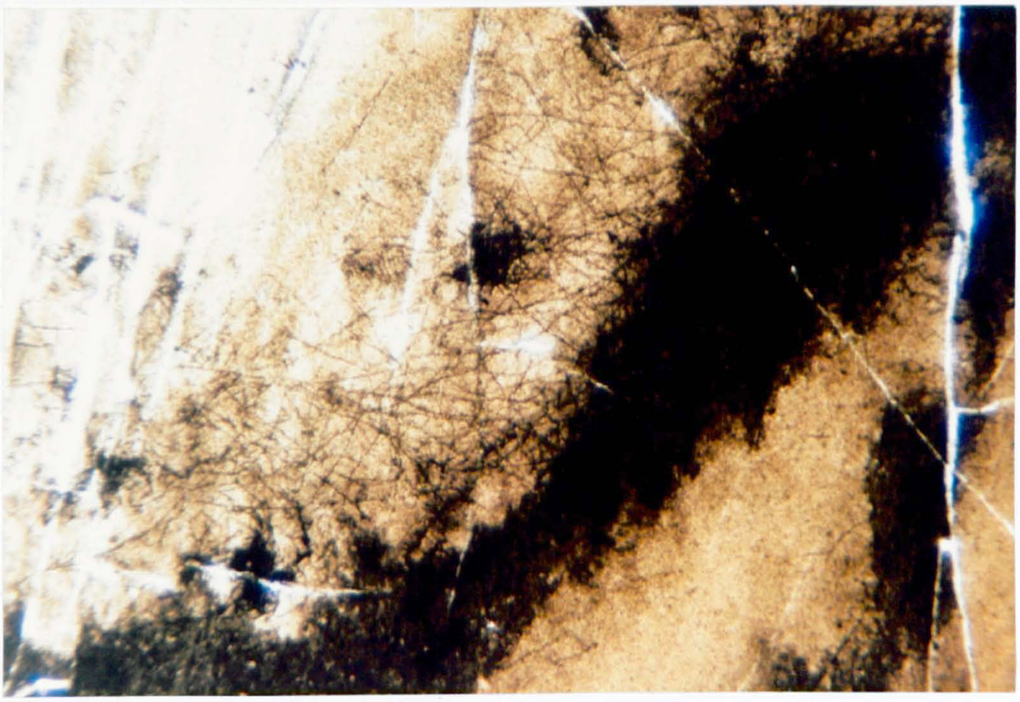


PHOTO A: Network of thin filaments overlying a dark lamina rich in disseminated organic material. Stratigraphic top to upper left corner, Thin section cut at low angle to lamination. Diagonal white stripes are veins. Plane polarised light. Green Lightspout Member. Scale bar = 1mm.

