PART 2

CHAPTER 6

FACIES

6.1 Introduction

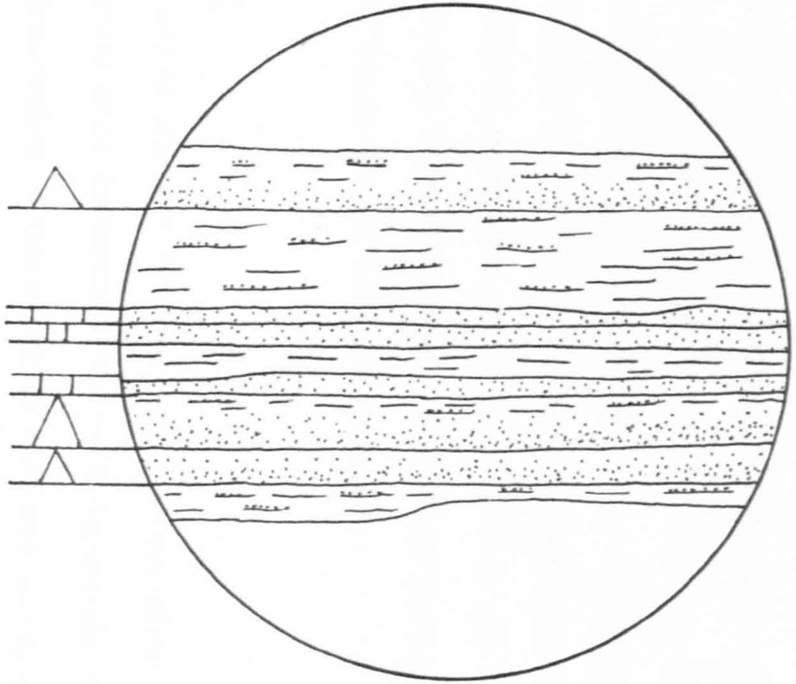
The facies are described in the order in which they occur stratigraphically, from the lowest to the highest. The main Longmyndian Supergroup has been divided into fifteen facies and four facies associations. The facies associations are (from the stratigraphically lowest to the stratigraphically highest): the turbidite facies association, the subaqueous delta facies association, the alluvial floodplain facies association and the braided alluvial facies association. Some of the facies in the alluvial floodplain facies association are similar. However, they can be separated as distinct stratigraphic units and it has been found that it is useful to differentiate them. Each facies is first described and then the interpretations are discussed. Where the facies are similar, then the description concentrates on the differences between them. The interpretation of the facies initially concentrates on the processes of deposition. The environments of deposition are considered in the sections which discuss the facies associations.

A key to the symbols used on the logs in this chapter is provided in the appendix (fig. 109). Note that the log numbers used on the figures are the same as those on geological maps 1 and 2 (in cover pocket).

6.2 THE THINLY LAMINATED MUDSTONE FACIES

This facies is comprised of thinly laminated, pale, light greenish grey siltstone and claystone. The laminae are on average 1mm thick and are between 0.5mm and 2.5mm thick. They are

FIG.35 SKETCH AND PHOTO ILLUSTRATING THE CHARACTERISTICS OF THE THINLY LAMINATED MUDSTONE FACIES



Sketch of a cross-section,normal to lamination,as seen under a binocular microscope.Laminae consist of very fine and fine grained siltstone (dotted),which may be normally graded or non graded.The relative grain size changes are shown on the left of the sketch.Claystone laminae contain very thin,discontinuous layers with silt grains.Thickness of specimen is 0.95cm.

Photo of the thinly laminated mudstone facies from the Stretton Shale Formation exposed in a trench at SO 4584 9327. Scale bar is graduated in decimetres and centimetres.

continuous, planar and parallel and they can be normally graded from siltstone to claystone or are non-graded. The bases of the siltstone laminae are usually non-erosional and abrupt and the upper surfaces may be abrupt or gradational with the overlying claystone. The claystone laminae usually contain very thin layers of very fine siltstone which are discontinuous, planar and parallel. This facies is illustrated by fig. 35. Some authors have noted that there are calcareous nodules in the Stretton Shale Formation, which is composed of this facies (Cobbold, 1900; James, 1952b and Whittard et al., 1953). These are discussed in section 4.4 and it is concluded that these are diagenetic and not of primary, sedimentary origin.

The Stretton Shale Formation is composed of this facies. There is no lateral or vertical change in this throughout its entire thickness of 605m. Some early authors divided the Stretton Shale Formation into the "Watling Shales", east of the Church Stretton Fault (e.g. Lapworth and Watts, 1910), and the "Brockhurst Shales", west of the Church Stretton Fault. However, no distinction was recognised by the author or by Greig et al. (1968, p.37) and it is concluded that there is no justification for separation. Similar mudstones occur in thin beds and laminae in the overlying Burway Formation, where they comprise part of the thin-bedded and thick-bedded turbidite facies.

Interpretation

The absence of current or wave-ripple cross-stratification and the fine grain size indicates that this facies was deposited in a low-energy environment. There are no features which are indicative of subaerial exposure. The lack of lateral and vertical facies changes indicates that these quiet water conditions were laterally extensive and persistent in time. Laminated mudstones may be deposited in a variety of subaqueous environments. However, laminated mudstones deposited on a shelf or slope are usually interbedded with, or pass laterally into, other facies. The characters of this facies could have been produced in a basin-plain environment. Since the overlying Burway Formation is interpreted as a progradational turbidite and deltaic sequence, which ends in the fluviatile Cardingmill Grit, then this interpretation is probable. If this progradational sequence is interpreted as a single event, then the minimum water depth, in which this facies was deposited, can be estimated from the thickness of the Burway Formation, which is 709m. This water depth is compatible with a basin-plain environment.

The normally graded laminae of silt might have been deposited by a current of waning velocity, such as a dilute, low-density turbidity current. Since the overlying Burway Formation is composed of turbidites, then this is possible. It is unlikely that this facies was deposited by a fluctuating, bottom traction current, since repeated, regular velocity/time relationships would have been required to produce the repeated, graded units of similar grain size. Stow and Bowen (1980) proposed that some laminated mudstones might have been deposited from low-concentration, slow moving, thick "turbidity flows", by a process of depositional sorting in the boundary layer at the base of a turbidity current. However, they recognised distinct, graded, laminated units (Piper, 1972), between 1cm and 10cm thick, in addition to continuous, laminated siltstone sequences. Graded, laminated units are not thought to occur in this facies.

The laminae of clay might have been deposited by the settling of clay from a turbid, near-bottom layer which was supplied with clay by dilute, low-density turbidity currents. Alternatively, clay could have been deposited from mid-water or near-surface, turbid layers supplied by rivers or storm processes. Some silt might have been deposited by settling from such layers.

6.3 THE RHYOLITIC FINE ASH TUFF FACIES

This facies is comprised of altered, pale grey, fine ash. In thin-section (fig. 37, photo B) the lithology consists of a very fine grained, cryptocrystalline and felsic groundmass with very fine mica and chlorite. Within the groundmass, occasional fragments can be recognised, including shards with straight, curved and "Y"- and "T"-shaped forms,and sand-sized, pale brown, blocky fragments. These fragments are interpreted as relicts of glass and it is proposed that the very fine grained groundmass could have been formed by the devitrification of these glassy fragments. Phenoclasts of subhedral plagioclase and anhedral, perthitic feldspar and feldspar-phyric volcanic clasts are occasionally present. A chemical analysis detailed by Greig et al. (1968, table 1, analysis 7, p.29) shows that the ash is of rhyolitic composition.

There are common, sand-sized, rounded patches, which are composed of radiate chlorite and/or polygonal, finely crystalline quartz and plagioclase feldspar. These are often zoned, with chlorite occupying the margins and quartz and feldspar occupying the cores. These rounded, subspherical bodies are similar to those discussed in section 4.7 and it was concluded that they are of diagenetic origin. An authigenic origin was also proposed by Greig

FIG. 36 LOG OF THE ENTIRE BURWAY FORMATION, ASHES HOLLOW

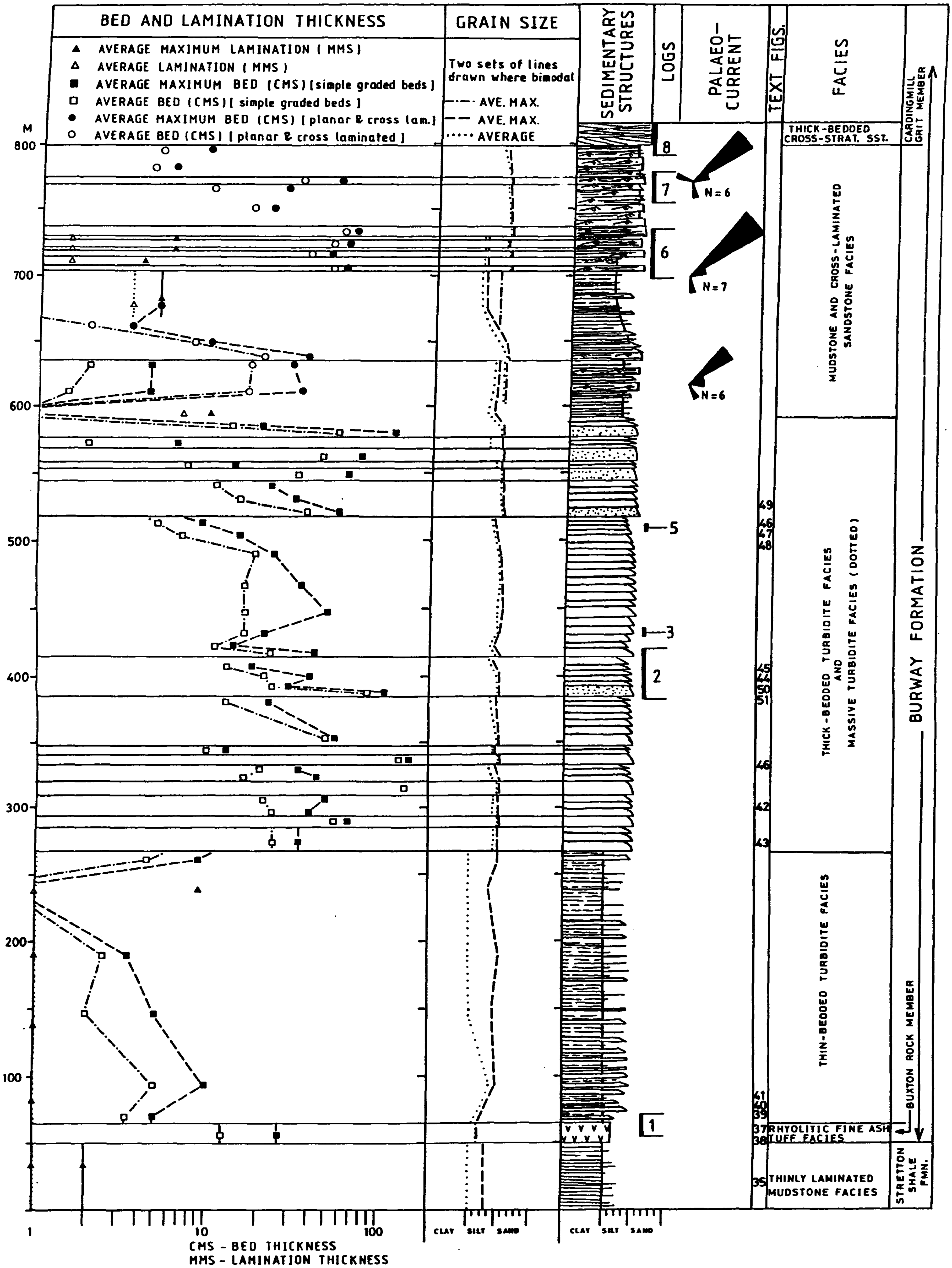


FIG:37 PHOTOGRAPHS ILLUSTRATING THE CHARACTERISTICS OF THE RHYOLITIC,FINE ASH TUFF FACIES.

PHOTO A: The Buxton Rock Member of the Burway Formation,exposed in Ashes Hollow at SO 4392 9266.Note the characteristic, lensoid bedding due to tectonic deformation.Breccia to left of scale is interpreted as a fault breccia.Scale bar (30cm long) is graduated in decimetres and centimetres.

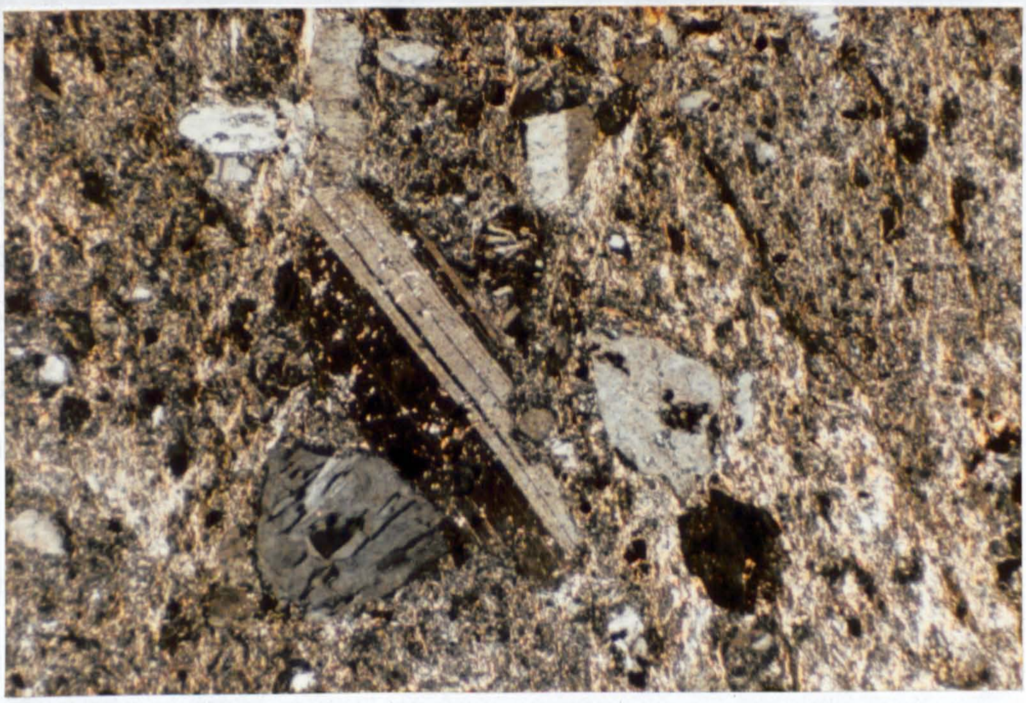


PHOTO B: Photomicrograph of a specimen from the base of the Buxton Rock Member near Church Stretton golf club house at SO 4518 9438.Note the phenoclasts of subhedral plagioclase and anhedral perthitic feldspar in a very fine grained,felsic groundmass with abundant fine mica and chlorite.Scale bar is 0.5mm long.

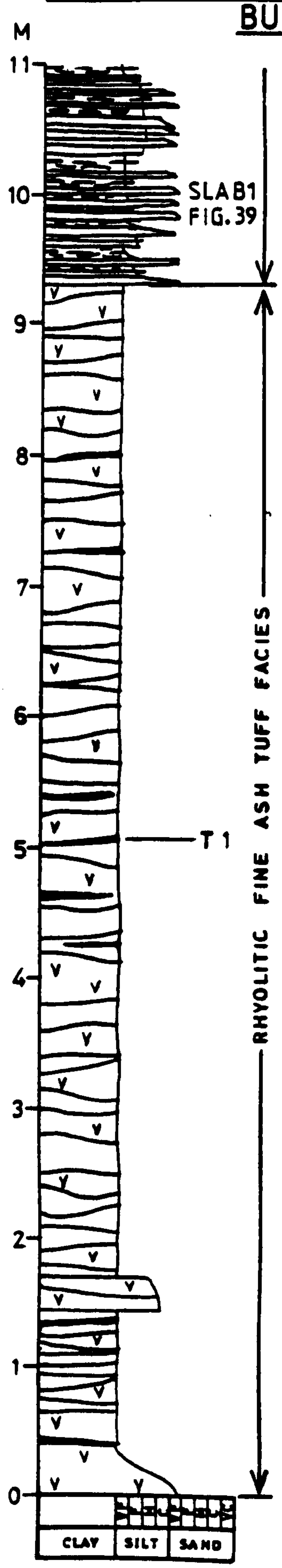
et al. (1968, p.40). One of these subspherical bodies is clearly authigenic since it cuts across a thin vein. Irregular patches of replacive calcite are common and there are occasional veins of chlorite, muscovite, calcite, quartz and epidote. These minerals may occur in zones within the veins.

The Buxton Rock Member of the Burway Formation is mostly composed of this lithology and is a homogeneous unit up to 15m thick (fig. 36). Throughout the outcrop, the member displays a lensoid bedding, 5cm to 40cm thick (fig. 37, photo A). Laminae of mudstone occur between some of these beds. The mudstone is very thinly laminated and these laminae are discontinuous and subparallel. The lithology is similar to the thinly laminated mudstone facies of the underlying Stretton Shale Formation and the mudstones in the overlying, thin-bedded, turbidite facies. The base of the Buxton Rock is visible near the Church Stretton golf club at SO 45189438. The base is abrupt and rests on the thinly laminated mudstone facies of the Stretton Shale Formation. The base is rich in crystals and volcanic fragments, some of which are rhyolitic. The top of the Buxton Rock is visible in Ashes Hollow at SO 4392 9266, where it is abruptly overlain by the thin-bedded turbidite facies of the Burway Formation. A log was constructed at this locality (fig. 38, log 1). Recrystallisation of the tuff has obliterated the original textures, however, in some places (at 0m and at 1.6m, log 1, fig. 38) normally graded units, 30cm to 40cm thick, can be recognised. A thin lamination is present in places, which is recognisable by slight colour changes, from pale, light grey to pale, medium grey. However, the majority of this facies

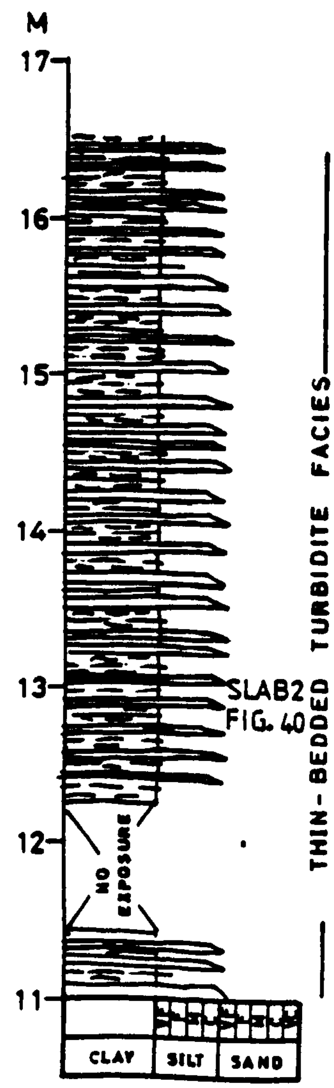
FIG:38

LOG 1 ASHES HOLLOW, GRID REF. SO 43929266

BUXTON ROCK



This log shows the rhyolitic fine ash tuff facies of the Buxton Rock, which is abruptly overlain by the thin-bedded turbidite facies. Note the lensoid nature of the bedding in the Buxton Rock, which is tectonic in origin, normally graded beds near the base of the Buxton Rock and the thick laminae of mudstone (dark layers). Note the thin beds of normally graded sandstone in the overlying, thin-bedded turbidite facies, together with thin interbeds of very thinly laminated mudstone. Two cut and polished specimens from this facies (slabs 1 and 2) are shown in figs. 39 and 40 .



appears to be homogeneous. Thin beds of tuff occur in the immediately overlying, thin-bedded turbidites of the Burway Formation at SO 4587 9554.

Interpretation

This facies is distinct from the turbidites of the Burway Formation. The distinctive characters are: the rhyolitic composition, the occasionally very poor sorting, the presence of very coarse to coarse lithic and crystal fragments and the presence of angular, delicate glass shards. The presence of laminae of thinly laminated mudstone within this facies suggests that the Buxton Rock Member is composed of a number of sedimentation events. The presence of thin beds in the immediately overlying, thin-bedded turbidite facies also suggests that discrete sedimentation events deposited this facies. Since there is a lack of significant thicknesses of other lithologies within the Buxton Rock Member, it can be argued that these sedimentation events were closely spaced in time and were probably genetically related. The presence of angular glass shards does not suggest that this facies has been extensively reworked by traction currents. It is later argued that the turbidites of the Burway Formation have been supplied with detritus by a fluviatile system which carried sediment with a maximum grain size of fine sand grade. It is unlikely that this facies was supplied with detritus by this system, since it contains angular glass shards and coarse to very coarse detritus, which are both absent in the deposits of this fluviatile system. Additionally, the turbidites of the Burway Formation are significantly less acidic in composition.

Immature, poorly sorted, volcaniclastic detritus suggests a pyroclastic origin. Pyroclastic activity of rhyolitic composition could have supplied all of the lithological components observed in this facies. This facies could have been deposited directly by pyroclastic activity or might have originated by reworking of a pyroclastic deposit. Since the lithology is immature, poorly sorted in places and contains delicate glass shards, any reworking could not have been extensive. The association with thin-bedded turbidites suggests a deep-water depositional environment. If this facies was deposited by settling through water, then the poorly sorted textures exemplified by the sample shown in fig. 37, photo B are difficult to explain. Additionally, deep-sea ash-fall deposits are usually less than 10cm thick (Wright and Mutti, 1981 and Fisher and Schmincke, 1984). The formation of the Buxton Rock would have required a number of unusually large, closely spaced subaerial eruptions. For example, Sparks et al. (1983) describe the deposits of a very large eruption, which produced $28km^3$ of ash. These include a distal, ash-fall unit only 5cm thick. Thicker deposits, up to 210cm thick, were deposited in a proximal location, however, these deposits contain resedimented ash deposited by turbidity currents (Sparks et al., 1983).

The closely spaced, discrete sedimentation events and the occasional normal grading could have been produced by turbidity currents, which resedimented ash from the slopes of a volcano. Turbidity currents might have been induced from slumps generated by the seismic or eruptive activity of the volcano. Layers rich in coarse crystals and lithic fragments could have been deposited from the head of a turbidity current (e.g. Sparks et al., 1983). The normal grading could have been produced by the waning flow of a turbidity current. The planar lamination observed in this facies may represent the "B" division of the Bouma turbidite sequence (e.g. Walker, 1984, p.173).

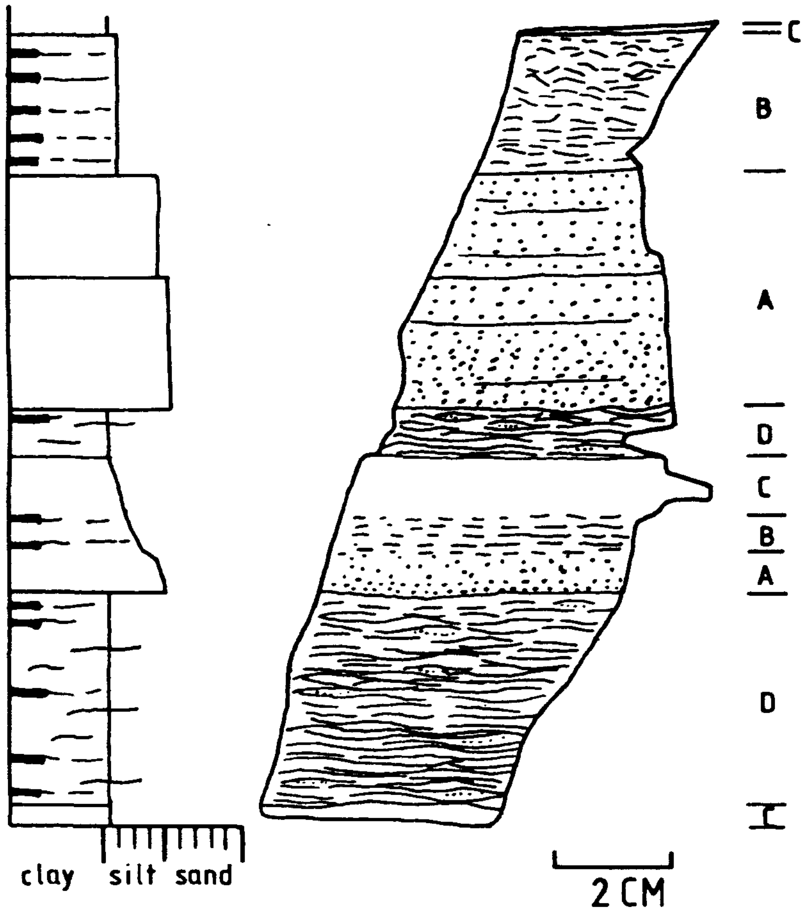
Alternatively, sediment gravity-flows may be generated directly from subaqueous pyroclastic flows or by the entering of a subaerial, pyroclastic flow into water (e.g. Suthren, 1985). The presence of glass shards in this facies does not preclude the possible production of the ash wholly by subaqueous, pyroclastic eruptions, since the volatile fragmentation depth for silicic magma, in which volatiles contribute to the disruption of the magma during eruption, is less than 500m (Fisher and Schmincke, 1984). However, the highly vesiculate nature of some of the shards in the Buxton Rock might suggest eruption into water depths considerably less than 500m. It is therefore possible that this ash was produced wholly by subaqueous, phreatomagmatic eruptions. Such an eruption would produce abundant, non-vesicular to poorly vesicular, fine grained fragments (Fisher and Schmincke, 1984). Collapse of a subaqueous eruption-column generates a subaqueous, pyroclastic flow (Fisher and Schmincke, 1984). Suthren (1985) notes that subaqueous, pyroclastic flows mostly deposit massive, unsorted, matrix-supported and pumice-rich material, but can pass distally into bedded tuffs deposited by thin, sediment gravity-flows, settling from suspension and reworking. Repeated phreatomagmatic explosions may occur at intervals varying from minutes to hours (Fisher and Schmincke, 1984). Consequently, it is inferred that a thick deposit may accumulate in a short time interval, which consists of a number of discrete, though genetically related, beds. These characteristics were noted in the rhyolitic, fine ash facies.

6.4 THE THIN-BEDDED TURBIDITE FACIES

This facies is characterised by thin, planar, parallel beds that have a maximum thickness of 10cm and an average thickness of 5cm or less (fig. 36). The maximum grain size is coarse silt to very fine sand and the average grain size is silt (fig. 36). The facies is composed of two main lithologies: thin beds of very fine grained sandstone or coarse siltstone and laminated mudstone. The sandstone and coarse siltstone is medium greenish grey and the mudstone is usually pale, light, greenish grey.

The thin beds of very fine sandstone and coarse siltstone from directly above the Buxton Rock Member of the Burway Formation display sequences of sedimentary structures (fig. 39, slab 1). Four units, composed of distinctive lithologies are recognised (A, B, C and D, fig. 39). Three of the units (A, B and C) are organised in an upward-fining sequence. The remaining unit (D) occurs as thin to very thin beds between these upward-fining sequences. Unit A rests with an abrupt, often loaded base on unit D. It is composed of medium greenish grey, coarse grained siltstone or very fine grained sandstone. The unit is very slightly, normally graded and can show planar, parallel lamination. Unit B may rest with an abrupt base on unit A or may have a gradational base. It is composed of coarse to very fine, light greenish grey siltstone and can be non-graded or normally graded. It contains numerous, very thin, discontinuous, wavy and subparallel laminae of clay. Unit C may rest with an abrupt or gradational base on unit B. It is composed of very fine grained, argillaceous siltstone, which is very pale, light greenish grey and is either homogeneous or has very faint, discontinuous laminae of clay. Unit D rests with an abrupt base on unit C and is abruptly

FIG. 39 SLAB 1, SKETCH OF THE THIN - BEDDED TURBIDITE FACIES, ILLUSTRATING THE SEQUENCE OF SEDIMENTARY STRUCTURES.



UNIT D: Very fine argillaceous siltstone, light to medium greenish grey, wavy, non-parallel laminated, occasionally with poorly sorted medium grained siltstone in lenses, abrupt base and top.

UNIT C: Very fine argillaceous siltstone, very pale, light greenish grey, homogeneous, or with very faint, discontinuous lamination, abrupt to gradational base.

UNIT B: Siltstone-light greenish grey, non graded or normally graded, numerous, very thin, discontinuous, wavy, parallel laminae of clay, abrupt or gradational base.

UNIT A: Siltstone or very fine grained sandstone, medium greenish grey, slight normal grading occasional planar, parallel lamination, occasional loading at base, abrupt base. Siltstone is coarse grained.

overlain by unit A. It is composed of laminated, light to medium greenish grey, very fine grained, argillaceous siltstone. The lamination is wavy, non-parallel and lenticular in appearance and poorly sorted, medium grained siltstone occasionally occurs in the lenses.

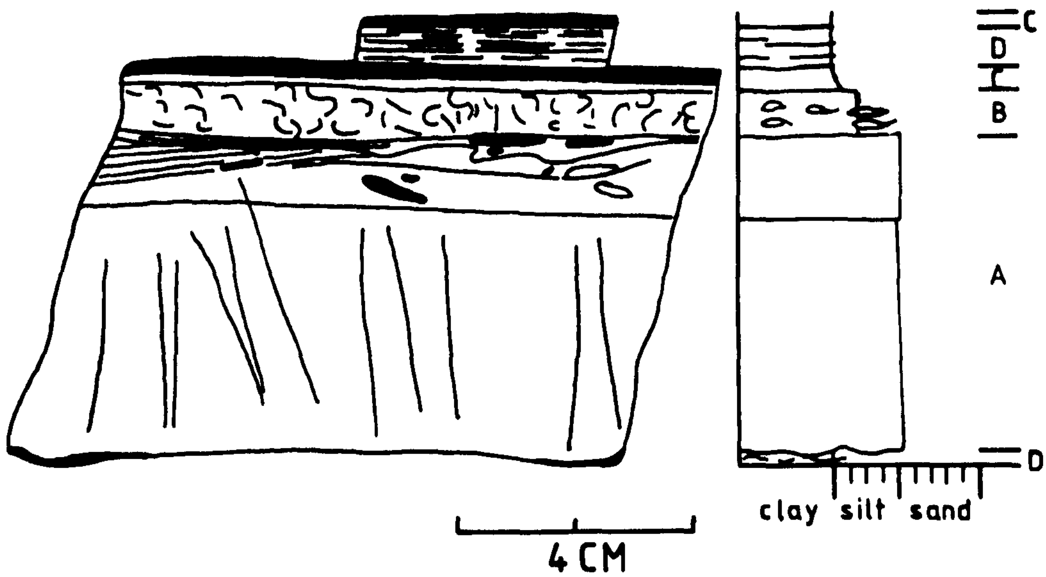
A number of samples of the thicker beds were cut and polished (figs. 40 and 41, slabs 2, 3 and 4). Some of these display additional features to those noted above. Most of the units described above can be recognised in these beds, however, in some cases, these units can be less easily differentiated. All of the beds display an abrupt and often flamed and loaded base on siltstone. The beds are composed of medium greenish grey, very fine grained sandstone, which is either non-graded or slightly, normally graded. This sandstone can be referred to unit A. The top of the sandstone is commonly abrupt and is overlain by mudstones with various types of lamination. These mudstones may be referred to units B, C and D.

In these thicker beds, planar, parallel lamination is commonly developed towards the base of unit A (fig. 41, slab 4). Towards the top of unit A, oblique lamination can be developed, which appears to be due to soft-sediment deformation and rarely, cross-lamination is developed in sets less than 1cm thick (fig. 40, slab 2 and fig. 41, slab 4). Clasts of mudstone, identical to the interbedded mudstone, can occur in unit A. These are commonly found towards, or at the very top of the unit (fig. 40, slab 2).

The immediately overlying, light greenish grey siltstone is often normally graded from coarse to fine grain size, is thinly, planar, parallel laminated and rests with an abrupt base on unit A.

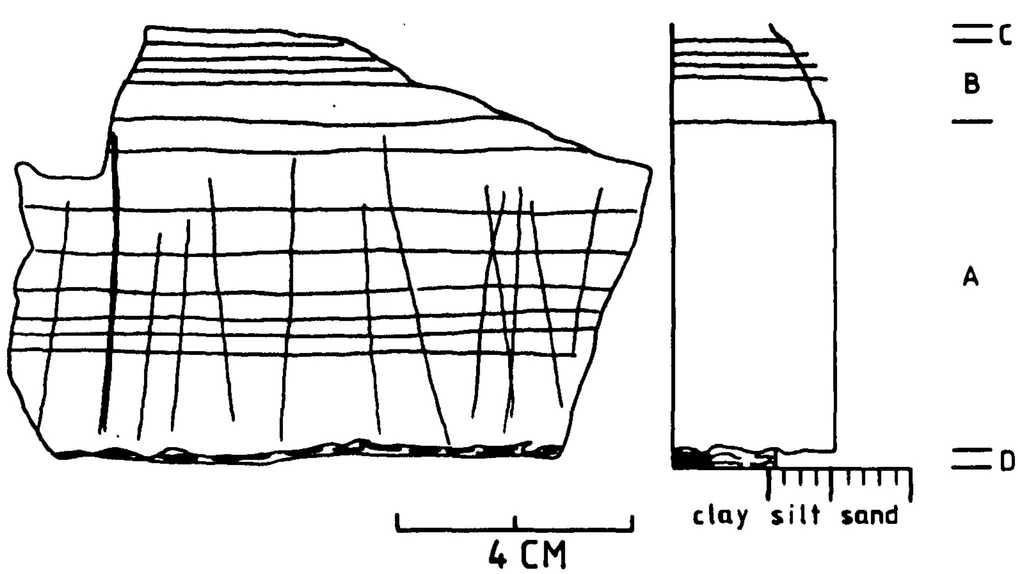
FIG.40 SLABS 2&3, SEDIMENTARY STRUCTURES IN THE THIN-BEDDED TURBIDITE FACIES

SLAB 2 BURWAY FORMATION



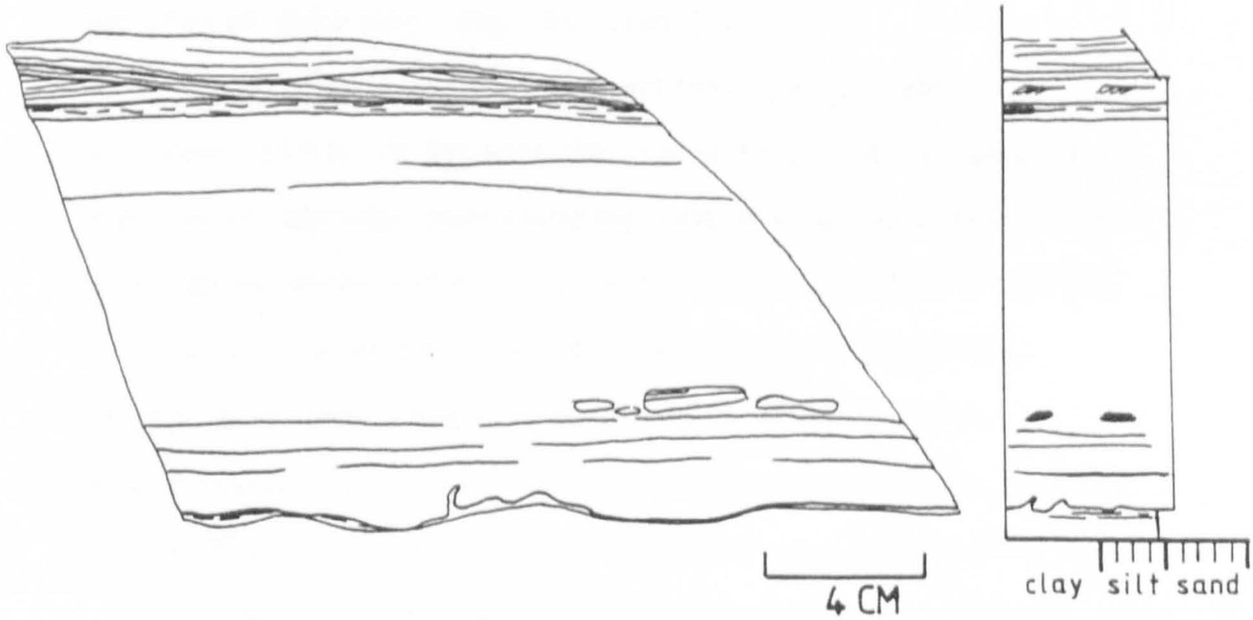
Note the loaded base to unit A;cross-laminated top to unit A with mudstone clasts;intermixed patches of claystone, siltstone and very fine sandstone in unit B and ? soft sediment deformation at top of unit A.Vertical lines are fractures.

SLAB 3 BURWAY FORMATION



Note flamed and loaded base to unit A and laminae of coarser siltstone in unit B.Vertical lines are fractures.

FIG.41 SLAB 4, THIN-BEDDED TURBIDITE FACIES



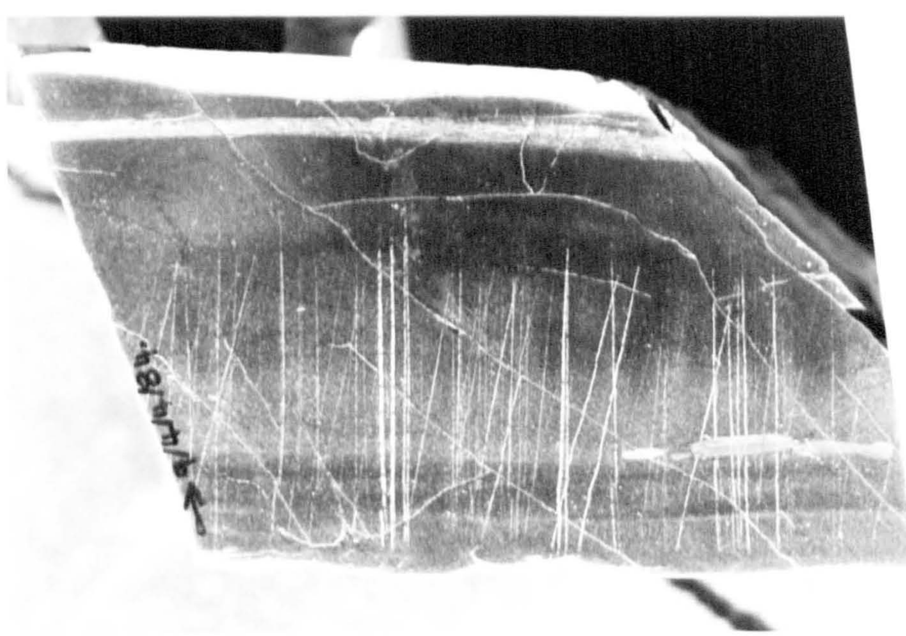


PHOTO AND SKETCH, SLAB 4. Note the flamed and loaded base to unit A; planar and parallel laminae and siltstone clasts towards the base of unit A; faint cross-lamination at the top of unit A; paler coloured siltstone of unit B. Unit A is only slightly normally graded. Vertical lines are fractures.

This siltstone can be referred to unit B. Occasionally, irregular patches and pods of coarse silt and sand occur within the fine silt and clay of this unit (fig. 40, slab 2).

A thin lamina of very fine grained, pale, light green siltstone, which can be normally graded to claystone, abruptly overlies or succeeds the preceding unit B by a rapid transition. This lamina may be referred to unit C (fig. 40, slabs 2 and 3).

Unit D is probably represented by very fine grained, argillaceous, pale, light, greenish grey siltstone with discontinuous, wavy, non-parallel laminae, which occurs beneath unit A (fig. 40, slabs 2 and 3) and above unit C (fig. 40, slab 2).

A plot of grain size and bed-thickness against stratigraphic level shows a systematic trend in the lower part of the Burway Formation of Ashes Hollow (fig. 36). Directly above the Buxton Rock Member of the Burway Formation, the bed-thickness increases and this is reflected in an increase in the average grain size and the average maximum grain size. Subsequently, the bed-thickness decreases gradually and this is accompanied by a decrease in the average grain size and a slight decrease in the average maximum grain size. These trends reflect the variation in the thickness of unit A and the variation in the percentage of sandstone and siltstone, resulting from the variations in the relative abundance of unit A. As unit A decreases in thickness, there is an accompanying decrease in the relative abundance of unit A. This results in a decrease in the average grain size. There is also a tendency for the maximum grain size of unit A to decrease slightly as the thickness of the unit decreases. However, this is not always the case. For example, one lamina of unit A is composed of fine-grained sandstone, which is slightly coarser than average, despite the thinness of the unit.

Where laminae rather than beds of unit A occur, these are associated with mudstones which usually exhibit abrupt contacts with the laminae and which are composed of light greenish grey, very fine grained and poorly sorted siltstone. This contains numerous, discontinuous layers of claystone and very thin, discontinuous laminae of fine to medium grained, poorly sorted siltstone. This lithology is similar to and probably referable to unit D. The thicker laminae of unit A are often very thinly, planar, parallel laminated.

The lower part of the Burway Formation is comprised of the thin-bedded turbidite facies. It is directly underlain by the Buxton Rock Member of the Burway Formation, which is comprised of the rhyolitic, fine ash tuff facies (fig. 36). Some thin units of thin-bedded turbidites occur in the thick-bedded turbidite facies and these are associated with the massive turbidite facies. These are similar to the thin-bedded turbidite facies and are discussed in the following sections.

Interpretation

The upward-fining sequence, composed of units A, B and C (fig. 39) was probably deposited from a current of waning velocity. The presence, in some of the beds, of ripple cross-lamination at the top of unit A and planar, parallel lamination at the base of unit A, suggests that unit A was partly deposited under upper-stage, plane-bed conditions with a transition at the top to lower flow-régime current ripples. However, unit A is mostly massive and lacks current structures. This suggests that the majority of unit A was deposited rapidly from suspension, with only occasional bed load transport. The ripple cross-lamination always occurs at the very top of unit A (figs. 40, slab 2 and 41, slab 4). There is a lack of climbing ripple cross-lamination and the cross-lamination occurs in single sets, some of which are form sets. This suggests that this cross-lamination might have been produced by the reworking of the upper parts of unit A by a current depleted in sand. There is usually an abrupt change from unit A to unit B, which implies a rapid change in flow velocity. That units B and C systematically follow unit A and are normally graded suggests that this sediment was supplied by the same current which deposited unit A. The lack of repeated grain size variation within units B and C, in the majority of cases, suggests that these units were not deposited by a fluctuating current or by a process of depositional sorting in the boundary layer of a current (e.g. Stow and Bowen, 1980). It is suggested that units B and C were mostly deposited by settling from suspension, but the presence of a planar, parallel lamination in some cases indicates current activity. Unit B is characterised by numerous, very thin, discontinuous, wavy, subparallel laminae of clay. These clay-rich laminae might have been produced by a dewatering mechanism and can be compared to dish structures. It is suggested that dewatering occurred immediately following the deposition of unit A and during the deposition of unit B. The disruption of unit B and the top of unit A in fig. 40, slab 2 might have resulted from dewatering, since there is no evidence for loading above unit B. Unit C can often be separated as a distinct unit by its grain size and colour. This was probably deposited from suspension by the slow settling of very fine silt and clay.

The upward-fining cycle, which consists of units A, B and C can be interpreted as representing the deposit of a single current and unit C represents the finest grain size which was deposited from this. Unit D appears to have a different mechanism of formation. Where unit A is thinner, then unit D forms a greater proportion of the lithology. This suggests that the argillaceous unit D represents background sedimentation which was interrupted by the currents which deposited units A, B and C. The fine grained, homogeneous nature of unit D suggests that it was deposited by settling from suspension. The fine grained sediment might have been supplied to the water column by the currents which deposited units A, B and C. Alternatively, the sediment might have been introduced into the water column by other mechanisms, such as in upper-water, turbid layers resulting from river floods or storms.

The currents which deposited units A, B and C can be compared to some subaqueous density-flows. The sedimentation of these units involved both traction and settling from suspension. Lowe (1982) interpreted this type of sedimentation to be due to a fluid turbulence, sediment support mechanism in a turbidity current. However, since suspension sedimentation is evidently dominant, fluidised flow and liquefied flow (Lowe, 1982) might have been important during the deposition of these units. Liquefied flows can deposit massive units of fine sand and coarse silt by rapid settling from suspension and the tops of these massive units may be reworked by water set into motion by shear at the flow surface (Lowe, 1982). Unit A shows similar characters to this type of deposit.

The sequence of structures in this facies can be directly compared with the classic Bouma sequence (e.g. Walker, 1984, p.173). Unit A is similar to the Bouma sequence division A. The planar lamination in unit A can be compared with the Bouma sequence division B. The cross-lamination at the top of unit A can be compared with division C,and units B and C can be compared with divisions D and E (t) (Walker, 1984, p.173) respectively. There are, however, significant differences between the Bouma sequence and the sequences observed in this facies. The top of unit A is commonly abrupt rather than gradational. Planar, parallel lamination is developed towards the base or in the middle of unit A, whereas in the Bouma sequence, this should be developed towards the top. The ripple cross-laminated units are very thin and this does not suggest that high sedimentation was coincident with the development of the cross-lamination. These differences may be due to the dominance of rapid settling from suspension during the deposition of unit A, as opposed to the importance of traction during the deposition of a classical Bouma sequence.

The sequence recognised in the beds of this facies is very similar to the facies C1 of Ricci-Lucchi (1984). Ricci-Lucchi (1984) recognises a massive, lower division followed by a thin division with a single set of cross-lamination. This is overlain by normally graded mudstone. Facies C1 was interpreted to have been deposited from a turbulent suspension by rapid settling, followed by non-depositional current drag (producing the thin, cross-laminated division) succeeded by slow settling from the tail of the current. These processes of deposition are similar to those deduced for the Longmyndian facies. Ricci-Lucchi (1984) interprets facies C1 as having been deposited from an immature, high density or concentrated turbidity current. Lowe (1982) maintains that the A division of the Bouma sequence, which is similar to unit A of this facies, was deposited by direct suspension sedimentation from high-density flows. High-density flows are characterised by sediment coarser than medium sand and particle suspension is dependent on concentration effects rather than by fluid turbulence alone (Lowe, 1982). However, the sediment grain size in the Longmyndian facies is not typical of high-density flows since it is of very fine sand to coarse silt size.

This thin-bedded turbidite facies can be compared with typical thin-bedded turbidites (e.g. Mutti and Ricci-Lucchi, 1978 and Ricci-Lucchi, 1984). These typically show base-absent Tbe, Tce, Tde and Te Bouma sequences (e.g. Ricci-Lucchi,1984). The Longmyndian facies is distinct from these sequences since the Ta and Tb divisions are dominant. Base-absent Bouma sequences in thin-bedded turbidites are usually considered to have been deposited by dilute (Mutti and Ricci-Lucchi, 1978) or low-density (Lowe, 1982) turbidity currents. For these reasons, the Longmyndian thin-bedded turbidite facies is not thought to have been deposited by dilute, low-density turbidity currents.

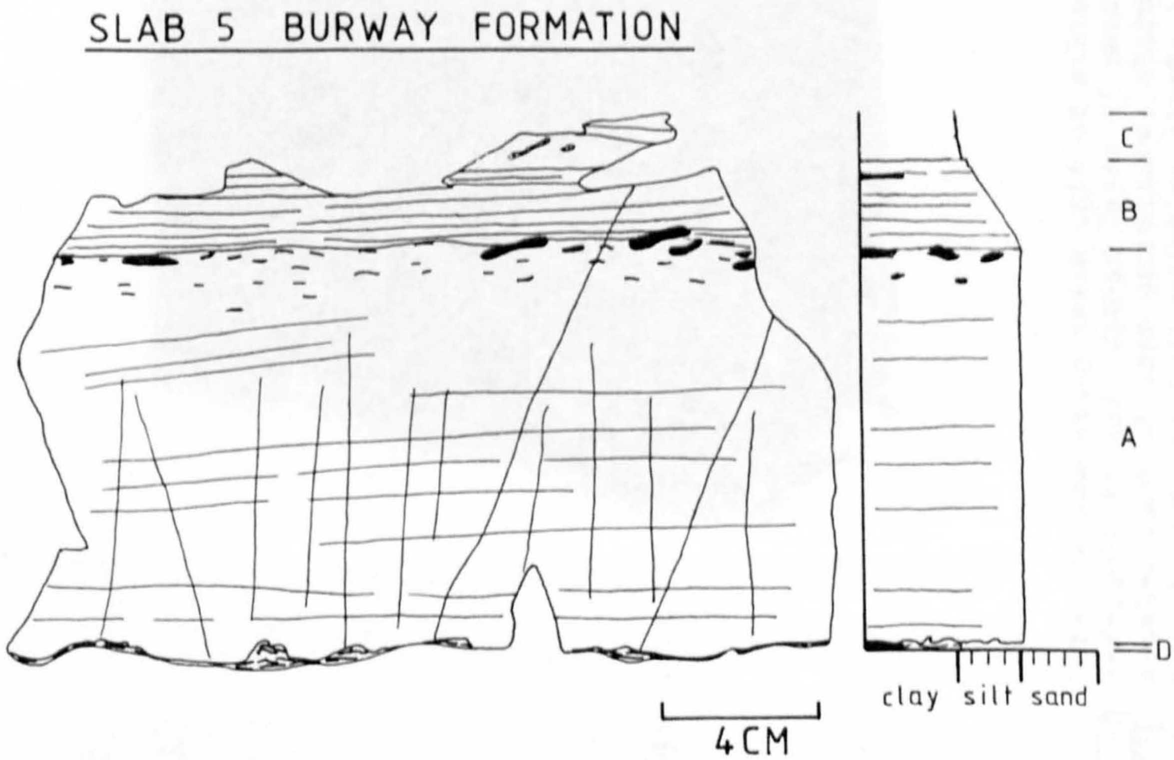
6.5 THE THICK-BEDDED TURBIDITE FACIES

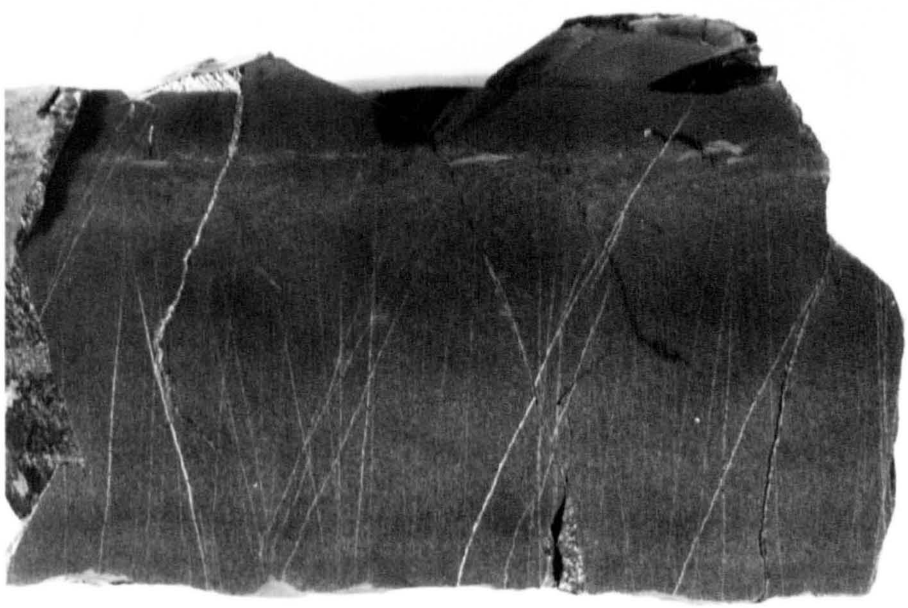
This facies is characterised by parallel-sided, planar beds, that have a maximum thickness of 1.5m, an average thickness of c.20cm and an average range of thicknesses between 10cm and 55cm (fig. 36). The maximum grain size is very fine sand and the average grain size is coarse silt to very fine sand (fig. 36). The facies is characterised by repeated, normally graded sandstone beds with intervening laminae of mudstone. There is a lack of thick beds of laminated mudstone throughout the majority of the facies. The proportion of sand in this facies is therefore higher than in the thin-bedded turbidite facies. This facies is illustrated by log 2, fig. 50.

Several samples of the beds were cut and polished. Some of these samples are illustrated in figs. 42, 43, 44 and 45. Four different types of lithology can be recognised in these samples and these are designated units A, B, C and D. These units are very similar to the units recognised in the thin-bedded turbidite facies. Unit A consists of greenish grey, very fine grained sandstone and rests with a loaded and often flamed base on the mudstone of unit D (fig. 43). There is an abrupt or rapid transition at the top into the mudstone of unit B (figs. 42 and 45). The unit is often planar laminated towards the base (fig. 45) and occasionally in the middle (fig. 42). Occasionally, convolute lamination is developed at the top. Small clasts of pale mudstone, similar to unit D, often occur at the base (figs. 44 and 45) and large clasts may occur at the top (fig. 42). Unit A is usually not graded or is very slightly normally graded (figs. 42, 44 and 45).

Unit B is composed of light greenish grey, normally graded, coarse to fine siltstone (figs. 42, 43 and 44). There is an abrupt or rapid transition at the base with unit A. Planar, parallel, continuous laminae are often present (figs. 42 and 43) and an oblique lamination can occur towards the middle of the unit (figs. 43 and 44) and at the top (fig. 44). Occasionally, discontinuous,

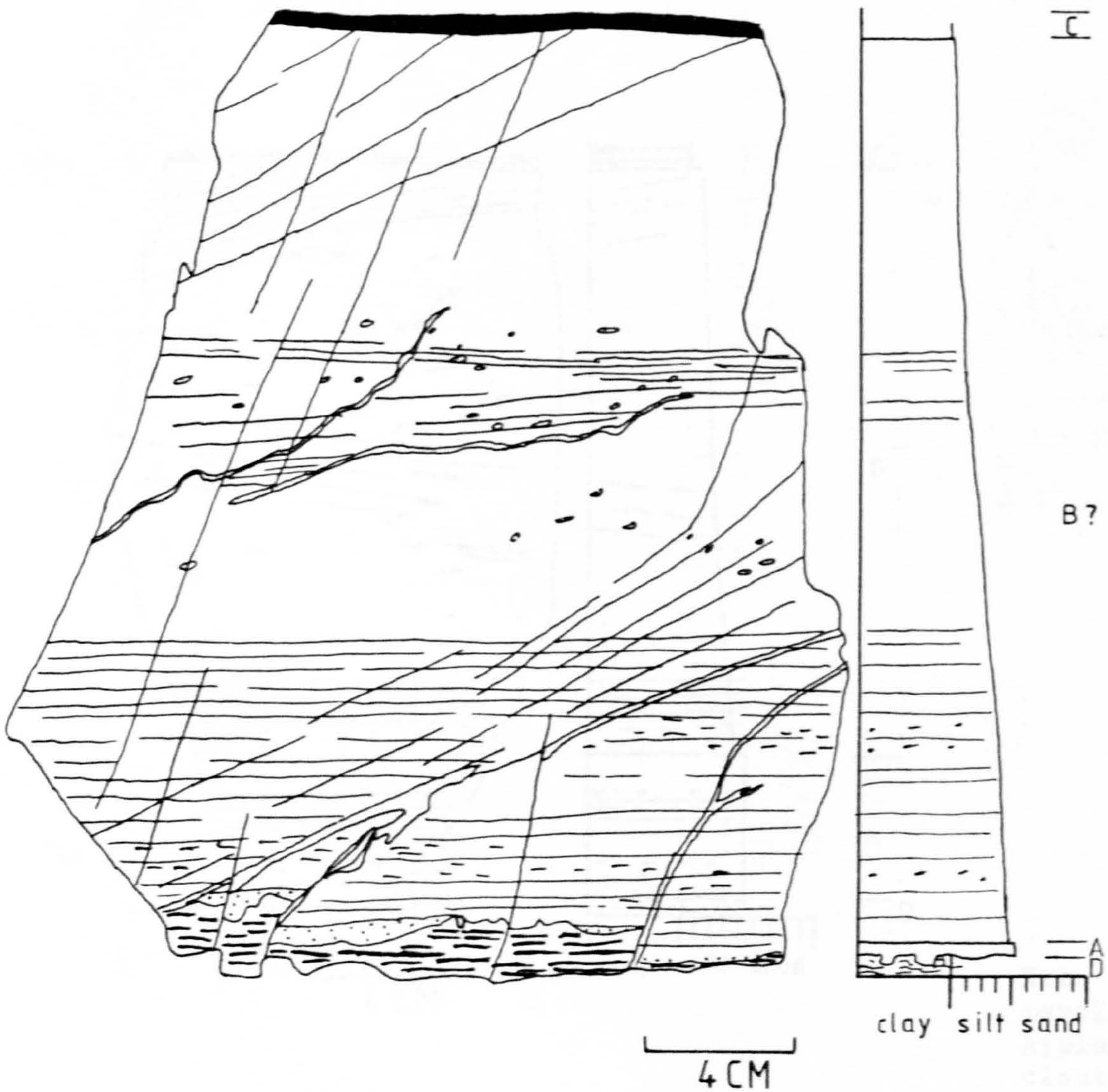
FIG.42 SLAB 5, THICK-BEDDED TURBIDITE FACIES





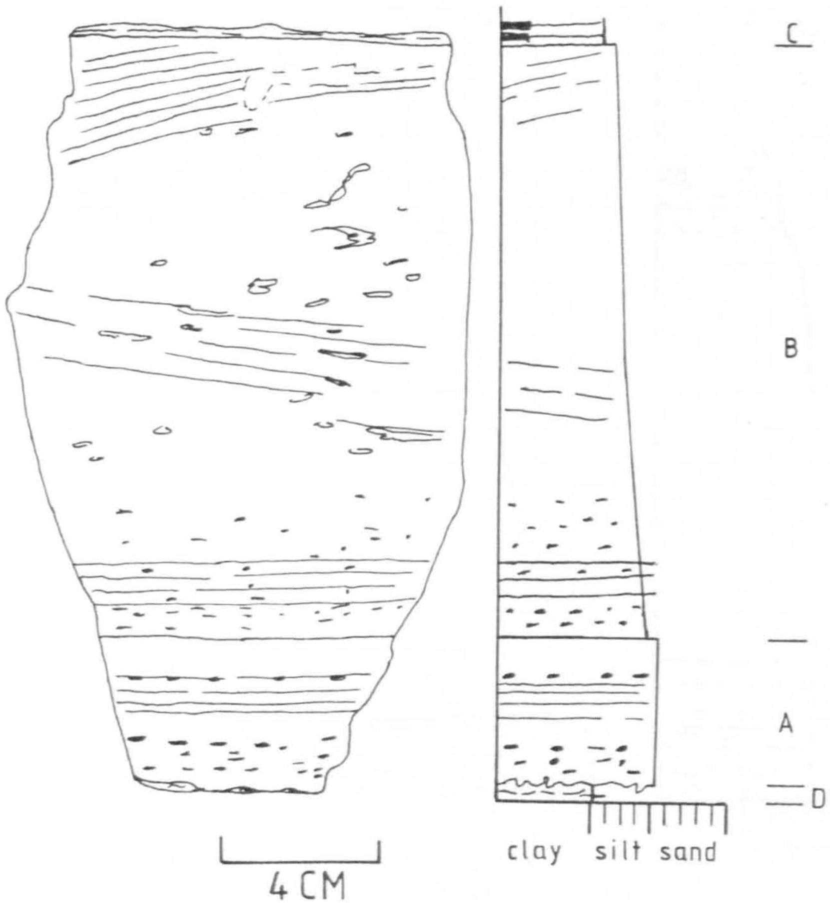
Note loaded base to unit A;planar,parallel lamination in unit A;lack of grading in unit A;mudstone clasts at top of unit A;abrupt base to unit B;planar,parallel lamination in unit B;lack of lamination in unit C and different tones shown by units C and D in photo.