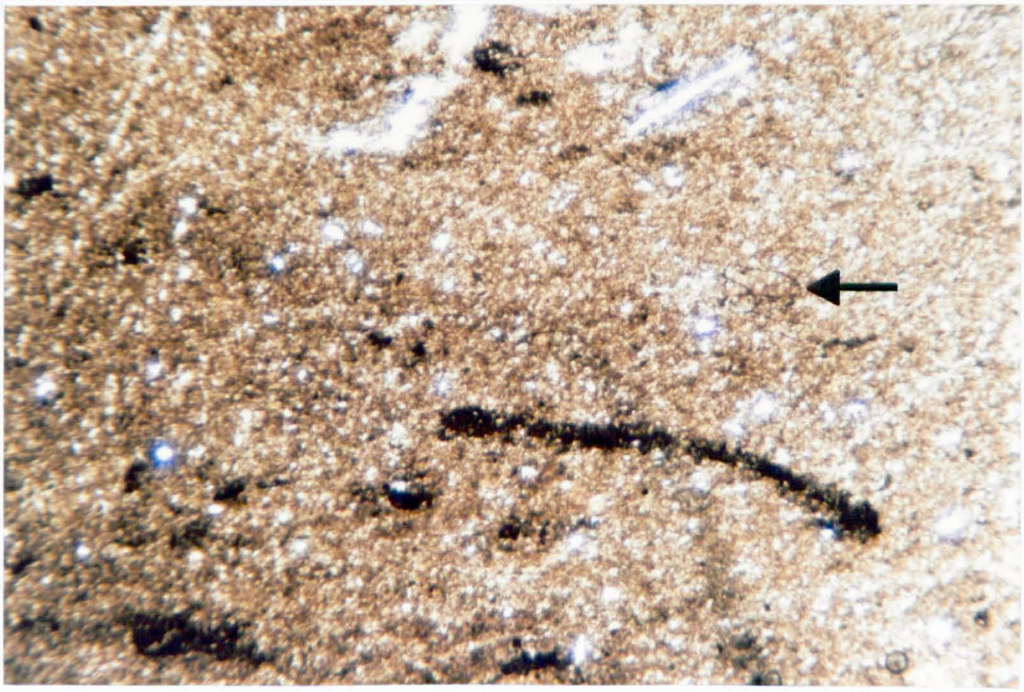


PHOTO B: Disseminated, dark brown, organic material in siltstone from the Portway Formation. Note contorted filament, right of centre, scattered clusters of organic material and broad ribbon (from centre to right of photo). Thin section cut at low angle to lamination. Plane polarised light. Scale bar = 1mm.

shallow,standing bodies of water, such as in abandoned channels, on the floodplain. This facies is symbolised by the letter G on logs 26 and 37 (in appendix).

Following the recommendations of Peat (1984a), thick, thin-sections were cut from the specimens at a low angle to the lamination. All of the above localities were sampled and all yielded narrow filaments of the type described by Peat (1984a). All of the samples contain disseminated, dark brown to black, very fine grained material, of inferred organic origin, which is occasionally concentrated in patches and laminae (fig. 34, photo B). Narrow filaments of the type described by Peat (1984a) were commonly found scattered throughout the specimens. These filaments are non-orientated and are commonly twisted and are thought to have suffered redeposition from mats. In many instances, the filaments were found organised in dense networks, associated with disseminated, fine organic material, commonly at the top of thin, normally graded laminae (fig. 34, photo A). These networks are comparable with the mats described by Peat (1984a, fig. 3a).

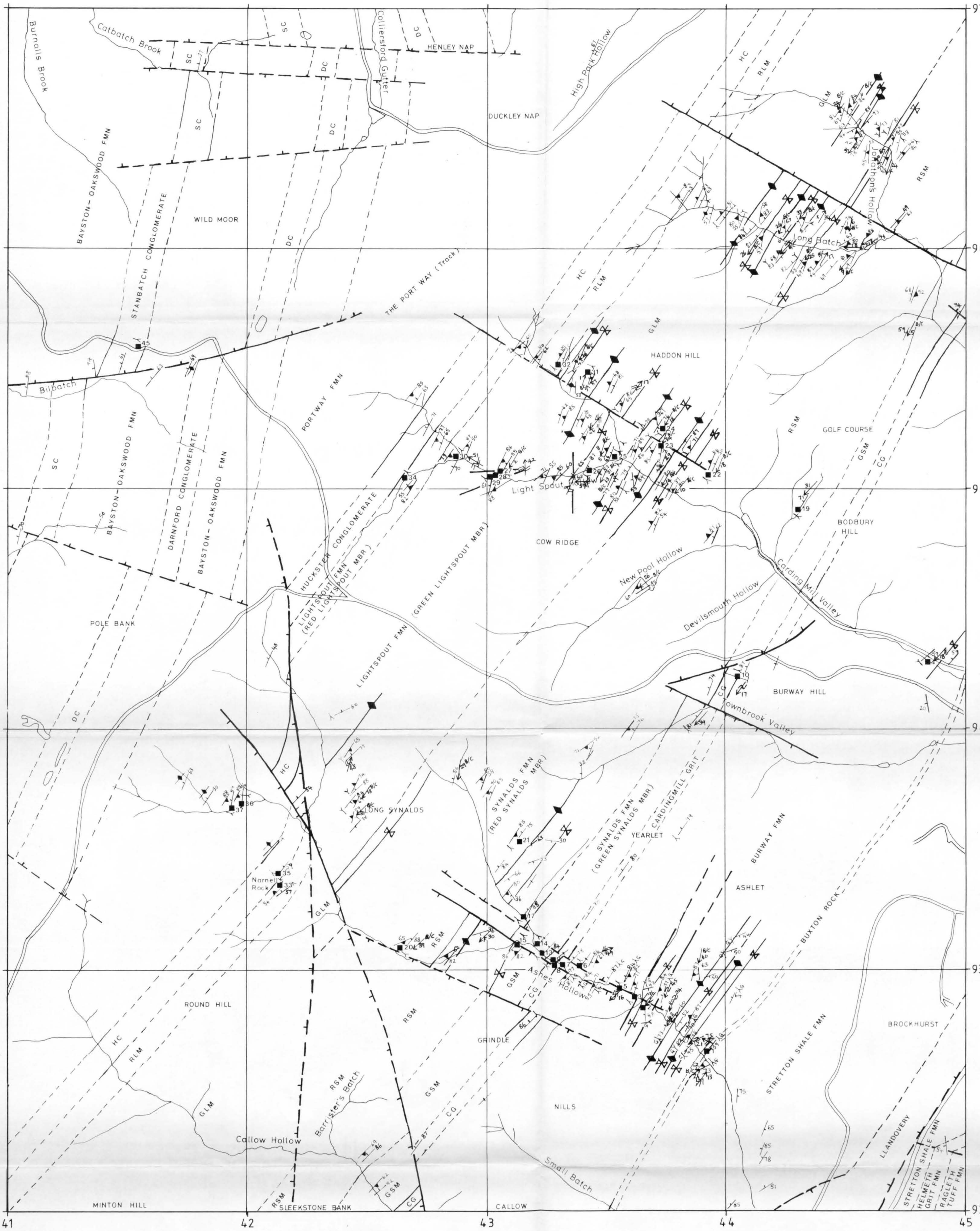
These laminated mudstones cannot be described as stromatolites. The latter are "an organosedimentary structure produced by the sediment trapping, binding and/or precipitation activity of microorganisms, primarily by cyanobacteria" (Awramik, 1984, p.2). The Longmyndian mats, however, do not appear to have trapped, bound or precipitated sediment, rather, they appear to have grown passively on the fine grained upper surfaces of graded laminae which were deposited by normal sedimentary processes.

No structures comparable to the broad, transversely striated, nematomorph cryptarch of Peat (1984a) were found. Small, circular structures, probably comparable to the sphaeromorph cryptarchs of

Peat (1984a), were noted within the disseminated, fine grained material and one structure, possibly comparable with the broad (25 μm to 155 μm wide) nematomorph cryptarchs of Peat (1984a, fig. 2b), was found in the Portway Formation (fig. 34, photo B).

Horodyski (1981) describes similar filaments from the Middle Proterozoic, Belt Supergroup of Montana. These range from 3-15 μm wide and are 20-200 μm long. They are thus comparable in size to the narrow filaments of the Longmyndian, which are 5 μm to 17 μm wide. These filaments occur in a similar facies, which is composed of finely-laminated, dark grey and black mudstones. Disseminated organic matter occurs in thin films parallel to the lamination. Spheroidal forms also occur, which are 1-20 μm in diameter. However, Horodyski (1981) interprets his structures as pseudomicrofossils consisting of 0.1 to 0.3 μm thick organic envelopes that encapsulate authigenic crystals of apatite, which occur isolated, in clusters and in filament-like aggregates. The organic envelope originated by the accretion around these crystals of organic material which was finely disseminated throughout the sediment (Horodyski, 1981).