FIG 43 SLAB 6, THICK-BEDDED TURBIDITE FACIES



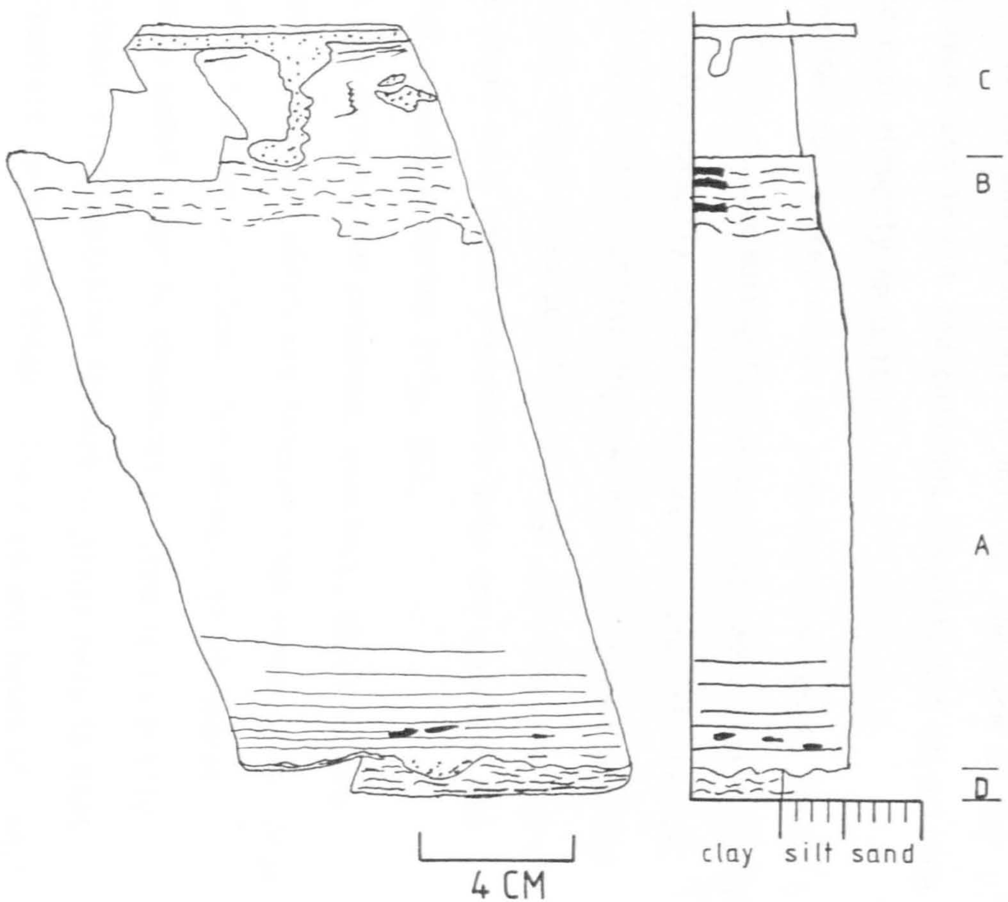
Note the unusually thick unit of siltstone probably referable to unit B; thin layer of sandstone of unit A; paler tone of unit D; fractures disrupting loaded base to unit A; small mudstone clasts in base of unit B; small rounded patches in middle of unit B are diagenetic.

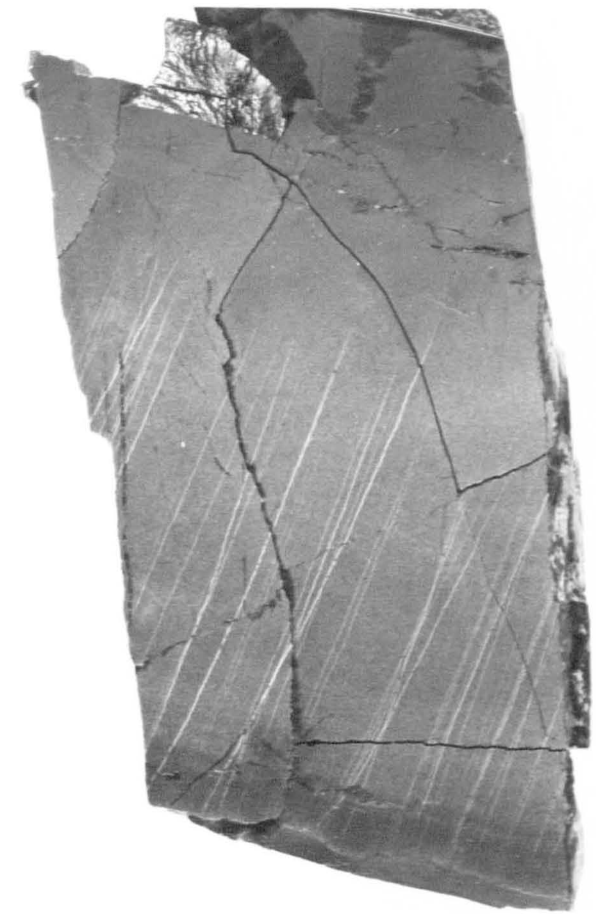
FIG.44 SLAB 7, THICK-BEDDED TURBIDITE FACIES



Note the flamed and loaded base to unit A;planar, parallel lamination and pale mudstone clasts in unit A;planar,parallel lamination and small mudstone clasts at base of unit B;oblique lamination at top of unit B;normal grading of unit B.Irregular wisps in unit B are diagenetic.

FIG.45 SLAB 8, THICK-BEDDED TURBIDITE FACIES





Note the loaded base to unit A;planar,parallel lamination and pale mudstone clasts at base of unit A;gradation from unit A into unit B; discontinuous lamination in unit B;homogeneous nature of unit C;lamina of sandstone with extreme loading in unit C.

subparallel laminae are developed (fig. 45). Small clasts of pale mudstone can occur in the basal sections (figs. 43 and 44). This unit usually shows an abrupt contact with the overlying unit C.

Unit C is composed of very fine, light greenish grey siltstone which is usually homogeneous but can be very thinly laminated. It often shows a slight normal grading (figs. 42 and 45). Unit D is composed of very fine grained, argillaceous siltstone and claystone and is characterised by numerous, very thin, discontinuous, wavy, non-parallel laminae. It is easily distinguished from the mudstone of units B and C by its much paler colour (fig. 43). It usually forms units less than 1cm thick between the sandstone and siltstone beds, but can occur in thin beds (up to 10cm thick). The mudstone clasts within units A and B are similar in colour and lithology to unit D. The base of the unit has not been seen in cut and polished samples, but at outcrop it appears to rest abruptly on unit C.

Thick-bedded turbidites occur in a sequence which is 625m thick in the middle of the Burway Formation (fig. 36). The base of this sequence rests abruptly on the thin-bedded turbidite facies. However, there is a slight thickening and coarsening-upward trend at the top of the thin-bedded turbidites (fig. 36). At the top of this sequence there is a rapid transition into the mudstone and cross-laminated sandstone facies (fig. 36).

Towards the base of the sequence, distinct, thick-bedded, non-amalgamated beds occur which are thicker than average and range in thickness from 55cm up to 1.55m. The normal, thick-bedded turbidites have a modal range in thickness of 10cm to 55cm (fig. 36). An additional distinguishing feature of these beds is that they occur in packets 7m to 10m thick. The tops and bases of these packets are usually abrupt, but thickening-upwards trends (c. 1.8m thick) can be recognised beneath them in some places (e.g. fig. 46, photo A). A similar thickening and coarsening-upward trend, 1m thick, occurs at c. 20m on log 2, fig. 50. Within these packets, thinner turbidite beds and occasionally, thin beds of laminated mudstone can occur. These packets can be distinguished from the massive turbidite facies in that there is a lack of amalgamation of the beds. Additionally, thinner turbidite beds and beds of laminated mudstone occur, which are absent in the massive turbidite facies. Although the massive turbidite facies commonly shows thinning and upward-fining sequences, these are not associated with the packets of thick-bedded turbidites. Additionally, thickening-upwards sequences do not occur beneath the massive sandstone facies.

Some thinning-upwards sequences occur, which are associated with the massive turbidite facies (e.g. fig. 49). These thinning-upwards sequences are of the order of 13m thick, including the basal massive turbidite facies. These sequences are discussed in section 6.6, which describes this facies.

A thick, thinning-upward sequence (28m thick) occurs in the region of 500m on fig. 36. The average bed-thickness decreases from 18cm to 5cm and the maximum bed-thickness similarly decreases from 24cm to 9cm (fig. 36). These changes are accompanied by a slight decrease in the average and maximum grain sizes. The beds at the top of this sequence are unusually thin when compared with the majority of the thick-bedded turbidite facies. These thin-bedded turbidites are illustrated in fig. 46, photo B. Similar, thin-bedded turbidites are interbedded with the massive turbidites in the region of 550m on fig. 36. These thin-bedded

FIG. 46 PHOTOGRAPHS SHOWING THICKENING-UPWARD SEQUENCE AND THIN-BEDDED TURBIDITES IN THE THICK-BEDDED TURBIDITE FACIES.

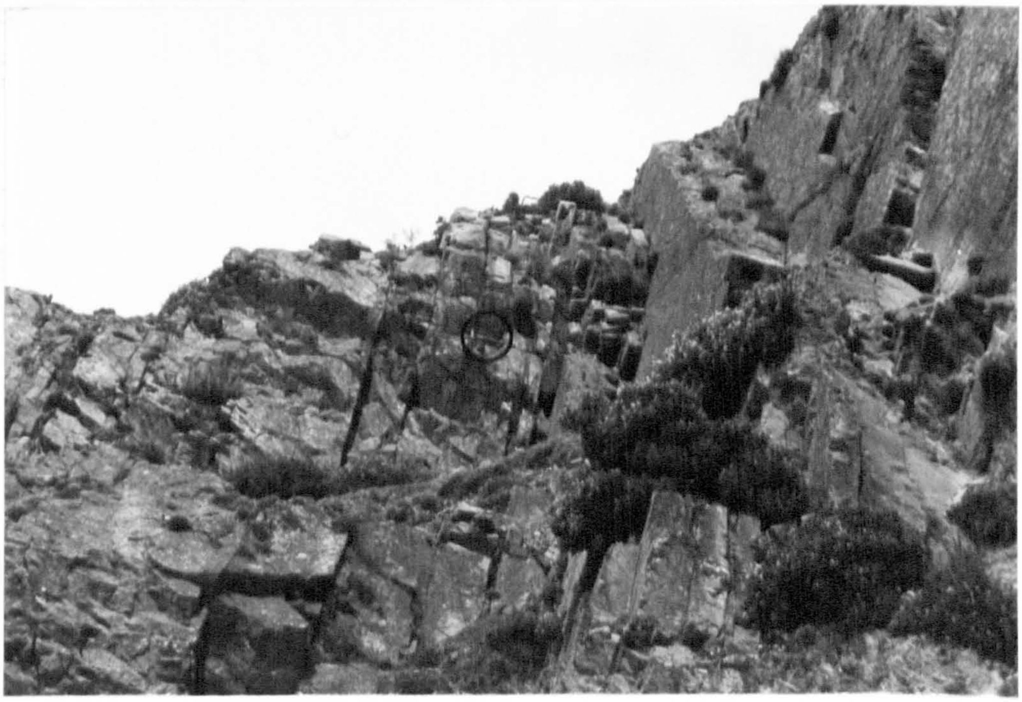


PHOTO A: Thickening-upward sequence in the thick-bedded turbidite facies, Ashes Hollow. Note the unusually thick turbidite bed on the left side of the photo, which is part of a packet of similar beds, 8.5m thick. The gradual thickening-upward trend is well displayed at the base of the photo and is 1.8m thick. Hammer is 28cm long (centre).

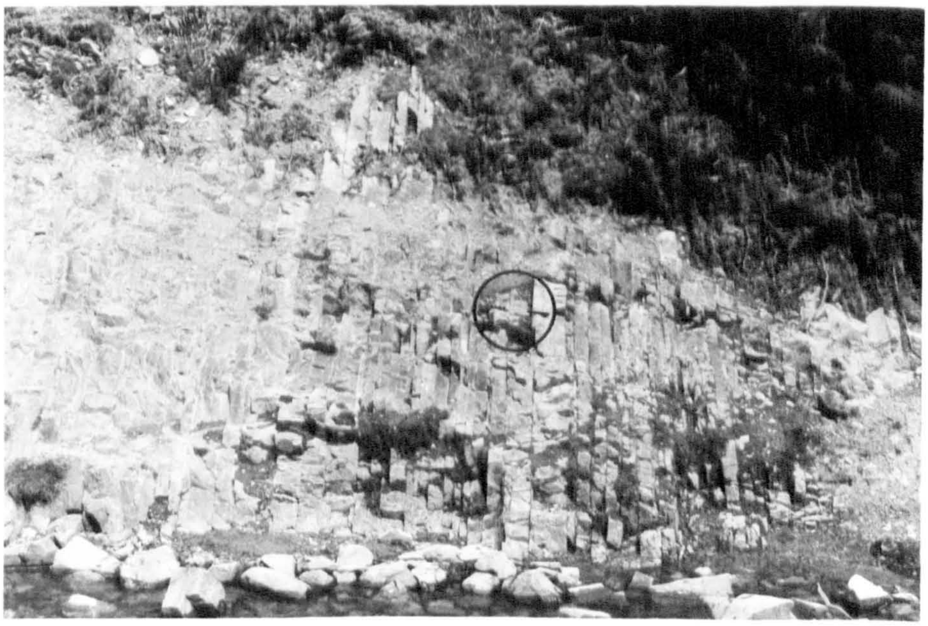
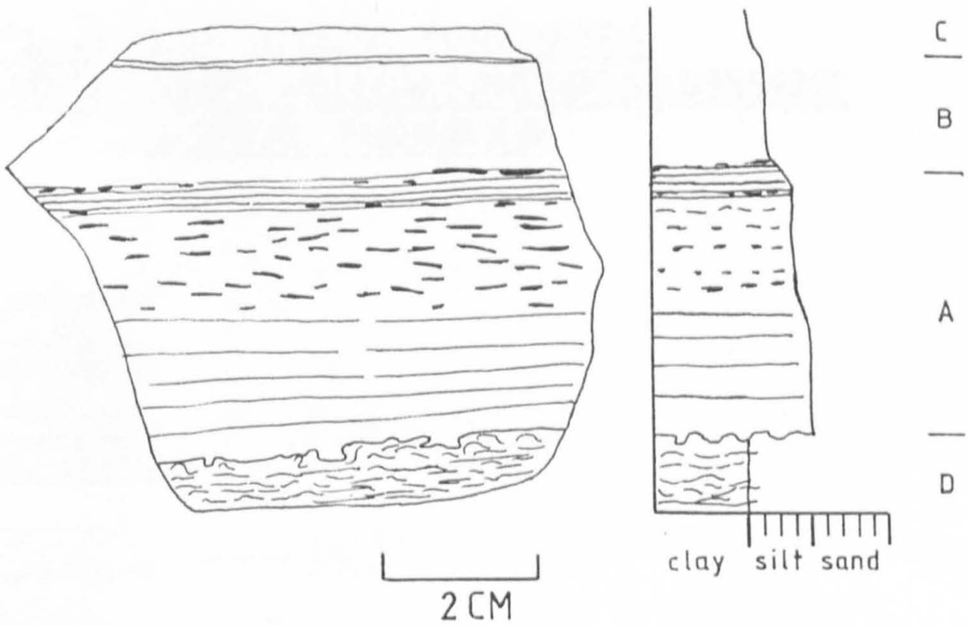
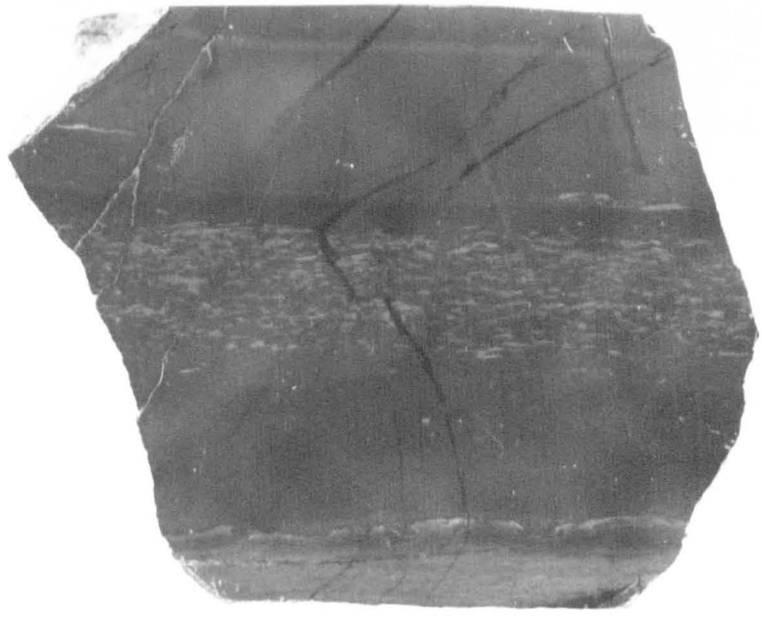


PHOTO B: Turbidites from the thick-bedded turbidite facies, which are thinner than average. These beds are directly overlain by massive turbidites. Scale is 30cm long and is graduated in decimetres and centimetres (centre of photo).

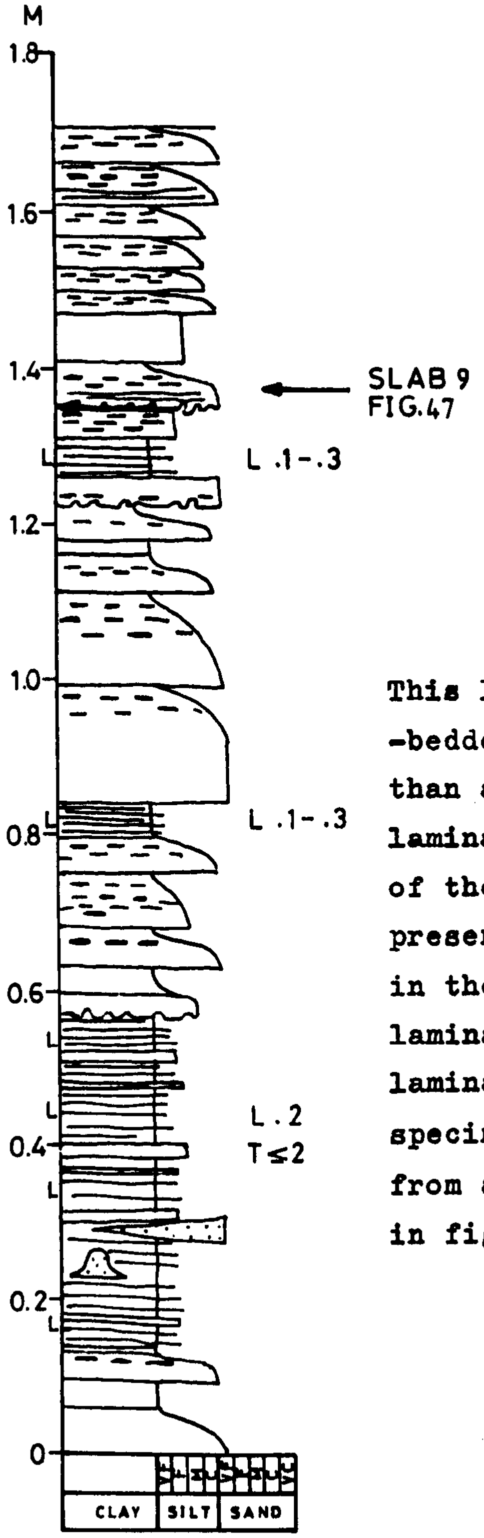
FIG.47 SLAB 9, THIN BED FROM THE THICK-BEDDED TURBIDITE FACIES





Note the very thin, discontinuous laminae of unit D; paler tone of unit D; flamed and loaded base to unit A; planar, parallel lamination at base of unit A; concentration of shale clasts at top of unit A; homogeneous nature of units B and C.

FIG:48 THIN-BEDDED TURBIDITES
LOG 5 ASHES HOLLOW, GRID REF, SO 43559291
BURWAY FORMATION



This log shows turbidites from the thick -bedded turbidite facies which are thinner than average.Note the high proportion of laminated mudstone compared with the majority of the thick-bedded turbidite facies;the presence of ubiquitous small mudstone clasts in the thin sandstone beds and the thick laminae of siltstone and sandstone in the laminated mudstone beds.A cut and polished specimen of the thin sandstone beds (slab 9) from approximately 1.4m on the log is shown in fig. 47 .

turbidites are usually 4.5cm to 10cm thick. The sequences of sedimentary structures in these beds are similar to those within the thick-bedded turbidites (fig. 47). Laminated mudstones are associated with these thin-bedded turbidites and these comprise a greater proportion of the lithology than in the majority of the thick-bedded turbidite facies (fig. 48, log 5).

There is a slight decrease in the average bed-thickness from the base to the top of the thick-bedded turbidite facies sequence in Ashes Hollow (fig. 36). At the top of the sequence (at c. 550m on fig. 36) there is a rapid decrease in the average and average maximum bed-thicknesses of the turbidites which are interbedded with the massive turbidite facies. The average bed-thickness decreases from c. 15cm to c. 2cm and the average maximum bed-thickness decreases from c. 31cm to c. 6.5cm, from 500m to c. 580m on fig. 36. This is accompanied by an increase in both the average and the average maximum bed-thickness in the massive turbidite facies.

Interpretation

The sequences of sedimentary structures and grain size changes within these beds are very similar to those observed in the thin-bedded turbidite facies. The four units that are recognised in these beds (units A, B, C and D) are directly comparable with the units recognised in the thin-bedded turbidite facies and similar hydrodynamic interpretations apply, for which the reader is referred to the interpretation of the thin-bedded turbidite facies. Some differences between the thinner beds and the thicker beds can be recognised. Unit B often shows planar lamination (figs. 42, 43 and 44), which suggests that currents were active during its deposition. An oblique lamination towards the middle and top of unit B (figs. 43 and 44) is suggestive of soft-sediment deformation, possibly due to dewatering. This was noted in some examples of the thin-bedded turbidites. No ripple cross-lamination was noted, either in the cut and polished specimens or at outcrop. This was occasionally noted in the thin-bedded turbidite facies.

The thick-bedded turbidites are interpreted as having been deposited by repeated waning currents. The common presence of planar and parallel lamination in units A and B suggests that traction was an important mechanism during their deposition. It was previously postulated that unit A, in the thin-bedded turbidite facies, was predominantly deposited by rapid settling from suspension, since it is often massive. Thus, traction might have been more important during the deposition of the thick-bedded turbidites than during the deposition of the thin-bedded turbidites. This difference could be due to higher current velocities during the deposition of the thick-bedded turbidites. Unit C is typically composed of very fine grained siltstone and silty claystone and is homogeneous. It is interpreted to have formed by the slow settling of clay and very fine silt. As noted in the discussion on the thin-bedded turbidites, unit D is distinct, in that it is argillaceous, is much paler in colour and is characterised by discontinuous, wavy, non-parallel lamination. This unit is similarly interpreted as having been deposited by slow settling from suspension and to represent background sedimentation.

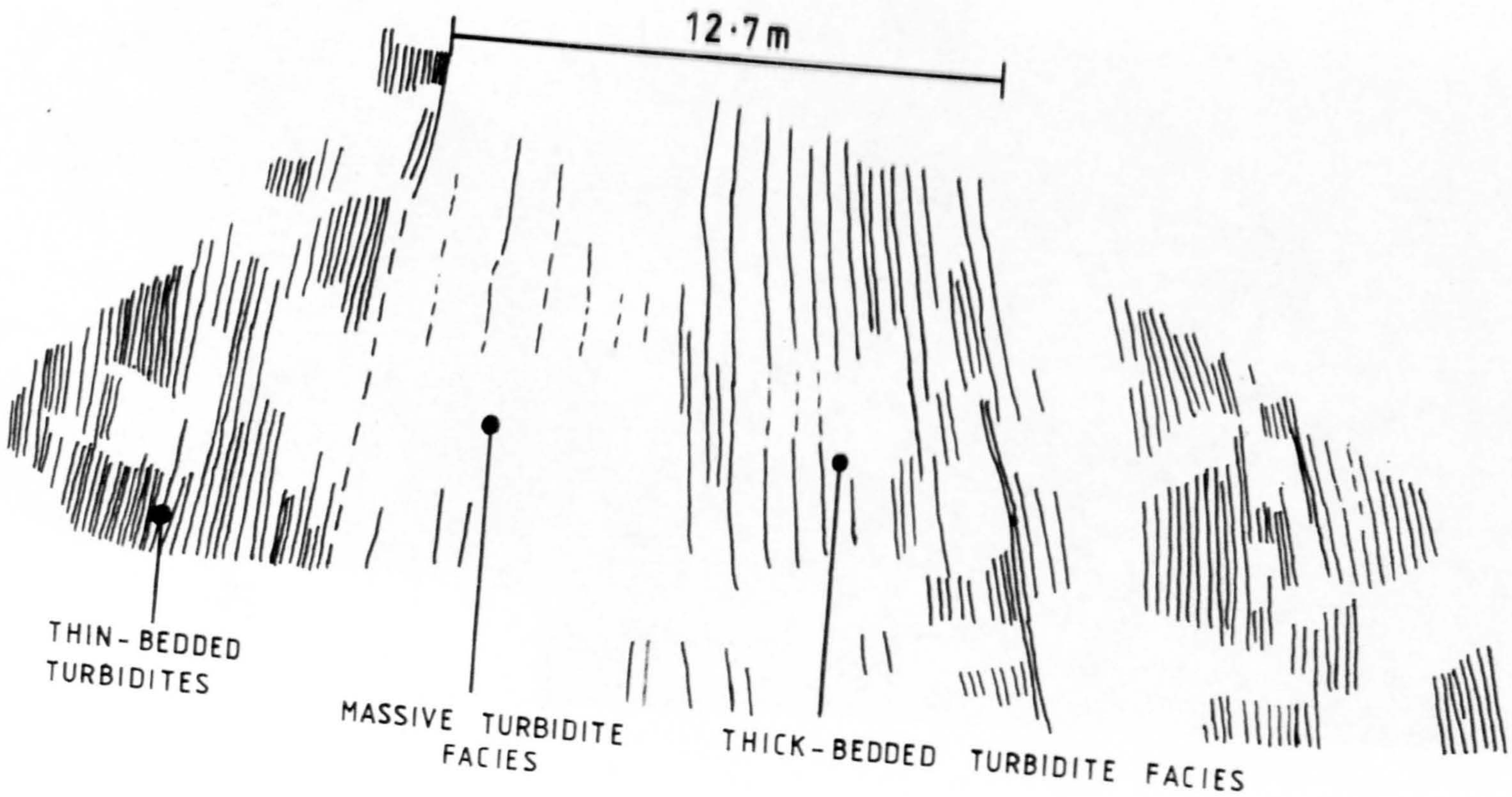
The deposition of these units involved both traction and settling from suspension. Lowe (1982) interpreted this type of sedimentation to result from a turbidity current. The sequence of structures in this facies can be compared with the classic Bouma sequence (e.g. Walker, 1984, p.173). Unit A is comparable with divisions A and B of the Bouma sequence. Unit B is comparable with division D, unit C with division E(t) and unit D with division E(h) (divisions after Walker, 1984, p.173).

This facies is comparable with facies C2 of Ricci-Lucchi (1984). This is interpreted as having been deposited from concentrated turbidity currents by traction and by settling from suspension and is compared with classic Bouma-sequence turbidites (Ricci-Lucchi, 1984). The Longmyndian turbidites are distinctive, however, in lacking a ripple cross-laminated division C and in showing distinct grain size changes at the unit boundaries, rather than a continuous gradation in grain size.

The significance of the trends in bed-thickness, the packets of non-amalgamated thick turbidites and the thinning-upward trends associated with the massive turbidite facies are discussed in section 6.7.

6.6 THE MASSIVE TURBIDITE FACIES

This facies is interbedded with the thick-bedded turbidite facies. It occurs as very thick beds, 2.5m to 9.6m thick, at the top of the sequence of thick-bedded turbidites, or more rarely within the middle of the sequence (fig. 36). This facies is characterised by massive or indistinctly thick-bedded, very fine to fine grained sandstone. The bases of the beds are abrupt and the tops can be abrupt or gradational with the thick-bedded turbidites. Where the tops are gradational, thinning-upwards sequences of turbidites can occur, which are of the order of 13m thick. One



12·7m
THIN-BEDDED TURBIDITES
MASSIVE TURBIDITE FACIES
THICK-BEDDED TURBIDITE FACIES

FIG:49 THINNING-UPWARD SEQUENCE, ASHES HOLLOW, SO 43559291

ESE

WNW

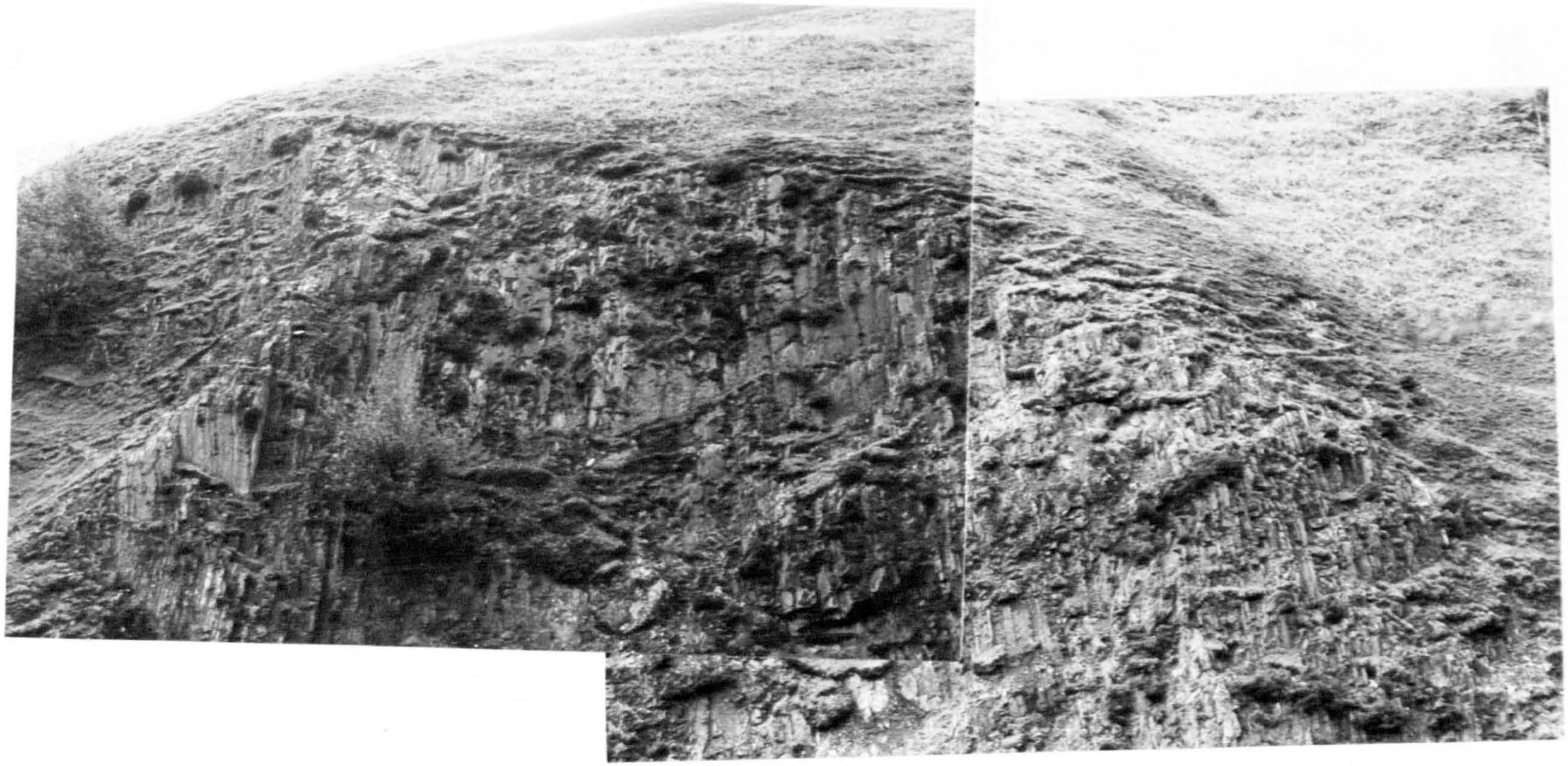


FIG:49 THINNING-UPWARD SEQUENCE, ASHES HOLLOW, SO 43559291

ESE

WNW

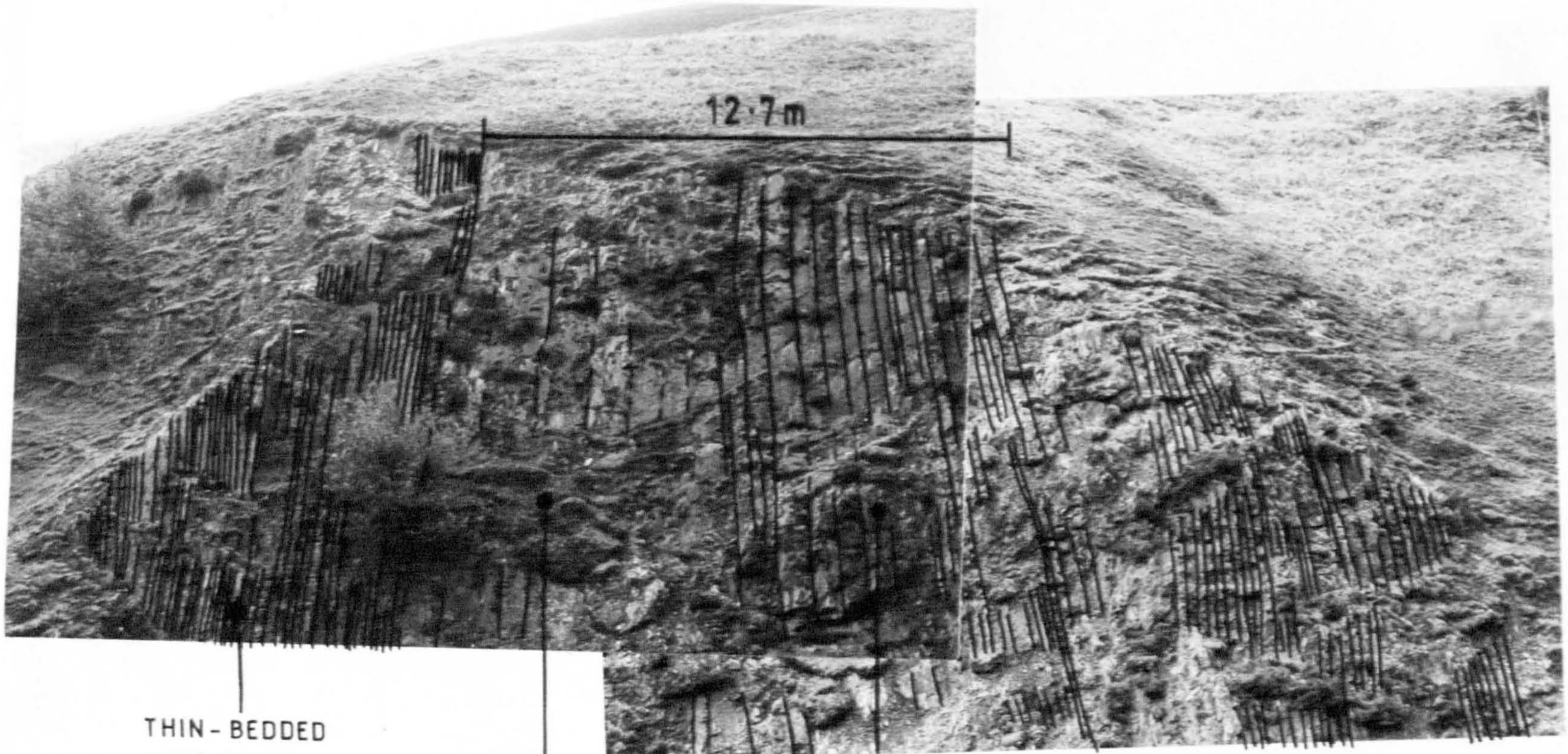
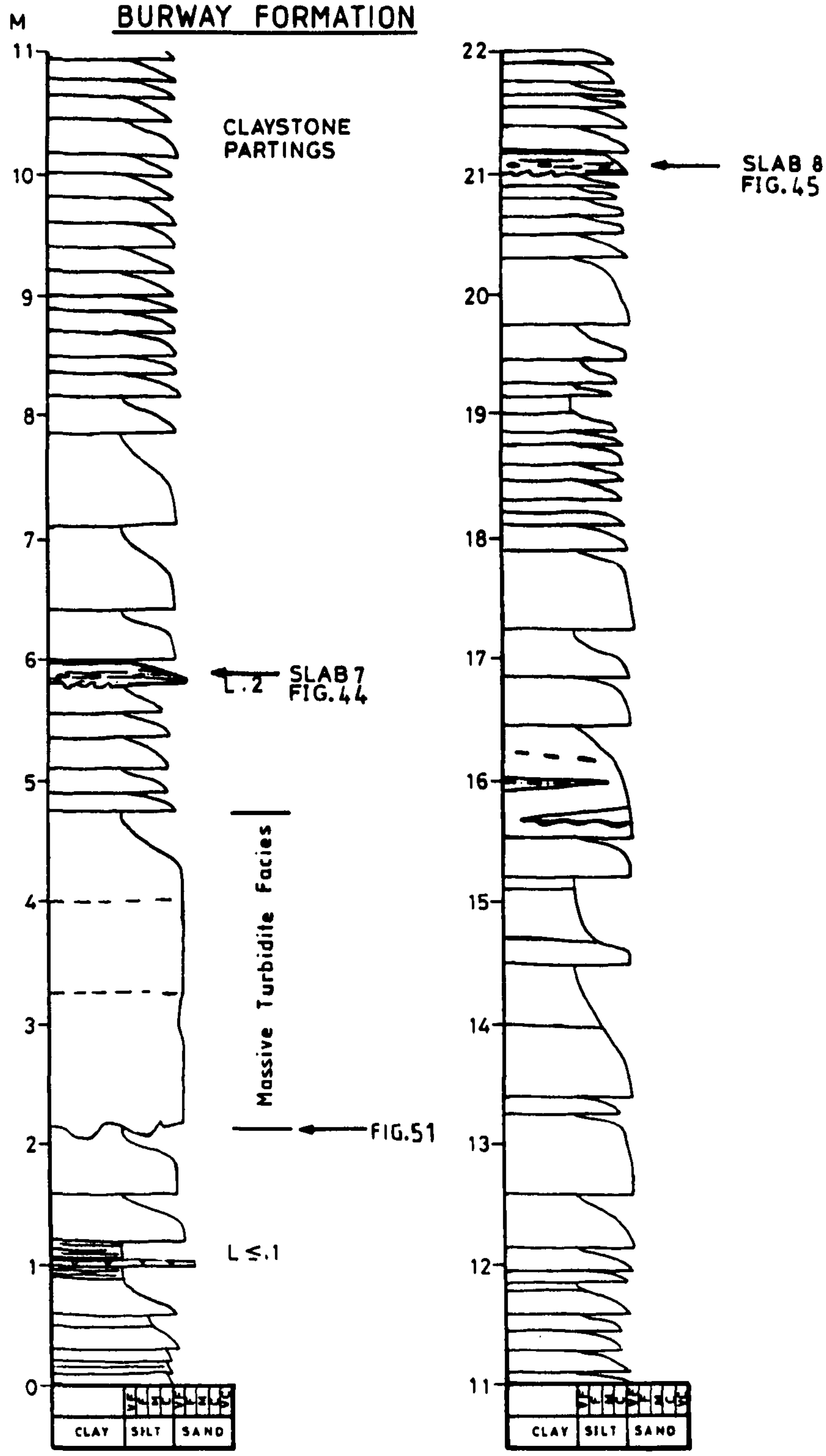
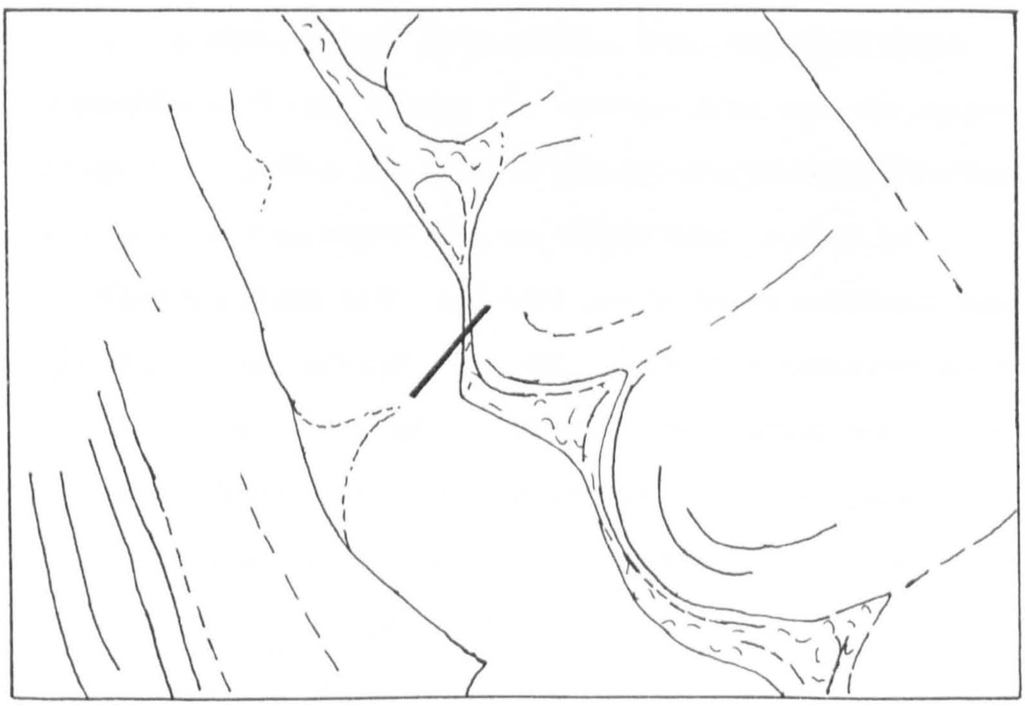


FIG:50 MASSIVE TURBIDITE & THICK-BEDDED TURBIDITE FACIES
LOG 2 ASHES HOLLOW, GRID REF, 43669285, SO
BURWAY FORMATION



This log shows the massive turbidite facies interbedded with the thick-bedded turbidite facies. Note the loaded base to the massive sandstone ,which is illustrated in fig. 51 and its homogeneous character. The top of the massive sandstone is overlain abruptly by the thick-bedded turbidite facies. Note the lensoid nature of the bedding at c.16m, within one bed, which may be due to sets of ripple cross-lamination. Note a thickening and coarsening upward sequence ,1m thick, at c.20m. Two cut and polished specimens of the thick-bedded turbidite facies (slabs 7 and 8) are shown in figs. 44 and 45 .

FIG: 51 PHOTO AND SKETCH SHOWING LOAD STRUCTURES AT THE BASE OF THE MASSIVE TURBIDITE FACIES ASHES HOLLOW, SO 43669285



Scale bar (centre) is 30 cm long

such thinning-upwards sequence is shown in fig. 49. A log through this facies and the associated thick-bedded turbidites is shown in fig. 50, log 2.

Several features are characteristic of this facies and enable it to be distinguished from the thick-bedded turbidite facies. Laminae of unit D are usually not present and the lithology mainly consists of very fine grained sandstone which is referable to unit A (see section 6.5). In places, indistinct bedding planes can be recognised (e.g. fig. 49), which suggests that this facies could have been formed by an amalgamation of thick beds. In some instances, bedding planes can be recognised, which are at a low angle to the overlying, thick-bedded turbidites. Apart from these bedding structures, the sandstone appears to be structureless. In some instances, large loads can be recognised at the base of the massive sandstones (fig. 51).

This facies occurs in four, very thick beds at the top of the sequence of thick-bedded turbidites. These beds are often interbedded with thin-bedded turbidites, which were discussed in section 6.5. Within the beds of the massive turbidite facies, some indistinct bedding planes can be recognised. A plot of bed-thickness shows that the thickness of these component beds increases up the section (fig. 36). This is accompanied by a thinning of the associated interbeds. The topmost bed of the massive turbidite facies is overlain by a thinning-upward sequence of turbidites which is then overlain by the mudstone and cross-laminated sandstone facies. No erosional features have been seen at the bases of the massive turbidite facies.

Interpretation

The recognition of bedding planes, in some instances, suggests that this facies was deposited by a number of events. The upward-fining sequence which occurs at the top of the massive sandstone shown in log 2 (fig. 50) shows that mudstone was capable of being deposited by these events. Therefore, it is probable that amalgamation occurred by the erosion of any mudstone deposited by the previous event. Alternatively, amalgamation could have occurred by a number of closely spaced current pulses carrying sand, which flushed any mud away from the site of deposition.

This facies is similar to unit A of both the thick-bedded and the thin-bedded turbidite facies. The lack of sedimentary structures, such as ripple cross-lamination and planar, parallel lamination suggests that this facies was deposited rapidly by settling from suspension. This is characteristic of liquefied flows and fluidised flows (Lowe, 1982). This facies is similar to division A of the classic Bouma sequence (e.g. Walker, 1984, p.173). It may also be compared with facies A1 of Ricci-Lucchi (1984). This is similarly texturally uniform, graded or crudely laminated and commonly has load structures at the base. It is interpreted to have been deposited by high-density turbidity currents or by liquefied flows or grainflows (Ricci-Lucchi, 1984). Facies A1 of Ricci-Lucchi (1984) is coarser than the massive turbidite facies of the Longmyndian, however, the lack of coarse grain sizes in the Longmyndian facies reflects the maximum grain size available at the source (the fluviatile Cardingmill Grit) and this is only very fine to fine grained sand.

6.7 THE TURBIDITE FACIES ASSOCIATION

This is comprised of the thinly laminated mudstone facies, the thin-bedded turbidite facies, the thick-bedded turbidite facies and the massive turbidite facies. These facies roughly occur in this stratigraphic order (fig. 36). This sequence shows an overall trend of increasing bed-thickness together with an increase in the average grain size. This latter trend reflects the increasing percentage of sandstone, since there is little change in the average, maximum grain size. The lack of significant change in the maximum grain size may be due to the lack of sediment, coarser than fine sand, available at the source. The source is thought to have been the Cardingmill Grit delta, the fluvial deposits of which are wholly fine-grained. This overall trend of increasing bed-thickness, together with an increase in average grain size and sandstone percentage, is similar to the hypothetical stratigraphic sequence developed during turbidite fan progradation (Walker, 1978, fig. 14, p.948).

The massive turbidite facies is interpreted to have been deposited in channels. This facies displays thinning-upwards sequences which are of the order of 13m thick. These channels must therefore have been of this order of depth. The occurrence of these channels at the top of the facies association is in agreement with the progradational fan model of Walker (1978, fig. 14, p.948). However, the sizes of the channels in the Longmyndian are significantly different from those of some feeder channels noted by Walker (1978, p.951), which are usually hundreds of metres deep. Shanmugam, Damuth and Moiola (1985) suggest that some of the channels observed in ancient fan exposures may actually be parts of much larger channel-levée systems. From the limited and dissected exposure of the Longmyndian, it is not possible to confirm if this hypothesis is applicable in this case.

The thick-bedded turbidites lack thinning-upwards trends and do not appear to have been deposited in channels. The thick-bedded turbidites are mostly invariant in bed-thickness and there is a lack of major thickening-upwards cycles. This suggests that the thick-bedded turbidites were not deposited on supra-fan lobes. Deposition in the mid-fan environment of the Walker fan model (1978) is dominated by thickening and coarsening-upward trends due to supra-fan lobe progradation. The Longmyndian turbidite system is therefore not similar to the Walker fan model (1978) in this respect. The thickening-upward sequences which do occur in the thick-bedded turbidite facies are rare and are only 1m to 1.8m thick. These are much thinner than the typical thickening-upwards cycles, formed by supra-fan lobes, which are usually 2m to 70m thick (Ricci-Lucchi, 1984). Alternatively, these sequences may be interpreted as compensation cycles (Mutti and Sonnino, 1981). These cycles are of a similar thickness to the Longmyndian sequences (usually being 0.6m to 6.5m thick) and are composed of a similar number of beds (usually up to 8), (Mutti and Sonnino, 1981).

The thick-bedded turbidite facies might have been deposited as widespread sheets rather than on supra-fan lobes. Such sheets may have resulted from sand being supplied by numerous, small channels rather than by a single, large channel, as postulated in the Walker fan model (1978). This mode of deposition is similar to that postulated by Heller and Dickinson (1985). They proposed that turbidite systems which are fed by multiple, shallow gullies in a prodelta slope, deposit turbidites which are characterised by random patterns of bed-thickness. In this turbidite system there are a lack of supra-fan lobes. The Longmyndian turbidite system is therefore comparable to the "submarine ramp facies model" of Heller and Dickinson (1985) in this respect. The similarities between the submarine ramp facies model of Heller and Dickinson (1985) and the Longmyndian turbidite system are that both systems are fed by a delta which has prograded to the shelf-slope break, both apparently lack a dominant feeder channel (however, it is possible that in the Longmyndian a major feeder channel might have existed outside of the area represented by the exposure), both show random patterns in bed thickness and lack predominant, asymmetric cycles and both are overlain by an abbreviated section of mud-rich slope deposits.

The packets of non-amalgamated, thick-bedded turbidites in the thick-bedded turbidite facies are of the same order of thickness as the beds of the massive turbidite facies and the bed-thicknesses in these packets are often similar to those in the massive turbidites. However, there is a lack of thinning-upward sequences associated with these packets. Their interpretation as channel-fills is therefore unlikely. It is thought that the unusually thick beds represent thicker turbidity currents which carried greater volumes of sediment. Such currents might have been generated as a result of increased seismic or volcanic activity in the magmatic arc source.

The Longmyndian turbidite facies association is distinct from the Walker fan model (1978) in that there are a lack of coarsening-up and thickening-up trends in the thin-bedded turbidite facies (fig. 36). Walker (1978) proposed that the progradation of the lower fan onto the basin plain would result in the formation of a coarsening-up and thickening-up sequence, with a gradual transition from thin-bedded turbidites of the lower fan to thicker-bedded turbidites of the mid-fan. However, in the Longmyndian turbidites, there is a clear distinction between the thin-bedded turbidite facies and the thick-bedded turbidite facies and there is a lack of gradation between them. The thick-bedded turbidite facies rests abruptly on the thin-bedded turbidite facies (fig. 36). This abrupt change suggests that there was a lack of gradual progradation from "distal" to more "proximal" parts of the turbidite system. This abrupt change might have been due to a rapid lateral shift in the source and could be the result of channel avulsion in the Cardingmill Grit delta.

No palaeocurrent data were obtained from the turbidite facies. However, since it is thought that the turbidites were sourced by the Cardingmill Grit delta, then they are likely to have been deposited by ENE directed currents, since this is the average palaeocurrent direction displayed by the Cardingmill Grit delta deposits. This direction is at an acute angle to the limit of the Uriconian Volcanic Complex, which is defined by the Church Stretton fault system. The proximity of the Uriconian Volcanic Complex which, on lithological grounds, is thought to have been the source for the Longmyndian sediments, is not suggested by the palaeocurrent data or by the lack of more proximal deposits in the turbidites. It is thus likely that the Uriconian Volcanic Complex has been tectonically juxtaposed with the deposits of this turbidite system. This is extensively discussed in the preceding chapters.

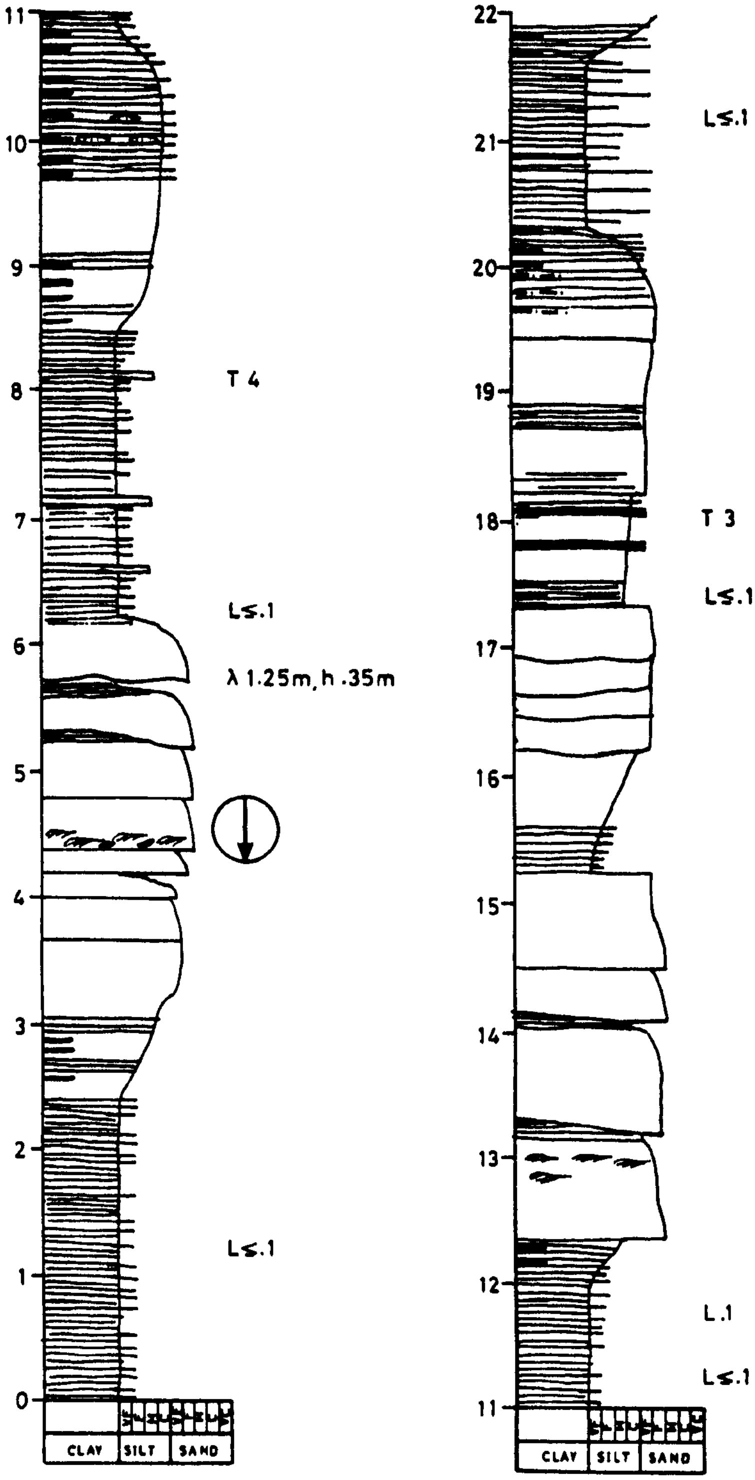
6.8 THE MUDSTONE AND CROSS-LAMINATED SANDSTONE FACIES

This facies occurs in the Burway Formation, directly above the interbedded massive turbidite and thick-bedded turbidite facies, with which it has a rapid, transitional contact. It is abruptly overlain by the thick-bedded, cross-stratified sandstone facies of the Cardingmill Grit Member of the Burway Formation (fig. 36). This facies is illustrated by log 6, fig. 52 and log 7, fig. 53.

Cross-laminated sandstone

This occurs in beds which are mostly 4cm to 70cm thick, as individual beds or in amalgamated units 3m to 7.5m thick (fig. 36). These sandstones are cross-laminated and planar, parallel laminated. The cross-lamination usually occurs in sets up to 3cm thick, but it can occur in sets up to c.10cm thick. The cross-lamination is usually of the trough type, is often climbing and indicates a unidirectional palaeocurrent which is consistently towards the NE (fig. 54). In some instances, ripple form-sets are preserved at the tops of some of the beds by an overlying lamina of mudstone (e.g. fig. 55, slab A). The bases of these sandstone beds are abrupt and the tops can be abrupt or gradational with the laminated mudstone. Commonly, the sandstones are normally graded and are gradational with the overlying laminated mudstone, but can be non-graded or very slightly normally graded, in which case the tops are abruptly overlain by mudstone. In the normally graded beds, a sequence of sedimentary structures occurs which is similar to that developed in some turbidites (fig. 56). The bases of the beds are faintly, planar, parallel laminated and are overlain by a thick sequence of current ripple cross-laminated sandstone, which

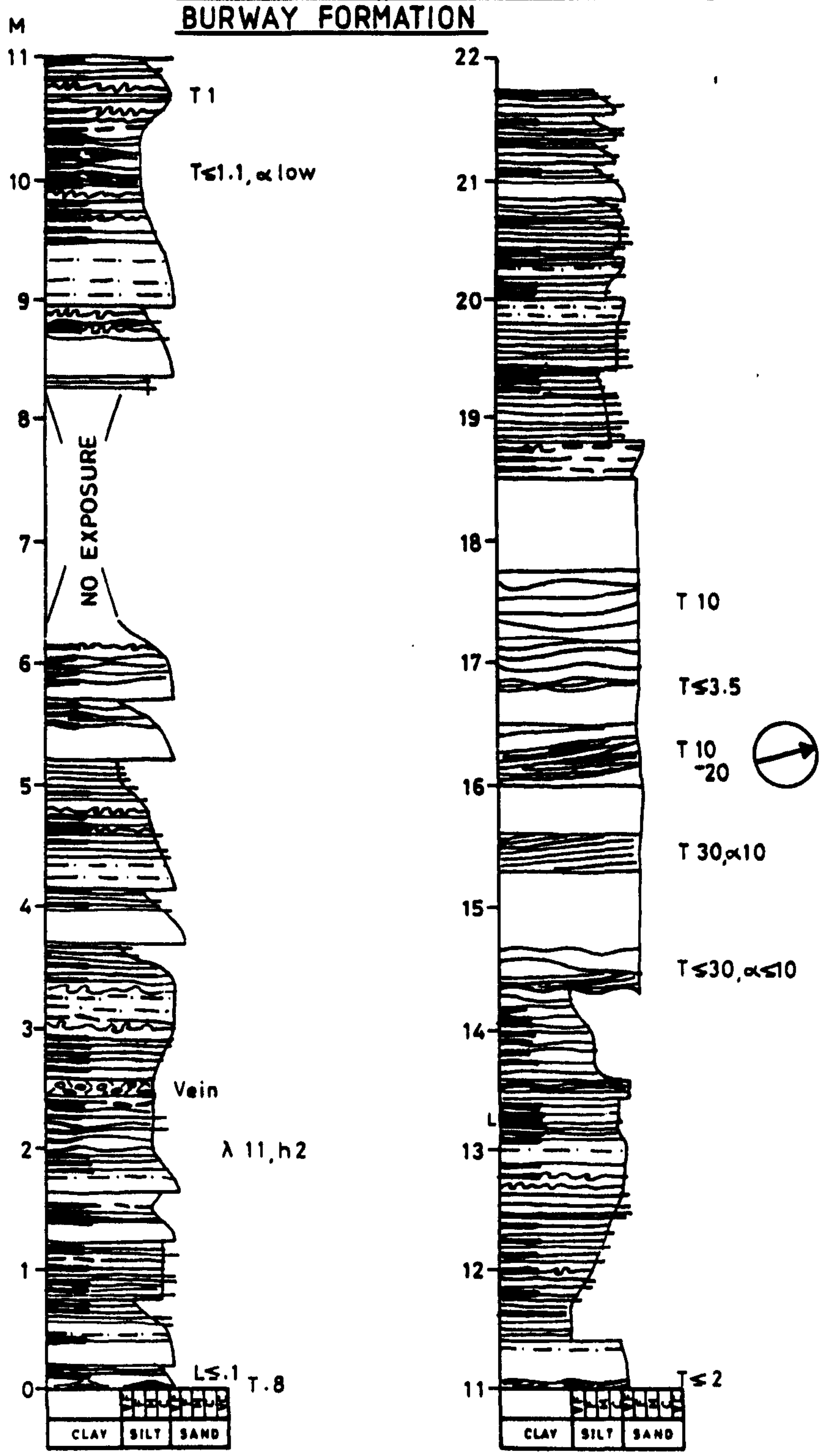
FIG.52 LOG 6 ASHES HOLLOW, GRID REF 43399302 SO
BURWAY FORMATION



This log shows the mudstone and the ripple cross-laminated sandstone facies. The mudstone is very thinly planar, parallel laminated and there are numerous ?biogenic or dewatering features on their surfaces (discussed in chapter 5). The sandstone beds are discrete or amalgamated and though apparently structureless at outcrop (and hence mostly blank on the log), some slabs show the presence of abundant ripple cross-lamination and parallel lamination.

FIG.53 LOG 7 ASHES HOLLOW, GRID REF, SO 43319303

BURWAY FORMATION



This log shows the mudstone and ripple cross-laminated sandstone facies. There are numerous, normally graded, thin sandstone beds. Note the 4.5m thick amalgamated sandstone bed with cross-bedding in sets 10cm to 30cm thick. The lamination is inclined at an apparently low angle, but the evaluation of the dip is hindered by the lack of 3-d exposure. Although the thin sandstone beds are apparently structureless on the log, slabs show ripple cross-lamination and this can be recognised in some of the beds at outcrop (e.g. at 10m and 11m on the log). Note the greater proportion of sandstone and coarse siltstone compared with the stratigraphically lower sequence shown in log 6.

FIG: 54 PALAEOCURRENT ROSE FOR THE MUDSTONE AND CROSS-LAMINATED SANDSTONE FACIES AND PHOTOGRAPH SHOWING AN AMALGAMATED UNIT OF SANDSTONE.

PHOTO; shows an amalgamated sandstone unit (left of photo), which abruptly overlies planar, parallel bedded sandstones and laminated mudstones. Scale bar (centre of photo) is at the contact between the two lithologies and is 30 cm long. Note the non-parallel beds in the amalgamated sandstone. This outcrop is shown in log 7, fig. 53 .

PALAEOCURRENT ROSE - MUDSTONE AND CROSS-LAMINATED SANDSTONE FACIES

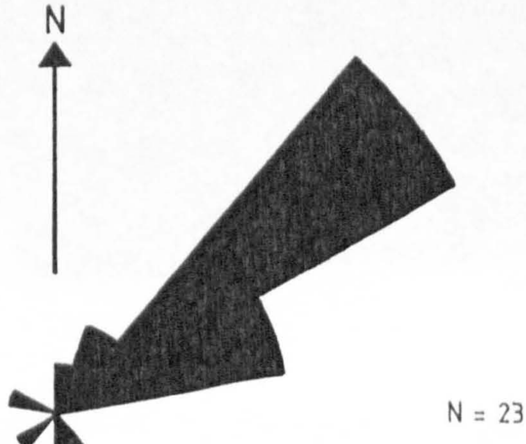
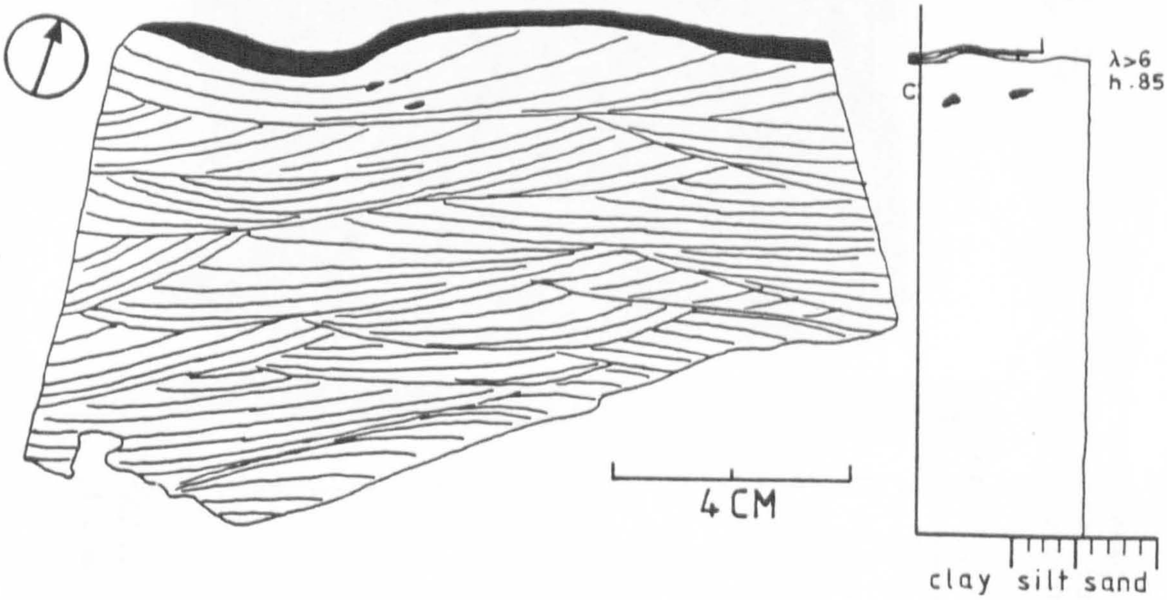
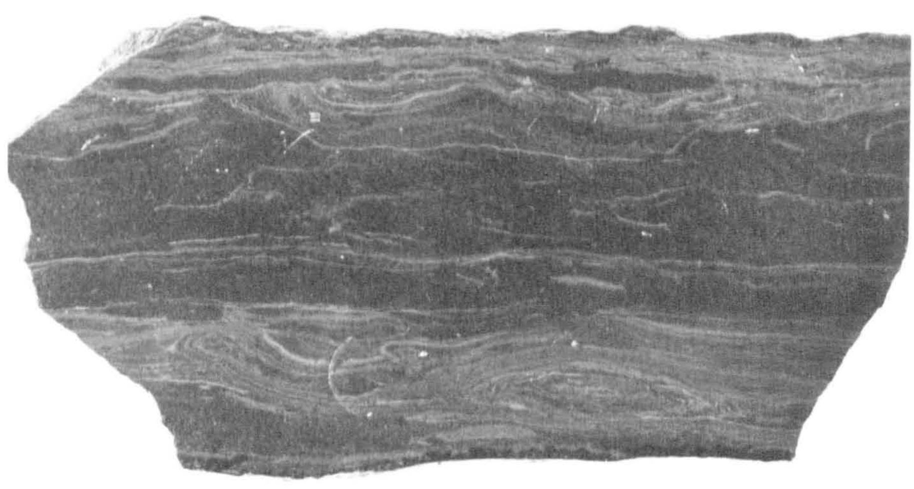


FIG:55 SLABS ILLUSTRATING SOME SEDIMENTARY STRUCTURES IN THE RIPPLE CROSS-LAMINATED SANDSTONE FROM THE TOP OF THE BURWAY FORMATION.



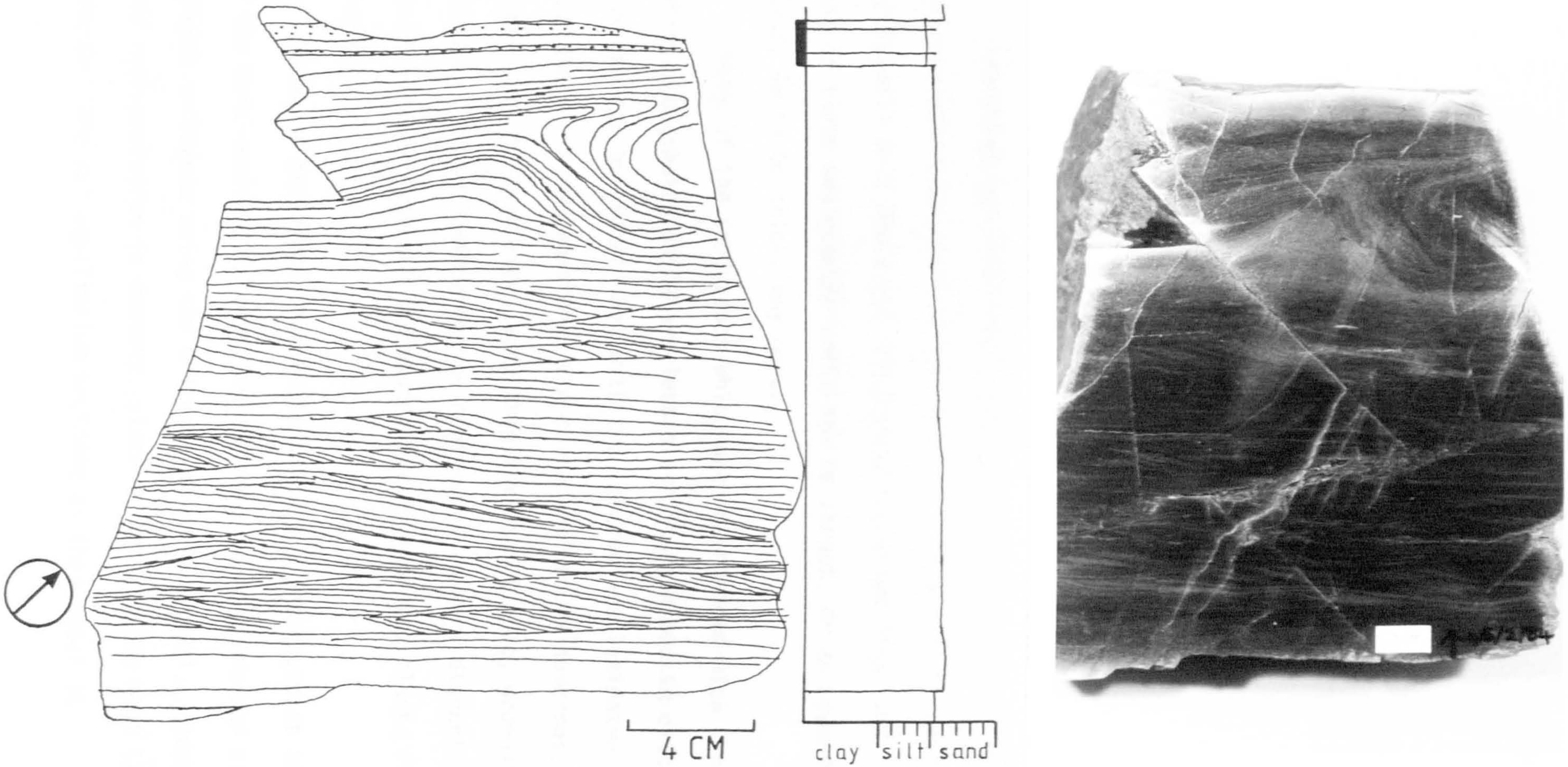
SLAB: A Note the extensive trough cross-lamination; lack of grading at the top of the bed and ripple form set preserved at the top with a claystone drape.



SLAB: B Discontinuous, subparallel and contorted laminae of mudstone in very fine grained sandstone. Note the contorted lamination at the base of the slab with overfolded laminae. Scale bar is 2cm long.

FIG:56 CROSS-LAMINATED SANDSTONE FROM THE TOP OF THE BURWAY FMN.

SLAB



Note the presence of climbing ripple cross-lamination (bottom left of slab);poorly laminated,slightly coarser base to the bed;slight normal grading;convolute lamination at the top of the sandstone unit and paler siltstone with sandstone laminae at the top.

is followed by convolute-laminated silty sandstone and finally by planar, parallel laminated siltstone. The sandstones often contain clasts of planar, parallel laminated, light greenish grey siltstone, which are similar to the interbedded siltstone.

The amalgamated sandstone units occasionally have thin interbeds of mudstone. They can be plane, parallel bedded or non-parallel bedded, giving a lenticular appearance (fig. 54, photo). The beds are from 10cm to 80cm thick and may be ripple cross-laminated or cross-bedded in sets up to 30cm thick. These amalgamated units are abruptly underlain by laminated mudstone with thin sandstone beds and they occur in the upper half of the stratigraphic sequence which is formed of this facies, beneath the Cardingmill Grit Member of the Burway Formation (fig. 36). The tops of these amalgamated units may be abrupt, or an upward-fining trend, 4m to 5m thick, may occur.

Many of the sandstones which are interbedded with these thick, amalgamated units are less clearly organised in distinct beds and can be gradational with the interbedded laminated mudstones. These sandstones are characterised by numerous, subparallel to contorted, very thin laminae of pale, greenish grey mudstone. These laminae may be continuous to very discontinuous and the latter approach the appearance of isolated clasts (fig. 55, slab B).

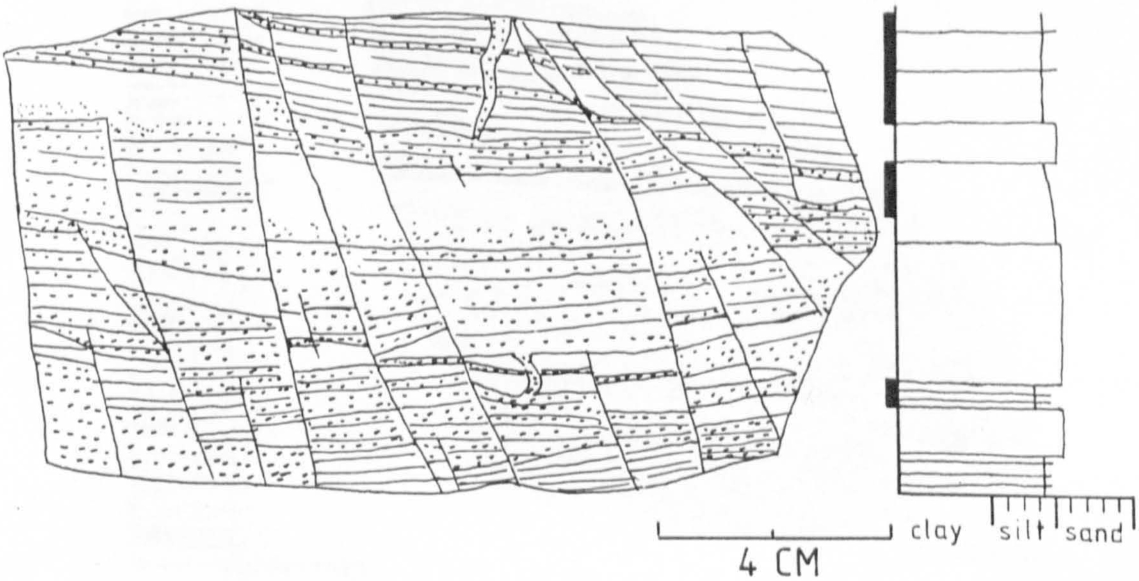
Some of the ripple form-sets which are preserved at the tops of the beds could be interpreted as being due to combined wave and current influence using the criteria of Harms (1969). However, this interpretation is dubious, since it is probable that these form-sets are not equilibrium bedforms as they occur at sandstone-shale boundaries and additionally, these criteria may not apply where there is significant net deposition of sand (Harms, 1969).

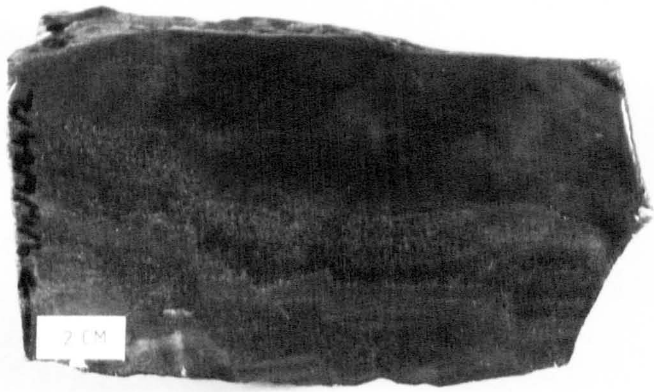
Mudstone

There are several distinctive types of mudstone present in this facies. At the base of the facies in Ashes Hollow, silt and clay couplets are developed. The resulting lamination is c.7mm and up to 10mm thick. The siltstone is planar, parallel laminated and can be non-graded or normally graded. The claystone is very thinly laminated and the laminae are discontinuous and irregular. This claystone is similar to the claystone which forms unit D in the underlying turbidite facies. Overlying this mudstone (at c.625m on fig. 36), normally graded, thin siltstone beds occur, which are on average 1.75cm thick and can be up to 5cm thick (fig. 36). This siltstone is coarse and is planar, parallel laminated and occasionally cross-laminated. Occasionally, similar beds of very fine grained sandstone occur (fig. 57). These are interbedded with very thinly to thinly, planar, parallel laminated, very fine to coarse siltstone. Some of the sandstone beds develop small injection features of sandstone into the interbedded mudstone and some microfaults are developed, which suggest that tensional deformation occurred whilst the sediment was not lithified, since there are thickness changes across the microfaults and some of the microfaults do not extend through the overlying laminae (fig. 57).

At c. 675m on fig. 36, planar, parallel laminated siltstones occur. The laminae are normally graded, four to five millimetres thick and are also very thinly internally laminated. These siltstones occasionally contain circular bedding plane features,

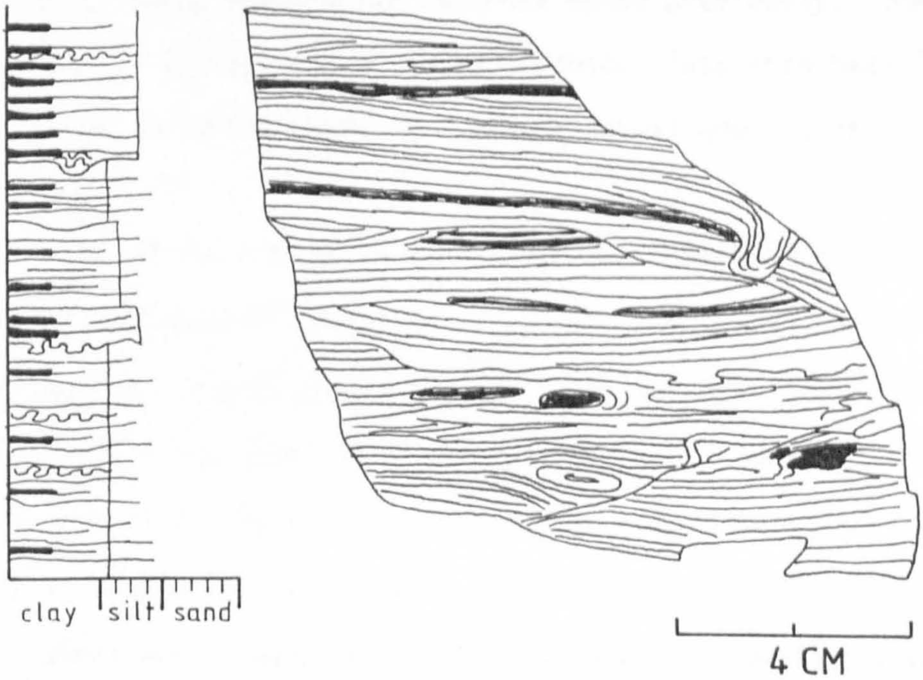
FIG:57 SLAB MUDSTONE WITH THIN SANDSTONE BEDS FROM THE MUDSTONE AND RIPPLE CROSS-LAMINATED SANDSTONE FACIES OF THE BURWAY FORMATION.

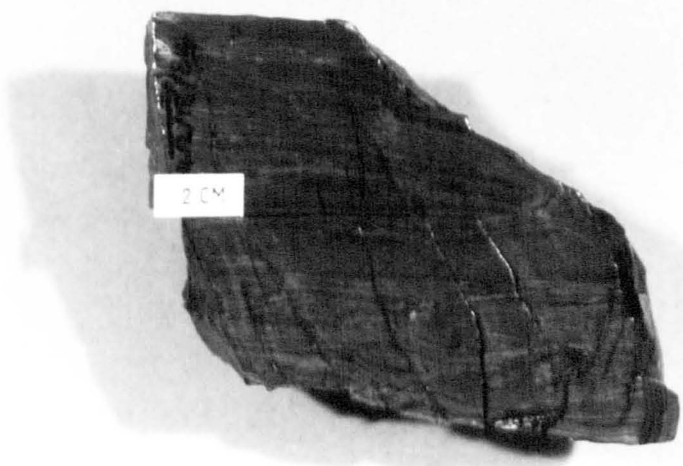




Note the numerous small microfaults, sandstone injection features and change in bed thickness accross faults (left side), suggesting soft sediment deformation The lithologies appear to be organised into a c.6cm overall fining-upward sequence (top of log). This slab is from the lower part of the mudstone and cross-laminated sandstone facies in Ashes Hollow.

FIG:58 SLAB MUDSTONE FROM THE MUDSTONE AND RIPPLE CROSS -LAMINATED SANDSTONE FACIES OF THE BURWAY FORMATION.





Note the heterolithic nature of the lamination,which is very contorted (lower part of slab) and also plane and parallel and the presence of small microfaults (bottom right corner).The laminated mudstone is organised into a c.3cm thick fining-upward sequence (middle of log).This slab is from a mudstone bed which is underlain and overlain by thick beds of amalgamated sandstone.

12mm to 16mm in diameter, which are discussed in section 5.2 and these are interpreted as dewatering features. These siltstones occasionally contain thin, normally graded beds of coarse siltstone, which are similar to those noted previously. These laminated siltstones appear to be organised into thin beds (3cm to 7cm thick), in which there is a slight normal grading of the siltstone.

Very thinly laminated siltstones are interbedded with the units of amalgamated sandstone at c.725m on fig. 36. This lithology is illustrated in fig. 25, photographs A and B, fig. 29, photographs A and B,and in fig. 58. The laminae vary rapidly in grain size from clay to very fine grained sand and are either normally graded or non-graded. They are planar, parallel laminated or slightly wavy, non-parallel laminated with pinch and swell structures and often the laminae are contorted (e.g. fig. 58). Numerous bedding plane markings occur which are discussed in chapter 5. These are classified as small ring-like features and blister-like features and the possibilities of a biogenic origin for both of these were discussed and it was considered that although a biogenic origin is possible, they could be interpreted as dewatering structures.

Siltstones occur directly beneath the Cardingmill Grit, where they are interbedded with the ripple cross-laminated sandstones as distinct beds or as parts of upward-fining sequences associated with the sandstone beds (fig. 53, log 7). They are commonly very thinly or thinly, planar, parallel laminated, but the laminae can be non-parallel, discontinuous and occasionally contorted. The lithologies of the laminae are very variable and sandstone, siltstone and claystone are often interlaminated. The silty claystone usually shows a very discontinuous, subparallel, slightly wavy lamination, which is similar to unit D in the underlying turbidite facies.

Trends throughout the facies

The base of the facies in Ashes Hollow is transitional with the underlying thick-bedded turbidite facies, with the development of a rapidly thinning- and fining-upwards sequence, which ends in laminated mudstones. This is abruptly overlain by laminated siltstones which contain thin, planar laminated and ripple cross-laminated sandstone and siltstone beds. These are abruptly overlain by an amalgamated sandstone unit which is 4.5m thick. This sequence may be interpreted as a thickening and coarsening-upward sequence. The sequence overlying the amalgamated sandstone is a thinning- and fining-upward one, 30m thick, which ends in laminated mudstone. The amalgamated sandstone beds which abruptly overlie this, display both an increase in the thickness of the amalgamated units (from 3m to 7.5m) and an increase in the thickness of the component beds (fig. 36). The overlying sequence may be interpreted as an overall thinning- and fining-upward sequence, which is punctuated by occasional, thick, amalgamated sandstones. This sequence is abruptly overlain by the thick-bedded, cross-stratified sandstone of the Cardingmill Grit.

Throughout the entire facies, an overall coarsening-upward trend is apparent, beginning with mudstones with thin sand beds at the base, followed by coarser siltstones, with thick amalgamated sandstone units and ending in the thick-bedded, cross-stratified sandstone facies at the top (fig. 36).

Interpretation

The thin sandstone and siltstone beds contain extensive ripple cross-lamination and were therefore mainly deposited by traction. High sedimentation rates are evident in many cases, since climbing ripple cross-lamination is common. These currents were short-lived, since these sand beds occur in a dominantly mudstone sequence. The sequence of sedimentary structures shown by the thin sandstone beds (e.g. fig. 56) indicates that the currents which deposited these beds waned in velocity. This is also suggested by the gradual upward decrease in grain size. However, many beds show little or no grading and are abruptly overlain by mudstone, which suggests that the currents rapidly decelerated in some cases. The presence of mudstone clasts in some of the beds indicates that the generating currents were often turbulent and of high velocity.

The amalgamated sandstone units display evidence for extensive traction, with the development of cross-bedding in some of the upper sandstone units and more extensive planar, parallel lamination and ripple cross-lamination in the lower sandstone units. These units imply that currents were continuous for longer periods than for the thin sandstone beds. Alternatively, some of the thick sandstone units (e.g. fig. 52, log 6) appear to have been formed by the stacking of a number of closely spaced, separate depositional events, since laminated siltstones separate some of the upward-fining sandstones.

Current directions from both the thin sandstone beds and the amalgamated sandstone units are unidirectional and consistently towards the NE (figs. 36 and 37). This direction is similar to that in the overlying fluviatile, thick-bedded, cross-stratified sandstone facies. Since the currents are unidirectional and since there is a lack of flaser bedding and reactivation surfaces, it is concluded that these sandstones were probably deposited by fluvially generated currents with a lack of tidal influence. Although there are some lenticular beds and some low angle cross-stratification, there is no evidence for structures which could be referred to as hummocky cross-stratification (e.g. Dott and Bourgeois, 1982). Additionally there are a lack of wave-generated structures and variable palaeocurrent directions which might indicate wave influence. These sandstones were therefore deposited by fluvially generated bottom currents in otherwise quiet water.

The laminated mudstones were probably deposited by a number of processes. Some of the coarse siltstones are ripple cross-laminated and were therefore deposited by traction. However, many of the finer siltstones and claystones were probably deposited by settling from suspension. The discontinuously, very thinly laminated claystones, which are similar to unit D in the underlying turbidite facies, may be similarly interpreted as representing background sedimentation by slow settling from suspension. The siltstones might have been deposited from turbid, surface layers introduced by the same fluvial currents which deposited the cross-laminated sandstones. Some of the laminated siltstones are organised into graded units 3cm to 7cm thick. A similar, normally graded, laminated mudstone bed is shown in fig. 58. These graded, laminated beds suggest that deposition occurred from a single waning current. Similar beds are described by Stow and Bowen (1980) and Piper (1972), who refer to graded, laminated beds. Stow and Bowen (1980) proposed that these beds were deposited from a single current by a process of depositional sorting in the boundary layer of a "turbidity current", which resulted in an interlaminated siltstone and claystone bed. The normally graded, laminated mudstone beds in the Longmyndian could have been deposited by a similar process. The depositional currents were referred to as slow, dilute flows, with velocities of 9-16cm.s^{-1}, by Stow and Bowen (1980). These flows might have been generated by fluvial processes rather than being turbidity currents.

The contorted lamination, which is commonly found in the mudstones which are interbedded with the thick, amalgamated sandstone units (e.g. fig. 58), might have been caused by dewatering processes initiated by the loading of the amalgamated sandstone onto the mudstone. However, some mudstones with disturbed lamination from lower in the sequence are not directly overlain by thick, amalgamated sandstone units. One slab (fig. 57) has numerous microfaults which appear to have formed whilst the sediment was not lithified, since there is a change in bed thickness across some of the faults and many of the faults do not pass through all of the beds. Additionally, these microfaults lack any mineralisation, which is usually associated with the late fractures, and there are associated sandstone injection features. They are all downthrown on the same side and are all normal faults. These faults might have been formed by the extension of unlithified sediment on an inclined surface, possibly during the development of a slide. In addition, some of the mudstones which are interbedded with the thick, amalgamated sandstone units (e.g. fig. 58) possess intensely swirled and overfolded, convolute laminae which may be indicative of horizontal movement, which possibly occurred during sliding.

The interpretation of the large-scale trends in this facies and the environments of deposition are considered together with the thick-bedded and cross-stratified sandstone facies during the discussion on the subaqueous-delta facies association (section 6.10).

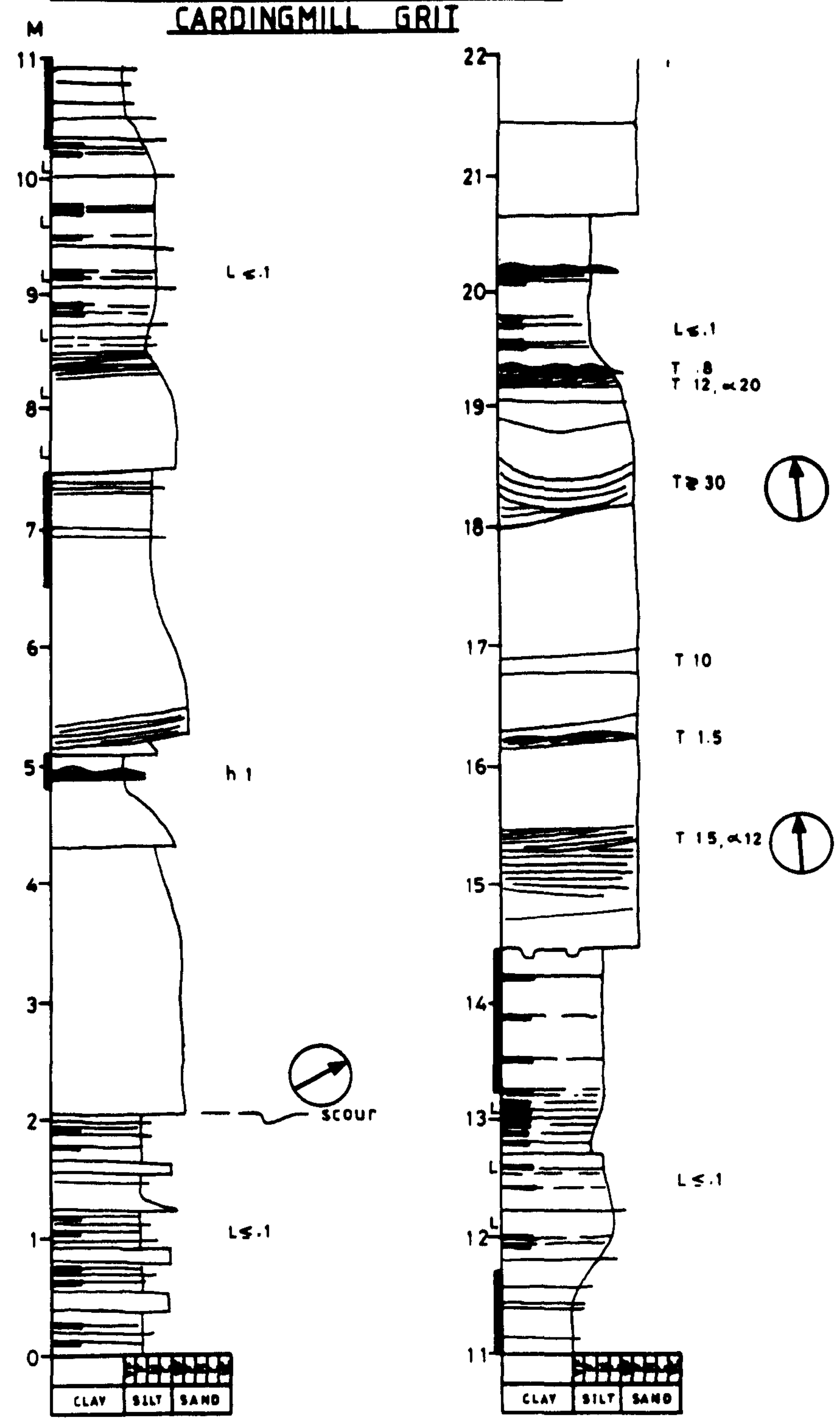
6.9 THE THICK-BEDDED AND CROSS-STRATIFIED SANDSTONE FACIES

The Cardingmill Grit Member of the Burway Formation, which is up to c.50m thick, is composed of this facies. The base of this member rests abruptly on the mudstone and cross-laminated sandstone facies of the Burway Formation and is overlain by the green mudstone and cross-laminated sandstone facies of the Green Synalds Member of the Synalds Formation. This facies is illustrated by log 9, fig. 59 and log 12, fig. 60.

The base of the Cardingmill Grit Member, although abrupt, cannot be demonstrated to be erosive or channelised. The base can contain large clasts of mudstone which are usually tabular in shape. These are purplish grey in colour and were therefore not derived from the underlying Burway Formation. Similar mudstones are found in the overlying Synalds Formation and are also interbedded with the sandstones of this facies.

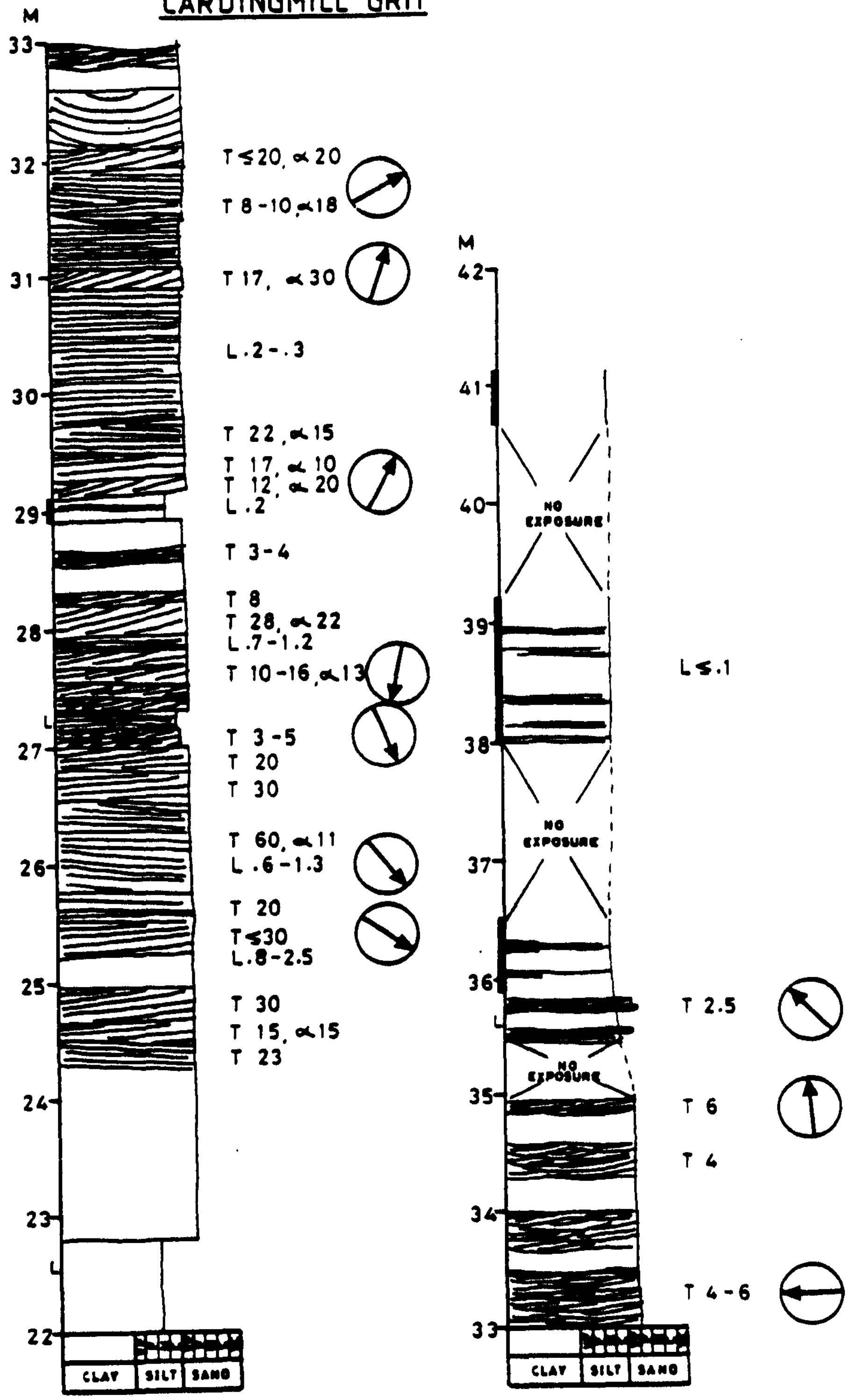
This facies is characterised by its uniform, fine grain size, homogeneous lithology and the presence of cross-bedding of various scales. The most common type of cross-bedding consists of sets which are usually less than 30cm thick and which are rarely up to 60cm thick (fig. 61, photo A). The type of cross-bedding is difficult to assess from the limited exposure, but appears to be trough cross-bedding in some instances. The bases to these sets are usually tangential and the observed angle of the

FIG.59 LOG 9 ASHES HOLLOW, SO 43279314
CARDINGMILL GRIT



This log illustrates the sediments of the thick-bedded and cross-stratified sandstone facies. Note the four upward fining sandstone sequences, each with an abrupt, occasionally loaded or scoured base, followed by cross-bedded, fine grained sandstone. This is overlain by planar laminated or ripple cross-laminated silty sandstone and thinly to very thinly laminated mudstone. Note that the mudstone is often purplish-grey (thick line along the log margin indicates red sediment). Thin sand beds (10cm-20cm thick) are interbedded with the mudstone at the base.

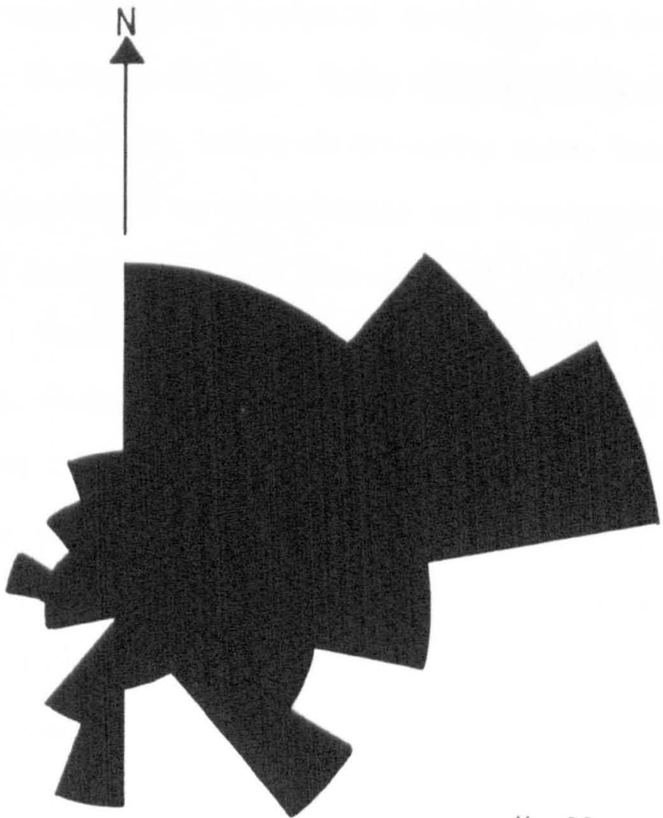
FIG.60 LOG 12 (cont'd) MOUNT GUTTER, SO 40308853
CARDINGMILL GRIT



This log illustrates the thick-bedded and cross-stratified sandstone facies. Note the lack of upward fining sandstone bodies. Cross-bedding and planar, parallel lamination are common. Note the thin interbeds of mudstone and the upward fining transition at the top of the log from the sandstone of the Cardingmill Grit to the mudstones of the Green Synalds Member. The mudstone is commonly red or purplish grey (the thick line along the margin of the log indicates red sediment). Note the different palaeocurrent directions for the two units of sandstone, separated by a mudstone bed and for the overlying mudstones.

FIG:61 PALAEOCURRENT ROSE FOR THE CARDINGMILL GRIT AND PHOTOGRAPH SHOWING A CROSS-BED SET FROM THE THICK-BEDDED AND CROSS-STRATIFIED SANDSTONE FACIES.

PHOTO;showing a cross-bed set which is typical of the thick-bedded and cross-stratified sandstone facies.Scale bar is 30cm long and is graduated in centimetres and decimetres.



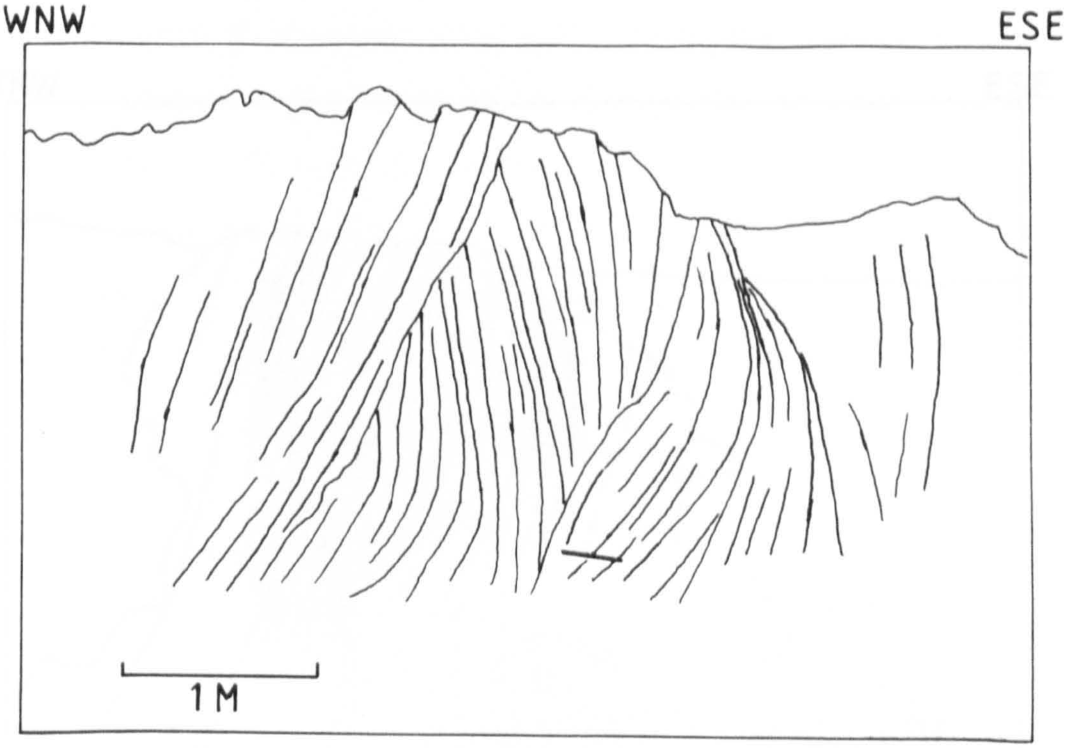
PALAEOCURRENT ROSE FOR THE CARDINGMILL GRIT MEMBER OF THE BURWAY FORMATION.

cross-lamination is usually low and in the region of 15 degrees, but can be up to c.30 degrees. In addition to this type of cross-bedding, larger scale cross-bedding occasionally occurs (figs. 62 and 63). The set shown in fig. 62, for example, is 1.2m thick.

The majority of the sandstone appears to be structureless, except for poorly developed, subparallel bedding planes. Their spacing is similar to the spacing of the cross-bedding set planes and this, together with the presence of a faint lamination in places, suggests that this sandstone may be cross-bedded. In addition to these structures, ripple cross-lamination and planar, parallel lamination is occasionally present. The ripple cross-lamination often occurs as single sets between sets of cross-bedding.

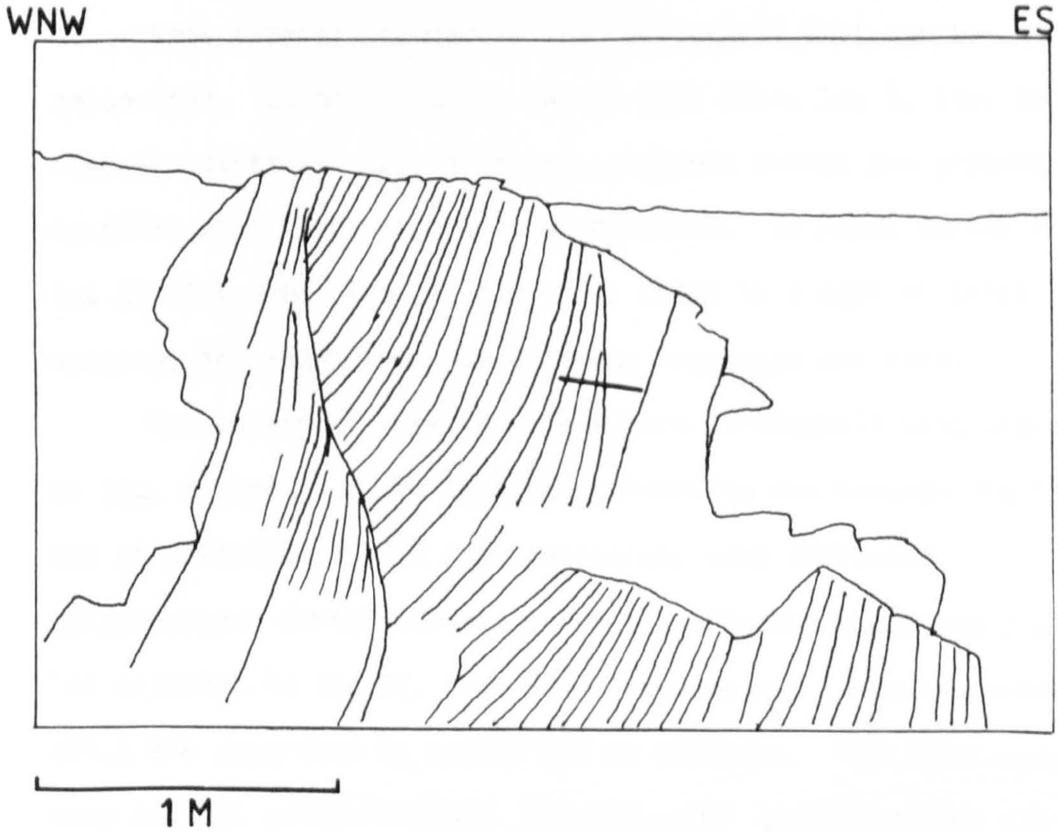
The majority of the sandstone shows little change in grain size, but some upward-fining bodies of sandstone are present (e.g. fig. 59, log 9, 14.5m to 20.5m). These upward-fining sequences begin with an abrupt base, which occasionally shows load and scour features. The sandstone is cross-bedded and fine-grained at the base and passes upward into very fine grained sandstone, which is planar, parallel laminated or ripple cross-laminated. In some instances, thick interbeds of mudstone are developed at the top of the upward-fining sequences. These are purplish grey and greenish grey and massive or thinly, planar, parallel laminated. They occasionally contain thin sand beds (10cm to 20cm thick) that have abrupt, often loaded, bases, concentrations of shale clasts at the base and which commonly fine upwards. Thin beds of mudstone may also occur within the sandstone. The upward-fining sequences are 2.2m to 6.5m thick and are, on average, 4m thick. The top of the

FIG: 62 LARGE SCALE CROSS-BEDDING IN THE THICK-BEDDED, AND CROSS-STRATIFIED SANDSTONE FACIES AT DEVIL'S MOUTH (SO 4404 9421).



Note the large set 1.2m thick, underlain by sandstone with large, curved basal surfaces and overlain by subparallel, even bedded sandstone. Scale bar in photograph (bottom centre) is 30cm long.

FIG: 63 LARGE SCALE CROSS-BEDDING IN THE THICK-BEDDED AND CROSS-STRATIFIED SANDSTONE FACIES AT PACKETSTONE HILL (SO 4215 9140).



Cardingmill Grit in Mount Gutter (at SO 4030 8853; log 12, fig. 60) shows the upward-fining transition from the Cardingmill Grit sandstone to the thinly laminated, purplish grey siltstones of the basal parts of the Green Synalds Member of the Synalds Formation.

Some lateral changes in the Cardingmill Grit can be recognised. In Ashes Hollow (at SO 4327 9314; log 9, fig. 59) several, distinct, upward-fining sandstone bodies are present, together with thick interbeds of mudstone. In Mount Gutter however (at SO 4030 8853; log 12, fig. 60), there is a lack of thick mudstone interbeds and upward-fining sequences are rare.

Palaeocurrent data for the entire Cardingmill Grit are shown in fig. 61 and indicate that the palaeoflow was towards the ENE and was unidirectional. In some instances, very different palaeocurrent directions are shown by different sandstone bodies. For example, in log 12, fig. 60, there are two sandstone bodies which are separated by a thin bed of mudstone. The lower sandstone body has SSE palaeocurrents, but the upper sandstone body has NNE palaeocurrents. The overlying mudstone has a different palaeocurrent direction to either of the sandstone bodies and this is towards the WNW.

Interpretation

The Cardingmill Grit occurs above a sequence of turbidites and below a sequence with evidence for subaerial exposure. This suggests that the Cardingmill Grit could have been deposited in a shallow marine or shoreline environment. The unidirectional palaeocurrents and the lack of reactivation surfaces and other structures indicative of fluctuating and periodically alternating high and low energy conditions implies a lack of tidal influence.

Similarly, there is a lack of wave-produced sedimentary structures and a lack of hummocky cross-stratification which are often associated with wave and storm dominated coastlines.

The presence of upward-fining cycles and unidirectional palaeocurrents suggests that the Cardingmill Grit might have been deposited in fluvial channels. The depths of these channels can be estimated from the thicknesses of the upward-fining cycles and therefore they must have been of the order of 2m to 6m deep. It is possible that these channels were meandering, however, no epsilon cross-bedding has been recognised in the Cardingmill Grit. The occurrence of large scale cross-bedding allows for an estimate of channel depth, which must therefore have been at least 1.2m deep in some instances. These large scale cross-bedding sets might have been deposited as bars.

The associated mudstones are interpreted as vertical accretion deposits. In one instance, the palaeocurrent directions in the mudstones are clearly different from those of the associated channel sandstones (fig. 60, log 12), which suggests that these mudstones were not deposited in the channel but by overbank flooding. The thin (10cm to 20cm thick) sand beds which are associated with the mudstones are interpreted as overbank flood deposits and possibly they are thin crevasse splay deposits.

The associated mudstones are often purplish red or purplish grey and are therefore different from those lower down in the succession, which are greenish grey. It is later argued that reddish colours in the Longmyndian are due to subaerial exposure (section 6.24). Therefore, some of the vertical accretion deposits which were formed by overbank flooding might have been subaerially exposed.

The lack or paucity of mudstones in many of the exposures indicates that the Cardingmill Grit was built by the amalgamation or stacking of a number of channel sandstones, together with the erosion of the associated mudstones. This erosion might have provided the purplish grey clasts which are found in the base of the Cardingmill Grit and occasionally within it. The variability in the percentage of mudstone in different vertical sections may be explained by the variability of channel erosion and stacking.

6.10 THE SUBAQUEOUS DELTA FACIES ASSOCIATION

This facies association consists of the mudstone and cross-laminated sandstone facies and the thick-bedded, cross-stratified sandstone facies. The term "subaqueous delta" is here used to indicate that the environment of deposition was below sea level and as such includes the prodelta, delta-slope and delta-front sedimentary environments, together with distributary channel deposits, where these are not associated with abundant fine grained delta-plain deposits.

The primary criterion for the interpretation of these facies as possibly deltaic, is that they form a thick, coarser clastic deposit which passes downward into deeper water facies and upward into subaerial deposits. In addition, these facies are organised into an overall upward-coarsening sequence which may be interpreted in terms of delta-progradation. Also, there are a lack of sedimentary structures which are typical of tidal and wave influenced deposits and there is a predominance of upward-fining cycles in the thick-bedded, cross-stratified sandstone facies, which may be interpreted as fluvial channel fills. A fluvial dominated interpretation is supported by the unidirectional palaeocurrents, which are displayed both by the thick-bedded and cross-stratified sandstone facies and the mudstone and ripple cross-laminated sandstone facies.

The palaeocurrents which are displayed by the ripple cross-laminated sandstones in the mudstone and ripple cross-laminated sandstone facies are identical in direction to those displayed by the thick-bedded and cross-stratified sandstone facies. This suggests that the currents which deposited the former were fluvial in origin. Such currents might have been generated during flood periods and were probably hyperpycnal flows (Wright, 1977). Similar, ripple cross-laminated sandstone beds are described from the delta-slope environment of the Roaches Grit by Jones (1980). These are attributed to river-generated density currents (Jones, 1980). These density-flows might have been driven by contrasts in temperature between the basin and river flood-water or by the concentration of suspended matter in the latter (McCabe, 1978), together with inertia as the flood entered the basin.

The thick units of sandstone represent a period of sustained traction. These conditions are prevalent in proximity to river mouths. It is possible therefore that some of these thick sandstone units represent distributary mouth-bars. The abrupt bases to these sandstone units may be the result of channel avulsion with the subsequent rapid development of a mouth-bar. The abrupt tops to these sandstone units may also be due to channel avulsion, however, in some cases, upward-fining trends occur which may be due to gradual channel abandonment.

The mudstones of the mudstone and ripple cross-laminated sandstone facies were probably partly deposited by settling from turbid surface layers generated by the river plumes. However, in some instances, graded, laminated beds are present, which might have been deposited by slow, dilute underflows, by a process of depositional sorting in the boundary layer (Stow and Bowen, 1980).

The sequence in Ashes Hollow (fig. 36) may be interpreted as representing one incomplete coarsening-upwards sequence from mudstones to mudstones with interbedded sandstones (from 590m to 640m on fig. 36), followed by one complete coarsening-upwards sequence from mudstones to interbedded mudstone and sandstone and then to sandstone (from 640m to 800m on fig. 36). Such sequences may be interpreted as representing delta progradation, with coarser deposits being deposited in closer proximity to the river mouths. The presence of two cycles may be the result of delta-lobe switching, in a manner comparable to the switching of the sites of deposition in the Mississippi delta (e.g. Coleman, 1981). The 30m thick fining-upwards cycle at c.650m on fig. 36 may be due to the gradual abandonment of the first lobe. Alternatively, these cycles may be due to an increase in sediment supply, resulting in delta progradation, and a decrease in sediment supply, resulting in delta regression.

The mudstones with thin sand and silt beds may be referred to the delta-slope environment. However, there is a lack of features such as gullies, slides, slumps and channels which are often associated with slope environments (Stanley and Unrug, 1972; Pickering, 1982 and Coleman, 1981). It is probable that the soft-sediment deformation features which were previously noted are the result of slope instability. The lamination in the sandstone of fig. 55, slab B is intensely overfolded and swirled and this suggests considerable flow in the horizontal sense. Additionally, the microfaulting present in the slab of fig. 57 suggests that synsedimentary extension and down faulting occurred, which was probably initiated by sliding on an inclined surface. These features, therefore, are compatible with the delta-slope environment. The delta-slope deposits are approximately 110m thick.

Since the base of the thick-bedded and cross-stratified sandstone is abrupt, it is possible that much of the delta-front deposits have been eroded.

6.11 THE RELATIONSHIPS BETWEEN THE DELTA AND THE TURBIDITE FAN

The turbidites of the Burway Formation occupy the prodelta environment with respect to the Cardingmill Grit delta. The thick-bedded turbidite facies occurs directly at the base of the prodelta slope. The turbidites, therefore, were likely to have been supplied with detritus by the Cardingmill Grit delta. Since the maximum grain size of the Cardingmill Grit is fine sand, then the maximum grain size of the turbidites is limited to this. This explains the lack of grain size variation exhibited by the turbidites (fig. 36).

No channels were found in the delta-slope sediments. Therefore, the mechanism by which the sediment was supplied to the turbidite fan from the delta is uncertain. There are similarities between the Longmyndian turbidite system and the submarine-ramp facies model for delta-fed, sand rich, turbidite systems (Heller and Dickinson, 1985). Heller and Dickinson (1985) suggest that this type of turbidite system was fed by a number of small gullies in the delta-slope rather than by a single large channel, as postulated by the Walker fan model (1978). However, the apparent lack of small channels in the Longmyndian delta-slope deposits does not conform with the model of Heller and Dickinson (1985). It is possible that the delta-slope deposits in the Longmyndian were deposited on a non-channelised part of the delta-slope, which was developed adjacent to an otherwise channelised slope which is not now exposed. Alternatively, some of the sandstones in the mudstone and cross-laminated sandstone facies may represent channel deposits.

6.12 THE GREEN MUDSTONE AND CROSS-LAMINATED SANDSTONE FACIES

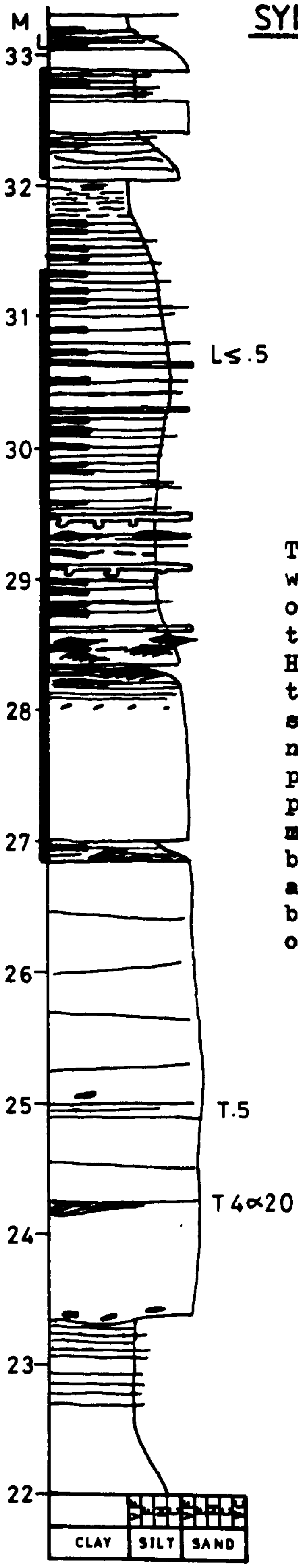
The Green Synalds Member of the Synalds Formation is composed of this facies. It is underlain by the thick-bedded and cross-stratified sandstone of the Cardingmill Grit Member of the Burway Formation and is overlain by the red mudstone and cross-laminated sandstone facies of the Red Synalds Member of the Synalds Formation. This facies is illustrated by log 13, fig. 64; log 15, fig. 65 and log 18, fig. 66.

This facies is characterised by three main lithologies: laminated mudstones, thick beds of cross-bedded sandstone and thin beds of cross-laminated sandstone.

Thick beds of cross-bedded sandstone

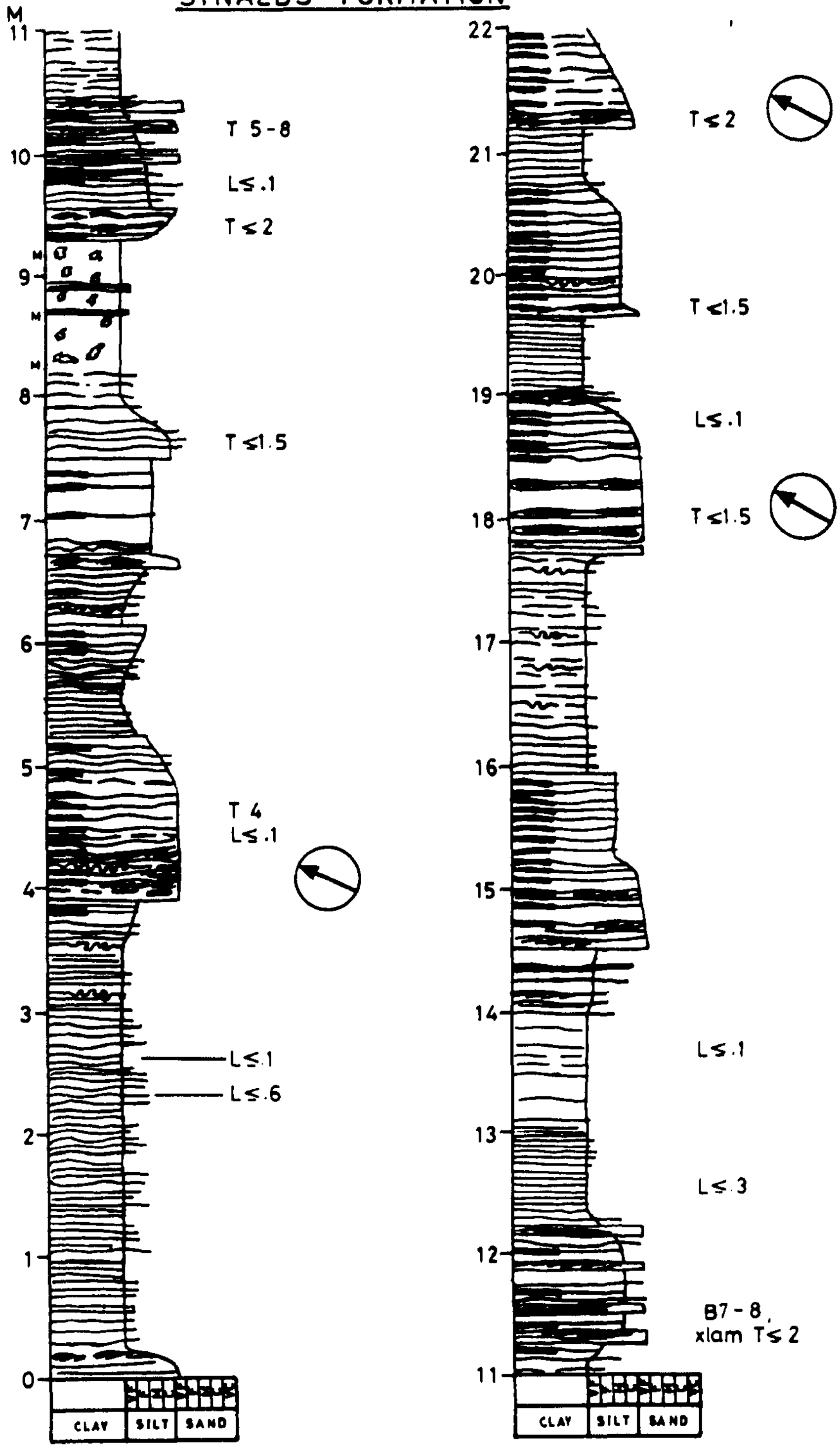
The thick beds of cross-bedded sandstone occur most commonly towards the base of the Green Synalds Member and are up to 5m thick, but are mostly only a few metres thick. Several of these beds are shown on log 18, fig. 66. The bases of these beds are always abrupt and the tops may be gradational or abrupt with the overlying mudstone. One upward-fining sandstone bed is shown in fig. 64, log 13. The sets of cross-bedding are of the order of 30cm thick. Many of the thinner sandstone beds appear to be

FIG 64 GREEN MUDSTONE AND CROSS-LAMINATED SANDSTONE
LOG 13 (cont'd) ASHES HOLLOW, SO 43239306
SYNALDS FORMATION



This log shows an 8.7m upward-fining sequence which is interpreted as a channel-fill and overbank flood sequence and which occurs at the base of the Green Synalds Member in Ashes Hollow. Note the presence of shale clasts at the base of the 5m thick, Cardingmill Grit-like sandstone, its thick-bedded (?cross-bedded) nature, the slight, normal grading and the presence of ripple cross-lamination and planar parallel lamination at the top. The overlying mudstone fines upward and contains thin, loaded beds of sandstone at the base, some of which are ripple cross-laminated. The thin sand beds between 32m and 33m on the log are probably overbank flood and crevasse-splay deposits.