The similarities between the Belt Supergroup pseudomicrofossils and the Longmyndian microfossils may merit further investigation. The Longmyndian microfossils do not appear to have apatite in their structure. However, this could be investigated by microprobe analysis.



KEY;

- ■10 LOGGED LOCALITY WITH LOG NUMBER
- FAULT (tick on downthrow side) — broken lines denote uncertainty
- GEOLOGICAL BOUNDARY (solid) — broken lines denote uncertainty
- BEDDING INCLINED (amount in degrees)
- BEDDING VERTICAL
- SLATY CLEAVAGE INCLINED (amount in degrees)
- SLATY CLEAVAGE VERTICAL
- YOUNGING DIRECTION
- PLUNGE - BEDDING/CLEAVAGE INTERSECTION
- PLUNGE - MINOR FOLD (sense of overturning indicated)
- PLUNGE - MINOR FOLD (undifferentiated)
- PLUNGE - KINK BAND AXIS (sense of overturning indicated)
- PLUNGE - KINK BAND AXIS (undifferentiated)
- AXIAL TRACE OF ANTICLINE
- AXIAL TRACE OF SYNCLINE
- PLUNGE - LINEATION (undifferentiated)
- PLUNGE - MULLIONS

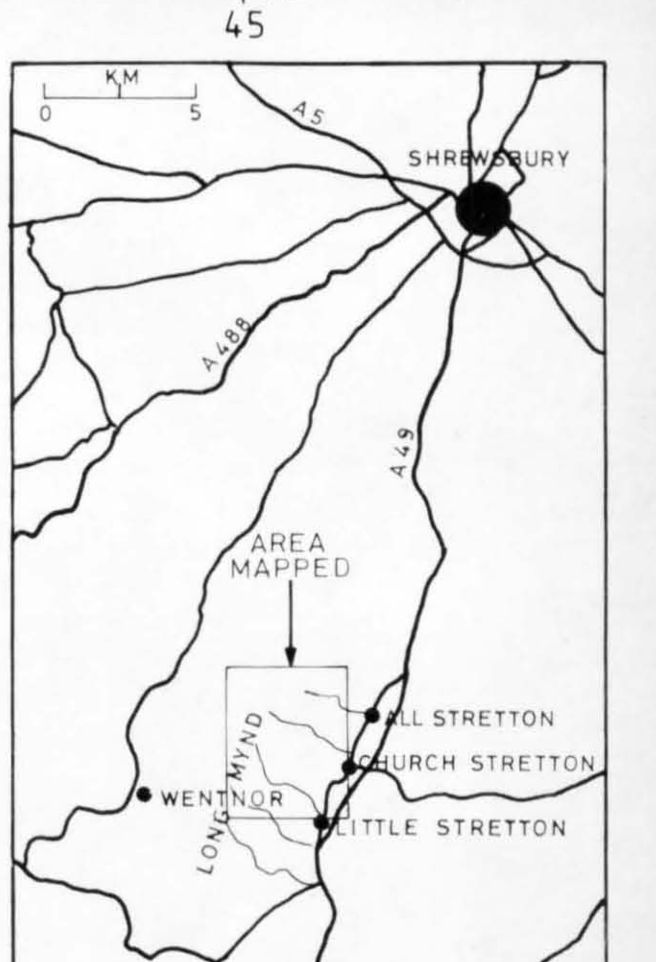
GEOLOGICAL MAP I

LONGMYND AREA

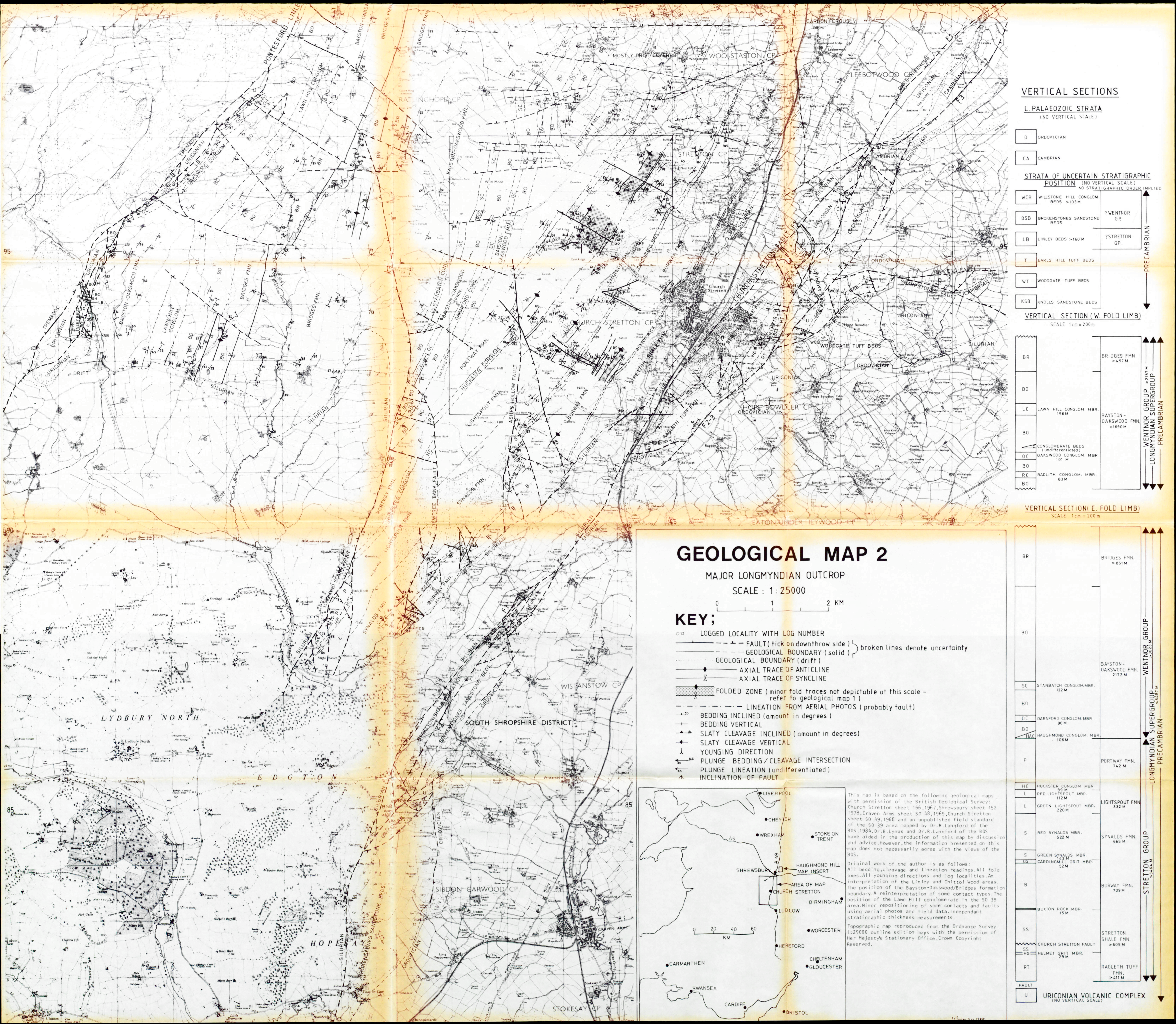
SCALE 1:10000

0 500 1000 m

J C Pauley Sept 85



INSERT - HAUGHMOND HILL AREA
UFFINGTON CP
UPTON MAGNA CP
PERMIAN
CARBONIFEROUS
Haughmond Hill
WESTBURY CP
SHREWSBURY AND ATCHAM DISTRICT
BAYSTON HILL CP
PONTESBURY CP
CONDOVER CP
CHURCH PULVERBATCH CP
SMETHCOTT CP
LONGNOR CP
WOOLSTASTON CP
LEEBOTWOOD CP
RATLINGHOPE CP
STRETTON CP