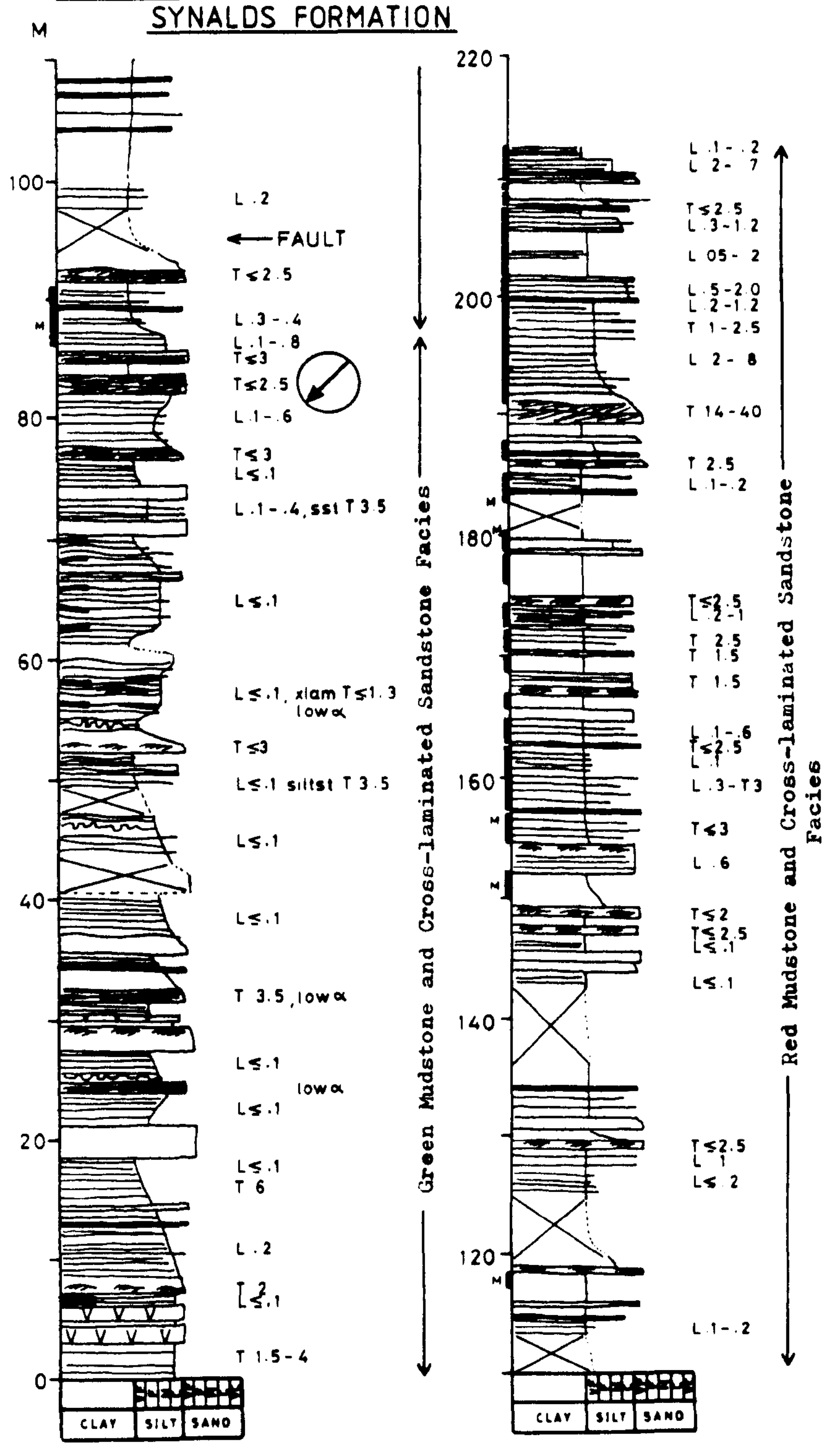
FIG:65 LOG 15 ASHES HOLLOW, GRID REF SO 43139310
SYNALDS FORMATION



This log shows the green mudstone and cross-laminated sandstone facies of the Green Synalds Member of Ashes Hollow. Note the thin sandstone beds (up to 1m thick) which are current ripple cross-laminated and have unidirectional palaeocurrents, which have a WNW orientation. The mudstone is thinly to very thinly laminated. Note the common, contorted laminae and the common, wavy and non-parallel and subparallel nature of the lamination.

FIG: 66 LOG 18 MOUNT GUTTER, SO 40178837

SYNALDS FORMATION



This log shows the transition from the green mudstone and cross-laminated sandstone facies to the red mudstone and cross laminated sandstone facies. Note the thick beds of upward-fining sandstone in the former (e.g. at 20m on log) and the wavy non-parallel lamination (c.50m to 70m on log). The red mudstone and cross-laminated sandstone facies is green and red at the base and becomes predominantly red upward (red sediment is indicated by a thick line on the left margin of the log). This facies is characterised by the presence of numerous thin (less than 20cm) sand beds which are indicated by a thick horizontal line on the log. Note the presence of the thick, cross-bedded and upward-fining sandstone bed at c.190m on the log.

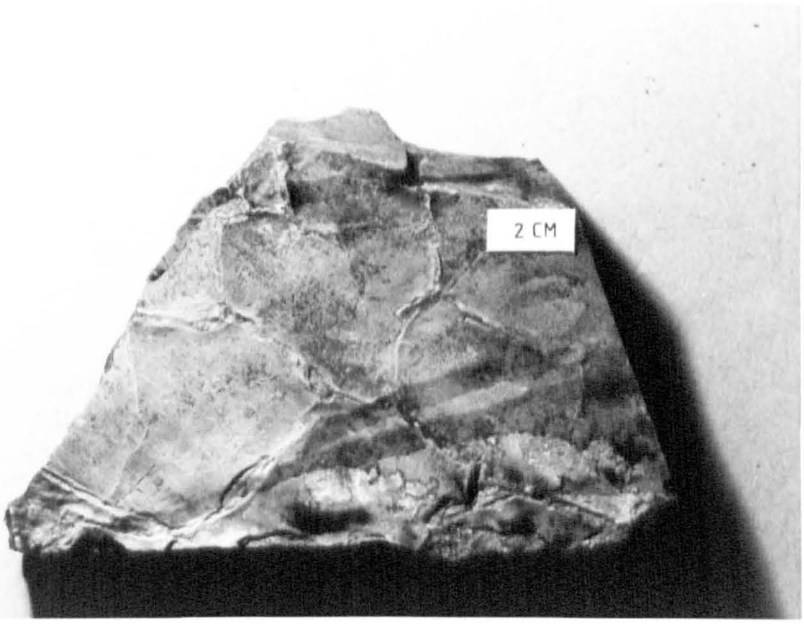
structureless. The finer grained tops of all these sandstone beds are ripple cross-laminated. Clasts of purplish red mudstone are occasionally found at the bases of these beds and occasionally within them.

Laminated mudstones

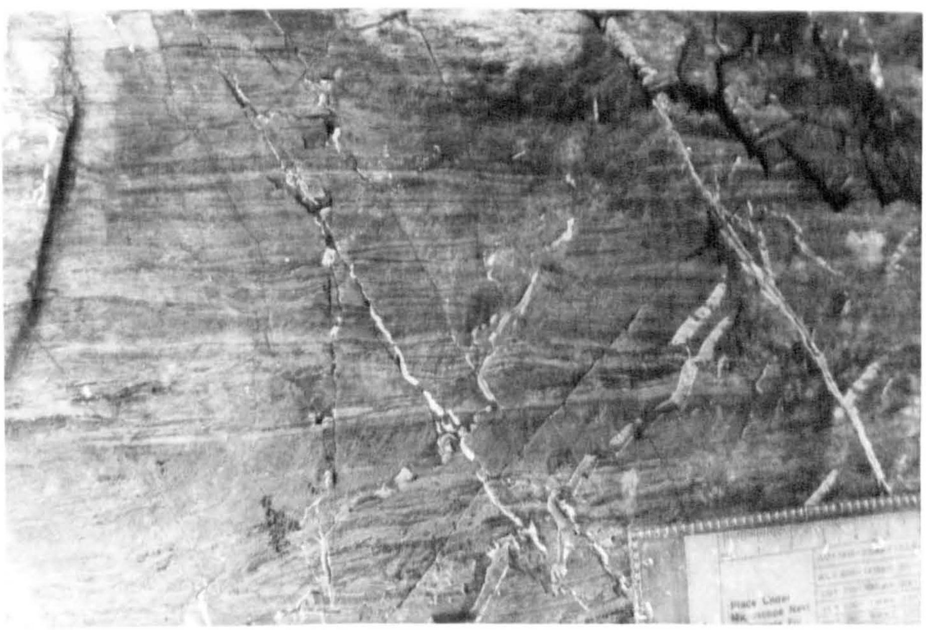
The mudstones are thinly to thickly laminated and are purplish red, purplish grey or greenish grey. The purple and red colours are developed towards the base of the Green Synalds Member where the mudstones are interbedded with the thick beds of cross-bedded sandstone and the greenish grey colours are predominant and often exclusively developed at the top of the member. In these mudstones, mudcracks are occasionally found and several horizons which contain mudcracks occur towards the top of the Green Synalds Member in Ashes Hollow at SO 4313 9313. One example of these mudcracks is shown in fig. 67, photo A.

The most common type of mudstone is thinly laminated siltstone (fig. 67, photo B). The laminae are continuous to discontinuous, subparallel and slightly wavy and contorted laminae are common. The lithologies of the laminae are very variable and claystone, siltstone and very fine grained sandstone are often interlaminated. The average grain size of the laminated mudstone is variable, from coarse siltstone to very fine siltstone. In the coarser lithologies, very thin beds and thick laminae, a few centimetres thick, of ripple cross-laminated and planar, parallel laminated coarse siltstone and very fine grained sandstone occur. These may be lenticular and are often normally graded.

FIG: 67 PHOTOGRAPHS SHOWING MUDCRACKS AND LAMINATED MUDSTONE FROM THE GREEN MUDSTONE AND CROSS-LAMINATED SANDSTONE FACIES OF THE GREEN SYNALDS MEMBER.



A; Mudcracks in thinly laminated siltstone. Ashes Hollow; SO 4313 9313.



B; Typical thinly laminated mudstone. Note the slightly wavy, subparallel, continuous laminae and the oblique and contorted laminae in the bottom left of the photo. Scale is graduated in cms and mms.

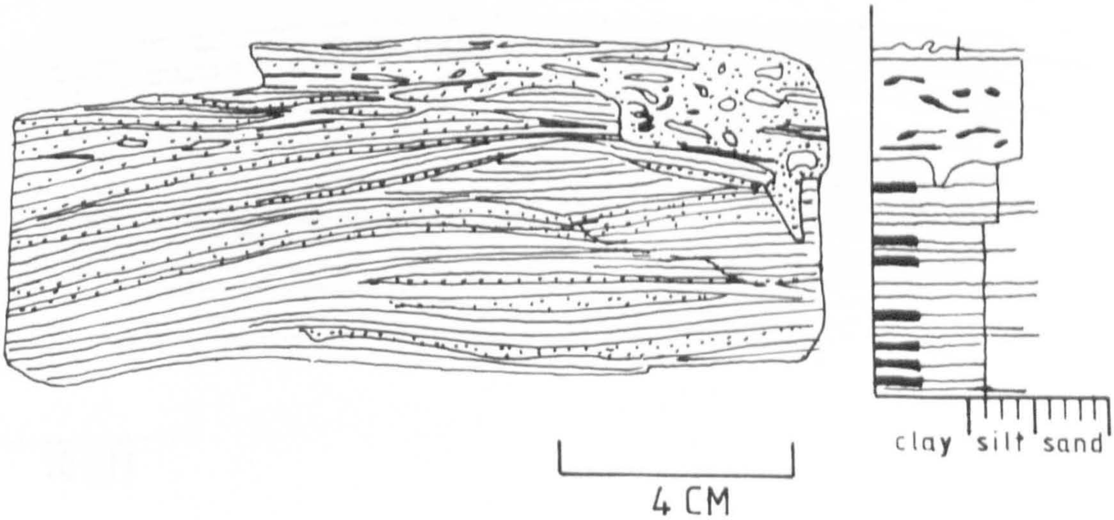
Soft-sediment deformation features are common. One type consists of a layer of highly inclined laminae of mudstone. The laminae are abruptly truncated both above and below by undisturbed, laminated mudstone. Layers of intensely contorted lamination are common. These deformed layers are composed of poorly sorted, coarse siltstone (fig. 68). The highly contorted nature results in the development of irregular "clasts" of mudstone, which have originated by the disruption of the laminae.

In the upper sections of the Green Synalds Member, bundles of laminae which are wavy and non-parallel are common (fig. 69). This structure is developed in coarse siltstone which has laminae of very fine grained sandstone and claystone. These structures are up to 40cm long and can be only a few centimetres long. A similar, small variety is shown in fig. 68. The larger varieties are superficially similar to hummocky cross-stratification (compare Dott and Bourgeois, 1982, fig. 2A with fig. 69).

The laminated mudstones often form the upper parts of upward-fining sequences which commence with either thick beds of cross-bedded sandstone (fig. 64, log 13) or thin beds of cross-laminated sandstone (fig. 65, log 15). In these sequences, sandstone laminae and thin beds, together with cross-lamination, are more common at the base of the upward-fining mudstone and the laminae become progressively thinner upward.

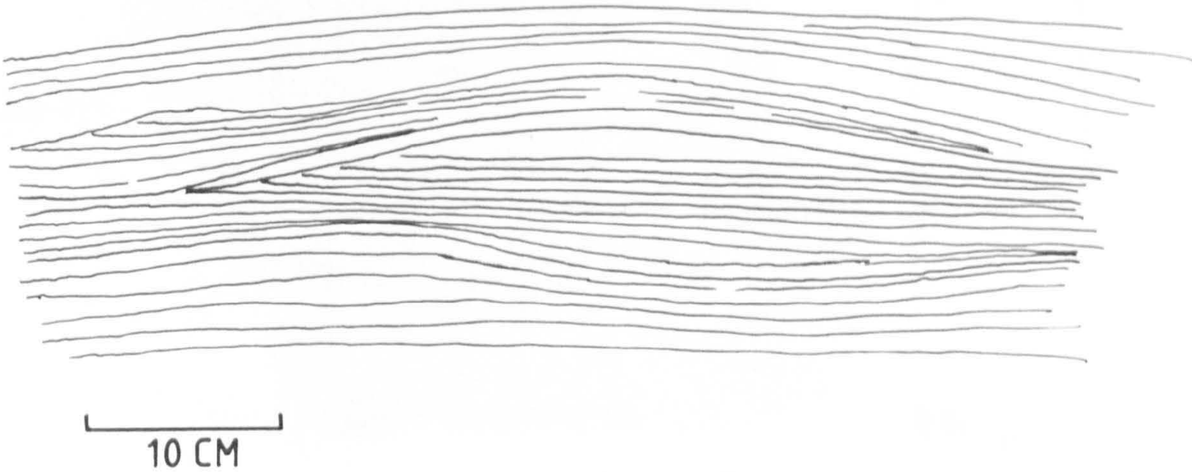
Distinctive, medium grey, very thinly laminated mudstones occur in Ashes Hollow at SO 4318 9311. These mudstones occur in a bed which is 1.4m thick. The top of this bed is abruptly overlain by sandstone and the base appears to be transitional with a normally graded ash which is 0.45m thick. The laminae of the mudstone are normally graded and are variably medium to dark grey

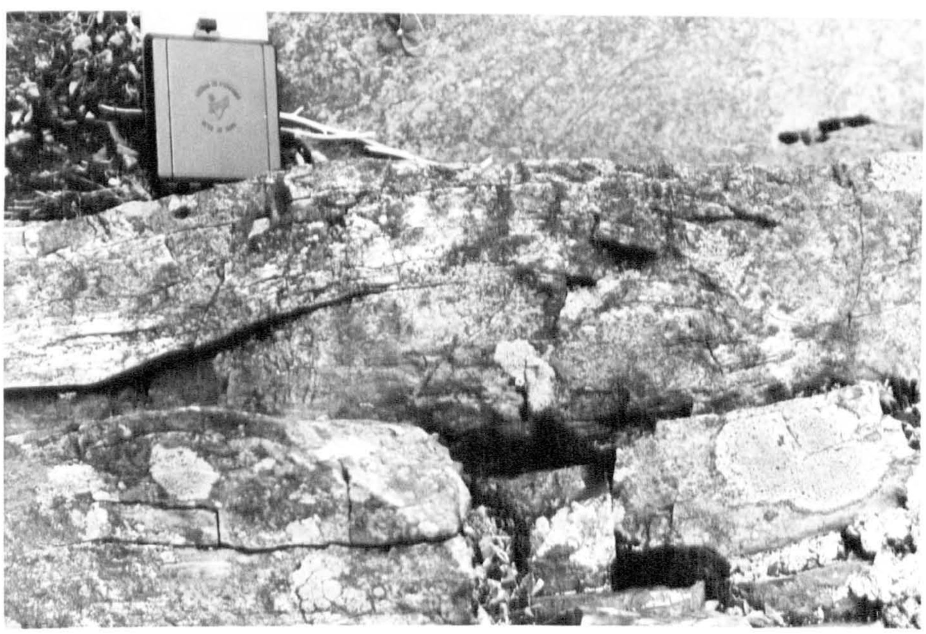
FIG: 68 SLAB LAMINATED MUDSTONE FROM THE GREEN MUDSTONE AND CROSS-LAMINATED SANDSTONE FACIES.



Note the thin,subparallel and continuous laminae of variable grain size,the contorted lamination (top right of centre) and the convex surface which truncates the underlying laminae (right of centre).The contorted layer lies above a discrete plane which separates it from the undeformed sediment.

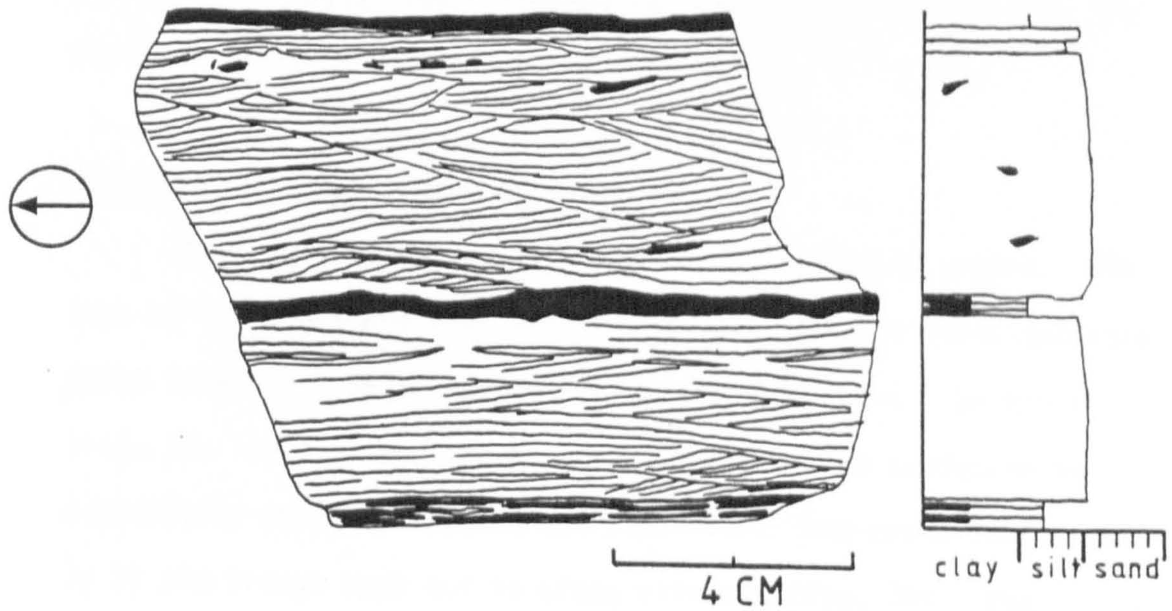
FIG: 69 WAVY, NON-PARALLEL LAMINAE BUNDLES IN THE GREEN MUDSTONE AND CROSS-LAMINATED SANDSTONE FACIES OF THE GREEN SYNALDS MEMBER.



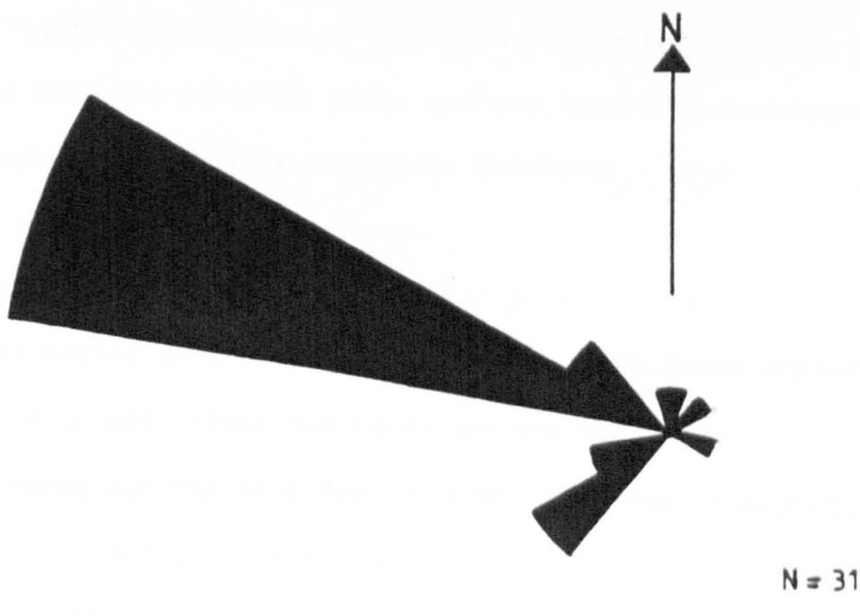


The lithology consists of coarse siltstone with thin laminae of sandstone and claystone. Compass/clinometer (upper left) is 10cm long.

FIG: 70 SLAB THIN CROSS-LAMINATED SANDSTONE BED FROM THE GREEN MUDSTONE AND CROSS-LAMINATED SANDSTONE FACIES OF THE GREEN SYNALDS MEMBER AND PALAEOCURRENT ROSE FOR THE ENTIRE SYNALDS FORMATION.



SLAB; Thin ripple cross-laminated sandstone bed from the Green Synalds Member. Note the presence of climbing ripple cross-lamination and trough cross-lamination.



Palaeocurrent rose for the ripple cross-laminated sandstones of the entire Synalds Formation (Green Synalds Member plus Red Synalds Member).

and greenish grey. The darker laminae contain disseminated organic matter and very small filaments which are interpreted as algal filaments (discussed in section 5.10). Similar filaments and a similar lithology, from different horizons, are shown in fig. 34, photos A and B.

Thin beds of cross-laminated sandstone

These beds are up to 1m thick and are normally graded. The tops of the sandstone beds are gradational with laminated mudstone, which forms part of an upward-fining sequence up to 1.5m thick (fig. 65, log 15), and the bases are abrupt. The sandstone is extensively current ripple cross-laminated. The cross-lamination is of the trough type and is often climbing (fig. 70). The palaeocurrents which formed this cross-lamination were unidirectional and towards the WNW. This direction is identical to that of the red mudstone and cross-laminated sandstone facies (fig. 70) and is at an obtuse angle to the palaeocurrent direction for the Cardingmill Grit Member of the Burway Formation. These thin sandstone beds are parallel sided and are laterally continuous over several metres with little change in thickness.

Interpretation

This facies must have been subjected to at least occasional subaerial exposure, since mudcracks are present. The palaeocurrents for the thin beds of cross-laminated sandstone are unidirectional and towards the WNW. The unidirectional currents and the lack of reactivation surfaces and other sedimentary structures characteristic of periodically alternating high and low energy conditions suggest that there was no tidal influence during the deposition of this facies. The WNW palaeocurrent direction is at an obtuse angle to the palaeocurrent direction for the fluviatile Cardingmill Grit Member of the Burway Formation. It is possible that the mudstones and the thin sandstone beds were deposited by overbank flooding of the rivers which deposited both the Cardingmill Grit and the thick sandstone beds in this facies.

The thick beds of cross-bedded sandstone are interpreted as representing channel fills, the palaeocurrent direction of which is not known, but it is suggested that these channels probably flowed ENE, in the same direction as the rivers which deposited the Cardingmill Grit.

There are some beds at the top of the Green Synalds Member with an internal lamination similar to hummocky cross-stratification (HCS) (fig. 69). The fluviatile nature of the associated sediment suggests that this type of lamination is unlikely to be HCS, which is usually found in shallow marine sequences which are affected by large storm waves (Dott and Bourgeois, 1982 and Walker *et al.*, 1983). Furthermore, the grain size of the beds in which this type of lamination occurs is finer than usual for HCS. Alternatively, it may be interpreted as a soft-sediment deformation feature. Oblique and disturbed lamination and subhorizontal planes of discontinuity, which are attributed to soft-sediment deformation, are common in this facies and this structure might have had a similar origin.

Since this facies is directly underlain by the Cardingmill Grit, which is interpreted as a delta distributary channel deposit and since it is mostly composed of floodplain deposits, then it is possibly referable to a delta-plain, interdistributary environment. The thick beds of sandstone are concentrated towards the base of the facies and there appears to be a gradual transition from the sandstones of the underlying Cardingmill Grit to the mudstones at the top of the Green Synalds Member, with a transitional interbedded sequence at the base. This transition is interpreted as representing a lateral shift away from an area dominated by distributary channels to one dominated by floodplain deposits. It thus appears that the distributary channels occurred in a belt which gradually moved laterally across the floodplain.

The beds of grey, very thinly laminated mudstones lack any indications of traction currents. The very thin, normally graded laminae were probably deposited by settling from suspension. The presence of undisturbed, delicate algal filaments (Peat, 1984a) attests to the quiet water conditions. Some of the filaments are contorted and solitary and were probably derived from the filament mats by gentle (?wind generated) currents. The environment of deposition thus appears to have been an isolated, standing body of water on the floodplain. Such a body of water could have occupied an abandoned channel.

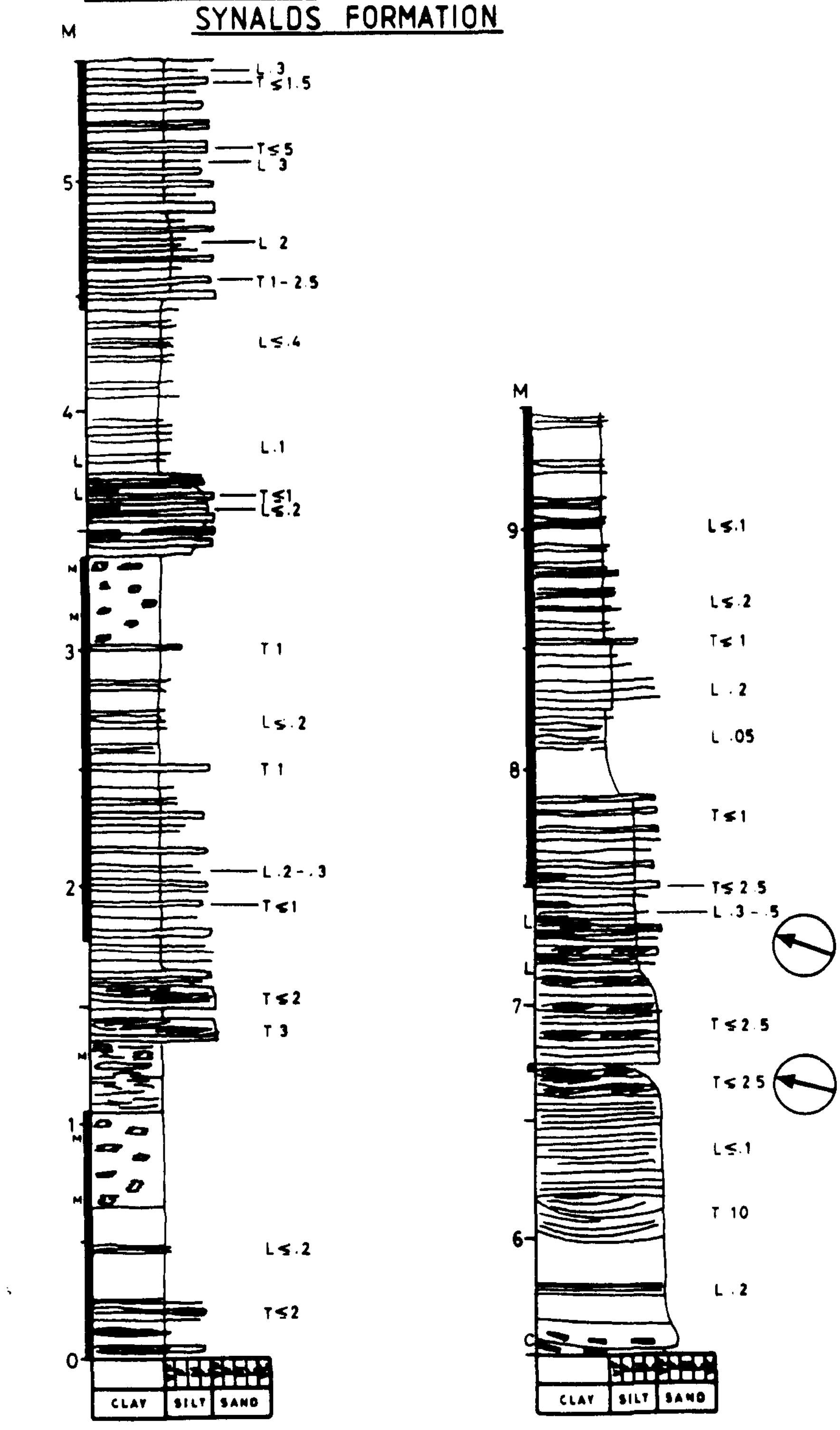
6.13 THE RED MUDSTONE AND CROSS-LAMINATED SANDSTONE FACIES

The Red Synalds Member is composed of this facies. This is underlain by the green mudstone and cross-laminated sandstone facies of the Green Synalds Member of the Synalds Formation and is overlain by the green mudstone and thick-bedded sandstone facies of the Green Lightspout Member of the Lightspout Formation. This facies is comprised of three main lithologies: red mudstone, thick beds of sandstone and thin beds of cross-laminated sandstone. In addition, several tuff beds occur. Those at the top of the Red Synalds Member are called the Batch Volcanics. This facies is illustrated by log 21, fig. 71 and log 22, fig. 72.

Mudstone

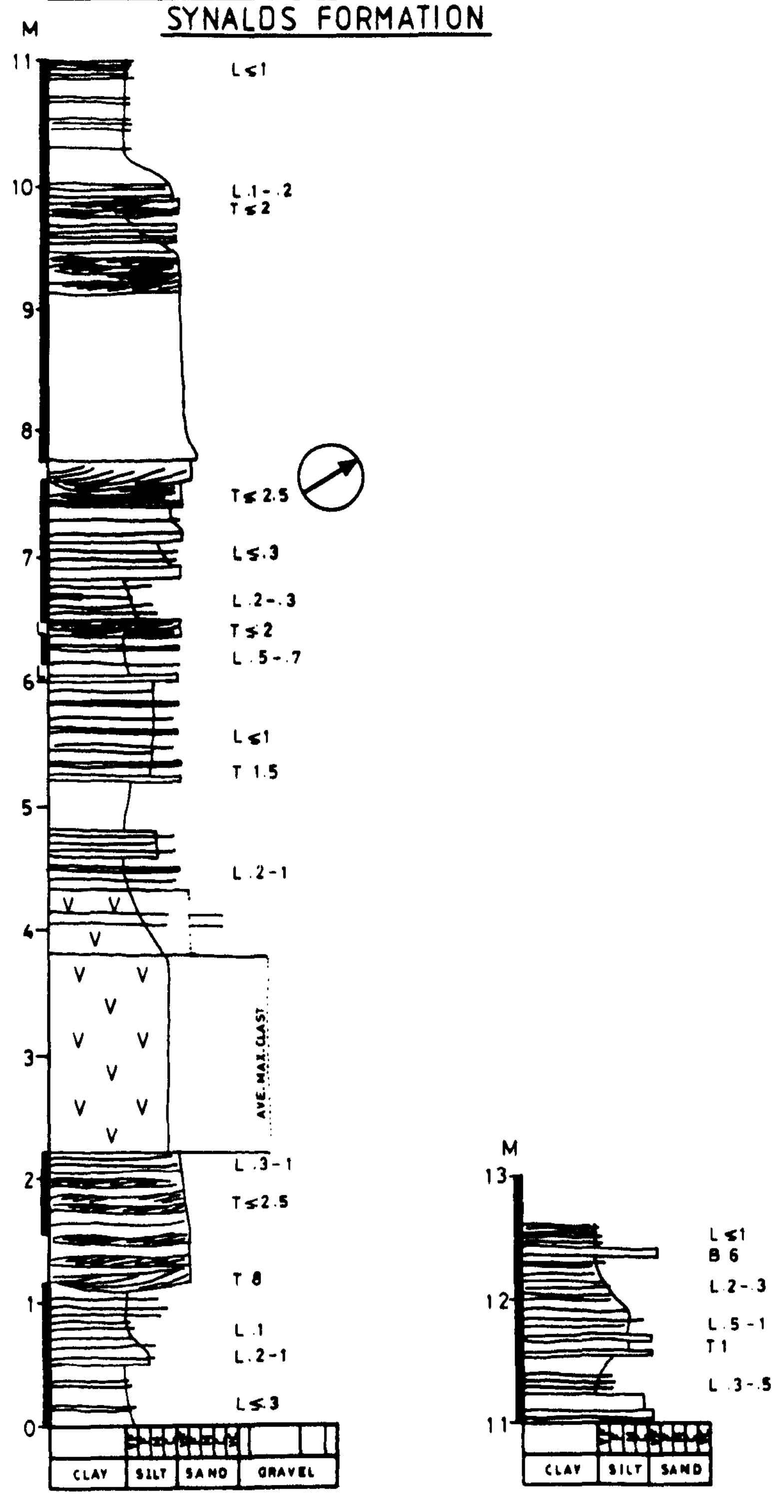
The mudstone is mostly purplish red and purplish grey in colour and is rarely greenish grey. Where a green colouration is developed, this usually occurs in patches which are irregular in shape and size and which have irregular margins (e.g. fig. 73). The mudstone is thinly to thickly laminated and the thicker laminae are of a coarser grain size. The average grain size is fine siltstone and thick laminae of coarse siltstone and very thin beds (less than 2.5cm thick) of ripple cross-laminated sandstone are common. The latter are often lenticular when viewed in cross-section. The laminae may be normally graded, inversely graded or non-graded. They are commonly planar, parallel and continuous, but due to tectonic deformation the laminae may be wavy. The bedding planes often display pseudo-ripples and these are discussed in section 3.11.1. The pseudo-ripple crests are coaxial with the fold axes and the pseudo-ripples were therefore formed during the folding of the Longmyndian. Mudcracks are present in the mudstones and examples of these are shown in fig. 74. They are infrequently found and this may be due to the lack of suitable exposures, since the beds are mostly nearly vertical and extensive bedding plane exposure is infrequent. Soft-sediment deformation is common. One type is shown in fig. 73. In this example, the siltstone laminae are broken and are injected by finer

FIG.71 LOG 21 ASHES HOLLOW, SO 43129338
SYNALDS FORMATION



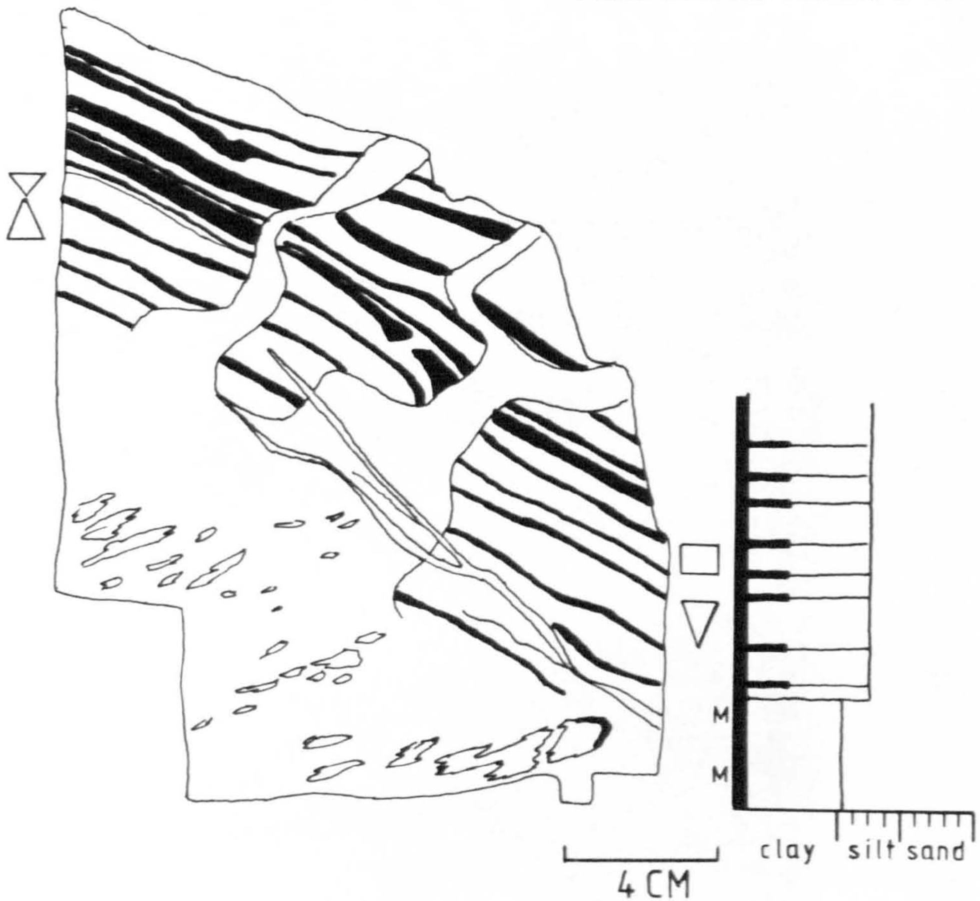
This log shows the red mudstone and cross-laminated sandstone facies. Note the thin, cross-laminated sandstone beds at c.1.5m and c.3.5m on the log and the thick bed of cross-bedded sandstone between 5.5m and 7.2m. Note the overall fining-upward profile of the thick sandstone bed and the sequence of structures, from cross-bedding to planar and parallel lamination and to ripple cross-lamination. The mudstone has numerous very thin beds of sandstone, which elsewhere are seen to be commonly current ripple cross-laminated.

FIG.72 LOG 22 CARDINGMILL VALLEY, SO 43939506
SYNALDS FORMATION



This log shows the White Ash (marked V and between 2.2m and 4.3m on the log) in the red mudstone and cross-laminated sandstone facies. Note the poorly sorted nature of the ash and the distinct, bedded and finer-grained top. The overlying mudstones are light grey and may represent reworked ash. Note the thick sandstone bed between 7.6m and 10.2m which has an overall upward-fining profile with cross-bedding at the base. Note also the probable interruption of channel deposition at c.2m by the White Ash.

FIG:73 SLAB LAMINATED MUDSTONE FROM THE RED MUDSTONE AND CROSS-LAMINATED SANDSTONE FACIES.



Typical laminated mudstone from the red mudstone and cross-laminated sandstone facies of the Red Synalds Member of the Synalds Formation. Note the injection features of fine silt-stone into the laminated fine to medium grained siltstone and the irregular patches of green in the otherwise purplish red siltstone.

FIG: 74 MUDCRACKS FROM THE RED MUDSTONE AND CROSS-LAMINATED SANDSTONE FACIES OF THE SYNALDS FORMATION

PHOTO A Base of a thin sandstone bed. Note the numerous, small blister-like projections (these are discussed in Chapter 5)

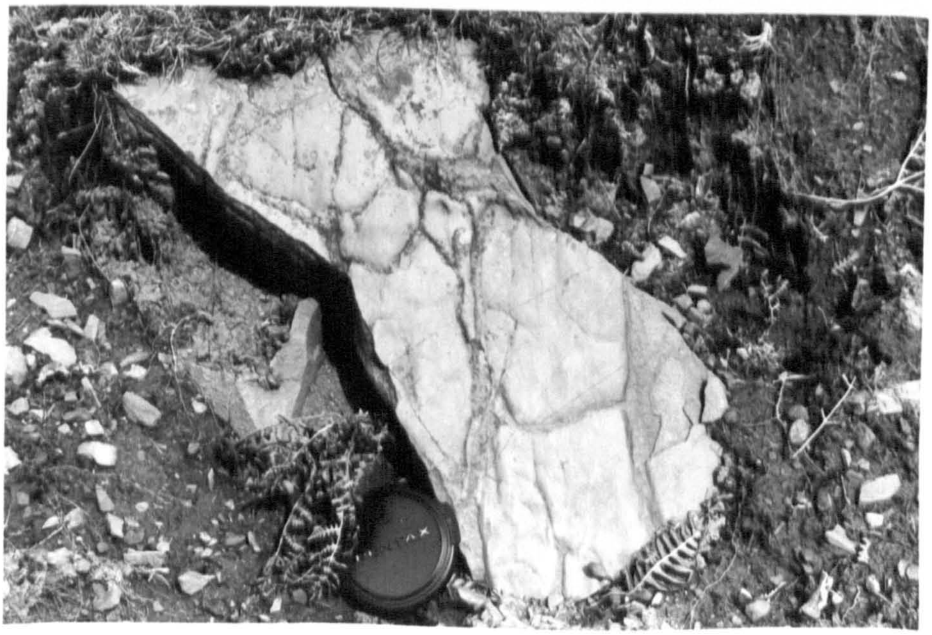


PHOTO B Top of mudstone bed. Lens cap diameter is 52mm.

grained siltstone. The very thin sandstone beds often show loading, dewatering and injection features and these are accompanied by disruption of the internal lamination.

There are numerous small pits on the bedding planes. These are discussed in section 5.5 and it was concluded that these are either biogenic or are due to air-escape. Several authors have referred to these pits as raindrop impressions (e.g. James, 1952b). However, this mechanism of formation is not thought to be probable (see section 5.5). No undisputable raindrop impressions have been found by the author. However, possible raindrop impressions are figured by Greig *et al*. (1968, plate 4, fig. D).

There are also common, subparallel, fine lines on many of the bedding planes. These are discussed in sections 5.8 and 5.9 and it was concluded that these are tectonic in origin. However, Bland (1984) considers them to be organic and refers them to the genus *Arumberia* (see section 5.9).

Thick beds of sandstone

These are up to c.2m thick and are composed of very fine grained sand. They are therefore much thinner and more fine grained than the thick beds in the green mudstone and cross-laminated sandstone facies. The bases of the beds are abrupt and occasionally erosional. The basal sandstone is fine grained and often contains clasts of purplish red mudstone which are tabular in shape. The sandstone is trough cross-bedded and planar, parallel bedded at the base and passes up into ripple cross-laminated and planar, parallel laminated sandstone at the top, with an accompanying gradual decrease in grain size. This

FIG.75 SANDSTONE BODY MARGIN FROM THE RED MUDSTONE AND CROSS-LAMINATED SANDSTONE FACIES OF THE RED SYNALDS MEMBER IN PIKE HOLLOW AT SO 3983 8830

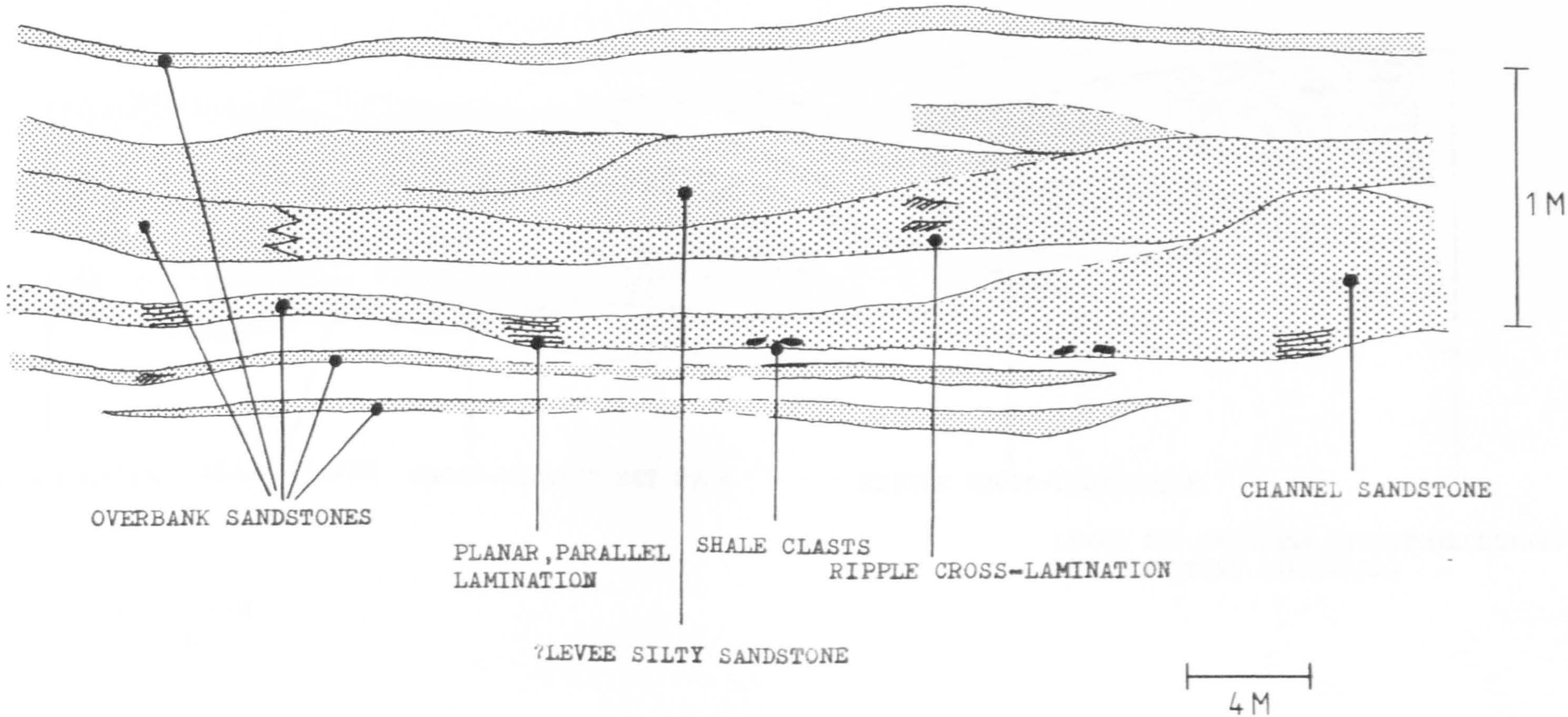
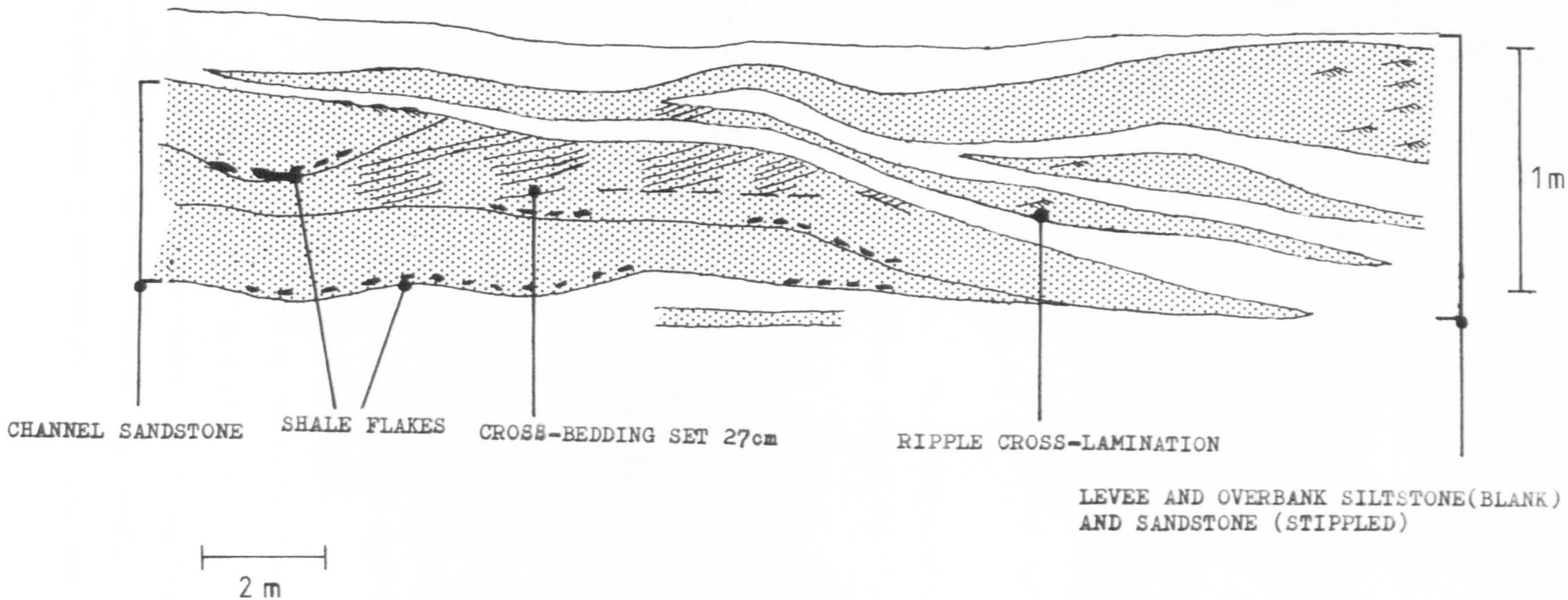


FIG. 76 SANDSTONE BODY MARGIN FROM THE RED MUDSTONE AND RIPPLE CROSS-LAMINATED SANDSTONE FACIES OF THE RED SYNALDS MEMBER IN PIKE HOLLOW AT SO 3987 8836



then grades into siltstone which contains numerous very thin beds (less than c.2.5cm thick) of sandstone which are ripple cross-laminated.

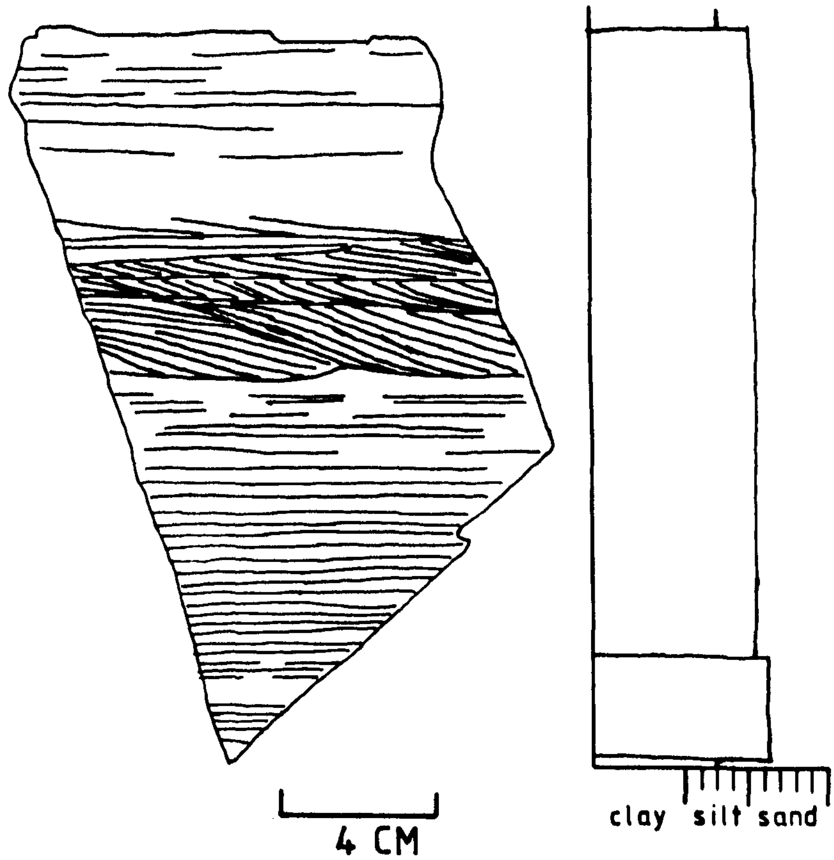
The beds mostly appear to be laterally continuous with little change in bed thickness. However, at two localities, the sandstone beds were observed to thin rapidly (figs. 75 and 76). At one locality (fig. 75), the sandstone passes into thin, laterally persistent sandstone beds and one sandstone bed passes laterally into a thick bed of silty sandstone. The thick sandstone bed in fig. 76 appears to be composed of three distinct units, each of which contains shale flakes at the base.

The thick beds of sandstone form a small percentage (less than 5%) of the total thickness of the Red Synalds Member of the Synalds Formation.

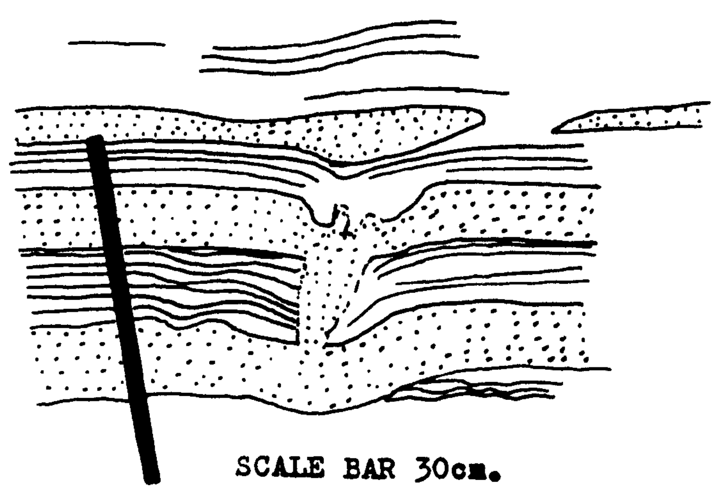
Thin beds of sandstone

These are common, laterally persistent, with little change in bed-thickness, up to c.40cm thick and are composed of very fine grained sandstone. The tops and bases of the beds are abrupt and the bed may be non-graded or normally graded at the top. The base of the bed can show flame and load structures and can contain small clasts of mudstone. A sequence of sedimentary structures is evident from a number of slabbed beds (fig. 77). The base is often planar, parallel laminated and passes up into ripple cross-lamination. This may be overlain by planar, parallel laminated coarse siltstone and very fine grained sandstone. Some of the slabs do not show all of these structures and one is wholly ripple cross-laminated and the cross-lamination is mainly of the climbing-ripple type. Some of the thin sandstone beds pass

FIG:77 SLAB: THIN SANDSTONE BED FROM THE RED MUDSTONE AND CROSS-LAMINATED SANDSTONE FACIES AND SANDSTONE DYKE FROM THE SAME FACIES.



SLAB: Complete sandstone bed. Note the sequence of sedimentary structures; planar and parallel lamination at the base, followed by ripple cross-lamination and overlain by planar, and parallel lamination



SANDSTONE DYKE; note the pinch and swell of the overlying sandstone bed.

laterally into thick sandstone beds (e.g. fig. 75), however this is not common. The palaeocurrent direction derived from these thin sandstone beds is WNW (fig. 70). The palaeocurrents are unidirectional and the direction (WNW) is identical to that for the thin sandstone beds in the green mudstone and cross-laminated sandstone facies and is similarly at an obtuse angle to the average palaeocurrent direction for the Cardingmill Grit, this being ENE, and to the average palaeocurrent direction for the Lightspout Formation, this being approximately NE.

Dewatering structures in these sandstones are common and the internal lamination may be contorted. The beds may exhibit pinch and swell features and occasionally, sandstone dykes occur (fig. 77).

Tuffs

Two main tuff beds occur towards the top of the Red Synalds Member. These, together with other minor, thinner tuff beds, are called the Batch Volcanics. The main tuff beds are called the White Ash and the Andesitic Ash, the former occurring below the latter. The White Ash is illustrated by log 22, fig. 72. It is approximately 2m thick and is a poorly sorted, light grey to white, lapilli tuff. The groundmass is very fine grained and is highly altered, with abundant mica, chlorite, epidote and sphene, which are the common alteration products of the Longmyndian sediments (see section 4.7). The lapilli and coarse ash fragments include crystals of quartz, plagioclase and interlaminated muscovite and chlorite. The latter has a chevron-style schistosity and may be metamorphic in origin. The majority of the fragments are lithic. These include flow-banded rhyolites, light brown, glassy volcanics and feldspar-phyric volcanic rock fragments. The most common lithic fragment is highly altered and is mainly composed of fine mica, chlorite, epidote and sphene. Patches of chloritised glass with plagioclase may occur in these fragments and a schistosity is occasionally developed with lepidoblastic chlorite, which is possibly replacive of biotite.

The base of the White Ash is abrupt and rests on a thick bed of sandstone in the Cardingmill Valley at SO 4393 9506 (log 22, fig. 72). This sandstone bed is unusually thin for a thick-bedded sandstone and an upward-fining sequence is lacking. It is possible that the deposition of the sandstone was interrupted by the deposition of the White Ash. The majority of the White Ash is massive. However, the top c.20cm is slightly finer grained and there is a fine bedding which is the result of grain size variations. The overlying siltstones are similar to the typical siltstones of the red mudstone and cross-laminated sandstone facies but are light grey in colour. These siltstones may therefore represent reworked ash.

The Andesitic Ash is a poorly sorted, mainly purplish red, lapilli tuff. The basal 30 centimetres are light greenish grey. In the exposure at the junction of Long Batch and Jonathon's Hollow, at SO 4463 9605, both the base and the top of the ash are abrupt and the ash is massive and lacks grading. The groundmass is highly altered, with fine chlorite, hematite, sphene and mica. Crystal fragments include quartz and plagioclase. Lithic fragments are predominant and include a feldspar-phyric volcanic rock in which the plagioclase commonly occurs in glomerocrysts. The groundmass of these fragments is mostly replaced by chlorite, epidote, sphene and mica. Other lithic fragments include light brown, aphyric, glassy volcanic rock fragments and fragments wholly replaced by epidote. The character and lithological composition of the Andesitic Ash is very similar to that of the White Ash, the main difference being the colour.

Interpretation

The mudstones are interpreted as having been deposited from suspension and by traction during the flow of an unconfined sheet of slow moving and turbid water. Evidence for traction is provided by the presence of common ripple cross-lamination in the thicker sand laminae and coarse silt laminae. The process of transportation envisaged is similar to that of "sheetflow" as defined by Hogg (1982). "Sheetflow" refers to an overland flow which occurs in a continuous sheet, is restricted to laminar flow conditions, is restricted to slopes of less than 5%, which has water depths of the order of millimetres and velocities of the order of centimetres per second (Hogg, 1982).

The thin beds of sandstone represent deposition from higher energy, short-lived, flood events. The currents which deposited these sandstones were unidirectional. Rapid deposition is indicated by the presence of climbing ripple cross-lamination and flame and load structures at the base. The currents often gradually waned in velocity, producing an organised sequence of sedimentary structures which begins with planar, parallel laminated sandstone, of probable upper flow-régime, which is followed by ripple cross-lamination and finally by planar, parallel laminated silty sandstone, both of probable lower flow-régime. The erosive power of these flood events is demonstrated by the presence of shale clasts at the bases of some of the sandstone beds.

The thick beds of sandstone are interpreted as channel-fills. Although a channel profile cannot be demonstrated, the margins of these channels are recognised at two localities (figs. 75 and 76). Since one of the thick beds of sandstone passes laterally into two thin beds of sandstone (fig. 75), then it appears probable that these and some of the associated mudstone beds might have been deposited by overflooding from the channel. The silty sandstone which is adjacent to one of the channel sandstones (fig. 75) could represent a levée, which was constructed during the overflooding of the channel. The occurrence, within one of the channels, of several erosive surfaces with shale flakes at the base, suggests that these channels were filled by several flood events. Since the number of thick beds of sandstone is small, then it can be concluded that channelisation of flow was of minor importance, with respect to sheetflow, in this environment. Channelisation of the flow was probably accomplished during higher magnitude floods of lower frequency.

Since the process of deposition is considered to be essentially by alluvial sheetflow and since there is evidence for subaerial exposure, the environment of deposition is interpreted to have been an alluvial floodplain. However, it is not certain if this floodplain environment represents the distal part of an alluvial braidplain or the overbank floodplain of a fluvial system which is represented by the thick-bedded and cross-stratified sandstone facies and the green mudstone and thick-bedded sandstone facies. These alternatives are discussed in section 6.17 on the alluvial floodplain facies association.

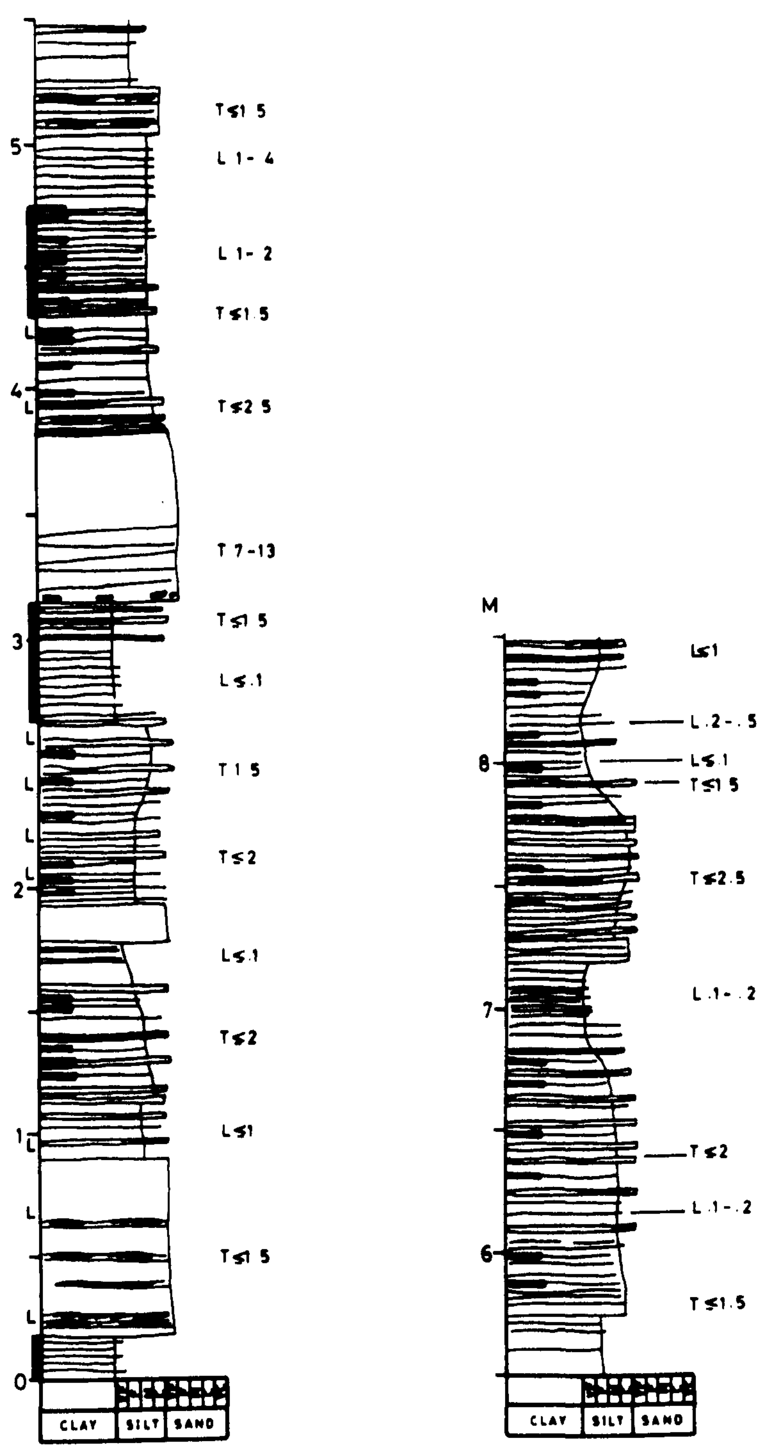
The tuffs which occur in this facies have the grain size characteristics of pyroclastic flow deposits, in that the sorting is poor and the grain size and sorting is invariable throughout the deposit (Walker, 1971). All of the observed tuffs are interpreted as single flow units. The bedded and finer grained top to the White Ash in the Cardingmill Valley (log 22, fig. 72) may be interpreted as an ash-cloud deposit (Fisher and Schmincke, 1984). The fine grained, laminated and light grey ash overlying this ash-cloud deposit may be interpreted as "reworked fallout tephra" (Fisher and Schmincke, 1984) or as "co-ignimbrite ash fall" (Suthren, 1985).

6.14 THE GREEN MUDSTONE AND THICK-BEDDED SANDSTONE FACIES

The Green Lightspout Member of the Lightspout Formation is composed of this facies. The stratigraphic thickness of this member is approximately 220m. This thickness is an estimate, since there is extensive minor folding and there is a lack of marker horizons. This facies is underlain by the red mudstone and cross-laminated sandstone facies of the Red Synalds Member of the Synalds Formation and is overlain by the red mudstone and cross-bedded sandstone facies of the Red Lightspout Member of the Lightspout Formation.

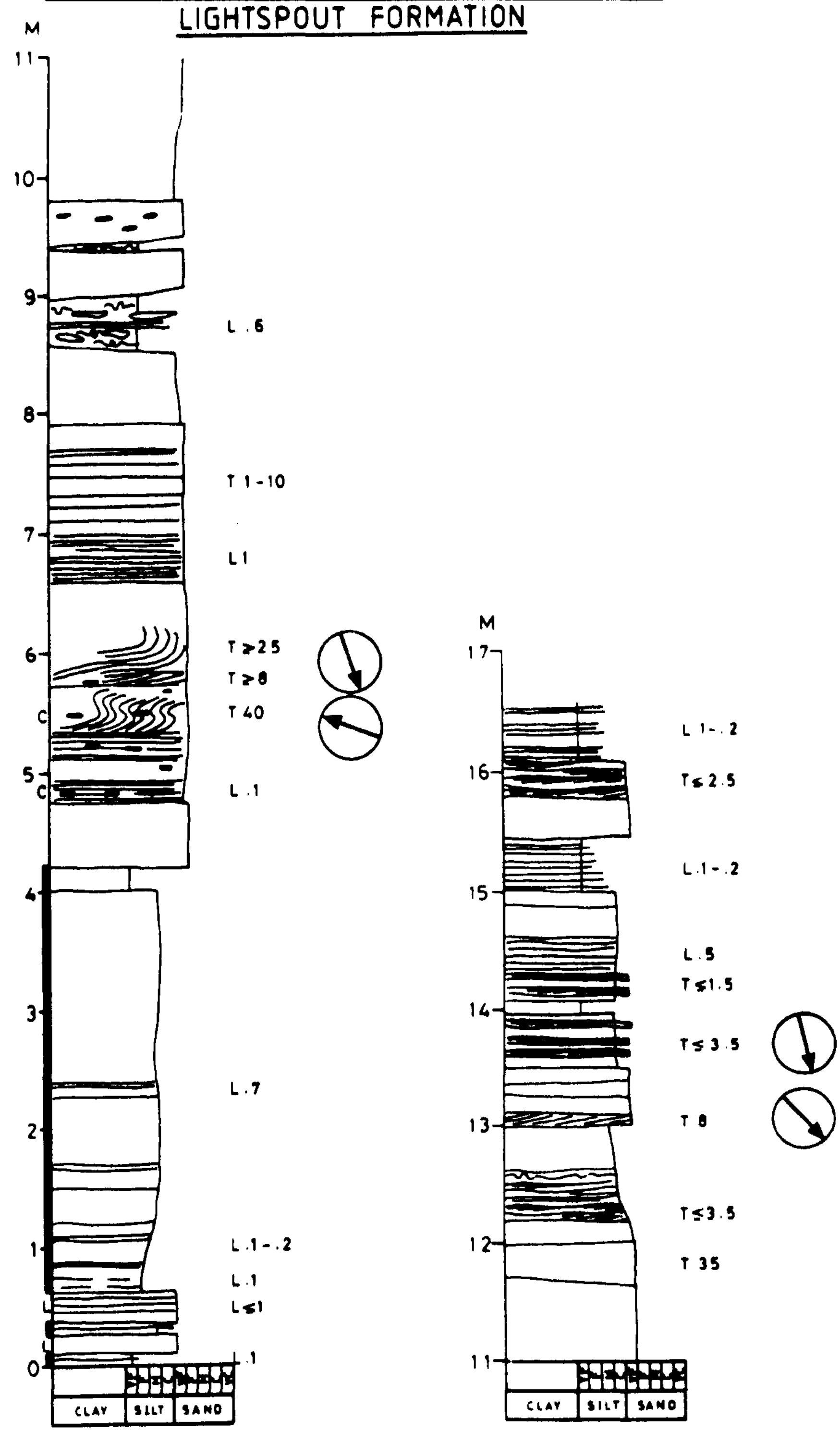
This facies is comprised of three main lithologies: laminated mudstone, thin beds of sandstone and thick beds of cross-bedded and cross-laminated sandstone. In addition to these lithologies there are occasional beds of tuff. This facies is illustrated by fig. 78, log 23 and fig. 79, log 24.

FIG.78 LOG 23 CARDINGMILL VALLEY SO 43739519
LIGHTSPOUT FORMATION



This log illustrates the green mudstone and thick-bedded sandstone facies. The average mudstone grain size is medium to coarse silt. There are numerous, very thin sandstone beds, which are less than 2.5cm thick, in which cross-lamination is occasionally visible. Note the two thick beds of cross-bedded and cross-laminated sandstone at c.0.5m and c.3.5m. Upward-fining sequences are present in the mudstones at c.1.5m and c.6.5m. Thin beds of cross-laminated sandstone occur at c.1.9m and c. 5.2m.

FIG.79 LOG 24 CARDINGMILL VALLEY, SO 43739525
LIGHTSPOUT FORMATION



This log illustrates the green mudstone and thick-bedded sandstone facies. Note the thick bed of cross-bedded and planar, parallel laminated sandstone between c.4.2m and c.12.2m. This appears to be composite, with a thin mudstone bed at c.8.8m. Note the contorted and overturned cross-bedding and the mudstone clasts which are concentrated towards the bases of the sandstone units. Note the thin sandstone beds at c.0.5m, c.13.2m and c.15.7m. The sequence above the thick sandstone bed appears to be a composite upward-fining sequence. The mudstone beneath the thick sandstone bed is purplish red.

Thick beds of cross-bedded and cross-laminated sandstone

These beds are similar to those at the base of the green mudstone and cross-laminated sandstone facies of the Green Synalds Member. The beds range from c. 0.7m to c. 8.5m in thickness. The cross-bedding sets are up to 48cm thick and the laminae are commonly deformed (fig. 80). Upward-fining profiles are often present and a sequence of sedimentary structures can be recognised which begins with predominant cross-bedding and ends with predominant current ripple cross-lamination. The thick sandstone beds are more numerous in this facies than in the green mudstone and cross-laminated sandstone facies.

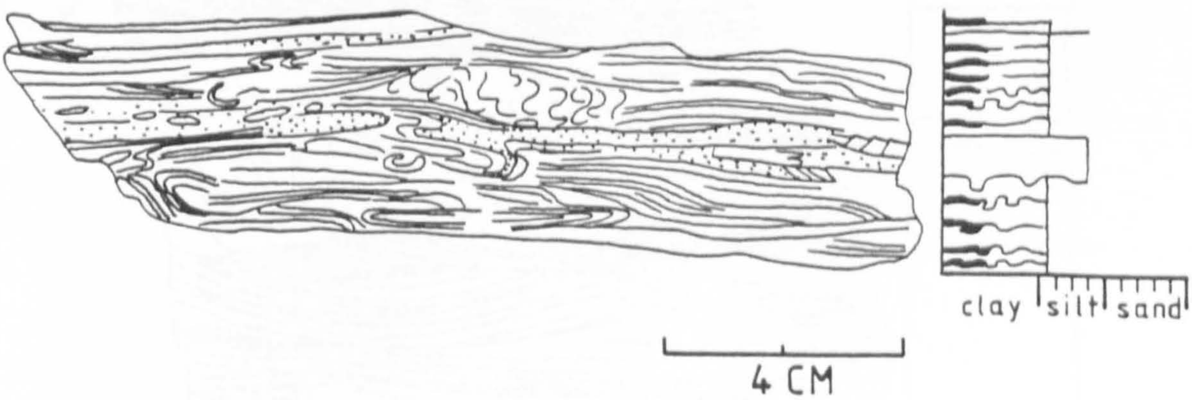
Thin sandstone beds

These beds are similar to those in the green mudstone and cross-laminated sandstone facies. They range in thickness from c. 10cm to c. 70cm. Many of the thinner beds show a sequence of sedimentary structures in addition to a normally graded profile (fig. 81). Current ripple cross-lamination with unidirectional palaeocurrents is common. The thin sandstone beds are more numerous in this facies than in the green mudstone and cross-laminated sandstone facies.

Mudstones

The mudstone is similar to that in the green mudstone and cross-laminated sandstone facies. However, the mudstone is generally coarser, being of medium to coarse silt grade and it contains more numerous, very thin beds and thick laminae of very fine grained sandstone. There is a lack of wavy and non-parallel bundles of laminae in this facies, which are common towards the top

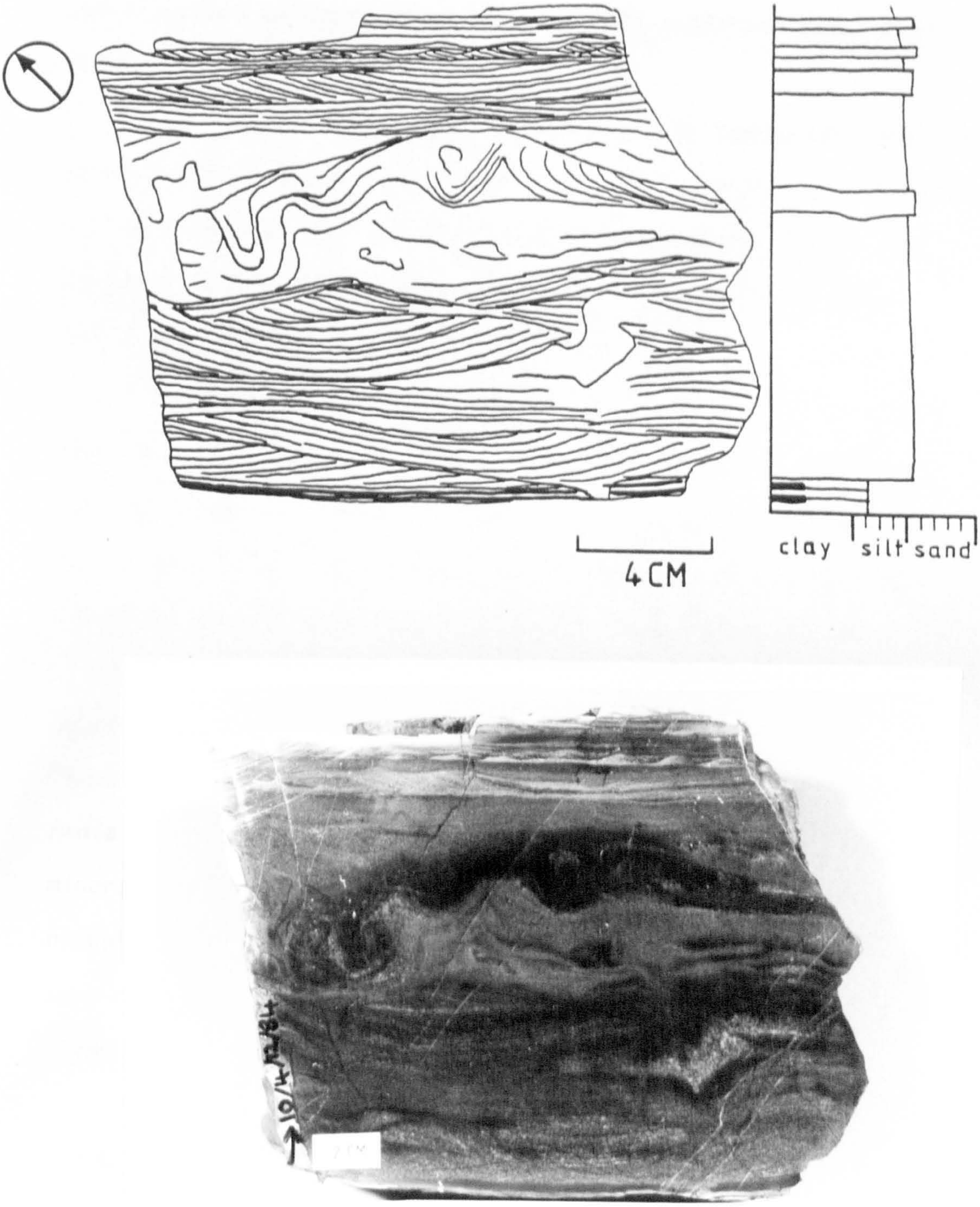
FIG:80 SLAB: MUDSTONE FROM THE GREEN MUDSTONE AND THICK-BEDDED SANDSTONE FACIES AND PHOTOGRAPH SHOWING DEFORMED CROSS-BEDDING FROM A THICK SANDSTONE BED IN THIS FACIES.



SLAB: Note the contorted and overturned lamination in the very fine-grained siltstone and the pinch and swell of the coarser siltstone lamina, together with small injection features and microfaults.

PHOTO illustrates a deformed thick set of cross-bedding in a loose block from a thick bed of sandstone. Length of hammer is 28cm.

FIG: 81 THIN SANDSTONE BED FROM THE GREEN MUDSTONE AND THICK-BEDDED SANDSTONE FACIES OF THE GREEN LIGHTSPOUT MEMBER OF THE SYNALDS FORMATION. SLAB .



Note the abrupt base;the upward-fining profile;the convolute lamination,probably caused by syndepositional loading and dewatering,in the centre of the bed and the laminated siltstone at the top of the bed.

of the green mudstone and cross-laminated sandstone facies. The mudstone laminae are often contorted and may be overfolded and the very thin sandstone beds and thick laminae may show pinch-and-swell and injection features (fig. 80), and small sandstone dykes may occur.

Distinctive, medium grey and very thinly laminated, very fine grained siltstone, which is organic rich and similar to that noted in the green mudstone and cross-laminated sandstone facies, was noted at SO 4342 9508 and SO 4286 9514. One example of this siltstone is shown in fig. 34, photo A.

Pseudo-ripples of tectonic origin are occasionally present in the mudstone. These are discussed in section 3.11.1. These may be distinguished from the sedimentary ripples by the orientation of their crests, which are coaxial with the fold axes and by their interlimb angles, which are often acute rather than obtuse.

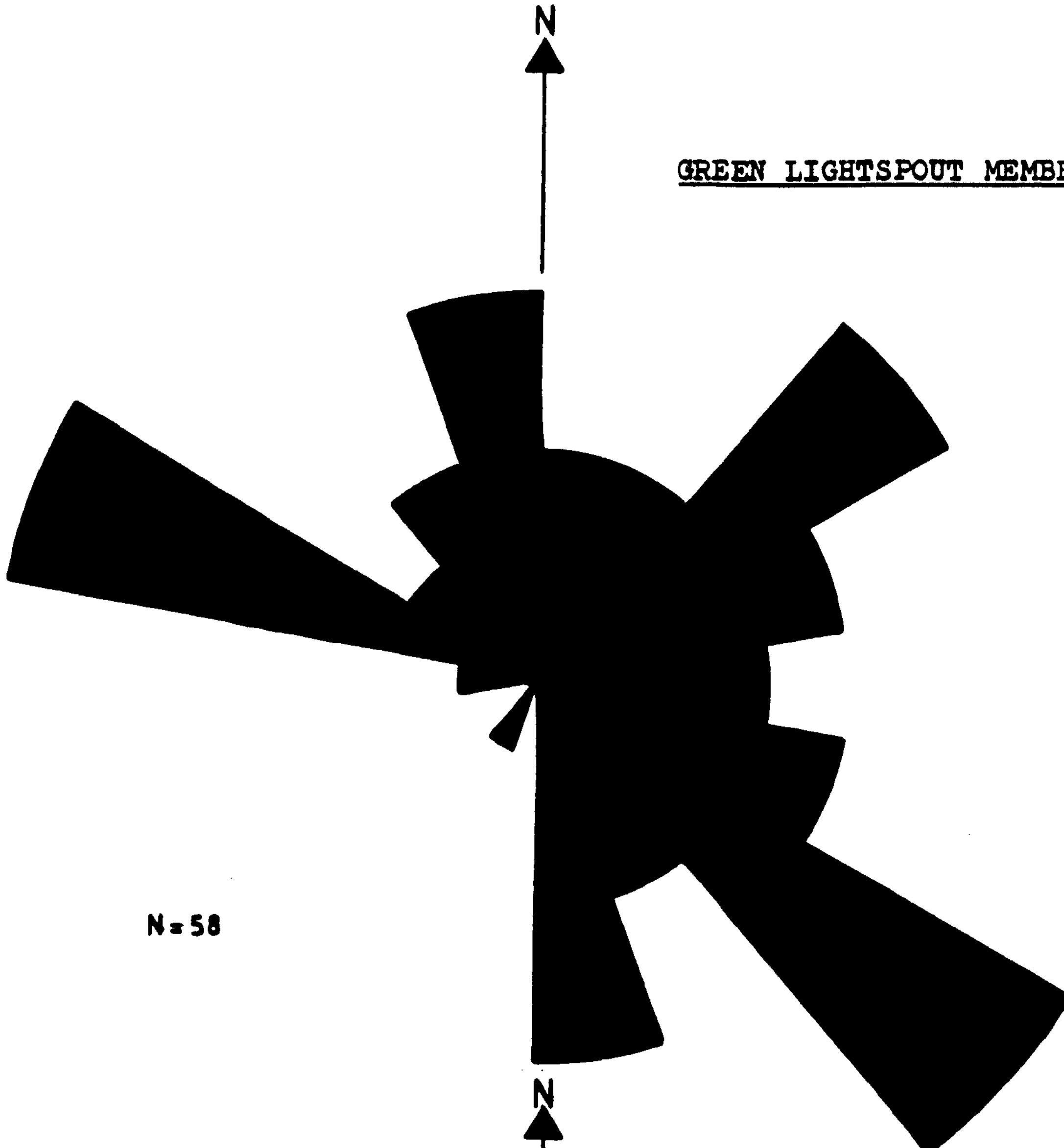
In some of the mudstones, irregular to subspherical, rounded patches and grains of medium to coarse sand grade are present. These commonly consist of anhedral, fine grained epidote and radiate chlorite. In the more subspherical patches, a mineralogical zonation is present. Radiate chlorite commonly occupies the core and epidote comprises the rim. These features were discussed in section 4.7 and it was concluded that they are diagenetic in origin.

Palaeocurrents

The palaeocurrents for the entire Green Lightspout Member of the Lightspout Formation are shown in fig. 82. The majority of the palaeocurrent readings were obtained from thick sandstone beds. From the limited number of readings obtained from the thin

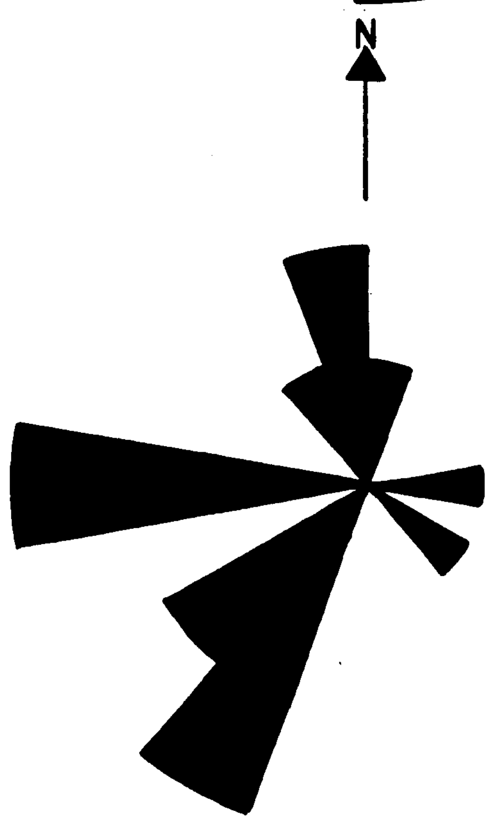
FIG:82 PALAEOCURRENT ROSES FOR THE ENTIRE GREEN LIGHTSPOUT MEMBER AND THE RED LIGHTSPOUT MEMBER.

GREEN LIGHTSPOUT MEMBER



N=58

RED LIGHTSPOUT MEMBER



N = 14

sandstone beds and from the mudstone lithologies, no differentiation of palaeocurrent directions could be made on the basis of different lithologies. The average palaeocurrent direction appears to be to the north-east. There is a wide scatter of the palaeocurrent directions, which may be explained by the tectonic deformation of the member. There are abundant moderately plunging, minor folds. In order to resolve the palaeocurrent direction, way-up had to be established, together with bed dip and a value for the local fold plunge, which was normally derived from the bedding/cleavage intersection lineation. The resolution of this data involved two stereographic rotations. Errors in the parameters used may explain the wide scatter of the palaeocurrent directions. Alternatively, the distribution may reflect a real scatter of the initial palaeocurrent directions.

Tuffs

A 1m thick tuff bed occurs within mudstones at SO 4302 9505 (at c. 2m on log 29, in appendix). This bed is normally graded from a poorly sorted crystal/lithic tuff with 15% lapilli to a fine ash tuff with pumice fragments. The lapilli and other fragments within the fine ash groundmass consist mainly of clusters of subhedral plagioclase crystals with adhering glass, together with glassy volcanic rocks, which are non-vesicular and which occasionally contain fine feldspar crystals. Towards the centre of the bed, armoured lapilli occur, which consist of clusters of plagioclase laths with rims of fine ash. These occur together with accretionary lapilli. The latter are shown in fig. 32, photo B. The majority of the bed lacks sedimentary structures, but the base is planar, parallel laminated.

Several thin beds in the region of the Lightspout Waterfall in Lightspout Hollow (at SO 4306 9507) may be interpreted as pyroclastic deposits or as slightly reworked pyroclastic deposits. They may be distinguished from the associated sediments by their coarser grain size, being medium to coarse grained, rather than very fine to fine grained, and by the greater abundance of subhedral to euhedral feldspar crystals, which in places may form up to 75% of the rock. The components of these beds are glassy volcanic rock fragments, which are non-vesicular, plagioclase crystals and epidote. These components are similar to those of the typical epiclastic (s.l.) deposits of the Lightspout Formation. Some of these beds are shown on log 27 (in appendix) at c. 5m and c. 6m. They may be normally graded and are either structureless or are planar, parallel laminated.

Interpretation

This facies is similar to the green mudstone and cross-laminated sandstone facies of the Green Synalds Member and the interpretations are therefore generally similar.

The thick-bedded sandstones are interpreted as fluvial channel-fills. The thin sandstone beds are interpreted as sheetflood deposits on an alluvial floodplain and the mudstones are interpreted as having been deposited from sheetflows on an alluvial floodplain.

The thick-bedded sandstones have an average north-easterly palaeocurrent direction, which is similar to the ENE direction obtained from the fluvial channel sandstones of the Cardingmill Grit. These sandstones are therefore considered to be part of the same fluvial system which deposited the Cardingmill Grit and the thick beds of sandstone at the base of the Green Synalds Member. The deformed cross-bedding sets within the thick sandstone beds are similar to those described by Allen and Banks (1972) and could be the result of liquefaction due to earthquakes in combination with current drag (Allen and Banks, 1972).

In comparison with the green mudstone and cross-laminated sandstone facies, the greater abundance of thick beds of sandstone in this facies suggests that channels were more abundant. Also, the coarser grain size of the associated mudstones and the more numerous thin sandstone beds reflect the greater abundance of these channels and hence the more proximal character of the alluvial floodplain with respect to the channel networks.

The medium grey, very thinly laminated, organic rich, very fine grained siltstones are interpreted as having been deposited from suspension in shallow, standing bodies of water, as discussed in section 6.12.

The thin tuff bed at SO 4302 9505 (at c. 2m on log 29, in appendix) is moderately sorted and normally graded. These features, together with the occurrence of accretionary lapilli suggest that this tuff is an air-fall deposit. The presence of armoured lapilli is indicative of hydroclastic eruptions (Fisher and Schmincke, 1984, p. 91). The presence of accretionary lapilli also suggests that the ash cloud was water rich. This water was probably generated by magma/water interaction during eruption.

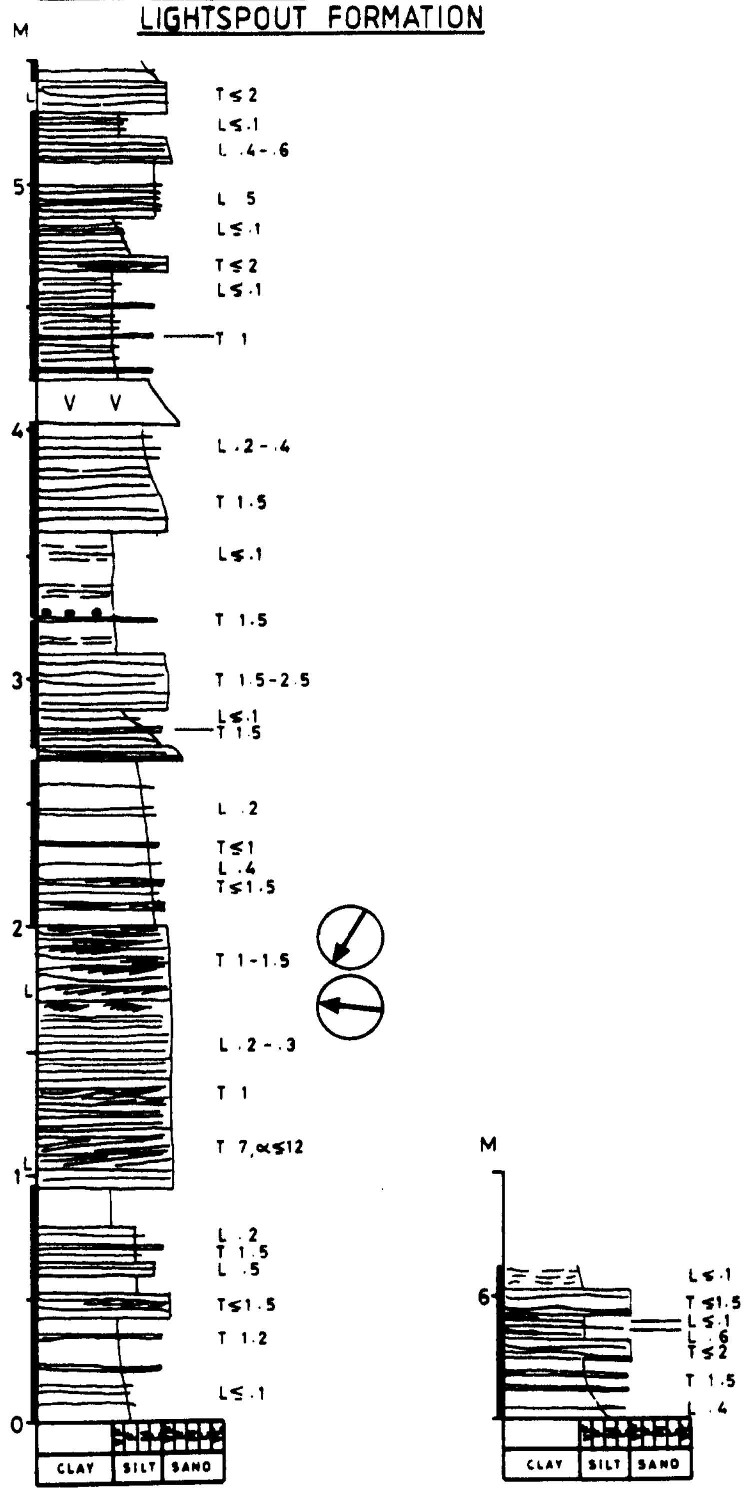
The thin crystal/lithic tuffs in the region of the Lightspout Waterfall in Lightspout Hollow (at SO 43069507) are moderately well sorted and normally graded. These features suggest that these tuffs are air-fall deposits. They are similar in character to the crystal/lithic-rich, basal parts of the tuff bed at SO 43029505. However, they lack the fine ash and accretionary lapilli which are present in the latter.

6.15 THE RED MUDSTONE AND CROSS-BEDDED SANDSTONE FACIES

The Red Lightspout Member of the Lightspout Formation is composed of this facies. This member is approximately 112m thick. It is underlain by the green mudstone and thick-bedded sandstone facies of the Green Lightspout Member and is overlain by the homogeneous, cross-bedded sandstone facies of the Huckster Conglomerate Member of the Portway Formation.

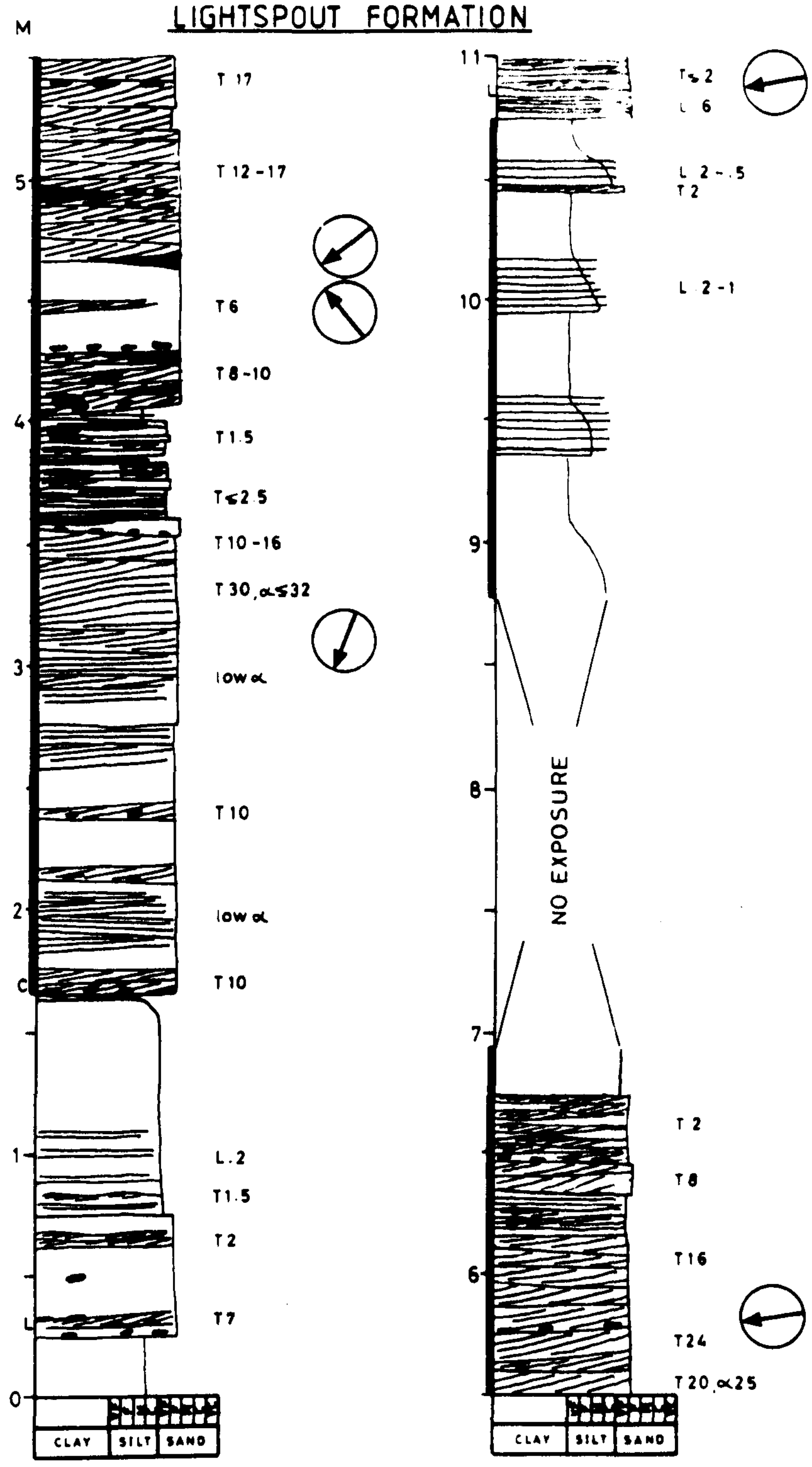
This facies is similar to the red mudstone and cross-laminated sandstone facies of the Synalds and Portway Formations, but differs from them in containing common, thick beds of cross-bedded sandstone. It is also similar to the green mudstone and thick-bedded sandstone facies. However, all of the lithologies in this facies are purplish red rather than greenish grey. In addition, the cross-bedded sandstones in this facies can be differentiated from the thick-bedded sandstones in the green mudstone and thick-bedded sandstone facies by the abundance of red mudstone clasts, their coarser grain size, the occurrence of basal erosion surfaces, and the westerly palaeocurrent directions, which differ markedly from the north-easterly palaeocurrent directions in the Green Lightspout Member (fig. 82). This facies is illustrated by log 32, fig. 83 and log 34, figs. 84 and 85. There are three principal lithologies: laminated mudstones, thin beds of planar laminated and cross-laminated sandstone and thick beds of cross-bedded sandstone.

FIG.83 LOG 32 CARDINGMILL VALLEY. SO 43309552
LIGHTSPOUT FORMATION



This leg illustrates the red mudstone and cross-bedded sandstone facies from near the base of the Red Lightspout Member. Note the thick bed of upward-fining sandstone at c.1.5m, which is overlain by an upward-fining mudstone sequence. There are numerous, thin beds of cross-laminated sandstone, e.g. at 0.5m and at 6m. These lithologies are similar to the red mudstone and cross-laminated sandstone facies of the Red Synalds Member.

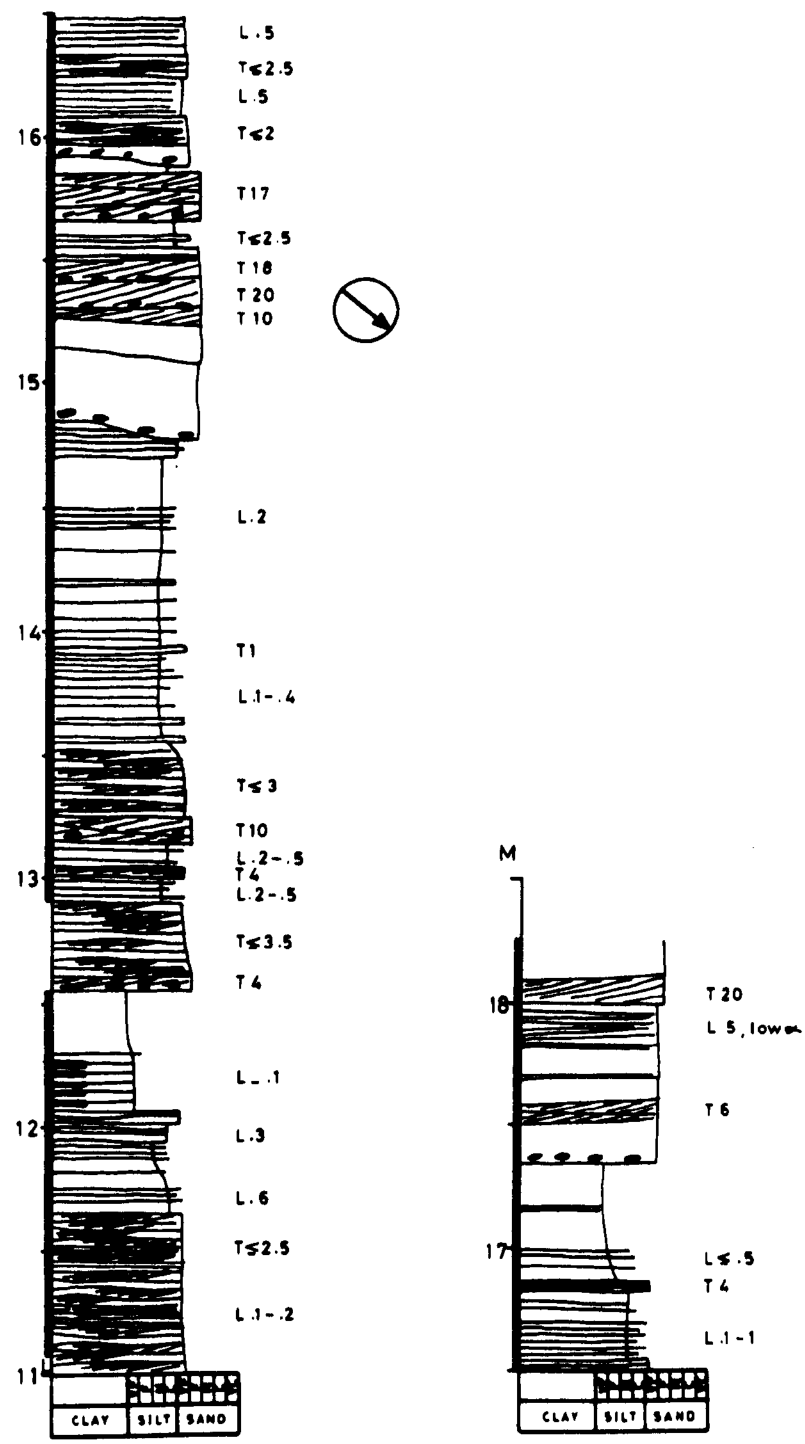
FIG.84 LOG 34 LIGHTSPOUT HOLLOW. SO 42669505
LIGHTSPOUT FORMATION



This log illustrates the red mudstone and cross-bedded sandstone facies from near the top of the Red Lightspout Member. Note the c.5.3m thick bed of trough cross-bedded sandstone which contains numerous mudstone clasts. Note the presence of mudstone lenticles, e.g. at c.5m and the upward-fining profile with thin siltstone beds between 3.5m and 4m. Laminated mudstones occur at c.1m and between 8.75m and 10.75m. This log is continued in fig. 85 on the next page.

FIG.85 LOG 34 (cont'd) LIGHTSPOUT HOLLOW SO 42669505

LIGHTSPOUT FORMATION



This log illustrates the red mudstone and cross-bedded sandstone facies from near the top of the Red Lightspout Member. Note the upward-fining and cross-bedded sandstone with an erosional base between 14.9m and 16.5m. Two, finer grained, thick beds of cross-laminated sandstone occur around 11.25m and 13m. Laminated mudstones form an upward-fining profile between 11.6m and 12.6m and also occur between 13.5m and 14.9m.

Laminated mudstones

These are similar to those in the red mudstone and cross-laminated sandstone facies of the Red Synalds Member. Soft-sediment deformation features are occasionally present (fig. 86).

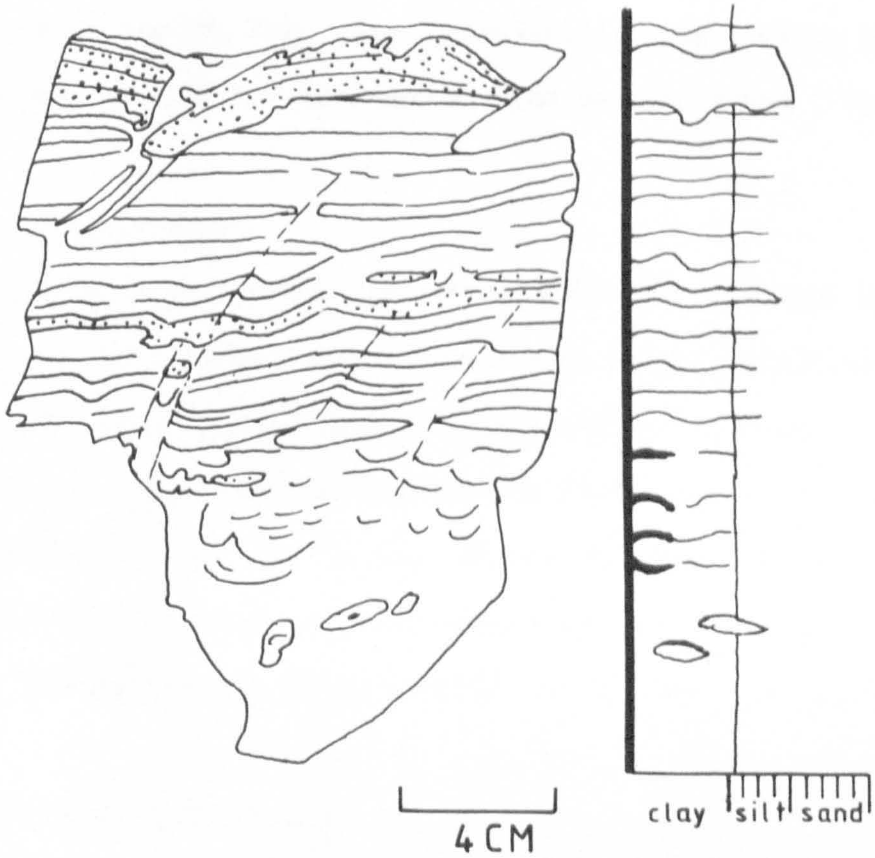
Thin beds of sandstone

These are commonly 6cm to 12cm thick and may be up to 22cm thick. The bases and tops of the beds are abrupt and the sandstone is either non-graded or normally graded. These sandstones are usually current ripple cross-laminated, but may be planar, parallel laminated.

Thick beds of sandstone

These are fine to occasionally medium grained. They are extensively trough cross-bedded with sets up to 24cm thick. Bed-thicknesses are usually in the range of 1m to 2m, however, thicker, homogeneous beds may occur, which are up to 5m thick. The beds may show a sequence of sedimentary structures from trough cross-bedding at the base to planar, parallel lamination and current ripple cross-lamination at the top (e.g. 15m to 16.5m on log 34, fig. 85). Occasionally, the beds may be predominantly planar, parallel laminated and ripple cross-laminated (e.g. bed at c. 1.5m on log 32, fig. 83). These latter beds are finer grained, being very fine to fine grained. The bases of the coarser beds often show erosional relief and contain concentrations of mudstone clasts. Clasts of purplish red mudstone are common and are commonly found at the set boundaries. In the thicker, homogeneous beds, there are occasional, thin beds of siltstone, which in places

FIG:86 SLAB: MUDSTONE FROM THE RED MUDSTONE AND CROSS-BEDDED SANDSTONE FACIES AND PHOTO ILLUSTRATING OVERTURNED LAMINATION IN A THICK SANDSTONE BED FROM THE SAME FACIES.



SLAB: Mudstone-note the thin laminae of coarser siltstone and the very thin bed of disturbed and laminated sandstone.

PHOTO:Overturned lamination in a thick sandstone bed.Scale bar is graduated in decimetres and centimetres.

occur within thin, upward-fining sandstone sequences (e.g. at c. 3.75m on log 34, fig. 84). Lenticles of mudstone, up to 5cm thick and 25cm wide, occasionally occur between the set boundaries (e.g. at c. 5m on log 34, fig. 84). Occasionally, the laminae in the cross-bedding sets are deformed and can be overturned (fig. 86).

Lithological trends

The base of the Red Lightspout Member is dominated by mudstones and thin sand beds, with minor, thick, very fine grained sandstone beds. These lithologies are similar to those of the red mudstone and cross-laminated sandstone facies and are illustrated by log 32, fig. 83. As the base of the Huckster Conglomerate is approached, thick beds of cross-bedded sandstone become common. These lithologies are illustrated by log 34, figs. 84 and 85.

Palaeocurrent directions

These are shown in fig. 82. All of the data presented on this palaeocurrent rose were obtained from the thick beds of sandstone. The average palaeocurrent direction appears to be approximately to the west. This direction is significantly different from the ENE and approximately NE palaeocurrent directions which were obtained from the thick beds of cross-bedded sandstone in the Cardingmill Grit and the Green Lightspout Member respectively. However, this direction is similar to the WNW direction obtained from the thin beds of cross-laminated sandstone in the Green Synalds Member and the Red Synalds Member. An average NNW palaeocurrent direction is shown by the Bayston-Oakswood

Formation, the homogeneous and cross-bedded sandstone facies of which is similar to some of the thick, cross-bedded sandstone beds (e.g. between c. 1.6m and c. 6.9m on log 34, fig. 84).

Interpretation

The mudstones are similar to those in the red mudstone and cross-laminated sandstone facies and they are similarly interpreted as having been deposited by overland sheetflow on an alluvial floodplain. The thin sandstone beds within the mudstones are similar to those within the red mudstone and cross-laminated sandstone facies and are interpreted as having been deposited by sheetflood events on the alluvial floodplain.

The thick beds of sandstone are interpreted as fluvial channel fills. Since the thick beds of sandstone are 1m to 2m thick, these channels were probably of this order of depth. The presence of mudstone wedges and thin siltstone beds in the thick (5m thick) sandstone bed of log 34, fig. 84 suggests that this bed was constructed by the amalgamation of a number of thinner sandstone beds. The palaeocurrent directions in the thick beds of sandstone are towards the west. This direction differs markedly from the NE to ENE directions obtained from the thick-bedded sandstones of the Green Lightspout Member and the Cardingmill Grit Member. It is similar to the NNW direction obtained from the homogeneous and cross-bedded sandstone facies. This fact, together with the observations that these thick beds of sandstone are more numerous towards the base of the homogeneous and cross-bedded sandstone of the Huckster Conglomerate Member and are similar to this, suggests that these thick beds of sandstone represent incursions of the braided fluvial system of the homogeneous and cross-bedded sandstone facies onto the alluvial floodplain. The overturned lamination illustrated in fig. 86 might have been produced by current drag in combination with liquefaction due to earthquake activity, as proposed for similar structures described by Allen and Banks (1972).

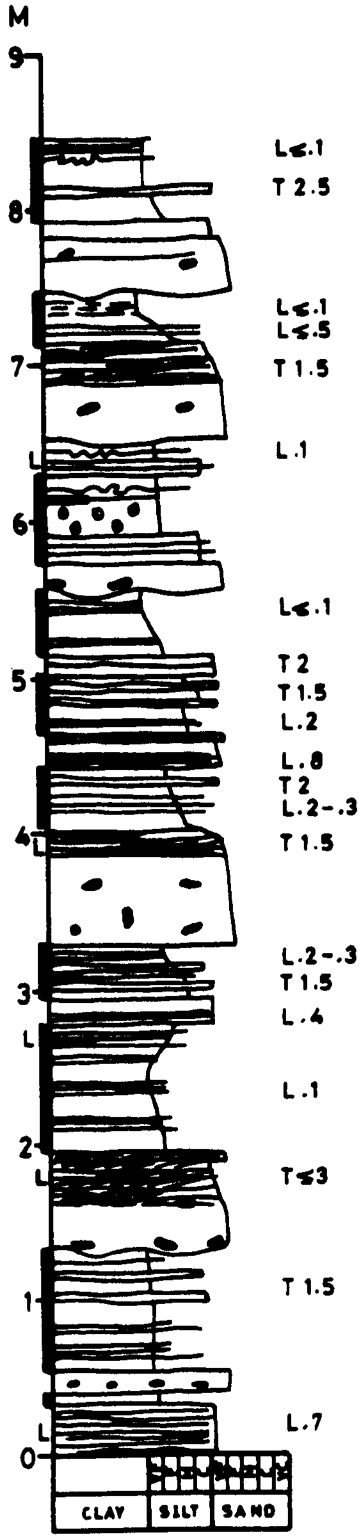
6.16 THE RED MUDSTONE AND CROSS-LAMINATED SANDSTONE FACIES OF THE PORTWAY FORMATION

The Portway Formation is predominantly composed of red mudstones and thin beds of cross-laminated sandstone. This facies is c. 643m thick and is underlain by the homogeneous and cross-bedded sandstone facies of the Huckster Conglomerate, which is the basal member of the Portway Formation, and is overlain by the homogeneous and cross-bedded sandstone facies of the Bayston-Oakswood Formation. The red mudstone and cross-laminated sandstone facies of the Portway Formation is similar to that of the Red Synalds Member of the Synalds Formation, however, several minor differences justify their separation. These differences are principally the slightly coarser grain size of the mudstones of the Portway Formation, the greater proportion of thin sandstone beds and the fine grained, rather than very fine grained nature of the latter. This facies is illustrated by log 36, fig. 87 and by log 38, fig. 88.

Thin sandstone beds

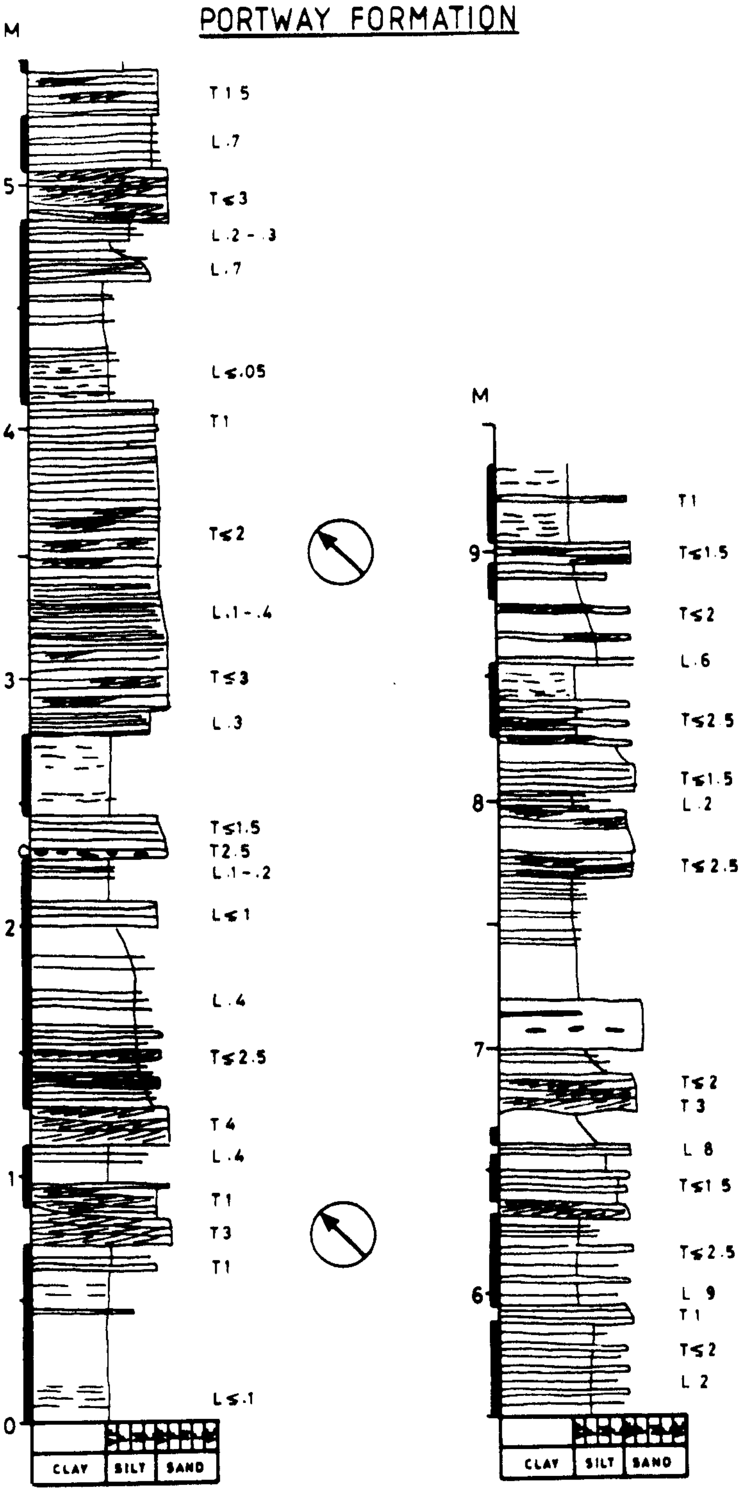
Thin sandstone beds are more abundant in the Portway Formation than in the Red Synalds Member and they commonly comprise 25% of the lithology. The majority of these beds are between 10cm and 25cm thick and the average thickness is c. 16cm. However, some

FIG.87 LOG 36 ASHES HOLLOW,SO 41979369
PORTWAY FORMATION



This log illustrates the red mudstone and cross-laminated sandstone facies of the Portway Formation.Note the common, thin beds of ripple cross-laminated sandstone with mudstone clasts and erosional bases.The bases of these appear to be structureless,but may be cross-bedded.Some upward-fining sequences occur in the mudstones (e.g. at c.5m).

FIG.88 LOG 38 HIGH PARK HOLLOW, SO 43959744
PORTWAY FORMATION



This log illustrates the red mudstone and cross-laminated sandstone facies of the Portway Formation. Note the common, thin beds of ripple cross-laminated sandstone (e.g. at 0.8m and at 6.8m). A thicker (1.2m thick) sandstone bed occurs at around 3.5m. Note the presence of numerous, very thin, ripple cross-laminated sandstone beds in the mudstones. Two upward-fining sequences occur in the mudstones between 1.3m and 2m and between 8.5m and 9.3m.

thicker beds occur which are up to 120cm thick. From the distribution of these bed thicknesses, it is not apparent that there are two distinct populations, rather, there appears to be a continuous variation in bed-thickness. All of the beds are predominantly fine grained and are thus coarser than the sandstones in the Synalds Formation, which are predominantly very fine grained. Unlike the thin sandstone beds in the Red Synalds Member, the bases of the beds may be irregular and erosional and this commonly occurs at the bases of the thicker beds. The relief shown by the base may be up to 10cm over short distances of c. 40cm. Clasts of red mudstone, which are usually tabular in shape, commonly occur towards the bases of the beds and may be scattered throughout the bed (fig. 89, slab B). The sequences of sedimentary structures are otherwise similar to those in the thin sandstone beds of the Red Synalds Member.

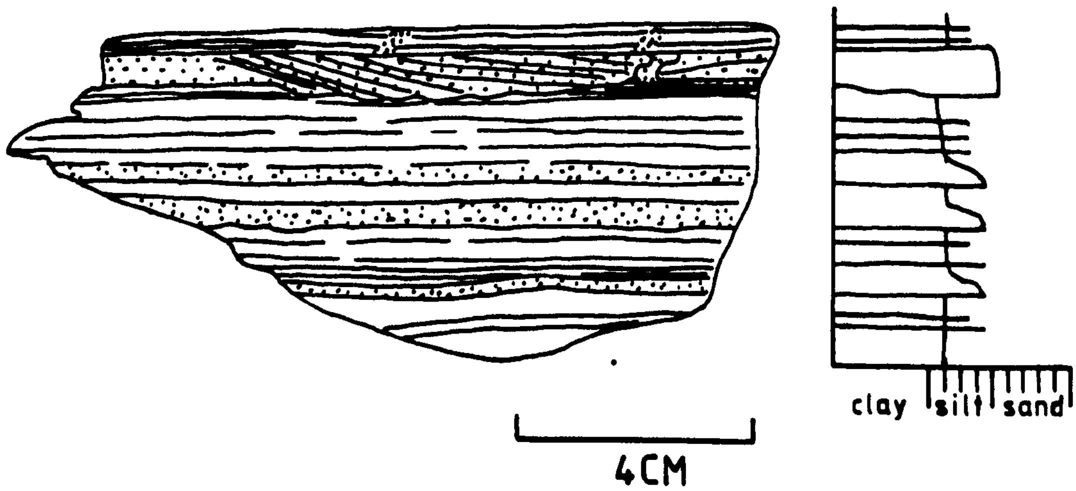
Thick sandstone beds

These are not common. They are up to c. 2.7m thick and are similar to the thick beds of sandstone in the Red Synalds Member, in their frequency and in the sequences of sedimentary structures which are developed.