MOSTLY DRIFT COVERED
PONTESFORD - LINLEY FAULT
BAYSTON-OAKSWOOD FMN.
BRIDGES FMN.
PORTWAY FMN.
STANBATCH CONG.
HUCKSTER CONGL.
LAWN HILL CONGLM.
URICONIAN
ORDOVICIAN
CAMBRIAN
Shrewsbury
Pontesbury
Condover
Dorrington
Longnor
Longden
Hanwood
Westbury
Minsterley
Snailbeach
Stiperstones
SJ
SO
VERTICAL SECTIONS
L. PALAEOZOIC STRATA
(NO VERTICAL SCALE)
O ORDOVICIAN
CA CAMBRIAN
STRATA OF UNCERTAIN STRATIGRAPHIC POSITION (NO VERTICAL SCALE)
NO STRATIGRAPHIC ORDER IMPLIED
WCB WILLSTONE HILL CONGLOM. BEDS >103M
BSB BROKENSTONES SANDSTONE BEDS
?WENTNOR GP.
LB LINLEY BEDS >160 M
?STRETTON GP.
T EARLS HILL TUFF BEDS
CAMBRIAN


GEOLOGICAL MAP 2
MAJOR LONGMYNDIAN OUTCROP
SCALE : 1:25000
0 1 2 KM
KEY;
LOGGED LOCALITY WITH LOG NUMBER
FAULT (tick on downthrow side)
GEOLOGICAL BOUNDARY (solid)
broken lines denote uncertainty
GEOLOGICAL BOUNDARY (drift)
AXIAL TRACE OF ANTICLINE
AXIAL TRACE OF SYNCLINE
FOLDED ZONE (minor fold traces not depictable at this scale - refer to geological map 1)
LINEATION FROM AERIAL PHOTOS (probably fault)
BEDDING INCLINED (amount in degrees)
BEDDING VERTICAL
SLATY CLEAVAGE INCLINED (amount in degrees)
SLATY CLEAVAGE VERTICAL
YOUNGING DIRECTION
PLUNGE BEDDING / CLEAVAGE INTERSECTION
PLUNGE LINEATION (undifferentiated)
INCLINATION OF FAULT