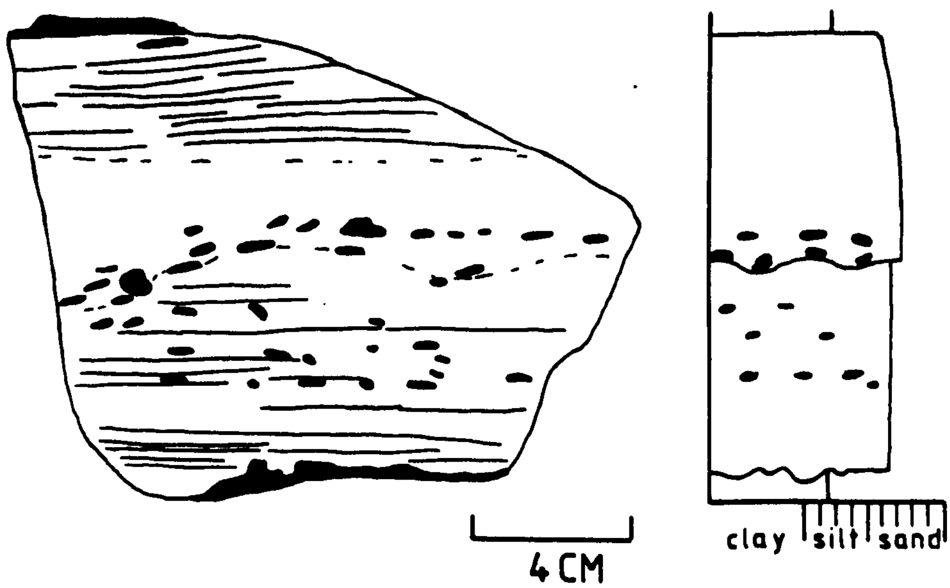
Mudstone

The mudstone is typically purplish red or purplish grey and, unlike the mudstone in the Red Synalds Member, it may be greenish grey. The average grain size is generally coarser than that of the mudstones in the Red Synalds Member, which are predominantly fine grained. Very thin beds and thick laminae of very fine grained sandstone are common and are more common in this facies than in the

FIG:89 <u>SLAB: A LAMINATED MUDSTONE AND SLAB: B THIN SANDSTONE BED FROM THE RED MUDSTONE AND CROSS-LAMINATED SANDSTONE FACIES OF THE PORTWAY FORMATION.</u>



SLAB: A Laminated mudstone. Note the upward-fining coarse siltstone laminae and the thinner laminae of medium to coarse siltstone. The latter are normally graded or non-graded and are 1 to 2mm thick. The thick lamina of sandstone is current ripple cross-laminated. Note the loaded base, small sandstone injection features and pull-apart structure.



SLAB: B Complete, thin sandstone bed. Note the loaded base, the ubiquitous mudstone clasts and the abrupt top. The bed is wholly planar and parallel laminated.

Red Synalds Member. In common with the Red Synalds Member, mudcracks occur, but these are infrequently found. A typical example of the mudstone is illustrated in fig. 89, slab A.

Unlike the Red Synalds Member, thick sequences of greenish grey laminated mudstones occur. In one of these sequences (illustrated by log 37, in appendix), two beds of medium grey, very thinly laminated and very fine grained siltstone occur, which are up to 1.5m thick. A photomicrograph of this lithology is shown in fig. 34, photo B. These mudstones contain disseminated organic matter and algal filaments and these are discussed in section 5.10.

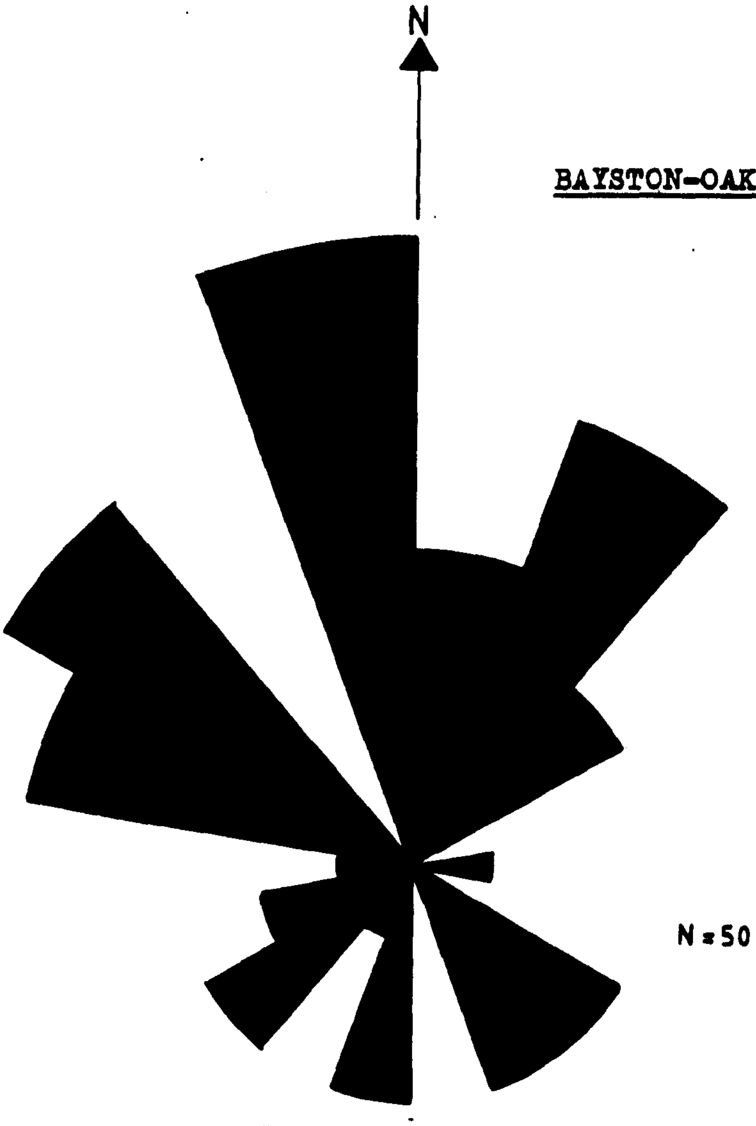
Some of the mudstones which overlie the thin sandstone beds have upward-fining trends. These upward-fining sequences are of the order of 75cm thick. Examples occur at c. 1.5m on log 38, fig. 88 and at c. 5m on log 36, fig. 87. These sequences can show an order of structures which is apparent in the sequence at c. 1.5m on log 38, fig. 88. Very thin beds of current ripple cross-laminated sandstone are concentrated towards the base of the sequence. As the grain size diminishes, the number of very thin sandstone beds decreases and the mudstone laminae decrease in thickness. The sequence ends in very thinly laminated, very fine grained siltstone. The laminae in this are subparallel and discontinuous.

Palaeocurrents

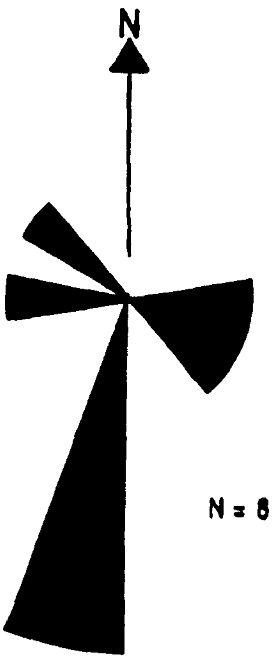
These are shown in fig. 90. These data were obtained from both the thin and thick sandstone beds, principally in Hawkham Hollow (SO 433 975) and in High Park Hollow (SO 438 973). The average palaeoflow appears to have been towards the south.

FIG:90 PALAEOCURRENT ROSES FOR THE BAYSTON-OAKSWOOD FORMATION (EXCEPTING BAYSTON HILL QUARRY) AND FOR THE PORTWAY FORMATION.

BAYSTON-OAKSWOOD FORMATION



PORTWAY FORMATION



However, the number of readings is small (only 8) and therefore the confidence in this direction, as representative of the Portway Formation, is low.

Interpretation

This facies is similar to the red mudstone and cross-laminated sandstone facies of the Red Synalds Member and the interpretations are therefore similar. The mudstones and thin sandstone beds are thought to have been deposited by sheetflows and sheetfloods on an alluvial floodplain. The thick sandstone beds are thought to be channel fills.

Compared to the Red Synalds Member, the generally coarser grain size of the mudstones and the thin sandstone beds, together with the greater frequency of the latter, suggest an increase in flow energy and the depositional environment of this facies appears to have been more proximal with respect to the source of the sediment. This increase in energy is also reflected in the common development of erosional bases to the thin sandstone beds and the common occurrence of rip-up clasts. The presence of erosional bases to the thin sandstone beds, together with the presence of upward -fining profiles in some of the overlying mudstones (e.g. at c. 1.5m on log 38, fig. 88), suggests that the sheetfloods and sheetflows might have been partially confined to broad channels. The upward -fining sequences in the mudstones may be the result of the gradual abandonment of these channels. Since the upward -fining profiles are only a few metres thick, then channel depths were probably of this order.

6.17 THE ALLUVIAL FLOODPLAIN FACIES ASSOCIATION

This association includes the green mudstone and cross-laminated sandstone facies of the Green Synalds Member, the red mudstone and cross-laminated sandstone facies of the Red Synalds Member and the Portway Formation, the green mudstone and thick-bedded sandstone facies of the Green Lightspout Member and the red mudstone and cross-bedded sandstone facies of the Red Lightspout Member. The thick sandstone beds in the Green Lightspout Member have north-easterly palaeocurrents. This direction is similar to the ENE direction which is displayed by the similar, thick-bedded and cross-stratified sandstones of the Cardingmill Grit. The thick sandstone beds in the lower parts of the Green Synalds Member may be considered to belong to this group of sandstones. In contrast, the thin and thick sandstone beds in the other facies have WNW or westerly palaeocurrent directions. Therefore, a distinction can be made between an ENE to NE flowing fluviatile system which is characterised by relatively deep channels and a W to WNW flowing system dominated by sheetflood and sheetflow processes. The former fluviatile system is associated with green mudstones and the latter system is characterised by red mudstones.

Interpretations of these systems depend upon the envisaged relationships between them and also their relationships with the braided fluvial system of the overlying Bayston-Oakswood Formation, which has a predominantly NNW palaeocurrent direction. One interpretation is that the W to WNW sheetflow and sheetflood system represents the distal facies of the braided fluvial system of the Bayston-Oakswood Formation and that this is distinct from the NE to ENE fluviatile system. An alternative interpretation is that the W to WNW system represents a lateral equivalent of the NE to ENE fluviatile system and has accumulated by the overbank flooding of the latter. A third interpretation is that both the NE to ENE fluviatile system and the W to WNW system are more "distal" equivalents of the braided fluvial system and that their differences are due to either extra-basinal control, such as a change in climate or sediment supply rate or to intra-basinal control, such as changes in slope or the height of the water-table.

The interpretation that the W to WNW system represents a "distal" facies of the braided fluvial system is favoured for these reasons:

1. There is a large, obtuse angle between the W to WNW system and the NE to ENE system, which is unlikely to be the result of overbank flooding.

2. The W to WNW palaeocurrents could be the distal equivalents of the NNW palaeocurrents in the braided fluvial system, since they are similar in direction.

3. The Huckster Conglomerate, which is a representative of the braided fluvial system, is developed within the W to WNW system and the Red Lightspout Member represents a transitional facies between them. Therefore, these two systems appear to be related.

4. There is a progradation from the more "distal" facies of the Red Synalds Member to the more "proximal" facies of the Portway Formation. This progradation culminates in the braided fluvial deposits of the Bayston-Oakswood Formation and therefore these facies appear to be related.

5. The red siltstone and cross-stratified sandstone facies of the Bridges Formation is similar to the facies of the W to WNW system. The facies of the Bridges Formation is interpreted as a longitudinally distal equivalent of the braided, fluvial system.

A possibly similar, ancient depositional system is the Precambrian Belt Supergroup (Winston, 1978). Winston (1978) suggested that there is a downslope transition from shallow, braided streams to extensive mudflats with an absence of meandering streams in the Belt Supergroup. The downslope transition is accompanied by a decrease in grain size and a decrease in the thickness of the sedimentation units. Graham (1983) records a similar transition from coarse -grained fluvial deposits of a low-sinuosity channel system to voluminous siltstones and very fine grained sandstones with a sheet-like character, which were deposited by vertical accretion on broad, flat, alluvial plains. Graham (1983) refers this system to a fluvial distributary system of terminal fan type. Friend (1978) suggests that terminal fan models may be applicable to many ancient river systems and are characterised by a gradual transition from braided rivers to clay playas with a gradual decrease in river depth and flow strength. Schumm (1968) suggested that, with a lack of vegetation, great thicknesses of sheet-like deposits, together with varve-like flood deposits, commonly constituted vast alluvial plains. The Longmyndian alluvial floodplain facies association may be referred to these models. However, the ENE to NE flowing channel system is not explained by these models.

The ENE to NE flowing channel system is associated with green sediments and it is later argued (section 6.24) that the green colouration is due to a high water-table. The high position of the water-table is most likely due to the inferred low topographic position of the environment of deposition, since the green mudstone and cross-laminated sandstone facies lies stratigraphically above the subaqueous delta facies and stratigraphically below the red mudstone and cross-laminated sandstone facies. Therefore, both the green mudstone and cross-laminated sandstone facies and the green mudstone and thick-bedded sandstone facies are thought to have been deposited near to sea-level. The development of deep channels in this environment may be related to the inferred low slope and the high water-table, which probably made the channel banks more cohesive. Therefore, it is not necessary to postulate that these facies were developed as a result of changes in extra-basinal conditions.

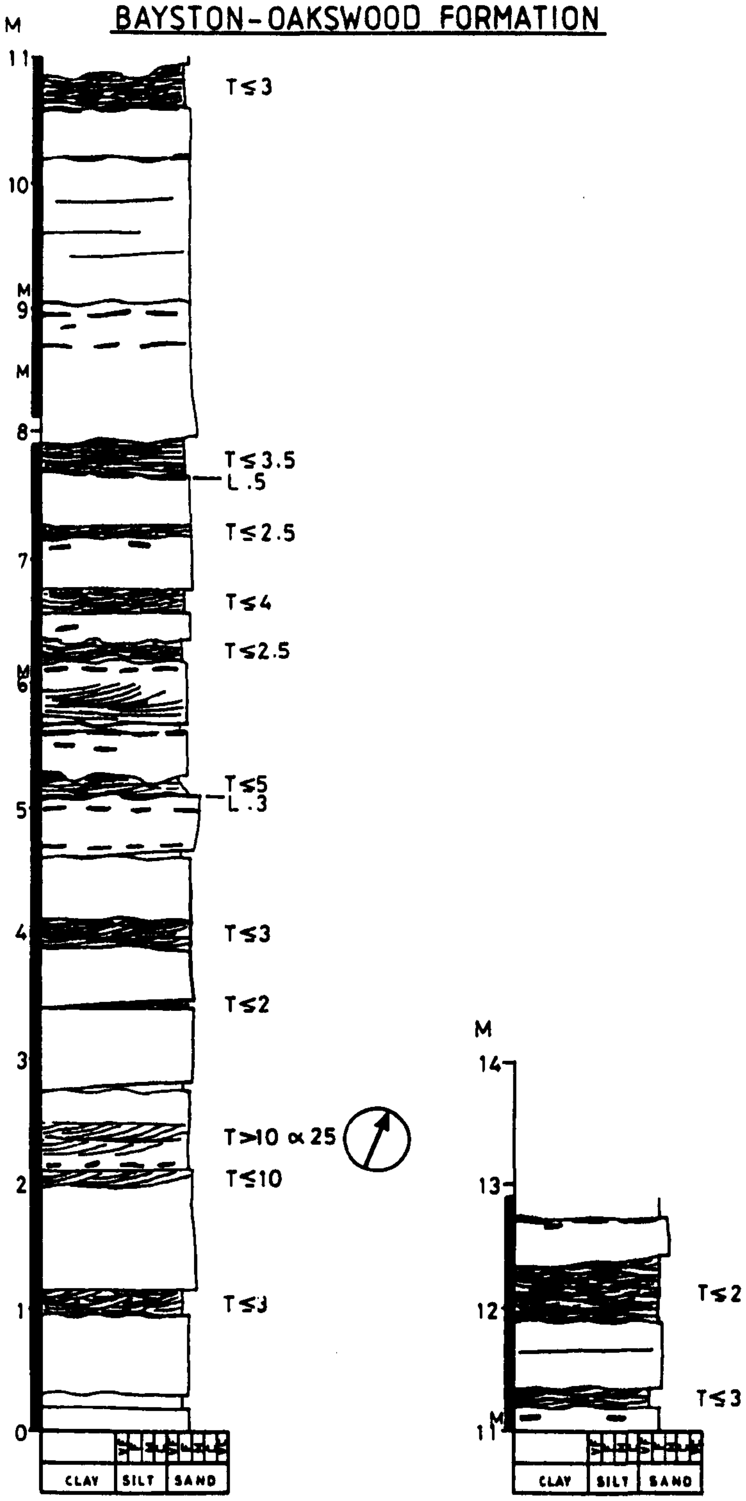
The high divergence in the palaeocurrent directions between the ENE to NE flowing channel system and the W to WNW alluvial system suggests that there are two separate systems: a longitudinal distributary system and a transverse, distal braidplain/floodplain.

The lack of pedogenic carbonate, the lack of evaporites and the lack of indications of aeolian activity, suggest that the Longmyndian alluvial floodplain facies were deposited whilst the climate was at least occasionally humid. The frequency of soft-sediment deformation features suggests that much of the sediment was water saturated. The abundant clay laminae probably prevented water-escape and pore-water overpressures might have been generated by rising groundwater or by compaction. Sandstone dykes and injection features, disrupted siltstone laminae and disturbed laminations in the sandstones were generated by water-escape consequent upon pore-water overpressure. It is probable that water-escape and sediment disruption might have been initiated by seismic activity. Similar water-escape and soft-sediment deformation features are described by Tunbridge (1984) from a Devonian, sandy ephemeral stream and clay playa complex.

6.18 THE HOMOGENEOUS AND CROSS-BEDDED SANDSTONE FACIES

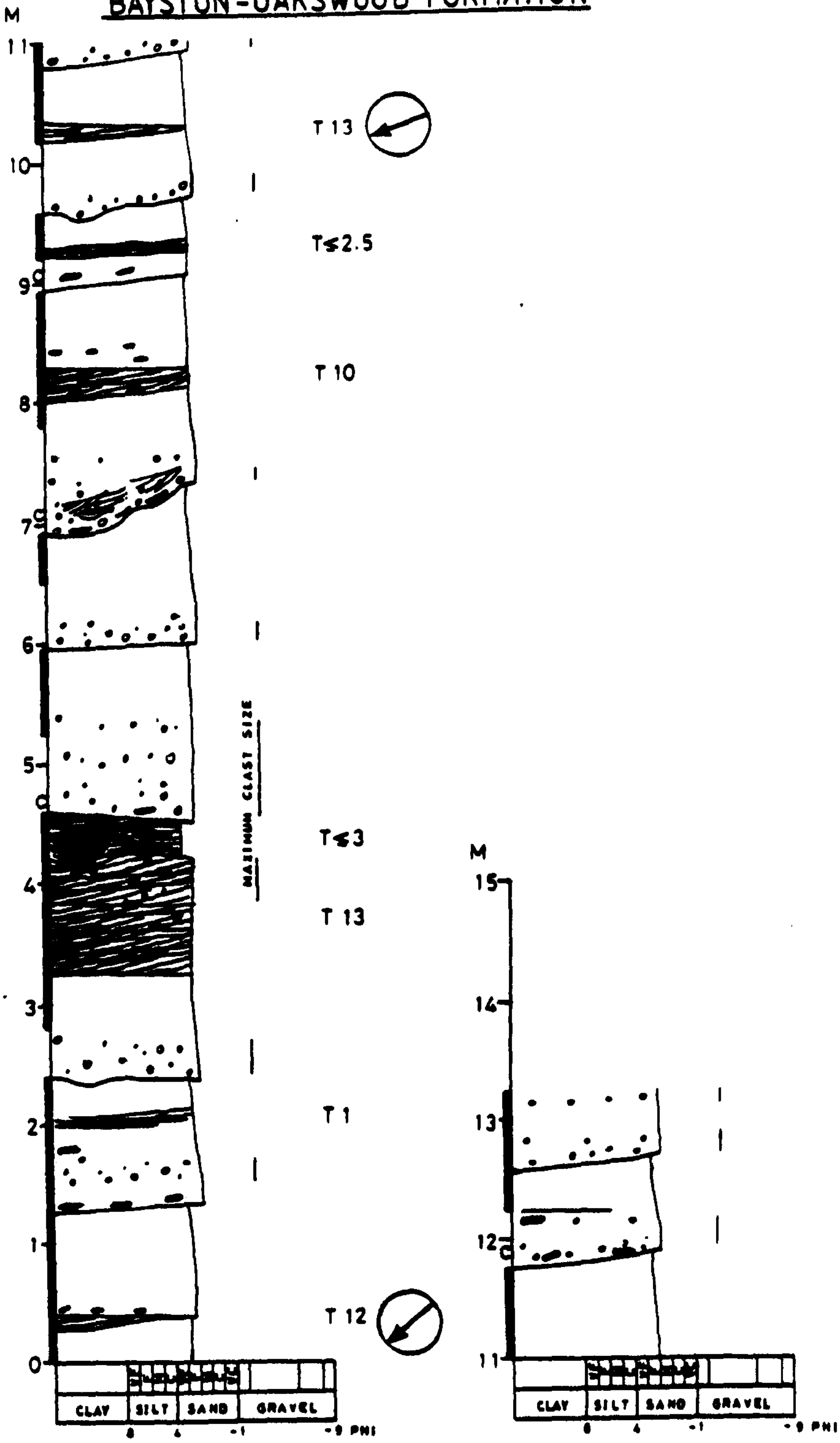
The Bayston-Oakswood Formation of the Wentnor Group is mostly composed of this facies. It overlies the red mudstone and cross-laminated sandstone facies of the Portway Formation and the contact is thought to be conformable, although previous authors have suggested that there is an unconformity. This is discussed in section 2.2. This facies is transitionally overlain by the red siltstone and cross-stratified sandstone facies of the Bridges Formation. Thick units of the conglomerate facies are interbedded with this facies and together these comprise the entire Bayston-Oakswood Formation, which is 2172m thick. The Huckster Conglomerate Member of the Portway Formation, which is 99m thick, is also composed of the homogeneous and cross-bedded sandstone facies. This transitionally overlies the red mudstone and cross-bedded sandstone facies of the Red Lightspout Member and is overlain by the red mudstone and cross-laminated sandstone facies of the Portway Formation. The Brokenstones sandstone beds are also mostly composed of the homogeneous and cross-bedded sandstone facies. This facies is illustrated by log 43, fig. 91 and log 40, fig. 92.

FIG. 91 LOG 43 LOWER DARNFORD, GRID REF. SO 41649781
BAYSTON-OAKSWOOD FORMATION



This log illustrates the homogeneous and cross-bedded sandstone facies of the Bayston-Oakswood Formation. Note the general homogeneity of the grain size. Some of the sandstone beds have slight upward-fining profiles which are c. 1m thick (e.g. at around 1.5m and around 8m). There are common thin cosets of trough cross-laminated very fine to fine grained sandstone which are abruptly overlain and underlain by trough cross-bedded sandstone. The upper contact is often erosional, resulting in lenticular beds. Note the presence of layers of mudstone flakes (e.g. at 5m) and occasional mudstone drapes. Some of the layers of mudstone flakes may be disrupted drapes (e.g. at 10.2m). The sandstone appears to be mostly structureless, but slabs show trough cross-bedding.

FIG. 92 LOG 40 HAWKHAM HOLLOW, GRID REF. SO 43279766
BAYSTON-OAKSWOOD FORMATION



This log illustrates a sandstone from near the base of the Bayston-Oakswood Formation, which is unusually rich in pebbles and granules. Note the presence of slight upward-fining profiles of the order of 1 to 2m thick. The bases of these are often erosional and the overlying sandstone contains pebbles, granules and mudstone flakes. Cosets of ripple cross-laminated sandstone are not common. The sandstone appears to be mostly structureless but slabs show trough cross-bedding.

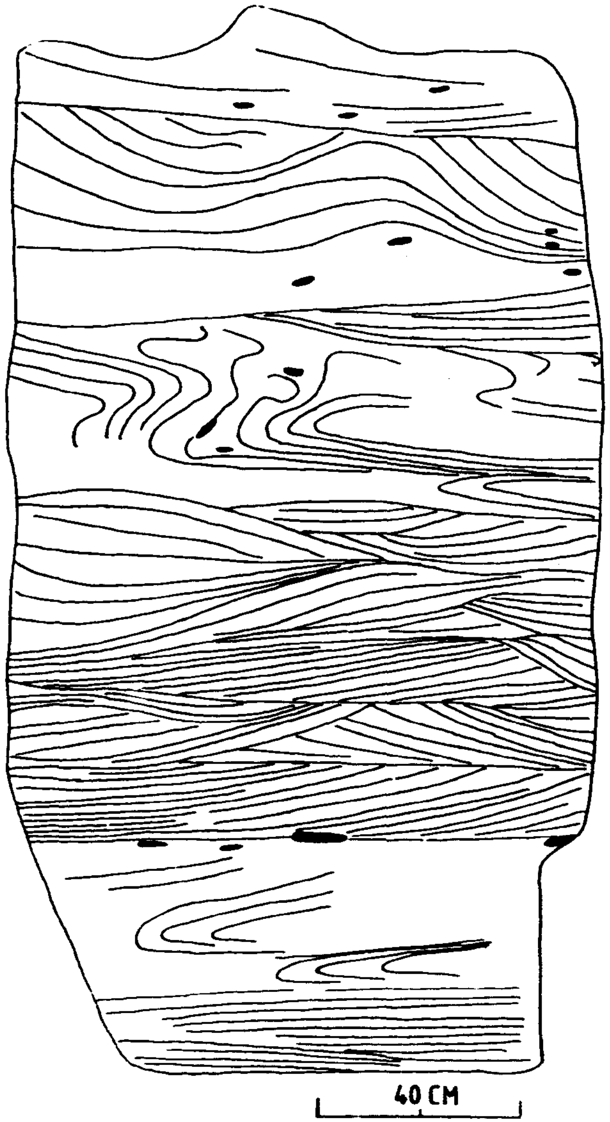
Lithology

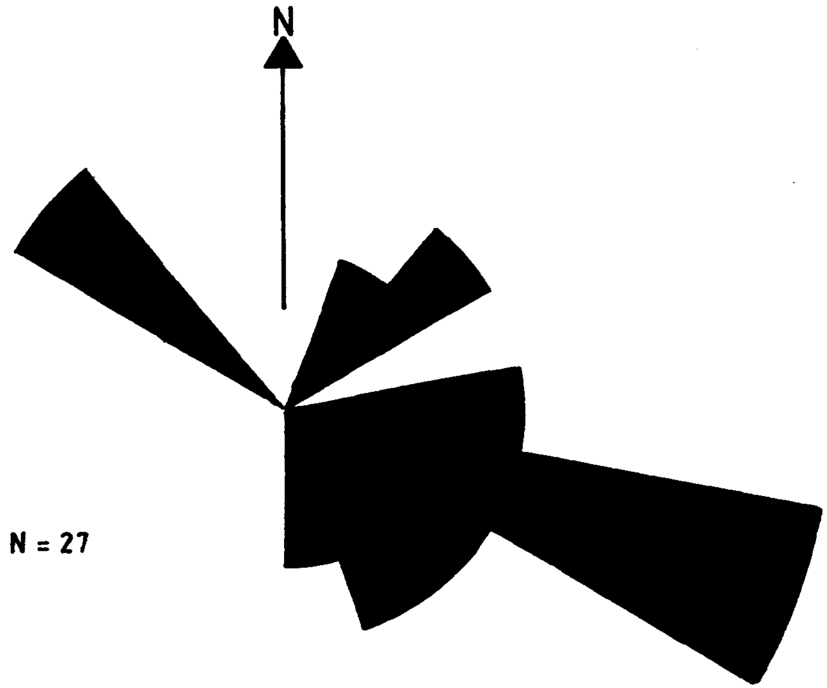
The sandstone is mostly fine to medium grained throughout. In the main outcrop, it is entirely purplish red, however in Haughmond quarry (SJ 542 146) and Bayston Hill quarry (SJ 494 092) it is mainly greenish grey. In places it is poorly sorted and contains granules and small pebbles. However, the majority of the sandstone is moderately well sorted. The grains are predominantly subrounded and the granules and small pebbles are mostly rounded. Pebbly sandstones are present towards the base of the Bayston-Oakswood Formation (log 40, fig. 92) and in the Huckster Conglomerate. The latter is conglomeratic in places.

Sedimentary structures

The most common type of sedimentary structure is trough cross-bedding, which occurs in sets which are 10cm to 15cm thick on average (fig. 93). The bases of the laminae are usually tangential and sigmoidal lamination is common. Planar, parallel lamination is very rare. Current ripple, trough cross-lamination is occasionally present and occurs in sets which are up to 5cm thick, which form thin cosets of very fine to fine grained sandstone. These cosets are 17cm thick on average. The current ripple cross-lamination is often climbing. Tabular cross-bedding occasionally occurs and this commonly forms sets which are c. 40cm thick. Where granules and small pebbles are present these usually occur at the bases of the cross-bedding sets. Deformed cross-bedding is fairly common and usually occurs in the thicker cross-bedding sets. The degree of deformation varies from oversteepened lamination to highly contorted and overturned lamination (fig. 93). The sense of overturning in adjacent sets is in the same direction. Rarely,

FIG: 93 TROUGH CROSS-BEDDING AND DEFORMED CROSS-BEDDING IN THE HOMOGENEOUS AND CROSS-BEDDED SANDSTONE FACIES OF THE BAYSTON-OAKSWOOD FORMATION AND PALAEOCURRENT ROSE FOR THIS FACIES IN BAYSTON HILL QUARRY.

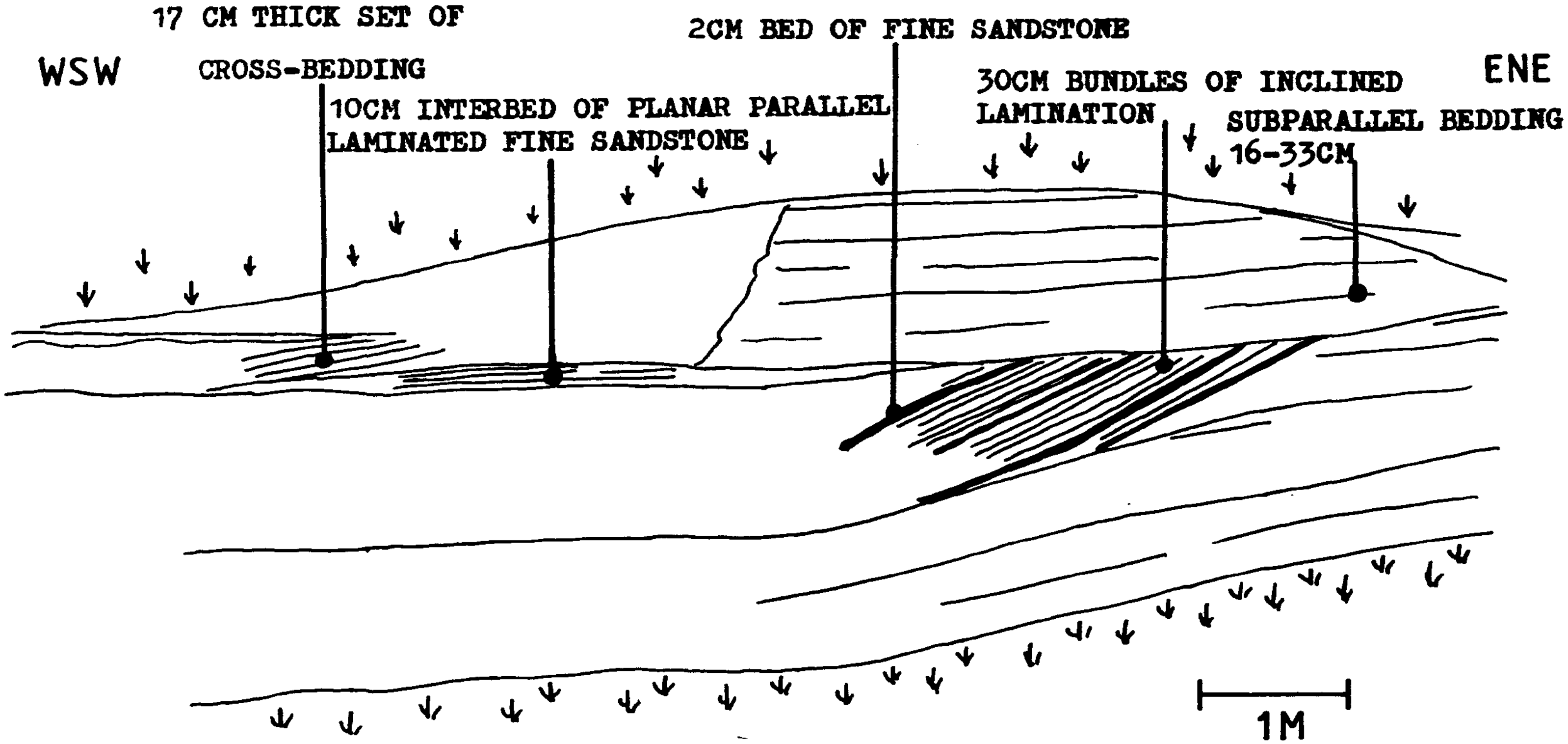




LEFT: Typical trough cross-bedding with some deformed cross-bedding sets. Homogeneous and cross-bedded sandstone facies of the Bayston-Oakswood Formation in Haughmond quarry.

ABOVE: Palaeocurrent rose for the homogeneous and cross-bedded sandstone facies of the Bayston-Oakswood Formation in Bayston Hill quarry.

FIG. 94 LARGE-SCALE CROSS-STRATIFICATION, BEACH FARM QUARRY, SO 35279426



NOTE THAT THIS EXPOSURE IS PROBABLY INVERTED larger scale cross-stratification, which may have formed on simple bars, can be recognised. This can be c. 1m thick (fig. 94) and can consist of gently inclined laminae. These occur in bundles 30cm thick, which are separated from each other by 2cm thick beds of finer grained sandstone. These beds are subparallel to the lamination in the bundles.

Mudstones

These occur most commonly as rip-up clasts within the sandstone. The clasts occur towards the bases of the cross-bedding sets, where they are associated with granules and small pebbles (fig. 95, photo A). Mudstone rarely occurs as thin beds within the sandstones and these beds are usually wedge-shaped or they can be lenticular (fig. 95, photo B). In these cases, the mudstone is massive and poorly sorted and the overlying sandstones are erosional. In some instances, mudstone forms a thin drape which overlies cosets of trough cross-bedding and which can underlie cosets of ripple cross-lamination (fig. 91, log 43). These mudstone drapes can be mudcracked (fig. 96, photo A). Some bedding surfaces are strewn with large polygonal mudstone clasts, which are interpreted as disrupted mudcracked drapes (fig. 96, photo B).

Sequences of sedimentary structures

Sequences of structures are not easily determined, since the structures are not easily seen. From log 43, fig. 91, it is apparent that cosets of ripple cross-lamination in the finer grained sandstone are abruptly overlain by fine to medium grained sandstone and the contact is often erosional. This sandstone is trough cross-bedded and there are often abrupt upper contacts

FIG: 95 PHOTOGRAPHS SHOWING MUDSTONE CLASTS AND A LENTICULAR BED OF MUDSTONE IN THE HOMOGENEOUS AND CROSS-BEDDED SANDSTONE FACIES OF THE BAYSTON-OAKSWOOD FORMATION.

PHOTO A: Mudstone clasts which are concentrated at the bases of trough cross-bed sets. Scale bar is graduated in decimetres and centimetres.

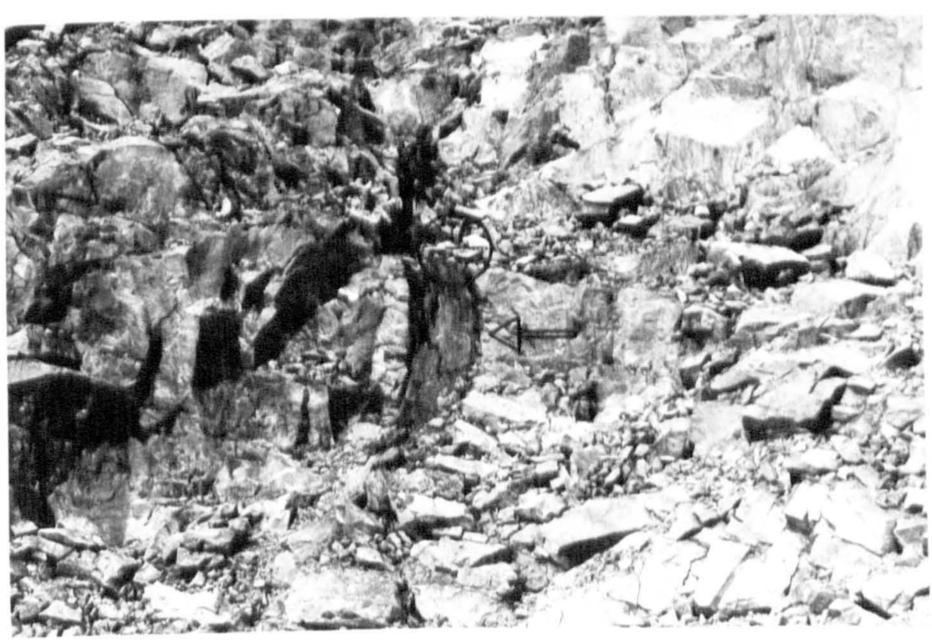


PHOTO B: Lenticular bed of mudstone in cross-stratified homogeneous sandstone. Scale bar (ringed) is graduated in decimetres and centimetres.

FIG: 96 PHOTOGRAPHS ILLUSTRATING MUDCRACKED DRAPES IN THE HOMOGENEOUS AND CROSS-BEDDED SANDSTONE FACIES OF THE BAYSTON-OAKSWOOD FORMATION.

PHOTO A: Mudcracked drape which overlies cross-stratified sandstone. Scale bar is graduated in decimetres and centimetres.

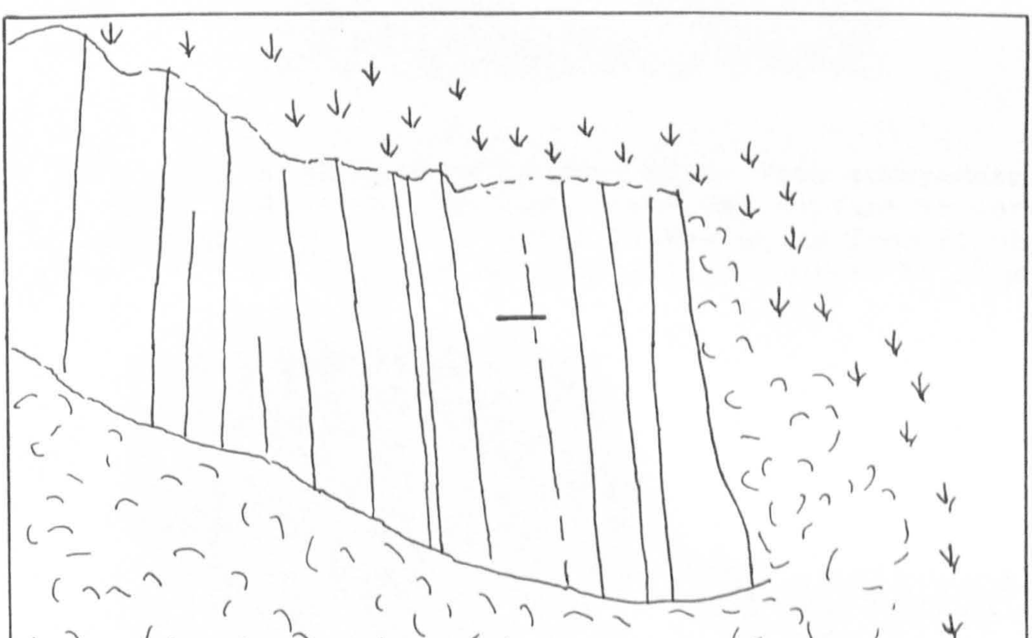
PHOTO B: Bedding plane which is strewn with polygonal mudstone clasts. These are interpreted as a disrupted mudcracked drape. Compass-clinometer is 10cm long.

between this and the overlying ripple cross-laminated sandstone. Occasionally, at this contact, there is a drape of mudstone which is 3mm to 5mm thick. The draped surface is often irregular and this, in some instances, may be due to preserved ripple forms. Some weakly defined upward-fining profiles, which are of the order of 1m thick, can occasionally be seen in the cross-bedded sandstones. From log 43, fig. 91 and log 44 (in appendix), the beds of cross-bedded sandstone are on average 55cm thick and range from 20cm up to 115cm thick. It appears that some of the beds are composite, since there are internal layers rich in mudstone clasts and granules and thin mudstone drapes occur which overlie ripple form sets. In some of the pebbly and granule-rich horizons (e.g. log 40, fig. 92), upward-fining trends are often present, which begin with erosional bases and these are overlain by medium to coarse grained sandstone rich in pebbles, granules and mudstone clasts. These upward-fining units are on average 120cm thick and range from 70cm to 180cm in thickness.

Bedding plane features

The most characteristic feature of the bedding planes is that in small outcrops they appear to be mostly flat and they define slightly wedge-shaped to parallel-sided beds (fig. 97). In the large quarries at Haughmond (SJ 542 146) and Bayston Hill (SJ 494 092), extensive bedding planes are exposed, which reveal that the beds are slightly wedge-shaped and the bounding surfaces are generally subparallel. Mudstone drapes can be traced as straight planes over many metres in some instances. However, some of the bedding planes are gently undulating with subsymmetrical, rounded, shallow troughs and crests (fig. 98, photo A). These can be

FIG: 97 BEDDING PLANES IN THE HOMOGENEOUS AND CROSS-BEDDED SANDSTONE FACIES OF THE BAYSTON-OAKSWOOD FORMATION AT WESTCOTT, SJ 4037 0128



Note that the bedding planes are subparallel and slightly wedge shaped. The way up is from right to left. Scale bar (centre of the photograph and the sketch) is 30cm long. This exposure is represented by log 44 (in appendix).

FIG: 98 UNDULATING BEDDING PLANE COVERED IN LARGE RIPPLES IN THE HOMOGENEOUS AND CROSS-BEDDED SANDSTONE FACIES OF THE BAYSTON-OAKSWOOD FORMATION, BAYSTON HILL QUARRY, SJ 494092.

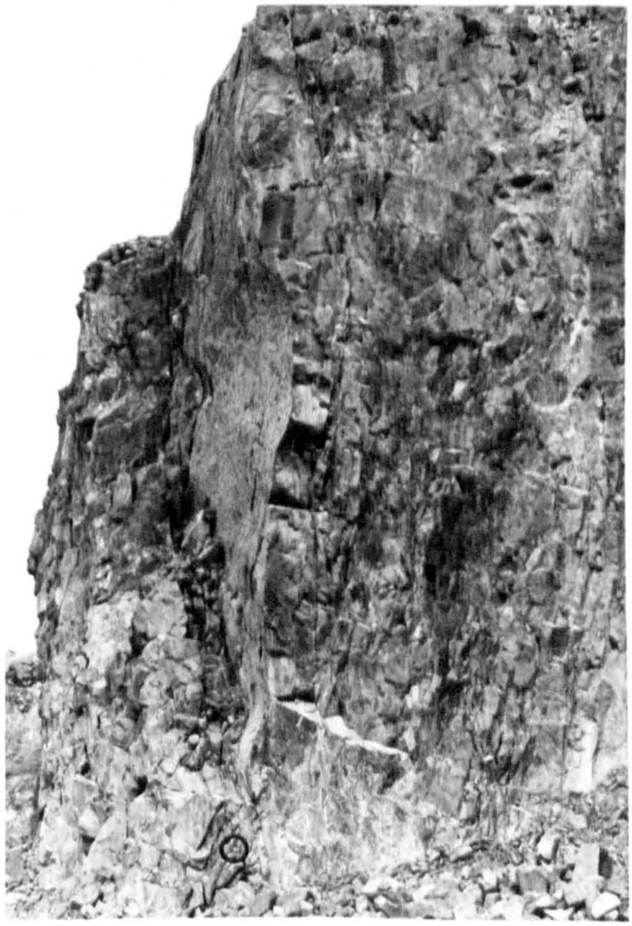


PHOTO A: Gently undulating bedding plane with subsymmetrical, rounded shallow troughs and crests. The surface is covered with large ripples (see photo B). Way up is from right to left. Hammer (ringed at bottom left of centre) is 28cm long.

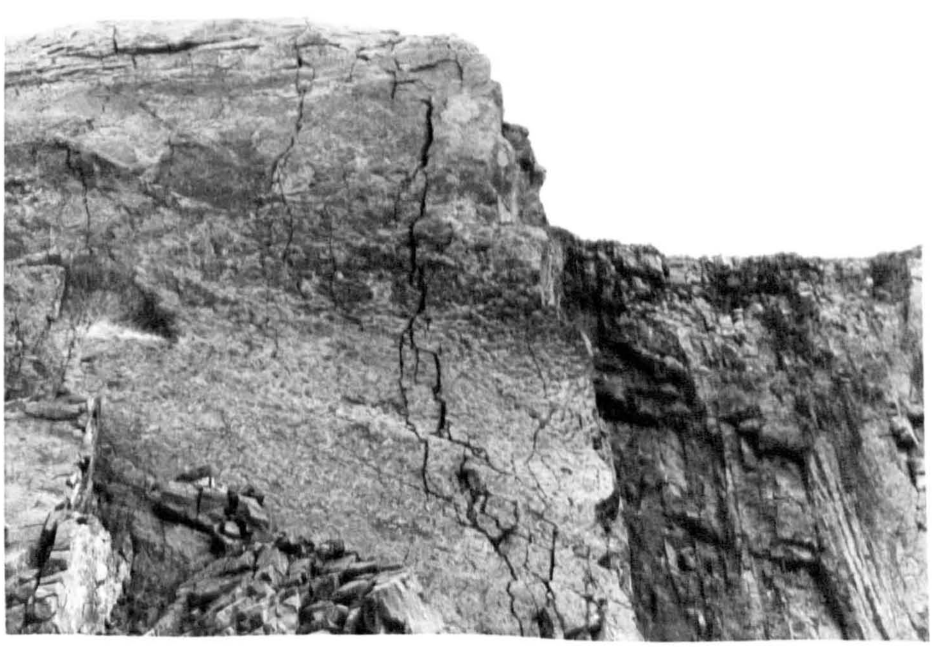


PHOTO B: Shows the top of the bedding plane shown in photo A.

FIG:99 PHOTOGRAPHS SHOWING MEGARIPPLES AND LINGUOID RIPPLES IN THE HOMOGENEOUS AND CROSS-BEDDED SANDSTONE FACIES OF THE BAYSTON-OAKSWOOD FORMATION.

PHOTO A: Undulating bedding plane which is covered with small megaripples,top surface of bed.Hammer (ringed,bottom centre of photo) is 28 cm long.Bayston Hill Quarry SJ 494092.

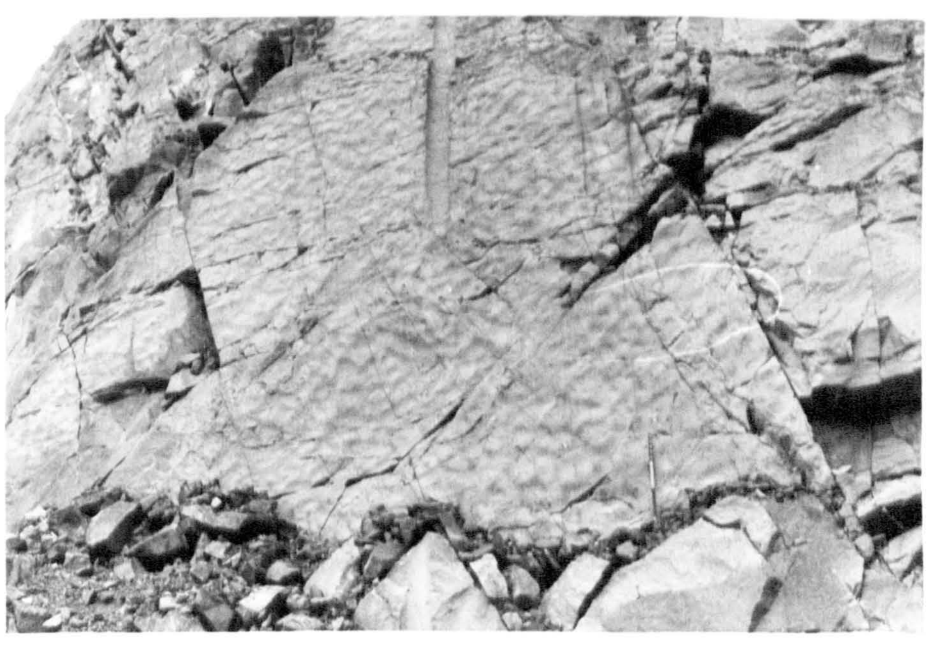


PHOTO B: Linguoid ripples covering flat bedding plane,top surface of bed.Scale (bottom right) is graduated in decimetres and centimetres.Haughmond Quarry SJ 542146.

covered with large ripples (fig. 98, photo B) or with sinuous crested megaripples (fig. 99, photo A). Rarely, almost flat bedding planes occur, which are covered with linguoid, small, current ripples (fig. 99, photo B).

Palaeocurrent data

All of the data from all of the palaeocurrent indicators, excepting the data from Bayston Hill quarry (SJ 494 092), are shown in fig. 90. The palaeocurrent rose shows that the palaeoflow was unidirectional, with an average flow towards the NNW. The data from Bayston Hill quarry (SJ 494 092) are anomalous, in that the unidirectional palaeoflow appears to have been towards the ESE (fig. 93). It is suggested that this is the result of a tectonic rotation between the outcrop at Bayston Hill and the main outcrop of the Long Mynd. The strike at Bayston Hill is subparallel to the adjacent major faults and is rotated 35° clockwise with respect to the strike of the main Long Mynd outcrop (geological map 2; in cover pocket). Although this rotation accounts for some of the apparent discrepancy in the palaeoflow, it does not account for all of it and this suggests that either there has been some rotation of the outcrops prior to, or during folding,or that there is a real difference in the palaeoflow.

Interpretation

The palaeocurrents are unidirectional and this suggests that this facies is wholly fluvial in origin. Deposition was entirely by traction, mostly during the migration of sinuous-crested megaripples and large ripples, which produced trough cross-bedding. In many instances, the flow decelerated rapidly and deposited thin cosets of current ripple cross-laminated finer sand. In many instances, settling from suspension was rapid, following the deceleration of the flow and climbing ripples were generated. The presence of finer grain sizes in the flows is implied by the occasional presence of mudstone beds and drapes, which were deposited following the very rapid deceleration of the flow. Mudstone and very fine grained sandstone is interpreted to have been mostly flushed away from the site of deposition. The presence of mudcracked drapes, disrupted mudcracked drapes and polygonal, mudstone rip-up clasts implies that the deposits were frequently subaerially exposed following flow deceleration. This suggests that the flow-depths were shallow. The deformed cross-bedding sets are similar to those described by Allen and Banks (1972) and could be the result of liquefaction due to earthquakes, in combination with current drag (Allen and Banks, 1972).

The lack of deeply eroded channels, the organisation of the beds into subparallel sheets and the presence of only shallowly incised, minor erosion surfaces beneath many of the sandstone beds, suggest that the flow occurred as widespread sheets that occupied very broad, shallowly incised channels in places. Undulating bed surfaces of little relief, which are covered in megaripples and large ripples, together with the rare presence of gently inclined sets of cross-bedding suggest that larger scale bedforms, possibly referable to sandwaves were generated. One type of larger scale bedform, probably referable to a simple bar is illustrated by fig. 94. Inclined surfaces with finer grained sandstone are interpreted as reactivation surfaces and these suggest that this bar periodically migrated during higher flood stages to generate 30cm thick sets of inclined parallel lamination.

Approximate water depths can be estimated from the thicknesses of the upward-fining units and from the height of the bar forms. The upward-fining units have a maximum thickness of 115cm to 180cm and the bar forms are up to 1m in height. This suggests that water depths were of the order of a few metres.

The sheet-like nature of the beds, the shallowly incised bases to these sheets, the great thickness of homogeneous sandstone, the inferred shallow water depths and the evidence for frequent emergence suggest that this facies was deposited by braided channels of low sinuosity. Since there is no evidence for lateral facies changes within these deposits, it appears that these braided channels were of a similar character over a wide area and a braidplain environment is thought to be the most likely depositional environment for this reason.

This facies is comparable with the sheet-braided fluvial deposits of Cotter (1978). The common characters are: the lack of major scour and fill channels, the sheet-like nature of the beds and the poor development of cyclical vertical sequences. Cotter (1978) noted that this sheet-braided style of deposition is characteristic of older Palaeozoic rocks and appears to be related to a lack of vegetation.

Several types of sandy braided river facies have been recognised by Miall (1978 and 1982). The Longmyndian homogeneous and cross-bedded sandstone facies differs from the Saskatchewan-type (Miall, 1978, 1982 and Cant, 1978) in that it lacks clearly defined, cyclic, vertical sequences of grain size and sedimentary structures. It also differs from the Bijou Creek-type, which is dominated by horizontally laminated and low angle cross-laminated sandstone. It resembles the Platte-type in that there is a lack of cyclicity and a predominance of cross-bedding, but differs from it, in that trough cross-bedding, rather than tabular cross-bedding, is predominant. This difference suggests that the homogeneous and cross-bedded sandstone facies might have been deposited by flows of higher velocity compared to the Platte-type. Miall (1982) notes that the Platte-type of facies is deposited by exceptionally broad, shallow rivers, which lack a well defined topography and which lack a clear distinction between active and inactive tracts. These characters are similar to those inferred for the homogeneous and cross-bedded sandstone facies.

6.19 THE CONGLOMERATE FACIES

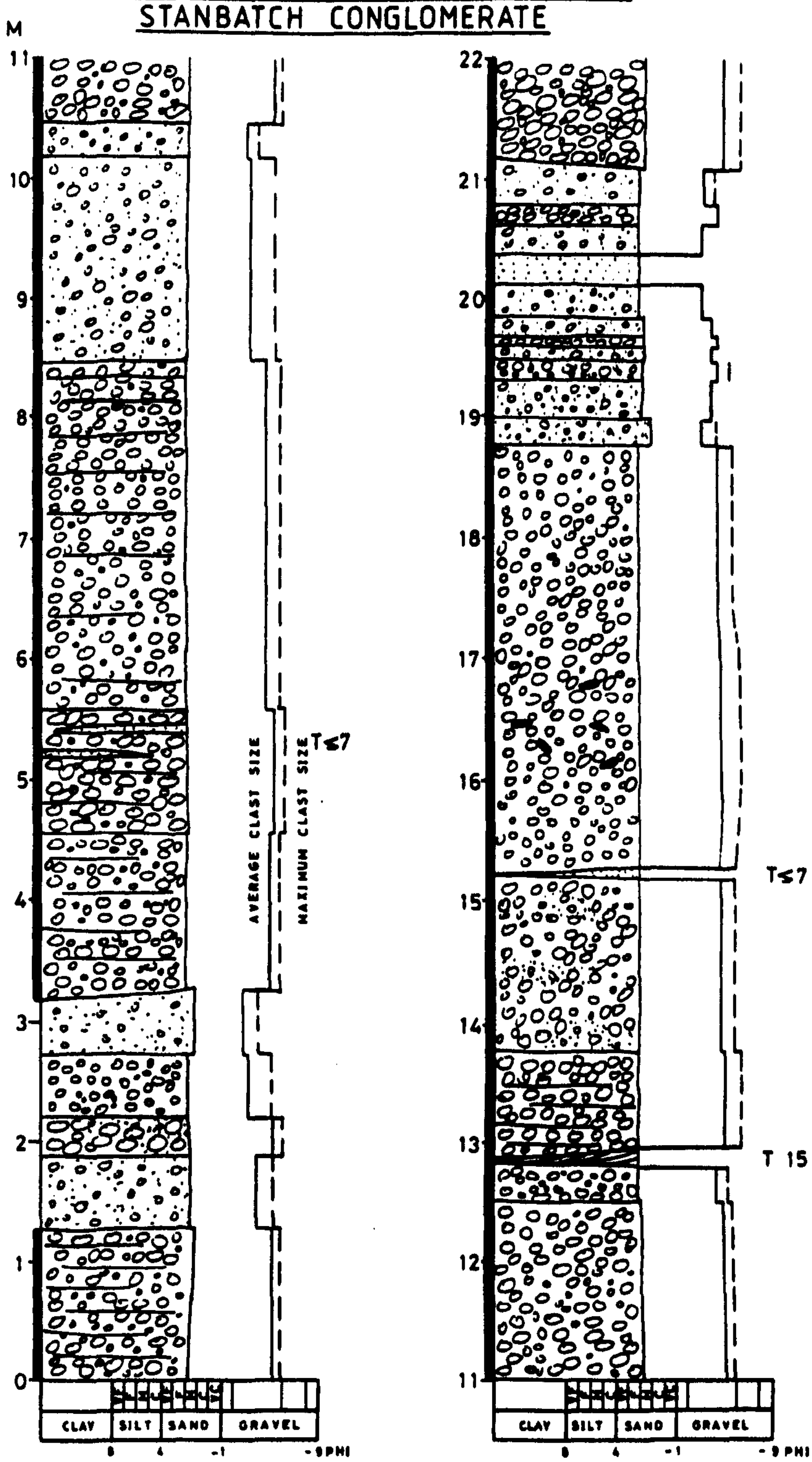
This facies is represented by several thick horizons in the Bayston-Oakswood Formation. On the eastern limb of the syncline, there are two main horizons: the Darnford Conglomerate Member (c. 90m thick) and the Stanbatch Conglomerate Member (c. 122m thick). The Haughmond Conglomerate Member is restricted to the areas around Haughmond Hill and Bayston Hill and is up to c. 106m in thickness. The Huckster Conglomerate Member of the Portway Formation is locally conglomeratic, but is mostly comprised of sandstone and pebbly sandstone. On the western limb of the syncline, there are three horizons: the Radlith Conglomerate Member (c. 83m thick), the Oakswood Conglomerate Member (c. 101m thick) and the Lawn Hill Conglomerate Member (c. 156m thick). In addition, there are other rare, impersistent conglomerate horizons in the Bayston-Oakswood Formation on the western limb of the syncline. It is likely that the conglomerate horizons on the western limb of the syncline are equivalent in part to those on the eastern limb. In particular, the Lawn Hill Conglomerate Member may be equivalent to the

Stanbatch Conglomerate Member and the Oakswood Conglomerate Member may be equivalent to the Darnford Conglomerate Member, however, this cannot be proved. The Willstone Hill conglomerate beds, which are associated with the Eastern Uriconian are also referable to this facies. This facies is illustrated by log 45, fig. 100.

Lithologies

The lithologies are very variable and there appears to be a continuous spectrum between clast-supported pebble conglomerates, with little sandy matrix (fig. 101, photo A), matrix-supported and very poorly sorted pebbly conglomerates (fig. 101, photo B) and pebbly sandstones. In addition, there are some poorly sorted granule conglomerates. The grain size distribution is generally polymodal, but often a bimodality is apparent, with a lack of grain size variation between medium sand grade and pebble grade. This is more apparent in the clast-supported conglomerates. The pebbles are rounded and subspherical to slightly elliptical. The average size of the pebble population is 12mm and the average maximum pebble size is approximately 47mm. The largest clasts occur in the Oakswood Conglomerate and these attain sizes up to 155mm, but these are not common. The sand-sized grains are subangular to subrounded. The lithologies of the pebbles are identical to the clasts in the rest of the Longmyndian sediments and some of the lithologies of the pebbles can be directly matched with those of the Uriconian Volcanic Complex. These conglomerates are therefore probably not polycyclic.

FIG.100 LOG 45 BILBATCH, GRID REF.SO 41549559
STANBATCH CONGLOMERATE



This log illustrates the conglomerate facies of the Bayston-Oakswood Formation.There are three grain size columns.The left column shows the average grain size of the matrix.The middle column shows the average clast size and the right column shows the average maximum clast size.The majority of the conglomerate is clast supported.Matrix supported conglomerate,pebbly sandstone and sandstone are symbolised by an increasing percentage of dots.Note the presence of upward-fining sequences (at 1.2m to 3.2m,8.5m to 10.5m and 18.7m to 20.3m) and thin sandstone beds (e.g. at 5.3m,12.9m and 15.2m).This conglomerate may be organised into larger scale upward-fining sequences (3.2m to 10.5m and 13m to 21.2m).

FIG:101 PHOTOGRAPHS ILLUSTRATING THE LITHOLOGIES OF THE CONGLOMERATE FACIES OF THE BAYSTON-OAKSWOOD FORMATION.

PHOTO A: Bimodal,clast supported conglomerate with rounded pebbles and sparse sand matrix.Top surface of bed.Oakswood Conglomerate.Scale bar is graduated in decimetres and centimetres.

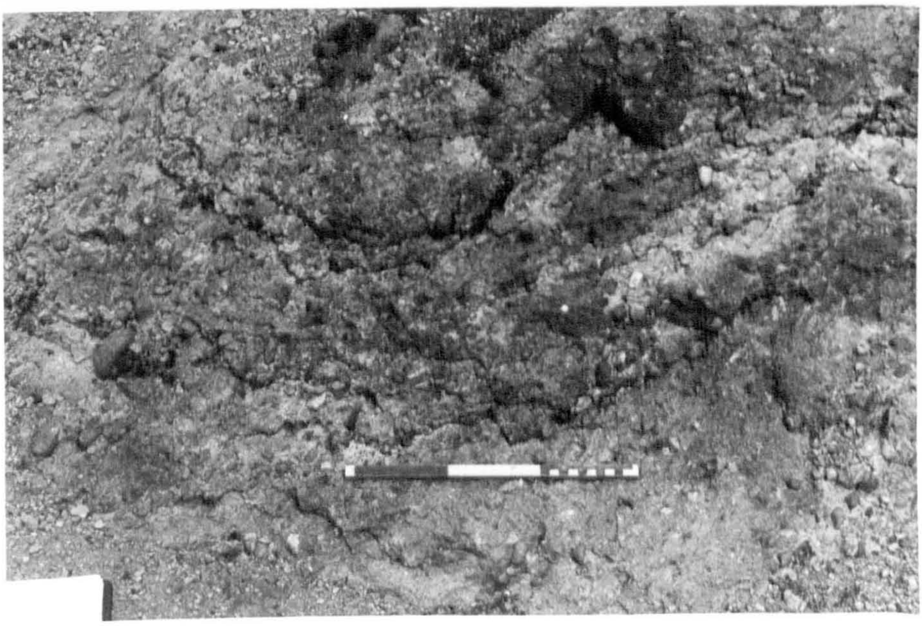


PHOTO B: Poorly sorted,polymodal,matrix supported conglomerate. View perpendicular to bedding.Stanbatch Conglomerate.Scale bar is graduated in decimetres and centimetres.

Sedimentary structures

The majority of the conglomerates appear to be structureless apart from a poorly developed subparallel, flat bedding, which is commonly on a 15cm to 50cm scale, and a planar, subparallel fabric (fig. 102, photo A). Where bedding planes are seen, these are usually flat (fig. 102, photo B). Cross-bedding, in sets 6cm to 30cm thick, rarely occurs, though it is more common in the pebbly sandstones and granule conglomerates. The pebbles in the clast-supported conglomerates occasionally show a slight imbrication.

Thin beds of sandstone commonly occur in the conglomerates. These are up to 50cm thick and have abrupt lower and upper contacts. The bases of these beds are often irregular and broadly trough-shaped (fig. 103, photo A) and the beds can often be seen to wedge out over distances of a few metres. The thinner beds are usually composed of fine to medium sandstone and are either structureless or planar, parallel laminated (fig. 103, photo B). The thicker beds are often pebbly and may contain thin layers of conglomerate. These beds may be trough cross-bedded or horizontally bedded. Homogeneous mudstone rarely occurs in beds 20cm to 30cm thick, which have abrupt bases and tops. Mudstone also rarely occurs in laminae.

Some upward-fining sequences occur which range from 25cm to 4m in thickness. The most common type of sequence shows a transition from matrix-supported conglomerate to pebbly sandstone. This sandstone can be trough cross-bedded or horizontally bedded. Some sequences show a transition from pebbly sandstone, which contains thin beds of conglomerate (4cm to 20cm thick), to

FIG:102 PHOTOGRAPHS ILLUSTRATING THE CHARACTERISTIC, FLAT PARALLEL BEDDING IN THE CONGLOMERATE FACIES OF THE BAYSTON -OAKSWOOD FORMATION.

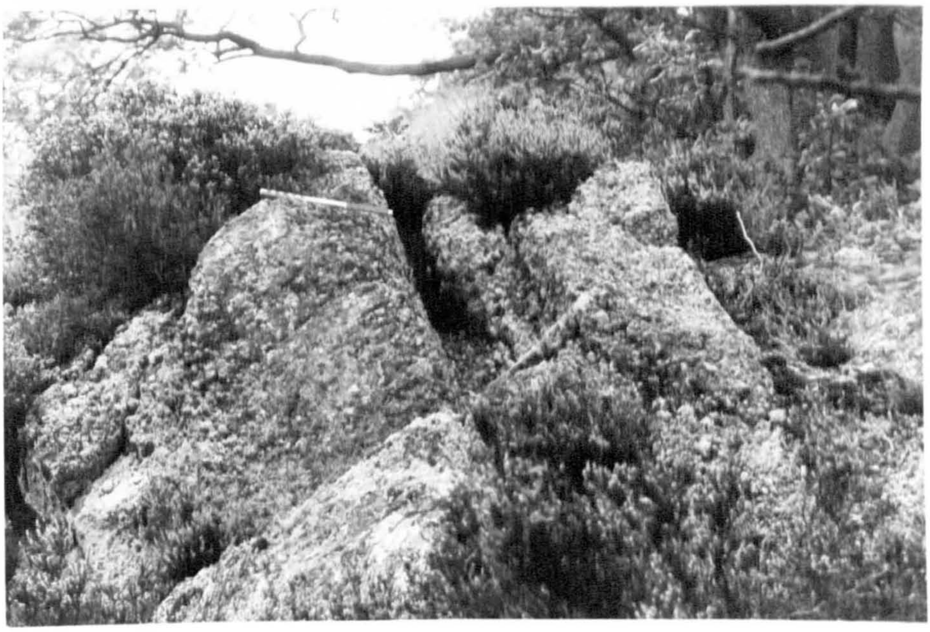


PHOTO A: Flat parallel bedding in the Oakswood Conglomerate. Scale bar is graduated in decimetres and centimetres.

PHOTO B: Flat bedding plane in the Darnford Conglomerate. Scale bar is graduated in decimetres and centimetres.

FIG:103 THIN SANDSTONE BEDS IN THE CONGLOMERATE FACIES OF THE BAYSTON-OAKSWOOD FORMATION.

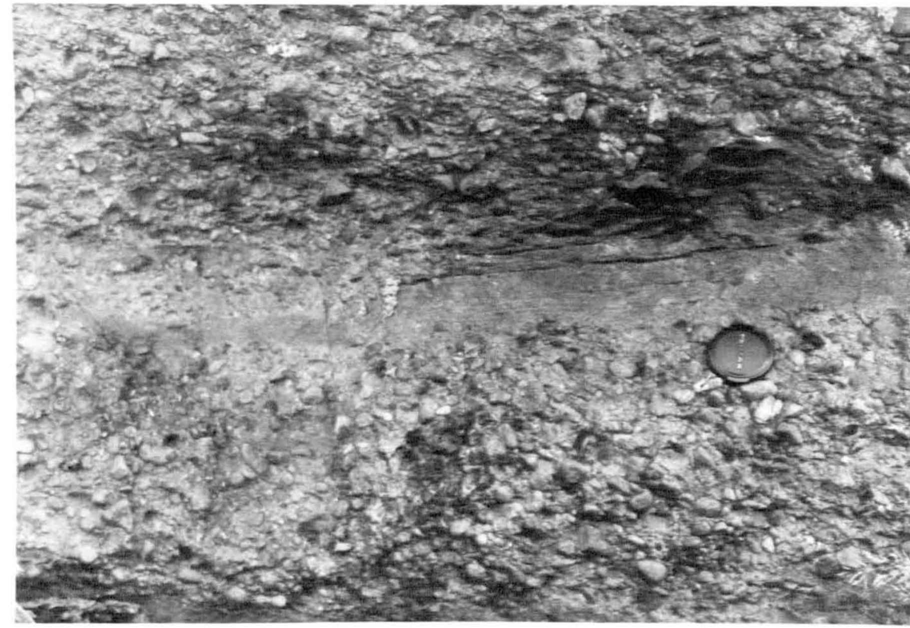


PHOTO A (left): Thin bed of trough cross-bedded,pebbly sandstone in the Stanbatch Conglomerate.Note the abrupt,irregular base, abrupt top and the rapid thinning of the bed.Way up is from right to left.Scale bar is graduated in decimetres and centimetres.

PHOTO B (above): Thin bed of planar parallel laminated sandstone in the Oakswood Conglomerate.Note the rapid thinning of the bed.Lens cap is 52mm in diameter.

homogeneous sandstone. In some instances, the conglomerate beds in these pebbly sandstones occur in shallow scours about 3m wide. Several upward-fining sequences are apparent on log 45, fig. 100.

<u>The relationships between the conglomerate facies and the homogeneous and cross-bedded sandstone facies</u>

The conglomerate facies can be traced as consistent horizons, with little change in character and thickness over several kilometres, and they cannot be demonstrated to occur within channels within the homogeneous and cross-bedded sandstone facies. The conglomerate horizons are parallel to the strike in the associated sandstones and therefore do not appear to be unconformable. The Haughmond Conglomerate Member is unusual in that it is rich in intermediate volcanic clasts and is laterally impersistent, since it is not represented in the area of the Long Mynd, except for a possibly equivalent pebbly sandstone at the base of the Bayston-Oakswood Formation in Hawkham Hollow (at SO 433 977). The contacts of the conglomerates with the homogeneous and cross-bedded sandstone facies are poorly exposed. However, from the topography and from the rare exposures, it appears that both the bases and the tops of these horizons are sharp. The Stanbatch Conglomerate Member consists of two conglomerate bands which are separated by c. 21m of homogeneous and cross-bedded sandstone. These three horizons can be traced by their topographic expression over several kilometres and little change is apparent in their thicknesses.

Palaeocurrent data

Very few palaeocurrent indicators were found in this facies and the data are insufficient to determine the palaeoflow. A pebble elongation lineation was noted in many of the conglomerates. This is discussed in section 3.9.1 and it is concluded that this lineation is due to tectonic strain and is not sedimentary in origin. The lineation is approximately coaxial with the major fold axis.

Interpretation

The predominant structure in these conglomerates is a flat, subhorizontal bedding and planar fabric. This structure is typically formed as a result of high fluid and sediment discharge, together with relatively shallow current depths, during the genesis of longitudinal bars in braided rivers (e.g. Collinson and Thompson, 1982, fig. 7.5, p. 111). There is a general lack of cross-bedded gravel, which suggests that slip-faces, associated with diagonal and transverse bars, were not generated. The cross-bedding that is rarely present occurs within pebbly sandstones, granule conglomerates and finer grained conglomerates and these are interpreted to have been deposited during lower flow stages, probably at the bar margins. Welded and gradational contacts between beds, which are differentiated on clast size and matrix content, suggest closely spaced, fluctuating and pulsating flow energies. The clast-supported conglomerates are interpreted to have been deposited during high flow-stages as diffuse gravel sheets from which sand was winnowed. Later infiltration of sand into the clast interstices would produce the bimodality of grain sizes in these clast-supported conglomerates. Since the matrix of the matrix-supported conglomerates is sandy and since there appears to be a continuous spectrum between clast-supported conglomerate, matrix-supported conglomerate and pebbly sandstone, it is suggested that the matrix-supported conglomerates were deposited by lower energy flows. Since the matrix in the matrix-supported conglomerates lacks stratification and is poorly sorted, it is likely that deposition was very rapid and that deposition was principally by settling from suspension during waning flow.

The pebbly sandstones are often trough cross-bedded or horizontally bedded and are interpreted to have been deposited between the longitudinal bars during low flood stages. The thin sandstone beds which are occasionally found within the conglomerates occupy broad shallow scours, which were probably generated on bar tops during low flow stages. Rapid shallowing of the water on the bar tops during low stages would have generated high flow velocities, which resulted in the formation of shallow scours filled with structureless or horizontally laminated sand deposited under upper flow-régime conditions.

Mudstone is very rare and is likely to have been eroded by the high energy flows. Preservation of the mudstone is likely to have occurred in abandoned channel tracts. The presence of mudstone beds testifies to the presence of finer grain sizes in the flows, which were usually flushed away from the site of deposition.

The sheet form of these conglomerates, the lack of channelled horizons and their lateral continuity over many kilometres, is compatible with their formation in low-sinuosity, braided channels. This facies may be compared with facies models for braided river deposits (Miall, 1978 and 1982 and Rust, 1978). It is directly comparable to facies Gm of Rust (1978) and Miall

(1978 and 1982). This is characterised by massive or horizontally bedded clast-supported conglomerate, which is similarly interpreted to have been deposited as longitudinal bars. The horizons which are composed of this facies in the Longmyndian are comparable to the Scott facies of Miall (1978 and 1982) and facies GII of Rust (1978). Both of these are dominantly composed of facies Gm and conglomerate forms greater than 40% of these deposits. Both Rust (1978) and Miall (1978) interpret the environment of deposition for these deposits to be a proximal braided river or alluvial plain. Since there is a lack of variation in the Longmyndian conglomerate horizons over many kilometres, the environment of deposition is thought to have been a proximal, braided alluvial plain rather than a proximal braided river.

6.20 THE BRAIDED ALLUVIAL FACIES ASSOCIATION

This association is comprised of the homogeneous and cross-bedded sandstone facies and the conglomerate facies, which together form the 2172m thick sequence of the Bayston-Oakswood Formation. This abruptly overlies the alluvial floodplain facies association. The contact between these associations has been interpreted as an unconformity by previous authors. This is extensively discussed in section 2.2 and it is concluded that, contrary to previous belief, a major unconformity is not present in the area of the Long Mynd. The probability that parts of the alluvial floodplain facies are distal equivalents of the braided alluvial facies is discussed in section 6.17 and it is concluded that this is likely. The transition from the alluvial floodplain facies association to the braided alluvial facies association can therefore be interpreted as a progradational coarsening-upward cycle, which is the culmination of the gradual progradational basin infil, which is represented by the turbidite facies association, the subaqueous delta facies association and the alluvial floodplain facies association. The red siltstone and cross-stratified sandstone facies of the Bridges Formation overlies the braided alluvial facies association and the contact is interpreted to be a rapid transition. This facies is interpreted as a more fine grained, distal equivalent of the braided alluvial facies association and it therefore appears that this association represents the culmination of the progradational cycle of the Longmyndian.

The Brokenstones sandstone beds and the Willstone Hill conglomerate beds are also representatives of this facies association. The Willstone Hill conglomerate beds are identical in character to the conglomerate facies of the Long Mynd and they therefore do not appear to be an intra-volcanic horizon as suggested by previous authors (e.g. Greig et al., 1968). Their stratigraphic position is discussed in section 2.5 and it is concluded that the contacts between the Willstone Hill conglomerate beds and the Uriconian Volcanic Complex are either unconformities or faults.

This facies association shows little change in character throughout the area and similar facies are found elsewhere in the Welsh Borders, at Old Radnor (Garwood and Goodyear, 1918), Pedwardine (Cox, 1912) and at Huntley (Callaway, 1900). The widespread occurrence of this facies association, with little change in character, suggests that the braided alluvial deposits were formed on a widespread braidplain, rather than in laterally restricted braided rivers. The presence of subspherical, rounded pebbles in the conglomerate facies suggests that considerable attrition of the clasts occurred and that the transport paths were long. The majority of the deposits are composed of moderately well sorted, medium grained sandstone with subrounded to subangular grains and there is a lack of very coarse rudite. These characteristics do not reflect the present day proximity of the proposed source rocks, which are represented by the Uriconian Volcanic Complex. Additionally, the NNW palaeoflow of these deposits does not suggest that they were locally derived from either the Western Uriconian or the Eastern Uriconian in their present outcropping configurations. It is considered probable therefore that these deposits have been tectonically juxtaposed with the Uriconian Volcanic Complex. This is extensively discussed in section 2.9.

The abrupt change from the alluvial floodplain facies association to the braided alluvial facies association is thought to be the result of a rapid, major uplift of the magmatic arc source. The repeated development of conglomerates, with a lack of transitional facies, could also be the result of repeated tectonic uplifts of less magnitude. These latter events could have been caused by faulting. The cause of the major tectonic uplift might have been a change in subduction rate or, alternatively, the initiation of major fault-scarps at the margin of the magmatic arc source area. Since the latter process is likely to have had a more rapid and immediate effect on the sedimentation, then the initiation of major faults is thought to have been the most likely cause. Although hypothetical, it is plausible that these faults were strike-slip faults, considering the evidence for strike-slip faulting presented in chapter 3 and the proposed forearc basin setting which is reviewed in chapter 7.

6.21 THE RED SILTSTONE AND CROSS-STRATIFIED SANDSTONE FACIES OF THE BRIDGES FORMATION

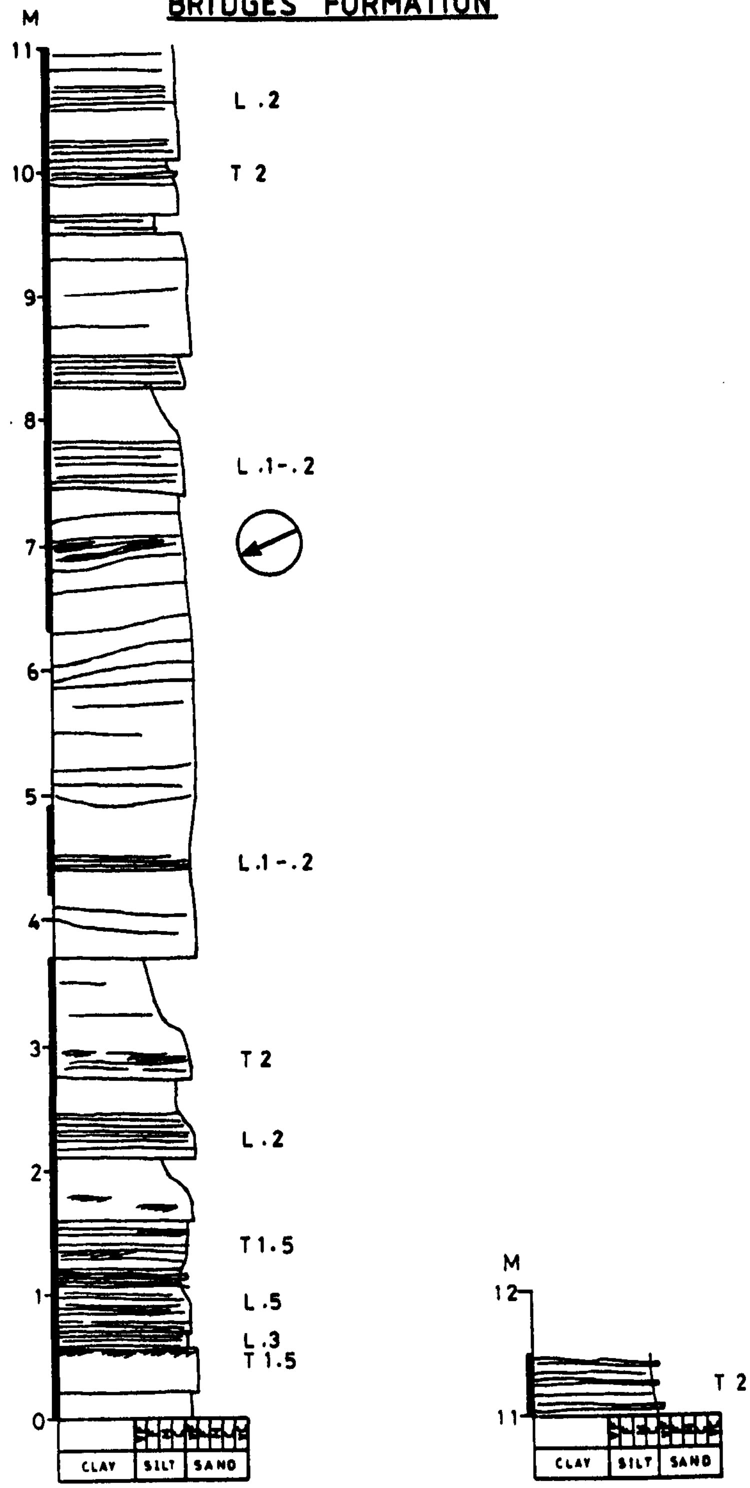
The Bridges Formation occupies the core of the major syncline. Thickness estimates are complicated by the presence of numerous minor folds, the lack of marker horizons and the probable presence of faults in the south-east of the outcrop. A thickness estimate of c. 500m is therefore very approximate. The top of this formation is not seen and the total original thickness is therefore not known. The formation overlies the homogeneous and cross-bedded sandstone facies of the Bayston-Oakswood Formation and there is a transitional contact. This transition results from the interbedding of siltstones in the homogeneous and cross-bedded sandstones and the gradual change from predominantly purplish grey, fine to medium grained sandstones in thick beds to thinner and fewer beds of very fine to fine grained, often slightly brownish grey sandstones. The transition is rapid and is not often seen because of the poor exposure. This facies has similarities to some of the facies of the alluvial floodplain facies association. In particular, it is similar to the red mudstone and cross-laminated sandstone facies of the Portway Formation and the red mudstone and cross-bedded sandstone facies of the Red Lightspout Member. This facies is illustrated by log 48, fig. 104 and log 47, fig. 105.

Siltstones

These constitute 40% to 50% of the facies. They are medium purplish grey to medium purplish red and are variably fine to coarse grained. Planar, parallel lamination is usually developed on a 1mm to 5mm scale and thick laminae and very thin beds (up to 2cm thick) of very fine grained sandstone are common (fig. 106).

FIG.104

LOG 48 RIVER EAST ONNY,SO 37889453
BRIDGES FORMATION



This log illustrates the red siltstone and cross-stratified sandstone facies of the Bridges Formation.There are numerous very fine grained,thin sandstone beds with abrupt or rapidly transitional tops (e.g. at 0.35m and 2.9m).The thick sandstone bed between 3.7m and 8.2m has a gradual upward-fining profile and is interpreted as a channel fill.

FIG.105 LOG 47 COTHERCOTT, GRID REF, SJ 41370103 BRIDGES FORMATION

This log illustrates the red siltstone and cross-stratified sandstone facies of the Bridges Formation. The thick sandstone bed between 1.9m and 5.7m is interpreted as a channel fill. Note the sequence of sedimentary structures from trough cross-bedding on a c.15cm scale to ripple cross-laminated sandstone. The associated siltstones are extensively planar, parallel laminated and contain numerous very thin beds and thick laminae of very fine grained sandstone. This type of lithology is shown in fig.106 .

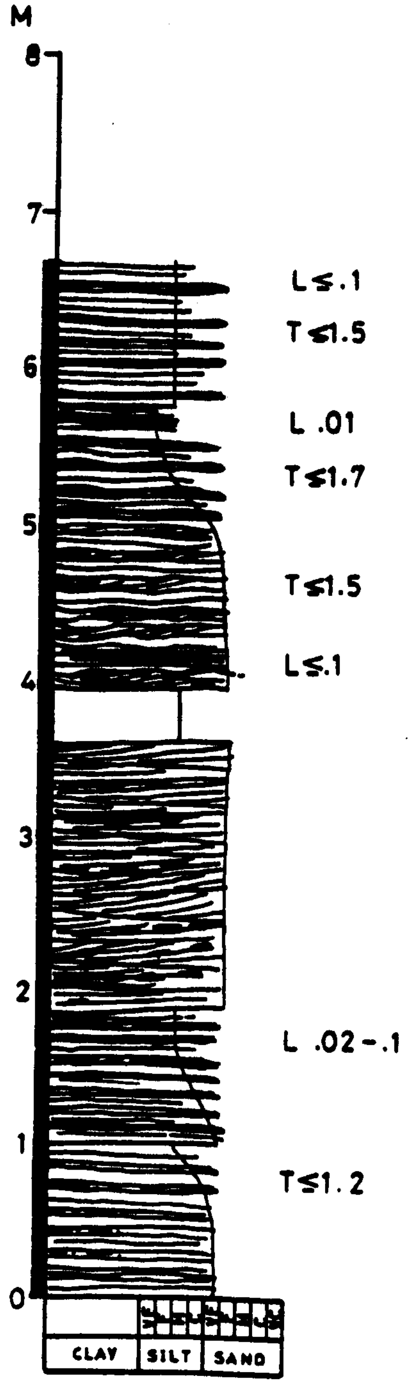
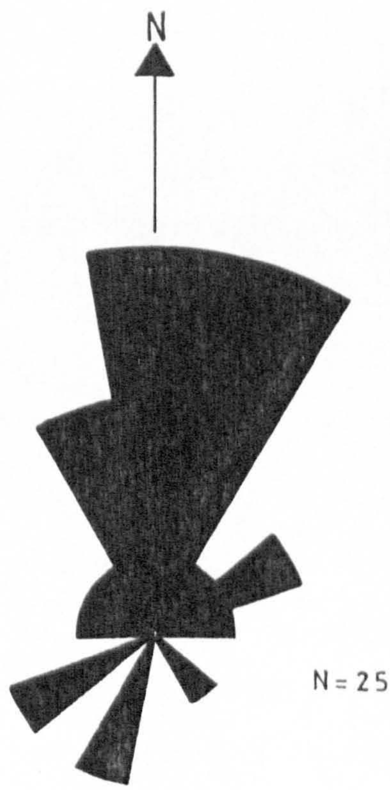


FIG:106 LAMINATED SILTSTONE FROM THE RED SILTSTONE AND CROSS-STRATIFIED SANDSTONE FACIES OF THE BRIDGES FORMATION AND PALAEOCURRENT ROSE FOR THIS FACIES.

PHOTO;shows laminated siltstone with thick laminae of sandstone Note the lensoid appearence of some of the laminae.Tape measure is marked in decimetres and centimetres.



PALAEOCURRENT ROSE for the red siltstone and cross-stratified sandstone facies of the Bridges Formation.

These are planar and parallel to slightly lenticular and are internally either very thinly parallel laminated or poorly cross-laminated. The bases of the thick sandstone laminae and very thin beds are occasionally loaded.

Sandstones

The sandstones are very variable in colour and may be grey, brownish grey, greenish grey or purplish grey. The grain size is very fine to fine, with the coarser grain sizes being shown by the thicker beds. The thickness of the beds is most commonly within the range of 10cm to 30cm, but can be up to c. 80cm. Additionally, there are a number of thicker beds which range from c. 1.3m up to c. 4m in thickness.

The thinner beds all have abrupt bases and either abrupt tops or a rapid gradation into siltstones. Slight upward-fining trends throughout the bed occasionally occur. Ripple cross-lamination, in sets up to 2cm thick, is commonly developed and this is usually of the trough type. Occasionally, a sequence of structures occurs, from planar, parallel lamination at the base to ripple cross-lamination at the top. These beds are characterised by their flat and parallel bedding. Soft-sediment deformation is commonly present. The deformation features include microfaults in the sandstone laminae, micro-loading features, which include chaotically intermixed sandstone and siltstone, and disrupted laminae which are injected by sandstone.

The thicker beds all have abrupt bases and a transitional top with siltstone. Slight upward-fining trends are occasionally present in the sandstone beds. Current ripple cross-lamination, in sets less than 2.5cm thick, is commonly developed and is usually of the trough type. Planar, parallel lamination is subordinate. Some beds have sets of trough cross-bedding which are up to 15cm thick. These occur towards the bases of the beds. Clasts of mudstone are often found in the basal parts of these sandstone beds. Siltstone occasionally occurs in thin beds, which are laterally impersistent, and also as drapes, which occasionally preserve ripple form-sets.

Occasionally, trough cross-bedded, fine to medium grained sandstone, which is similar to the homogeneous and cross-bedded sandstone facies of the Bayston-Oakswood Formation, occurs in thick units, which are occasionally greater than 6m thick. This type of sandstone outcrops along the banks of the River East Onny from SO 3875 9597 to SO 3908 9621.

Palaeocurrents

A palaeocurrent rose diagram (fig. 106) shows that the palaeoflow was unidirectional and towards the NNE.

Interpretation

This facies is similar to the red mudstone and cross-laminated sandstone facies and the red mudstone and cross-bedded sandstone facies of the alluvial floodplain facies association. The interpretations are therefore similar. The siltstones and thin sandstones are interpreted to have been deposited by sheetfloods and sheetflows on an alluvial floodplain. The thick sandstone beds are interpreted as channel-fills. This facies differs from the others in generally being more coarse grained, with more abundant thin sandstone beds and more abundant laminae and very thin beds of sandstone in the siltstone.

The NNE palaeoflow is broadly similar to the NNW palaeoflow of the underlying homogeneous and cross-bedded sandstone facies of the Bayston-Oakswood Formation and the red siltstone and cross-stratified sandstone facies is thought to be a longitudinally distal equivalent of this.

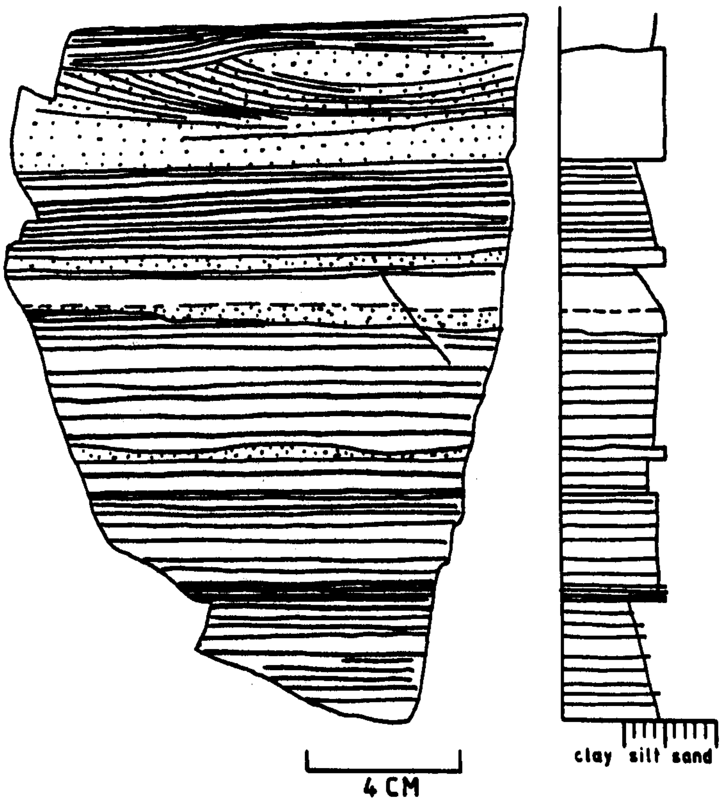
The development of this facies represents a regression from a sandy braidplain to a "distal" floodplain and this is atypical, since the rest of the facies in the Longmyndian are organised in a progradational sequence. The cause for this regression might have been the gradual retreat and denudation of the magmatic arc source terrain and this suggests that magmatic and tectonic activity in the source area had mostly ceased prior to the deposition of the Bridges Formation. The time at which this might have occurred was during the postulated major source uplift and faulting episode, which immediately preceded the development of the homogeneous and cross-bedded sandstone facies of the Bayston-Oakswood Formation.

6.22 THE GREEN SILTSTONE FACIES OF THE LINLEY BEDS

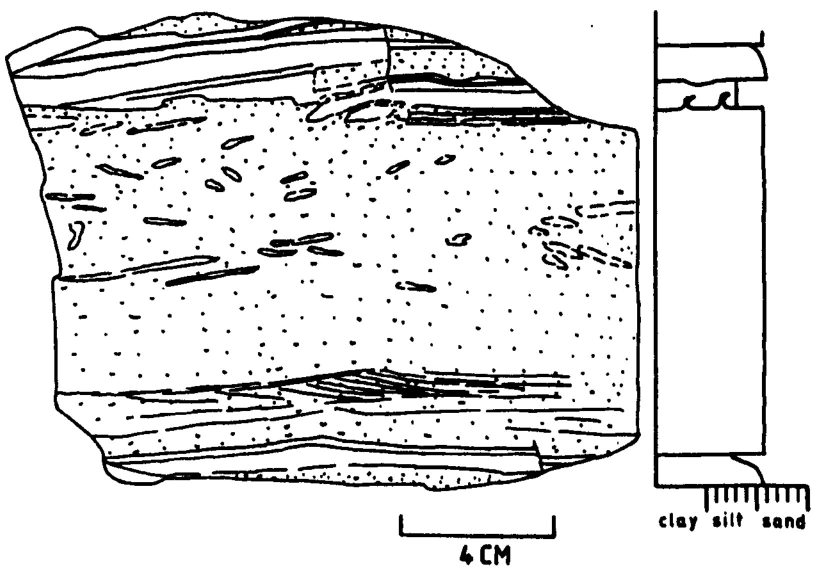
The Linley beds are poorly exposed close to and within the Pontesford-Linley fault system in the SW of the area. The contacts between the Linley beds and the adjacent Bayston-Oakswood Formation and Uriconian Volcanic Complex are poorly exposed, but are thought to be faulted. The stratigraphic position of these beds is therefore uncertain (see full discussion in section 2.6). The minimum thickness of these beds is estimated to be 160m.

The facies is characterised by siltstones and silty shales, which are predominantly light greenish grey in colour and are thinly to very thinly planar, parallel laminated. Occasionally, thick laminae and very thin beds of very fine grained sandstone

FIG:107 SLAB A MUDSTONE AND SLAB B THIN SANDSTONE BED FROM THE GREEN SILTSTONE FACIES OF THE LINLEY BEDS.



SLAB: A Mudstone. Note the presence of a ripple cross-laminated very thin bed of sandstone (top), the occasional graded laminated units and the pervasive, planar, parallel lamination



SLAB: B Thin sandstone bed. Note the lack of grading and the massive nature of the bed, apart from a single set of current ripple cross-lamination at the base of the bed. The patches towards the top of the bed are diffuse areas of siltstone. These and the injection features at the top of the bed may have resulted from dewatering. Note the presence of syn-sedimentary microfaults.

LOG 54 LINLEY, SO 34509414
LINLEY BEDS

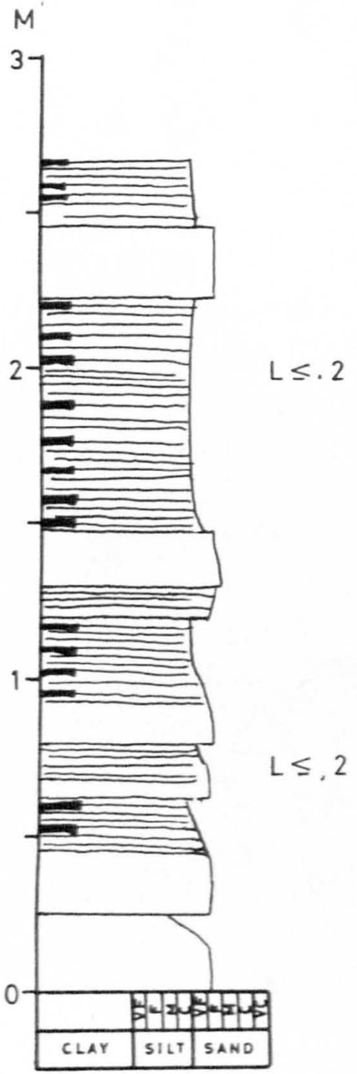


FIG:108 SLAB AND LOG 54 ILLUSTRATING THE GREEN SILTSTONE FACIES OF THE LINLEY BEDS.

PHOTO(above) SLAB ;illustrates the very thinly and planar parallel laminated lithology, which is typical of the Linley beds. The darker laminae are composed of very fine grained sandstone and the paler laminae are composed of siltstone.

LOG 54 (left) illustrates the coarser lithologies of the facies. Note the extensive planar, parallel lamination in the coarse grained siltstones and very fine grained sandstones and the presence of thin, apparently massive sandstone beds.

occur, which can be current ripple cross-laminated (fig. 107, slab A). Rarely, the siltstones are purplish grey. Occasionally, the silty shale is homogeneous and lacks lamination. In some instances, thin units (2-3cm thick) occur, which are normally graded and internally laminated and which begin with a thin lamina of sandstone (fig. 107, slab A).

Occasionally, thin beds of medium greenish grey, very fine to fine grained sandstone occur within the siltstones (fig. 108, log 54). These beds are 8cm to 22cm thick. They have abrupt bases and gradational to abrupt tops. They are mostly massive or planar, parallel laminated and are rarely current ripple cross-laminated (fig. 107, slab B). Sandstone injection features may occur at the top of the bed and diffuse patches of siltstone may occur towards the top of the bed, within massive and ungraded sandstone (fig. 107, slab B). These features suggest that extensive dewatering of the bed has occurred, probably during and immediately following deposition.

Rarely, thick beds (up to 2m thick) of very fine to fine grained greenish grey sandstone occur. These can be very thinly laminated with paler, light greenish grey siltstone and silty claystone (fig. 108, slab). At Chittol (SO 3496 9495) is an exposure of lapilli tuff, which appears to be associated with the more usual siltstones. There are no other tuff beds in the Linley beds.

Interpretation

Since this facies is greenish grey, it is probable that it was deposited either under marine conditions or close to sea-level, where the water-table was high (see section 6.24). The laminated siltstones were probably mostly deposited by settling from suspension. However, the presence of ripple cross-laminated sandstones indicates that some of the sediment was deposited by traction. The graded laminated units might have been deposited by a waning current, by a process of depositional sorting in the boundary layer (Stow and Bowen, 1980).

The thin sandstone beds appear to have been deposited rapidly from suspension, since they are often massive and have evidence for extensive dewatering. Planar, parallel lamination and occasional current ripple cross-lamination suggest that traction was occasionally operative. These beds might have been deposited by turbidity currents or by fluvially generated density-underflows.

This facies is similar to some parts of the mudstone and ripple cross-laminated sandstone facies of the upper part of the Burway Formation. The common characteristics are:the development of planar, parallel lamination, the greenish grey colouration, the presence of graded laminated units, the occurrence of thin sandstone beds, which are similar to turbidites, and the presence of thick sandstones with planar, parallel lamination. For these reasons, this facies is tentatively thought to have been deposited in a similar delta-slope to delta-front environment and the Linley beds are possibly stratigraphically equivalent to the upper parts of the Burway Formation. However, since there are also broad similarities between this facies and the green mudstone and cross-laminated sandstone facies of the Synalds Formation, these correlations are uncertain. There are many differences between this facies and the red mudstone and cross-laminated sandstone facies of the Portway Formation. These differences are expounded in section 2.6. Therefore, the identity of the Linley beds with the Portway Formation, as proposed by Cave et al. (1985) and Langford and Lynas (1985) is considered to be unlikely.

6.23 THE FACIES OF THE RAGLETH TUFF FORMATION (INCLUDING THE HELMETH GRIT MEMBER)

The Ragleth Tuff Formation is extensively discussed in section 2.3. This account therefore summarises the main features of this facies. The majority of this facies is comprised of mudstones. These are predominantly greenish grey and are occasionally brownish grey or purplish red. Planar, parallel lamination is commonly developed and there are occasional laminae of sandstone.

There are common thin beds of greenish grey to brownish grey, very fine to fine grained sandstone, which are usually 15cm to 30cm thick, but can be up to 60cm thick. These have abrupt bases and abrupt or gradational tops and appear to be structureless (fig. 4, log 53). Occasionally, there are some thick sandstone beds, which are common in the Helmeth Grit Member of this formation, and their greater abundance in the Helmeth Grit enables it to be distinguished from the rest of the formation. Cobbold and Whittard (1935), with the availability of better exposure, report four or five thick beds of sandstone, from 0.5m to 4m thick, in the Helmeth Grit Member. These are interbedded with thick units of green and purple shales and the sandstones contain angular shale fragments up to c. 10cm in length, which are indistinguishable from the interbedded shales (Cobbold and Whittard, 1935). In one 3m thick sandstone bed in the Ragleth Tuffs, there appears to be irregular patches of siltstone (fig. 4). This disturbed bedding might have been due to slumping or alternatively may be a further manifestation of dewatering and/or loading, which is common in all of the sediments of the Longmyndian Supergroup. Greig et al. (1968) record the presence of slump structures and current bedding in the Ragleth Tuffs.

Interpretation

The interpretation of this facies is uncertain. The thin, apparently structureless sandstones might have been deposited by turbidity currents. However, thick units of laminated mudstone are not commonly developed in the turbidite facies of the Longmyndian Supergroup. The lithology is commonly highly altered, with abundant epidote and mica. This alteration has probably obscured the majority of the sedimentary structures. Rarely, current ripple cross-lamination is visible in the sandstones. The development of purplish red colours in some of the mudstones might suggest that this facies was occasionally subaerially exposed (see section 6.24). In this case, the thin sandstone beds might have been deposited by sheetflood processes on an alluvial floodplain. If slumping is present in this facies, then a deep water origin is favoured. However, this apparent "slumping" might have been produced by soft-sediment deformation during dewatering and/or loading. The thick sandstone beds with rip-up clasts might have been deposited in either environment and are not diagnostic.

6.24 THE COLOUR OF THE LONGMYNDIAN SEDIMENTS

The Longmyndian sediments are usually either greenish grey or purplish red to purplish grey. Brownish grey colours are also rarely developed in the Bridges Formation. 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 and 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.

The different colours of the sediment are related to particular facies and this suggests a depositional control on their development. All of the marine facies are greenish grey. In contrast, the majority of the alluvial facies are purplish red. Greenish grey colours are developed in the sediments which are associated with the thick beds of sandstone, which were deposited in ENE to NE flowing channels. In contrast, the purplish red colours are developed in the sediments which were deposited by an approximately westerly to northerly flowing braided, sheetflow and sheetflood alluvial system. The colour difference is most likely due to the position of the water-table, with greenish grey colours being developed beneath the water-table and purplish red colours being developed above it (Morad, 1983; Van Houten, 1973 and Friend, 1966). This would suggest that the green mudstone and cross-laminated sandstone facies and the green mudstone and thick-bedded sandstone facies were deposited where the water-table was relatively high. The high position of the water-table was most likely due to the inferred lower topographic position of the environment of deposition, since the green mudstone and cross-laminated sandstone facies lies stratigraphically above the subaqueous delta facies and stratigraphically below the red mudstone and cross-laminated sandstone facies. The position of the water-table is reflected in the occurrence of mudcracks. These are usually found in the red sediments, but they rarely occur in the green sediments. Where they do occur in the green sediments, they occur just below the transition to red sediments.

CHAPTER 7

THE EVOLUTION OF THE LATE PRECAMBRIAN OF ENGLAND AND WALES - A REVIEW BASED UPON THE EVIDENCE FROM THE LONGMYNDIAN AND PRECAMBRIAN OF SHROPSHIRE

7.1. The results of this study which are of regional importance

Several of the conclusions of this study are considered to be important in modelling the evolution of the late Precambrian of England and Wales. These are as follows:

1. The Longmyndian represents a progradational basin infil, approximately 6500 m thick. This progradational sequence is represented by basin plain mudstones, turbidites, deltaic deposits, alluvial floodplain deposits and sandy braidplain deposits in ascending order.
2. There is no major unconformity or evidence for a major stratigraphic break in the sequence.
3. The sediments are all volcaniclastic and were mainly derived from an undissected magmatic arc of Uriconian type.
4. The presence of schistose quartzites and garnet implies that a metamorphic terrain was present in the source area.
5. The characteristics of the sediments suggest that the source area was not proximal to the site of deposition and it is therefore likely that if the Uriconian Volcanic Complex was the source for the Longmyndian, then it has been later juxtaposed with it along the Church Stretton and Pontesford-Linley fault systems. There is also the possibility however that a volcanic

complex, similar to, but not of the same age as the Uriconian Volcanic Complex, acted as the source for the Longmyndian, in which case the Longmyndian could unconformably overlie the Uriconian Volcanic Complex.

6. All of the Longmyndian/Uriconian boundaries, except for the Willstone Hill conglomerate beds/Eastern Uriconian contact, are interpreted as faults. The Willstone Hill conglomerate beds, which are correlated with the Wentnor Group of the Longmyndian, may rest unconformably on the Eastern Uriconian.

7. Sinistral strike-slip appears to have been important during the deformation of the Longmyndian.

8. It is likely that the Church Stretton and Pontesford-Linley fault systems are major wrench faults (e.g. Woodcock, 1984a and 1984b and Lynas et al., 1985). This is suggested by the characteristics of these fault systems and by the evidence for sinistral strike-slip in the Longmyndian.

9. The appearance of coarse fluvial clastics in the sequence suggests that major uplift of the magmatic source area occurred immediately prior to the deposition of the Wentnor Group.

10. The palaeocurrent directions suggest that the source area for the Longmyndian Supergroup lay to the south, since the directions are all between WNW and ENE. The deposits of the Wentnor Group have NNW and NNE palaeocurrent directions and these strongly suggest a southerly source.

11. Movement along the Church Stretton fault system probably occurred prior to the deposition of the Lower Cambrian Wrekin Quartzite, which rests unconformably on the Eastern Uriconian.

12. The folding of the Longmyndian occurred before or during uplift, which is dated by radiometric means as c. 530 Ma (Naeser et al., 1982) and probably occurred prior to the Lower Cambrian, Wrekin Quartzite transgression over the adjacent Uriconian in the Church Stretton fault system.

13. The lack of an Ediacaran fauna suggests that the Longmyndian may be Varangerian to early Ediacarian, which is dated as c. 700 to c. 640 Ma (Glaessner, 1984a).

14. The Uriconian Volcanic Complex is probably overlain unconformably by the Willstone Hill conglomerate beds, which are probably equivalent to the Wentnor Group of the Longmyndian. Since there are pyroclastic horizons in the Longmyndian and since the Longmyndian appears to have been mainly sourced by an undissected magmatic arc of Uriconian type, it is considered that the Uriconian Volcanic Complex is in part broadly equivalent in age to and in part slightly older than the Longmyndian and is therefore probably c. 700 Ma old. This allows for the correlation of the Uriconian Volcanic Complex with the similar, calc-alkaline, Stanner-Hanter, Malvernian and Johnston Complexes, which are dated as 700 Ma to 643 Ma old (Thorpe et al., 1984).

15. The Rushton Schists have a radiometric age date of 667 ± 20 Ma (Beckinsale *et al.*, 1984). This is approximately coincident with magmatic arc activity.

16. The mineral cooling date of 536 ± 18 Ma obtained from the Rushton Schists (Patchett *et al.*, 1980) and the age of the post-Uriconian Ercall granophyre intrusion (533 ± 12 Ma, Patchett *et al.*, 1980) are both similar to the uplift ages obtained from the Longmyndian of c. 530 Ma (Bath, 1974 and Naeser *et al.*, 1982). This suggests that significant tectonic activity may have occurred at this time.

7.2. The nature of late Precambrian sedimentation in England and Wales

The Longmyndian records the infilling of a deep basin which was sourced by a magmatic arc. A comparable late Precambrian basin is represented by the Charnian. Moseley and Ford (1985, p. 3) note: "... the Charnian sediments accumulated in a NNE-SSW trending basin sited to the immediate south-east of a volcanic centre." They interpret it as an active zone with intermittent volcanicity and earthquake activity. The sediments include volcanic and sedimentary breccias, slump breccias and greywackes with graded bedding and load structures (Moseley and Ford, 1985 and Moseley, 1979). Some of these deposits may be interpreted as turbidites from the descriptions given by Moseley (1979). The presence of breccias and slumps, together with volcanic breccias and brecciated intrusives (Moseley and Ford, 1985) suggests that the Charnian may represent a more "proximal" facies than the Longmyndian in places. However, the Charnian and Longmyndian appear to be similar in that their sediments are volcaniclastic and both were deposited during magmatic arc activity and were initially deep water basins. Palaeocurrents for the Charnian are to the NNE (Moseley, 1979) and this direction is compatible with the southerly source postulated for the Longmyndian. In common with the Longmyndian, the clasts in the Charnian include quartzite, schist and acid plutonics and Moseley (1979) postulated that a Malvernian-type basement provided these. A Malvernian-type source for some of the Longmyndian clasts is considered to be probable.

Other late Precambrian deposits in England and the Welsh Borders are only preserved in small areas and it is difficult to interpret their significance. The late Precambrian sediments of Carmarthen (Cope, 1977, 1979 and 1982) have been examined by the author and are interpreted as shallow water epiclastics which may be equivalent to the delta-front deposits of the Burway Formation in the Longmyndian.

Small inliers occur at Old Radnor, Pedwardine and Huntley Quarry. These are comparable to the Bayston-Oakswood Formation of the Longmyndian (Garwood and Goodyear, 1918; Cox, 1912 and Callaway, 1900). The sandstones at Old Radnor have been examined by the author and are considered to be representatives of the homogeneous and cross-bedded sandstone facies of the Longmyndian and to have been similarly deposited on an alluvial braidplain. Compositionally, the sandstones of Old Radnor are almost identical to those of the Bayston-Oakswood Formation (fig. 18).

Of probable similar age to the Longmyndian is the Monian Supergroup of Anglesey and Lleyn. This is comprised of turbidites and other deep water deposits. The presence of chert, manganiferous shale and mafic and ultramafic bodies, including pillow lavas are thought to indicate the existence of oceanic crust (e.g. Thorpe et al., 1984). The Gwna mélange is interpreted as an olistostrome and contains blocks of shallow water limestone and quartzite in addition to pelagic shale and pillow basalt (e.g. Thorpe et al., 1984). Orthoquartzites also occur within the turbidites of the New Harbour Group (the Holyhead Quartzite) and these are interpreted as the deposits of density flows (Shackleton, 1975). Facies changes suggest that deeper water lay to the south-east and a source area lay to the north-west (Shackleton, 1975).

The Cullenstown Formation of south-east Ireland is thought to be part of the Mona Complex. It is principally comprised of "greywackes" and some slumps occur. Quartzites and quartzitic siltstones occur as turbidites, as exotic blocks of bedded, white, quartzitic conglomerate and in slump sheets (Max, 1975).

7.3 Conclusions regarding the nature of late Precambrian sedimentation in England and Wales

1. The Longmyndian, Charnian and Mona Complex have thick sequences of deep water deposits. Only in the Longmyndian is there a record of gradual change from deep water to subaerial conditions.
2. The Charnian, Longmyndian and the inliers of Longmyndian type at Old Radnor, Pedwardine, Huntley Quarry and Carmarthen, were mainly derived from a magmatic arc which lay towards the south. Schistose quartzites occur in these deposits and it is suggested that a sheared plutonic complex, similar to the Malvernian, was the probable source for these.
3. The Monian Supergroup contains orthoquartzite, oolitic limestones and stromatolitic limestones. These appear to have been derived from a shelf region towards the NW.

7.4 The late Precambrian magmatic arc

Igneous rocks of late Precambrian age are present in Pembrokeshire (the Dimetian Complex, Pebidian volcanics, Johnston Complex and Bentonian volcanics), along the Church Stretton fault system in the Welsh Borderland (the Uriconian Volcanic Complex, the Stanner-Hanter Complex and the Carmarthen rhyolites),in the Malvern Hills (the Malvernian Complex and Warren House volcanics) and in Leicestershire (the Charnian and S. Leicestershire diorites). Igneous rocks and some sediments are also known from several boreholes in central and eastern England (Dunning, 1975). Thorpe et al. (1984) consider these igneous rocks to have formed above a south-eastward dipping subduction zone in an island arc setting. The Uriconian Volcanic Complex has the chemical characteristics of a continental margin island arc (Thorpe, 1972a, 1974 and 1979). The Warren House Group of the Malverns is chemically different from the calc-alkaline volcanics of the island arc in being tholeiitic and Thorpe et al. (1984, p. 525) consider that this group "... might, therefore, represent a tectonically emplaced fragment of ocean floor derived, for example, from a marginal basin."

If the age of the Uriconian Volcanic Complex is considered to be in the region of 700 Ma, as is proposed by this study, rather than 558 ± 16 Ma as proposed by Thorpe et al. (1984) and Patchett et al. (1980), then all the dated Precambrian igneous rocks in South Wales, the Welsh Borders and England can be differentiated into two groups. The Stanner-Hanter Complex, Malvernian Complex, Johnston Complex and Uriconian Volcanic Complex fall within the age range of c. 700 Ma to c. 640 Ma. They can therefore be considered to have formed during a single magmatic episode and within the same magmatic arc province. Later intrusions are represented by the southern Diorites of the Charnian (540 ± 57 Ma), the St. Davids Granophyre (587 + 25, - 14 Ma) and the Ercall granophyre (533 ± 12 Ma), (age data from Thorpe et al., 1984).

Modern arc-trench gaps are usually greater than 100 km and are commonly between 100 km and 200 km (e.g. Dickinson and Seely, 1979). The distance between Anglesey and the proposed magmatic arc is c. 100 km and this distance was most likely to have been much greater before tectonic shortening and within the range of an arc-trench gap. Therefore, it is possible that there was a trench in the region of Anglesey, which was contemporaneous with the late Precambrian magmatic arc to the south, as proposed by Thorpe et al.

(1984), Shackleton (1975), Dewey (1969), Wood (1974) and others. An alternative view is presented by Gibbons (1983), who considers that the evidence for palaeo-subduction in the Anglesey region is weak and he offers alternative explanations for the formation of the blueschists, "ophiolites" and mélange in the Mona Complex.

7.5 Precambrian metamorphic rocks

The evidence for a metamorphic basement in England and Wales is conflicting. Chadwick et al. (1983) and Whittaker and Chadwick (1984) argue that there is a crystalline basement to southern Britain which consists of gneiss. This is thought to occur between depths of 10 km and 16 km and appears to be overstepped by the late Precambrian of variable thickness. These arguments are based on deep seismic reflection profiles. However, Thorpe et al. (1984) argue, from the age data and from the chemical data obtained from the late Precambrian igneous rocks, that there is little evidence for the presence of a basement older than c. 900 Ma below England and Wales and that the crust is composed solely of late Precambrian igneous rocks and their associated sediments. Hampton and Taylor (1983), from Sr and Pb isotopic data derived from granites, also argue that the basement in south Britain is c. 800 Ma and definitely less than 1200 Ma old.

Gibbons (1983) argues that the Coedana granite (with its hornfels), Monian gneisses and Sarn Complex "... represent slices of continental crust of different ages sheared into the Monian Supergroup." (p. 152). However, all these have ages between c. 600 Ma and c. 540 Ma and from the Sm-Nd model ages, field evidence and these Rb-Sr isochron ages, Thorpe et al. (1984) argue that the Monian gneisses do not represent a gneissose continental basement and they conclude: "... all these lines of evidence are consistent with derivation of the gneisses by metamorphism of part of the bedded succession." (p. 523).

Max (1975) argues, from unpublished Rb-Sr isochron data, that the metamorphic rocks of the Rosslare Complex in south-east Ireland are at least 1600 Ma old. However, Winchester and Max (1982) report a Rb-Sr isochron age of only 650 Ma from the synkinematic St. Helens Gabbro. The antiquity of these metamorphic rocks is therefore considered to be dubious.

There are no gneissose fragments in the Longmyndian. Garnet and schistose quartzite fragments in the Longmyndian might have been derived from a sheared plutonic complex similar to the Malvernian. Moseley (1979) also suggests that similar fragments in the Charnian were derived from a Malvernian-like source. Lambert and Holland (1971) argue that garnet-mica schists and other garnetiferous rocks in the Malvernian "... suggest the existence of a former greywacke-type metamorphic basement into which the Malvernian diorites were intruded." (p. 345).

Similar garnetiferous rocks probably acted as a source for the garnet in the Longmyndian. Similar garnetiferous rocks are represented by the Rushton Schists in Shropshire, which have a Rb-Sr isochron age of 667 $\pm$ 20 Ma (Thorpe et al., 1984). Thorpe et al. (1984) argue, from Sm-Nd data, that the Rushton Schists have a maximum model age of 1600 Ma and from the initial Sr ratios, they might have been formed by the metamorphism of sedimentary rocks as old as c. 950 Ma. The age date of c. 667 Ma is compatible with the initiation of magmatic arc activity at c. 700 Ma. It is possible that these garnetiferous rocks might therefore represent the uplifted roots of the early magmatic arc which, together with sheared plutonics, contributed detritus to the late Precambrian sedimentary basins.

From the evidence presented, there does not appear to be an old crystalline basement to England and Wales. The only undisputed metamorphic basement is represented by the Pentevrian of Brittany and the Channel Islands, which is at least 900 Ma old (e.g. Thorpe, 1982).

7.6. The evidence for strike-slip

Evidence for strike-slip in the Palaeozoic of the Welsh Basin, mostly unpublished at present, is gradually accumulating (Welsh Basin sedimentation and tectonics conferences, 1985 and 1986). Evidence has been presented for a component of sinistral strike-slip in the deformation of the Longmyndian and it is postulated that both the Pontesford-Linley and Church Stretton fault systems are major wrench faults (section 3.8.4). Woodcock (1984a, p. 326) considers that "... there is considerable local evidence of strike-slip displacement throughout Wales" and he interprets the Pontesford Lineament and the Church Stretton fault system as strike-slip faults (Woodcock, 1984a and 1984b). Gibbons (1983) notes that the Mona Complex is pervaded by steeply dipping faults and mylonite zones and he interprets the Mona Complex as a tectonic collage which "... may represent part of an exotic terrain moved approximately to its present position during the Cambrian." (p. 147). In addition, Gibbons (1983) considers that the NW source terrain, which supplied shelf detritus to the Monian Supergroup, may be missing as a result of strike-slip faulting. In SE Ireland, the Cullenstown Formation is separated from the Rosslare Complex by a NE-SW trending mylonite zone (Max, 1975 and Winchester and Max, 1982) which might be comparable with the mylonitic Penmynydd schists in Anglesey, which are considered by Gibbons (1983) to have been involved in transcurrent movements. Arthur (1982) has identified areas of unexposed as well as exposed Precambrian igneous rocks in South Wales and the Welsh Borders from aeromagnetic maps. All the Precambrian anomalies strike NE-SW and Arthur (1982, p. 108) concludes: "... the NE-SW striking magnetic anomalies in South Wales have been sinistrally displaced relative to each other in an approximately NE-SW direction, parallel to their individual strikes by a system of faults ...".

All of the exposed Precambrian rocks of Wales, the Welsh Borders and SE Ireland appear to be associated with major NE-SW trending faults, some of which are mylonitic and there is an overall NE-SW tectonic grain. From the previous discussion, it appears that many of these faults may be interpreted as strike-slip faults (Woodcock, 1984a, 1984b and Gibbons, 1983). The Malvernian appears to be unique in that it is associated with a N-S trending major fault system rather than a NE-SW trending one.

The age of these NE-SW major faults is important for plate tectonic models and the determination of their age of initiation is made complex by the possibility that these are long-lived faults which have been rejuvenated at various times. The sense of movement across these faults is also important. However, the sense of movement on any individual fault may change with time, may not reflect the overall stress field of the orogeny, and adjacent faults may display different senses of movement depending on the competence of the rocks and the local stress field.

7.7. The age of the faulting

In Shropshire, it is thought that major movement occurred along the Church Stretton fault system in late Precambrian times, since folded and faulted Eastern Uriconian volcanics are overlain unconformably by the Lower Cambrian Wrekin Quartzite. Movement along the fault system was probably penecontemporaneous with the folding of the Longmyndian, which occurred prior to c. 530 Ma from the radiometric age data and the folding probably predated the Lower Cambrian Wrekin Quartzite transgression. It has been suggested that major uplift of the magmatic arc occurred immediately prior to the deposition of the Wentnor Group. This event might have been related to the development of strike-slip faulting in the magmatic arc and to a change from a subduction orogen to a strike-slip orogen. The Charnian accumulated in a NNE-SSW trending basin (Moseley and Ford, 1985) and the Longmyndian has approximately northerly to ENE directed palaeocurrents, which might reflect control by NNE to ENE trending faults. This evidence suggests that major faulting along NE-SW lines was initiated in the late Precambrian. Lynas (1985) similarly suggests that "... the Precambrian 'basement' in Wales and perhaps the 'Midland Platform' is built up by a collage of accreted tectonostratigraphic terranes like those now recognised in California and Alaska." (p. 935) and suggests that the later upward-propagation of the strike-slip junctions generated the fault patterns in the Lower Palaeozoic.