This map is based on the following geological maps with permission of the British Geological Survey: Church Stretton sheet 166,1967,Shrewsbury sheet 152 1978,Craven Arms sheet SO 48,1969,Church Stretton sheet SO 49,1968 and an unpublished field standard of the SO 39 area mapped by Dr.R.Langford of the BGS,1984.Dr.B.Lynas and Dr.R.Langford of the BGS have aided in the production of this map by discussion and advice.However,the information presented on this map does not necessarily agree with the views of the BGS.
Original work of the author is as follows:
All bedding,cleavage and lineation readings.All fold axes.All younging directions and log localities.An interpretation of the Linley and Chittol Wood areas. The position of the Bayston-Oakswood/Bridges formation boundary.A reinterpretation of some contact types.The position of the Lawn Hill conglomerate in the SO 39 area.Minor repositioning of some contacts and faults using aerial photos and field data.Independant stratigraphic thickness measurements.
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VERTICAL SECTIONS
L. PALAEOZOIC STRATA (NO VERTICAL SCALE)
STRATA OF UNCERTAIN STRATIGRAPHIC POSITION (NO VERTICAL SCALE)
VERTICAL SECTION (W. FOLD LIMB)
VERTICAL SECTION (E. FOLD LIMB)
URICONIAN VOLCANIC COMPLEX
LIVERPOOL
CHESTER
WREXHAM
STOKE ON TRENT
SHREWSBURY
HAUGHMOND HILL MAP INSERT
AREA OF MAP
CHURCH STRETTON
BIRMINGHAM
LUDLOW
WORCESTER
HEREFORD
CHELTENHAM
GLOUCESTER
CARMARTHEN
SWANSEA
CARDIFF
BRISTOL
LYDBURY NORTH
EDGTON
SOUTH SHROPSHIRE DISTRICT
CHURCH STRETTON FAULT