The Penmynydd schists in Anglesey and Lleyn are thought to be younger than the Coedana Granite and the Sarn Complex (Gibbons, 1983). Consequently, they are younger than 603 ± 34 and 549 ± 19 Ma respectively. Gibbons (1983) argues that the "Penmynydd tectonometamorphic event" took place sometime during the early Cambrian at c. 520 Ma. However, the Mona Complex is usually considered to have been folded and metamorphosed during the Precambrian (e.g. Thorpe et al., 1984) and it is possible that some strike-slip might have occurred during the late Precambrian in association with this.

The age dates of 533 ± 12 Ma for the Ercall Granophyre (Beckinsale et al., 1983) and the Rb-Sr biotite cooling age for the Rushton Schists of 536 ± 8 Ma (Patchett et al., 1980), both of which occur within the Church Stretton fault system, may be related to major movement along the Church Stretton fault system, accompanied by uplift of the Longmyndian, which is dated at c. 530 Ma.

Woodcock (1984a and 1984b) suggests that there is evidence for strike-slip displacements in Wales from the latest Ordovician through to the post-Carboniferous. The major movement is thought to have been dextral during the latest Ordovician to earliest Silurian.

7.8. The direction of movement across the major faults

The evidence from the Longmyndian suggests that sinistral displacements were important during its deformation. Arthur (1982) proposes that several magnetic anomalies in South Wales, which are interpreted as Precambrian igneous rocks, have been displaced sinistrally along NE-SW faults. This is the only direct evidence from the Precambrian of England and Wales for the sense of any strike-slip movement, at present, and it is suggested that the sinistral movement which is recorded in the Precambrian is probably more indicative of movements during the late Precambrian than the evidence from the Palaeozoic. During the Palaeozoic, the principal movements along some of the major faults are thought to have been dextral (Woodcock, 1984a and 1984b and Lynas et al., 1985).

7.9. A model for the evolution of the late Precambrian of England and Wales

The late Precambrian igneous rocks of South Wales, England and the Welsh Borders are interpreted as representing a magmatic arc which is c. 700 Ma to 640 Ma old (e.g. the Uriconian Volcanic Complex). The chemical and isotopic evidence indicates that this magmatic arc could have been at a continental margin or could have been intra-oceanic. There is no evidence for appreciable crystalline basement. The subduction zone apparently dipped towards the SE and a trench might have existed in the region of Anglesey.

A forearc basin developed in the arc-trench gap and this was filled with volcaniclastics and occasional pyroclastics which were derived from the magmatic arc to the south (e.g. the Longmyndian). The forearc basin was initially deep and was infilled with turbidites. The basin gradually shallowed and, in places, deltaic sediments were deposited, which were overlain by alluvial floodplain deposits (e.g. the Stretton Group). Local uplift and shearing of the plutonic roots of the magmatic arc provided some plutonic and metamorphic detritus (including garnet and schistose quartzites). Contemporaneously with the evolution of the magmatic arc, shallow water limestones and orthoquartzites accumulated on a passive continental margin to the north-west of the subduction zone. In the region of the trench (e.g. Anglesey), turbidites and pillow lavas accumulated.

In the later stages, volcanic activity probably waned and the magmatic arc was uplifted, during which time the forearc basin was infilled with coarse fluvial detritus (e.g. the Wentnor Group). In the region of the trench (e.g. Anglesey) olistostromes formed by mass movement of quartzite and limestone blocks which were derived from a passive continental margin to the north-west.

Before the end of the Precambrian, the Mona Complex was deformed, the Longmyndian was folded and there was probably major strike-slip displacements of segments of the magmatic arc and forearc basin along NE-SW faults (e.g. the Pontesford-Linley and Church Stretton fault systems). The driving force for this deformation might have been collision between the magmatic arc and the passive continental margin.

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APPENDIX

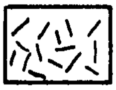
LOGS

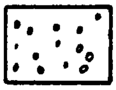
The following logs have been used in the facies analyses. Some are referred to in the main text, but others are not referred to. All the log numbers are the same as those used on geological maps 1 and 2 (in cover pocket).

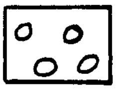
FACIES AND FORMATION	LOG NUMBERS
BURWAY FORMATION	
Thick-bedded turbidite	2, 3, 4
Mudstone and cross-laminated sandstone	6, 10
CARDINGMILL GRIT MEMBER OF THE BURWAY FORMATION	
Thick-bedded and cross-stratified sandstone	8, 11, 12
SYNALDS FORMATION	
Green mudstone and cross-laminated sandstone	13, 14, 15, 16
Red mudstone and cross-laminated sandstone	17, 19, 20
LIGHTSPOUT FORMATION	
Green mudstone and thick-bedded sandstone	25,26,27,28,29,30,31
Red mudstone and cross-bedded sandstone	33
PORTWAY FORMATION	
Huckster Conglomerate Member - homogeneous and cross-bedded sandstone facies	35
Red mudstone and cross-laminated sandstone	37, 39
BAYSTON-OAKSWOOD FORMATION	
Homogeneous and cross-bedded sandstone	41, 42, 44
Conglomerate	45
BRIDGES FORMATION	
Red siltstone and cross-stratified sandstone	46, 49, 50, 51, 52
RAGLETH TUFF FORMATION	53

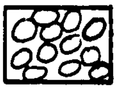
FIG:109 KEY TO THE SYMBOLS USED ON THE LOGS.

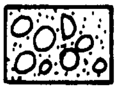
 Tuff

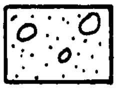
 Intrusion

 Granules

 Pebbles

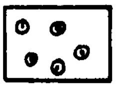
 Clast supported conglomerate

 Matrix supported conglomerate

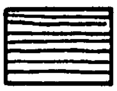
 Pebbly sandstones (in the conglomerates)

 Irregular sandstone patches

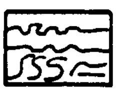
 Diagenetic patches

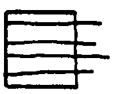
 Zoned spheroidal diagenetic patches

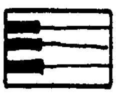
 Mudstone clasts

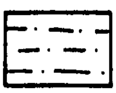
 Continuous planar parallel lamination

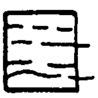
 Continuous wavy non-parallel lamination

 Convolute lamination

 Laminae of coarser grain size

 Laminae of claystone

 Laminae of siltstone (in sandstone)

 Discontinuous lamination (of various types)

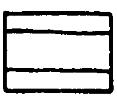
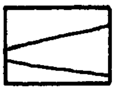
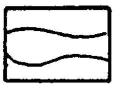
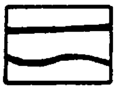
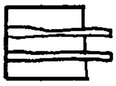
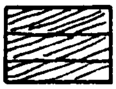
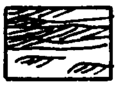
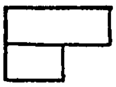
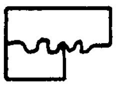
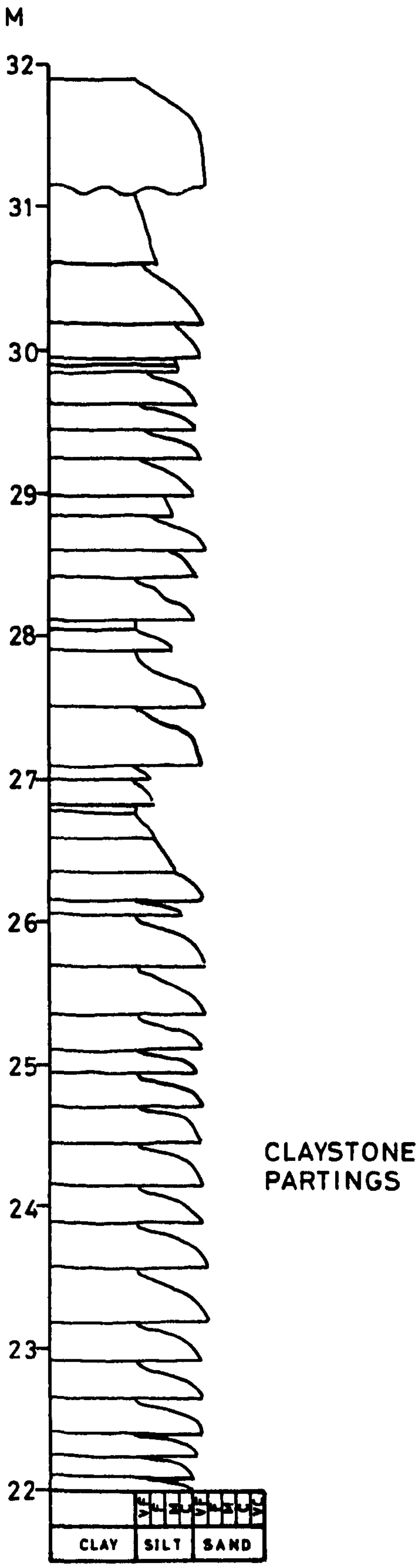
 Planar parallel bedding

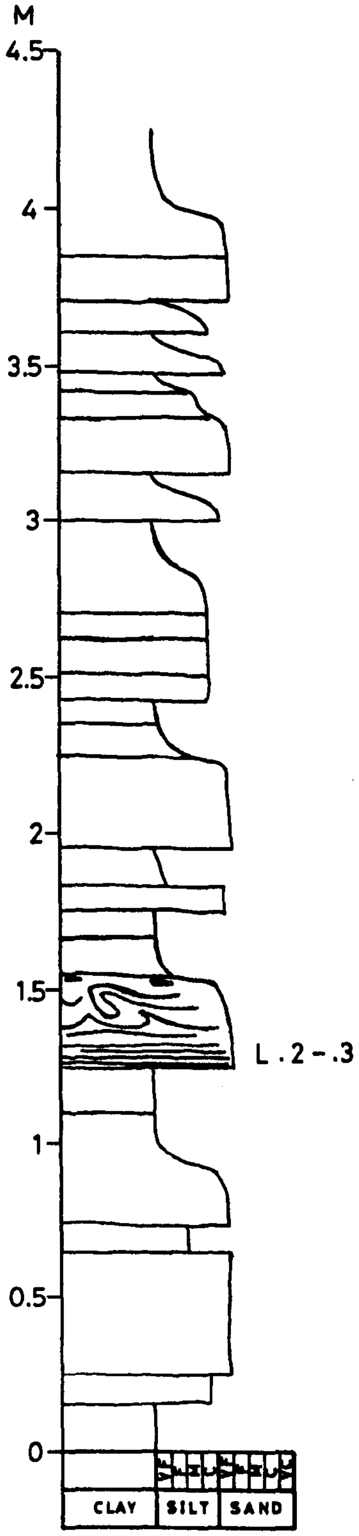
FIG:109 (continued) KEY TO THE SYMBOLS USED ON THE LOGS

	Planar non-parallel bedding
	Wavy non-parallel bedding
	Thin mudstone beds
	Very thin beds and thick laminae of coarser grain size
	Trough cross-bedding
	Tabular cross-bedding
	Current ripple cross-lamination
	Abrupt base
	Abrupt loaded base
	Erosional base
L.1-.3	Lamination thickness 1mm to 3mm
T10-18	Bedding or cross-bedding set thickness 10cm to 18cm
α15	Angle of cross-stratification is 15°
Low α	Angle of cross-stratification is low
B10,T1.5	Individual bed thickness is 10cm,with 1.5cm sets of cross-lamination
λ10,h.3	Bedform wavelength is 10cm and height of bedform is 3mm.
G	Grey colours
	Purplish red or brownish colours
	Greenish grey colours
M	Colour mottling
L C	Lamination (L) and clasts (C) are purplish red in greenish grey sediment.
L C	Lamination (L) and clasts (C) are greenish grey in purplish red or brownish sediment.
	Palaeocurrent direction

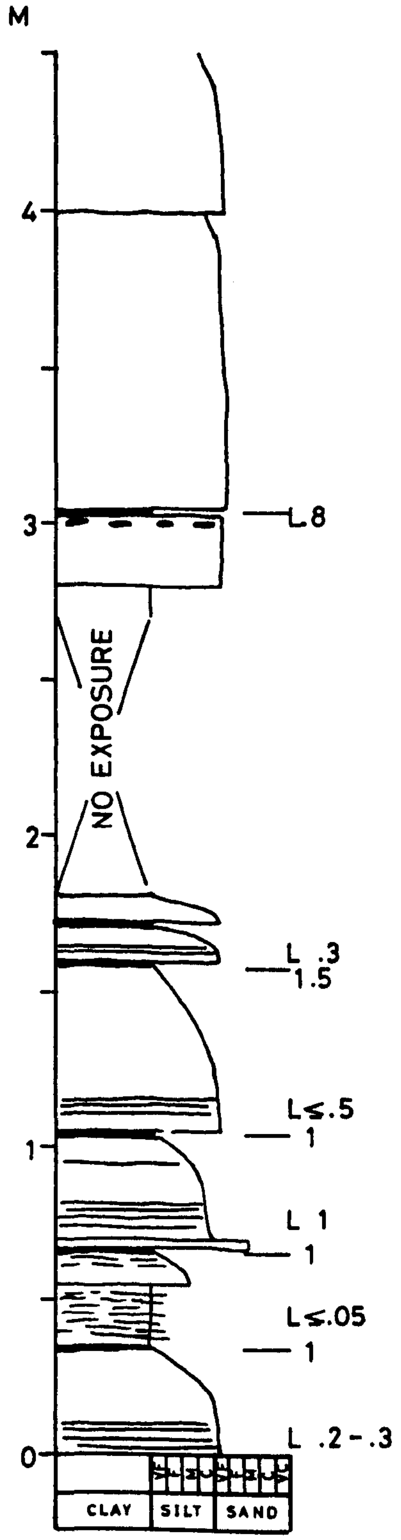
LOG 2 cont'd, ASHES HOLLOW, GRID REF, SO43669285
BURWAY FORMATION



LOG 3 ASHES HOLLOW, GRID REF, SO 43619288
BURWAY FORMATION

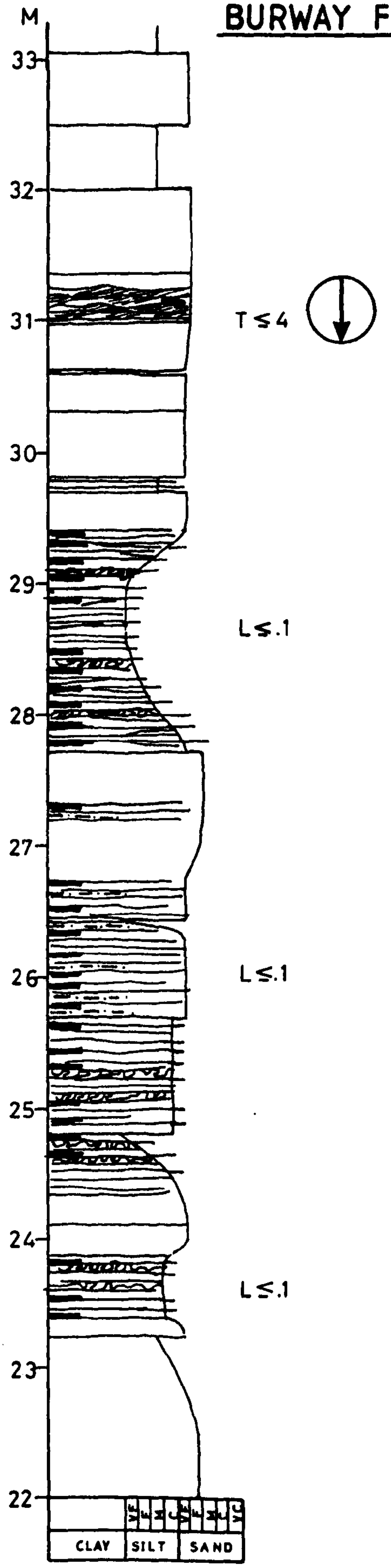


LOG 4 CARDINGMILL VALLEY, SO 44849427
BURWAY FORMATION

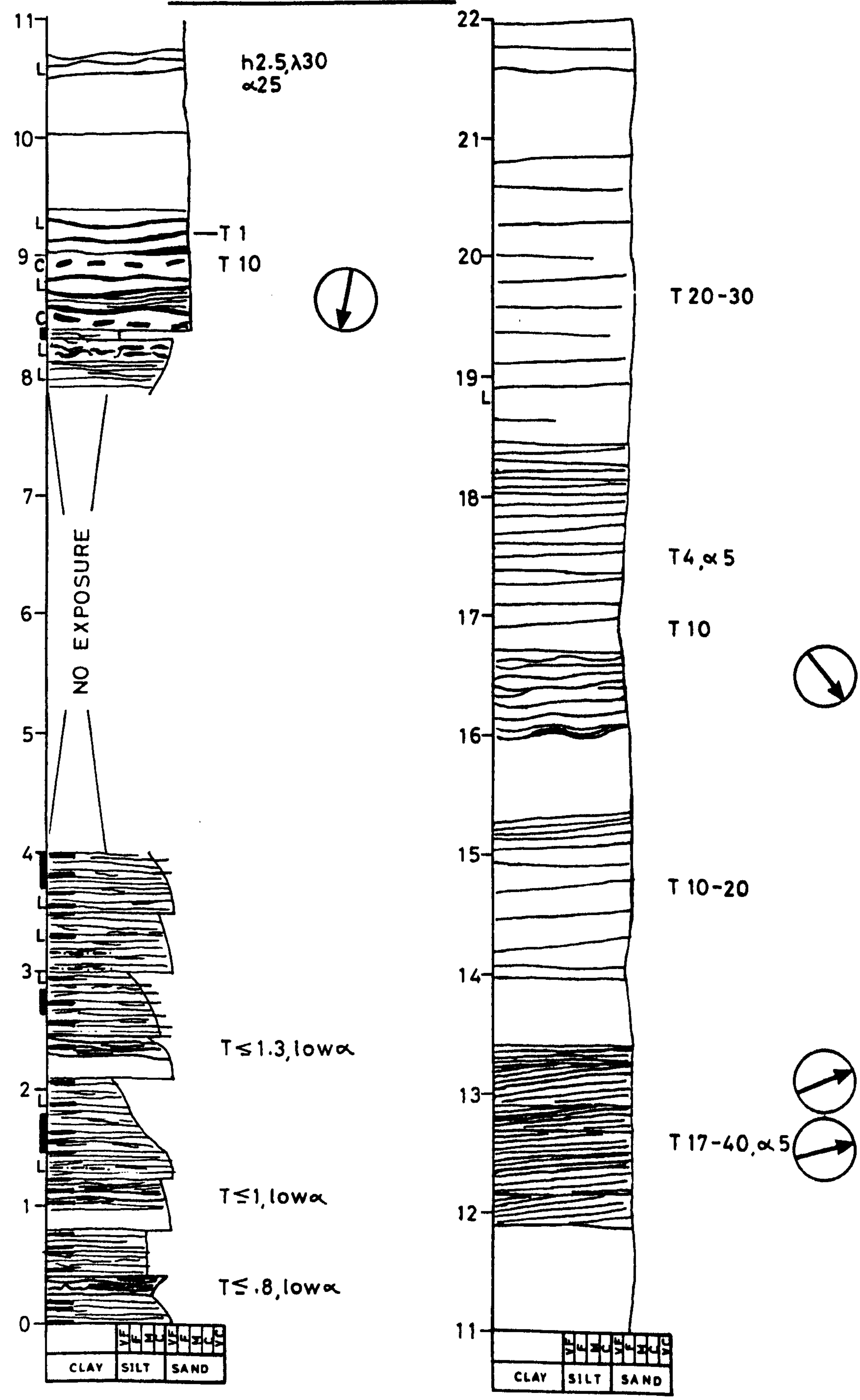


LOG 6 cont'd ASHES HOLLOW, GRID REF.43399302 SO

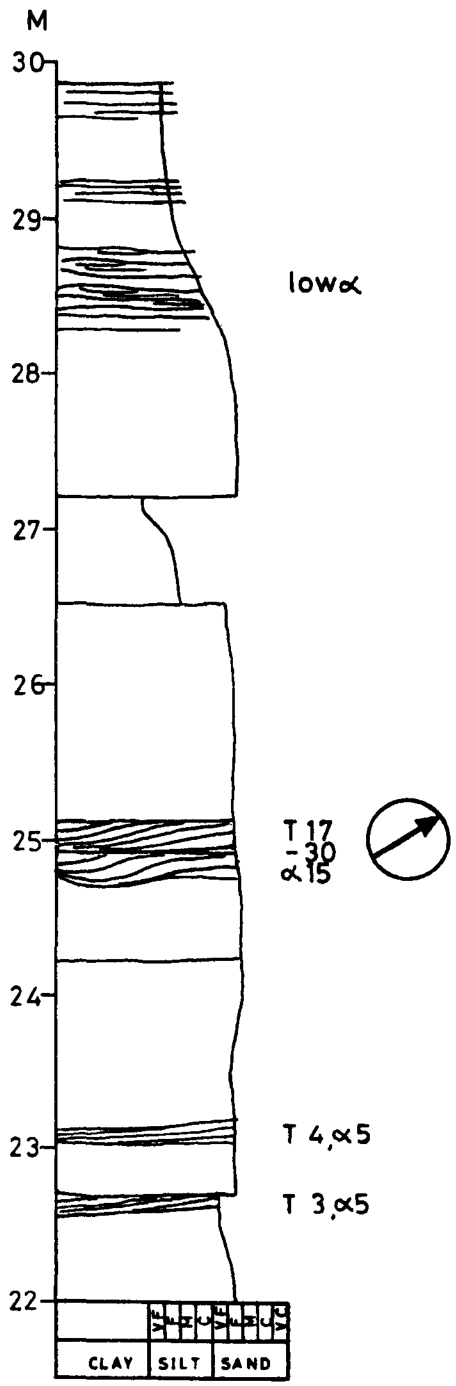
BURWAY FORMATION



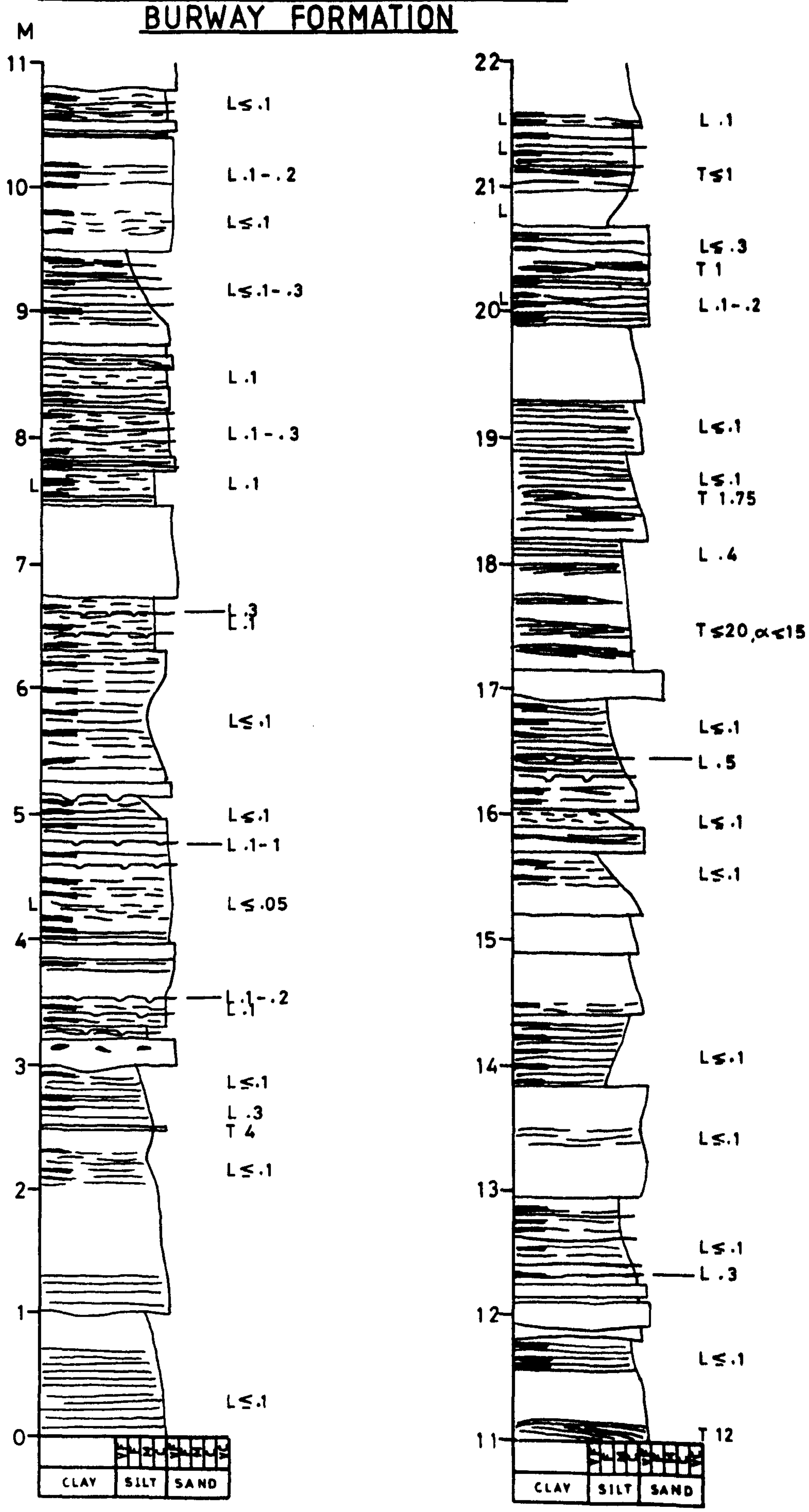
LOG 8 ASHES HOLLOW, GRID REF.43289302,SO
CARDINGMILL GRIT



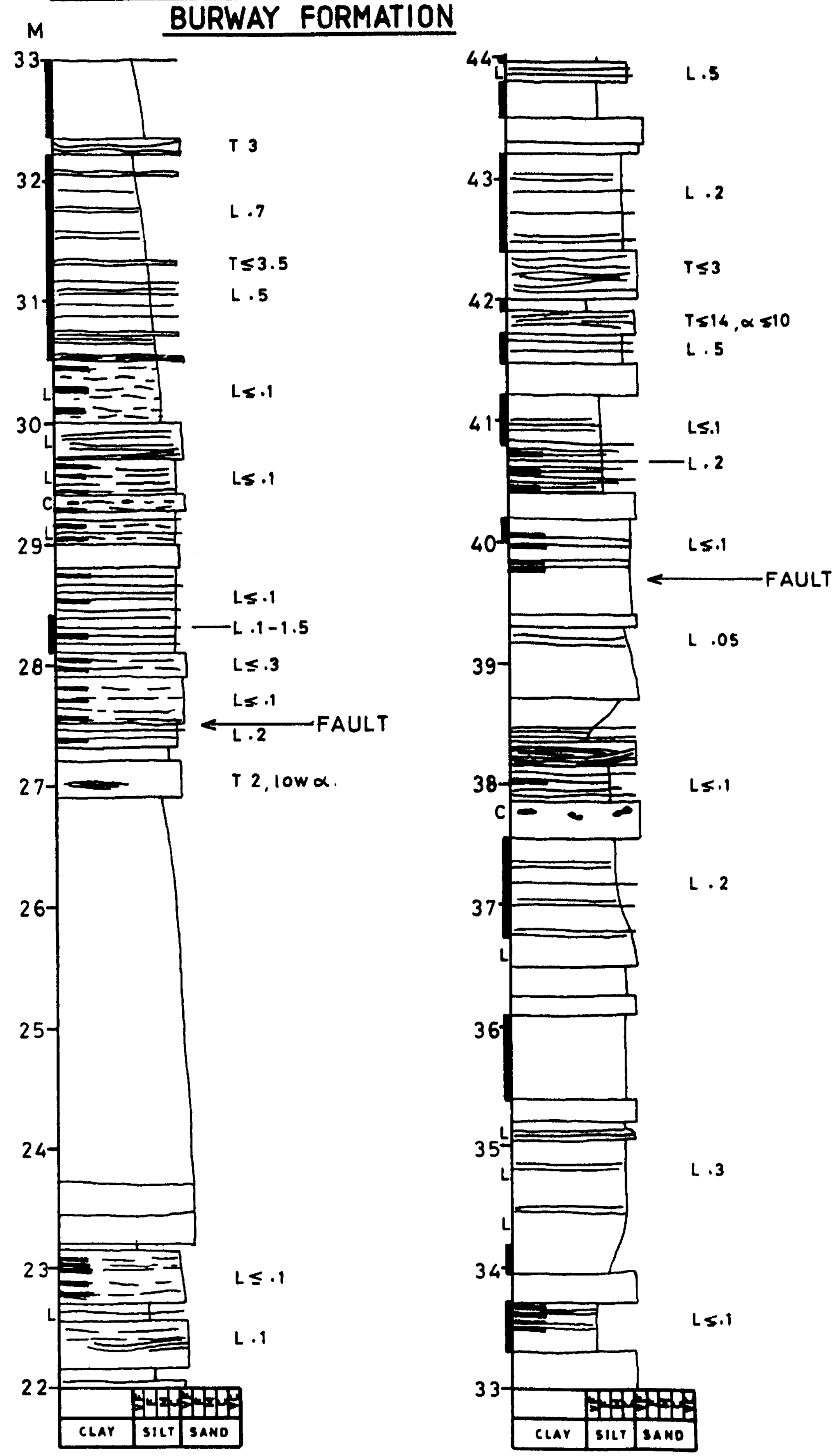
LOG 8 cont'd ASHES HOLLOW, GRID REF, 43289302, SO CARDINGMILL GRIT



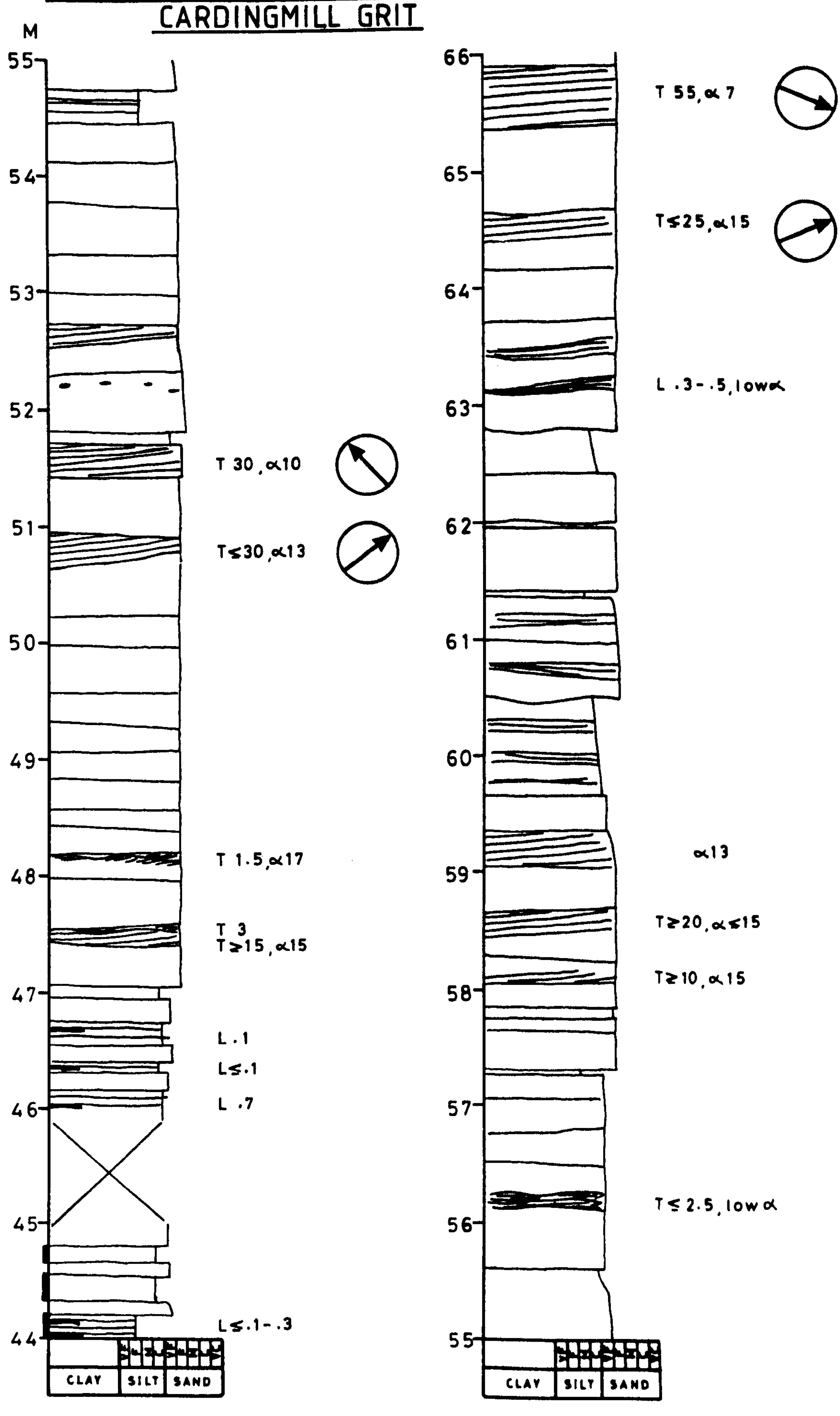
LOG 10 BURWAY HILL, SO 44059422
BURWAY FORMATION



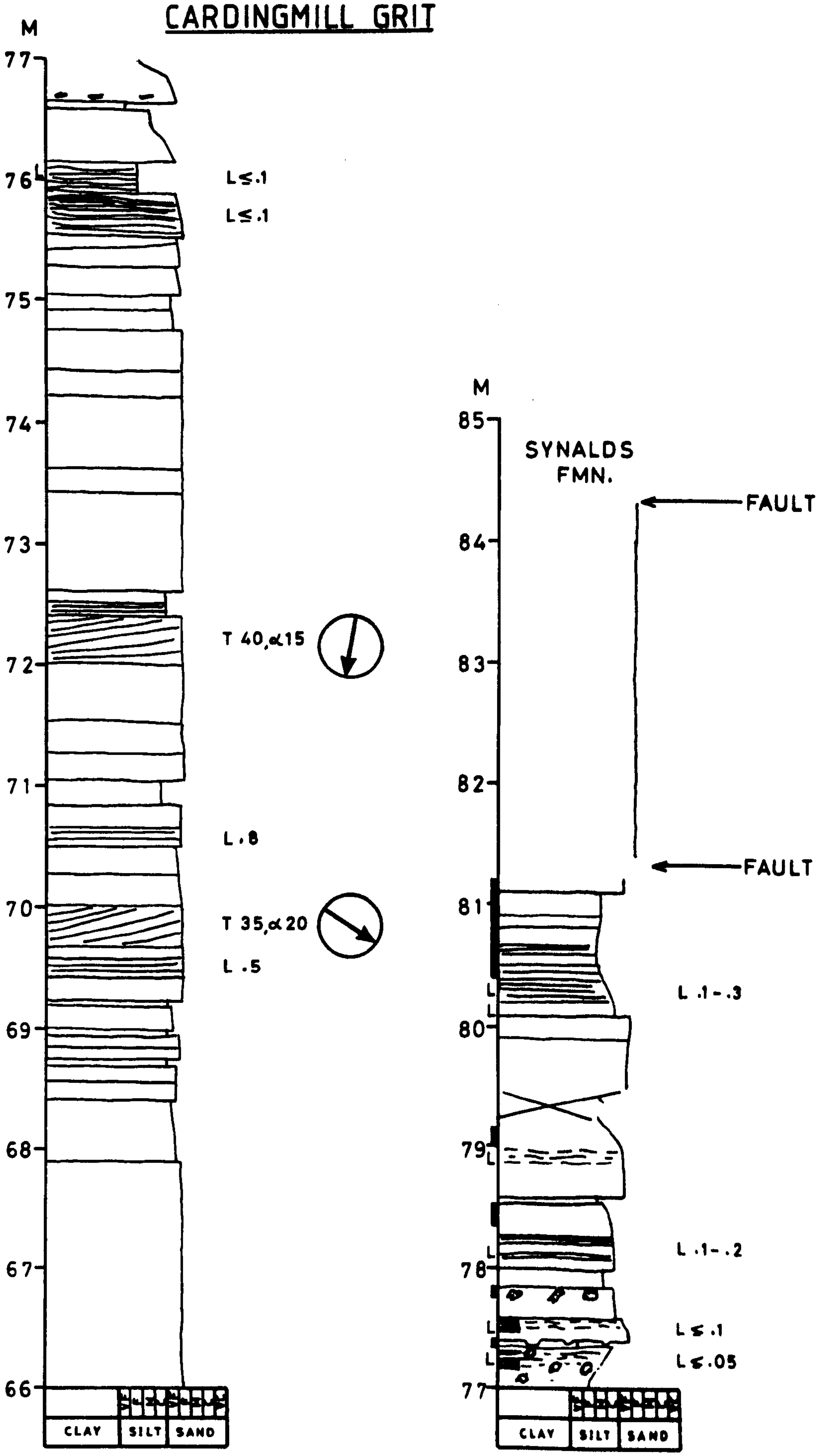
LOG 10 (cont'd) BURWAY HILL, SO 44059422
BURWAY FORMATION



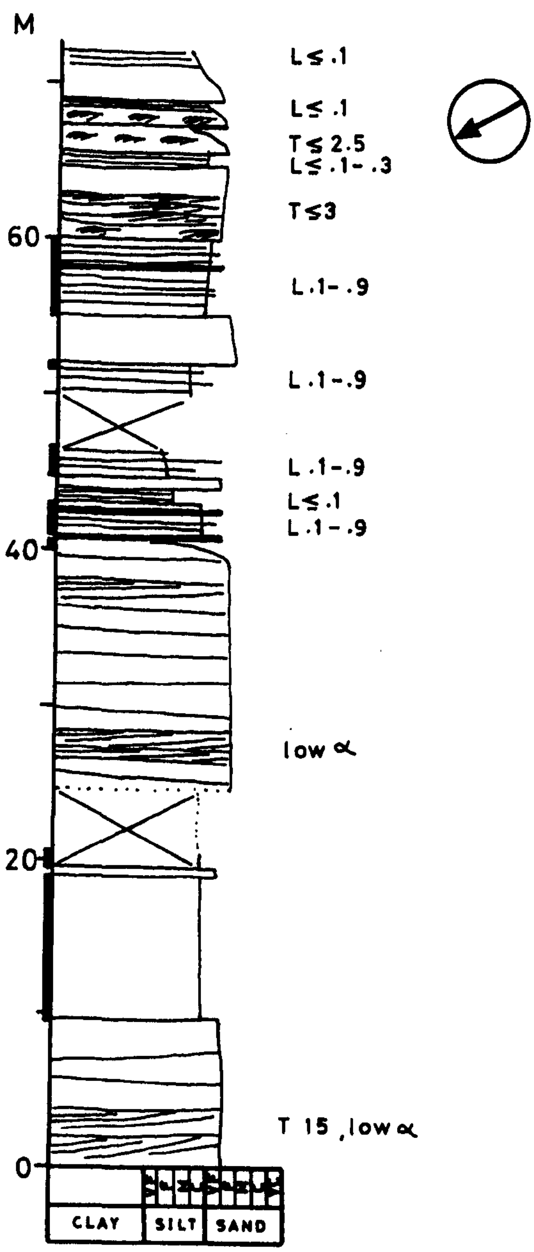
LOG 10 (cont'd) BURWAY HILL, SO 44059422
CARDINGMILL GRIT



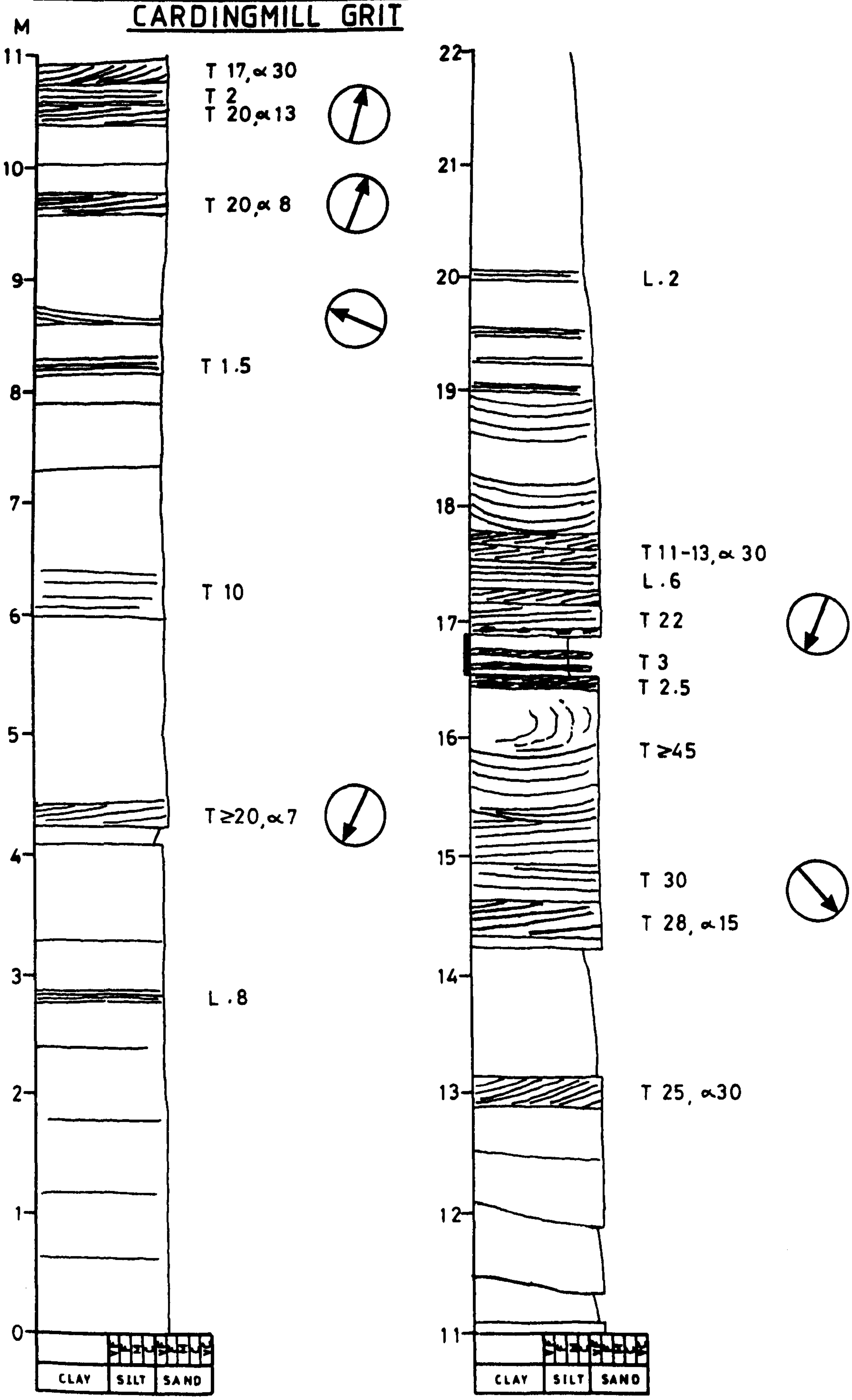
LOG 10 (cont'd) BURWAY HILL SO 44059422
CARDINGMILL GRIT



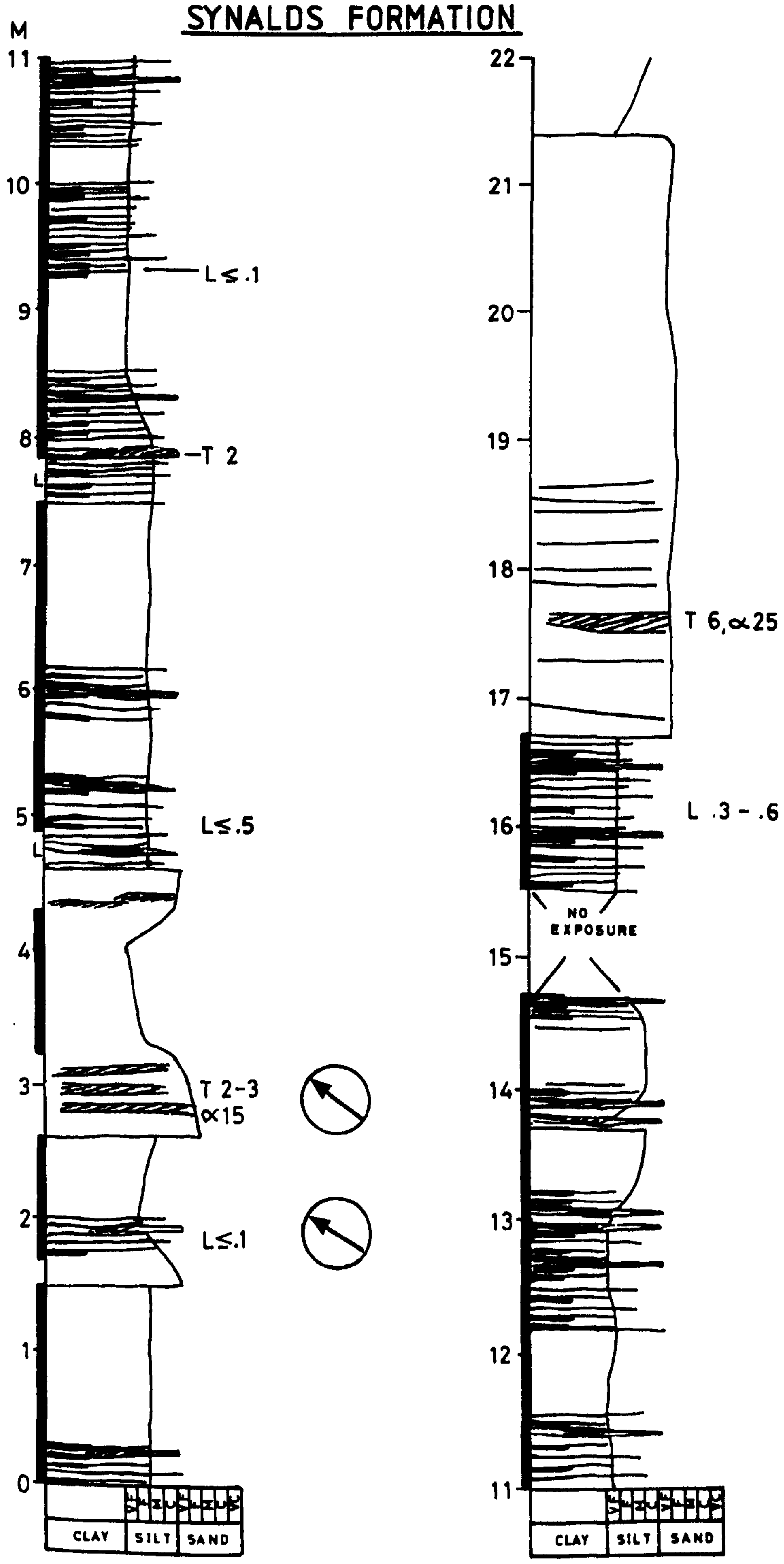
LOG 11 MOUNT GUTTER, SO 40258827
CARDINGMILL GRIT



LOG 12 MOUNT GUTTER, SO 40308853
CARDINGMILL GRIT

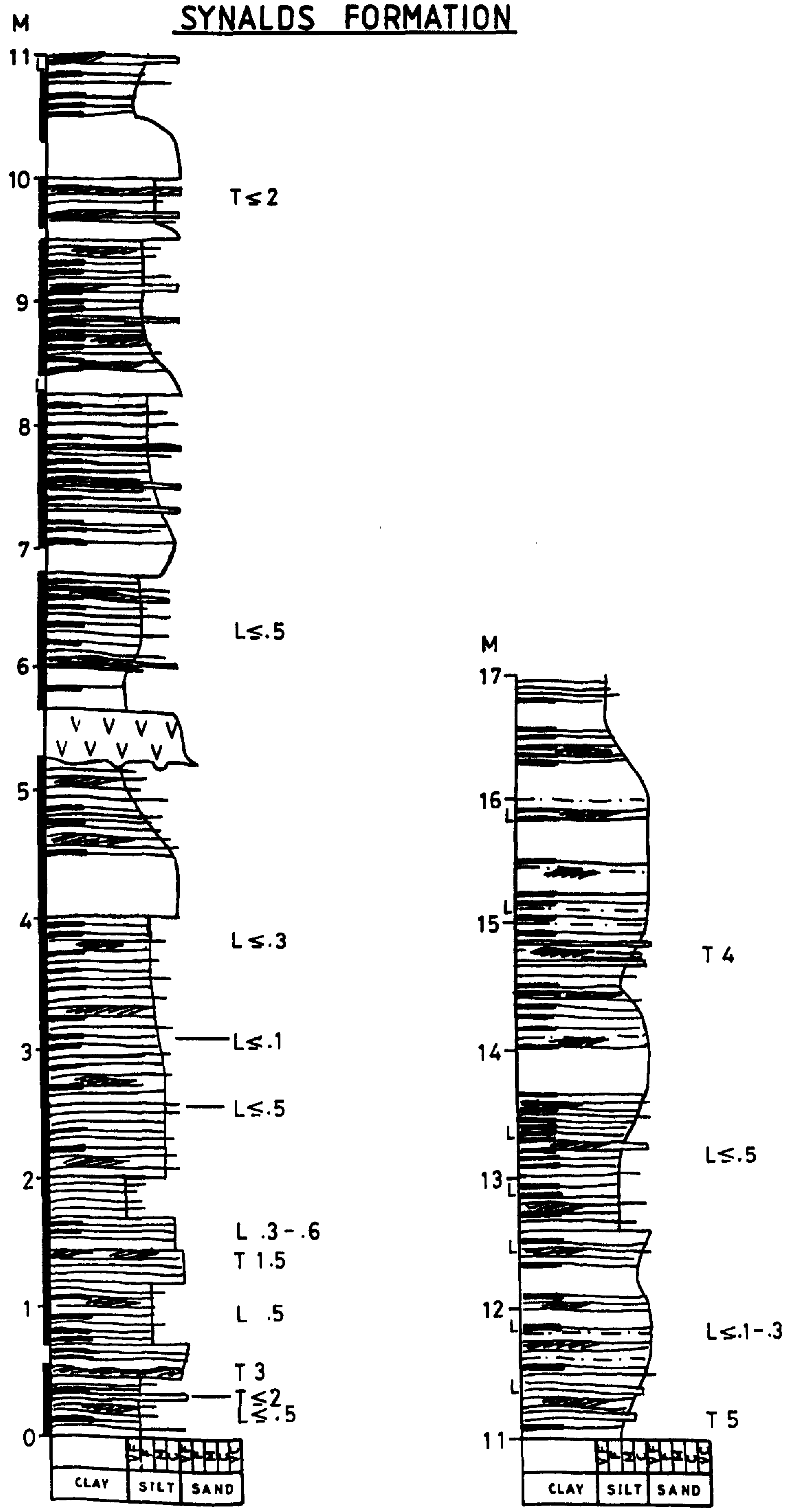


LOG 13 ASHES HOLLOW, GRID REF. SO 43239306
SYNALDS FORMATION

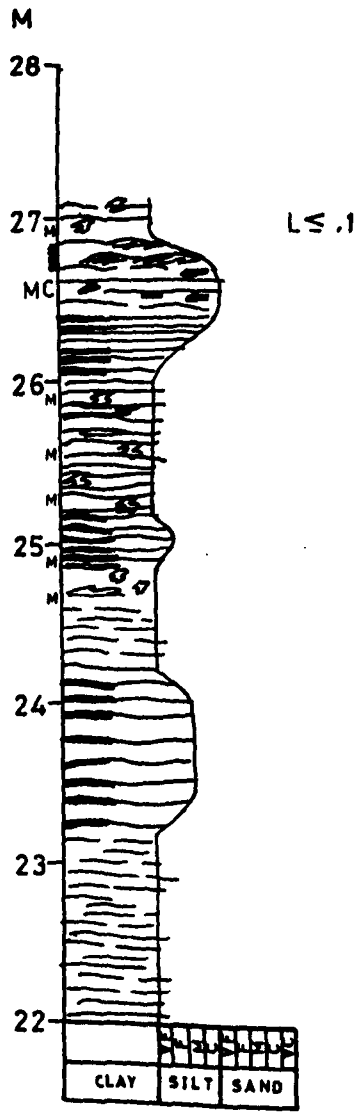


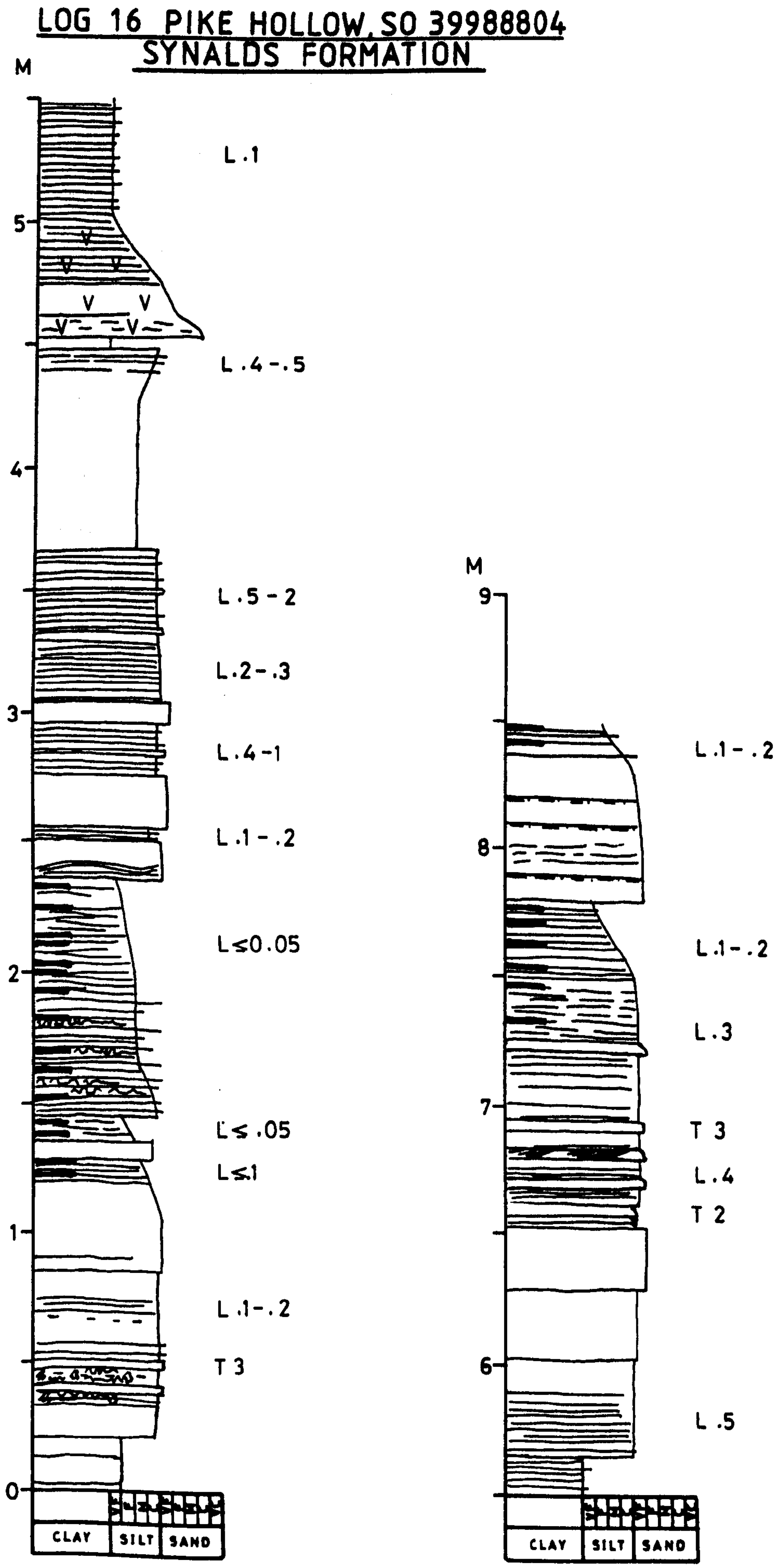
LOG 14 ASHES HOLLOW, GRID REF SO 43209310

SYNALDS FORMATION

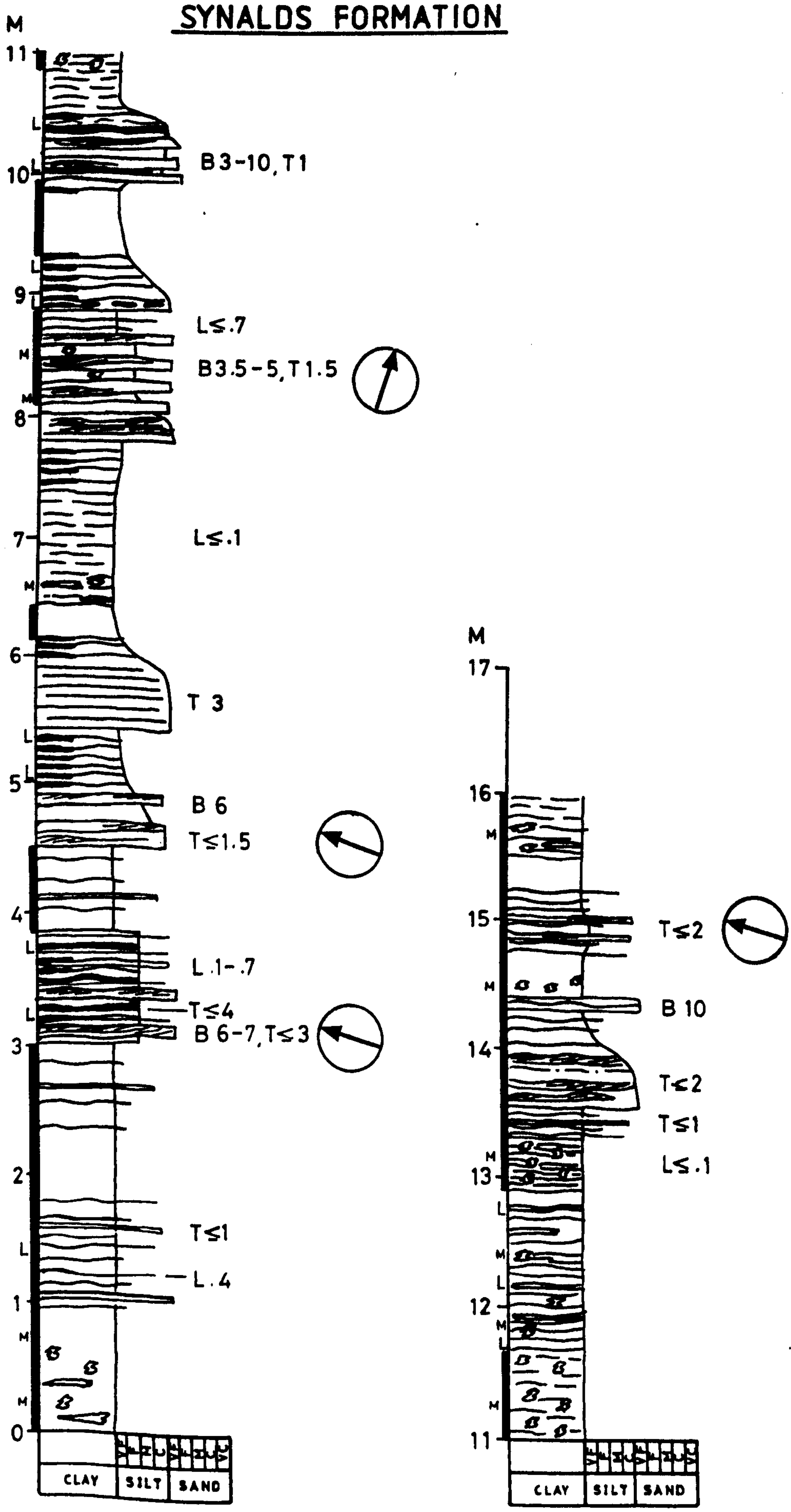


LOG 15 (cont'd) ASHES HOLLOW, SO 43139310
SYNALDS FORMATION

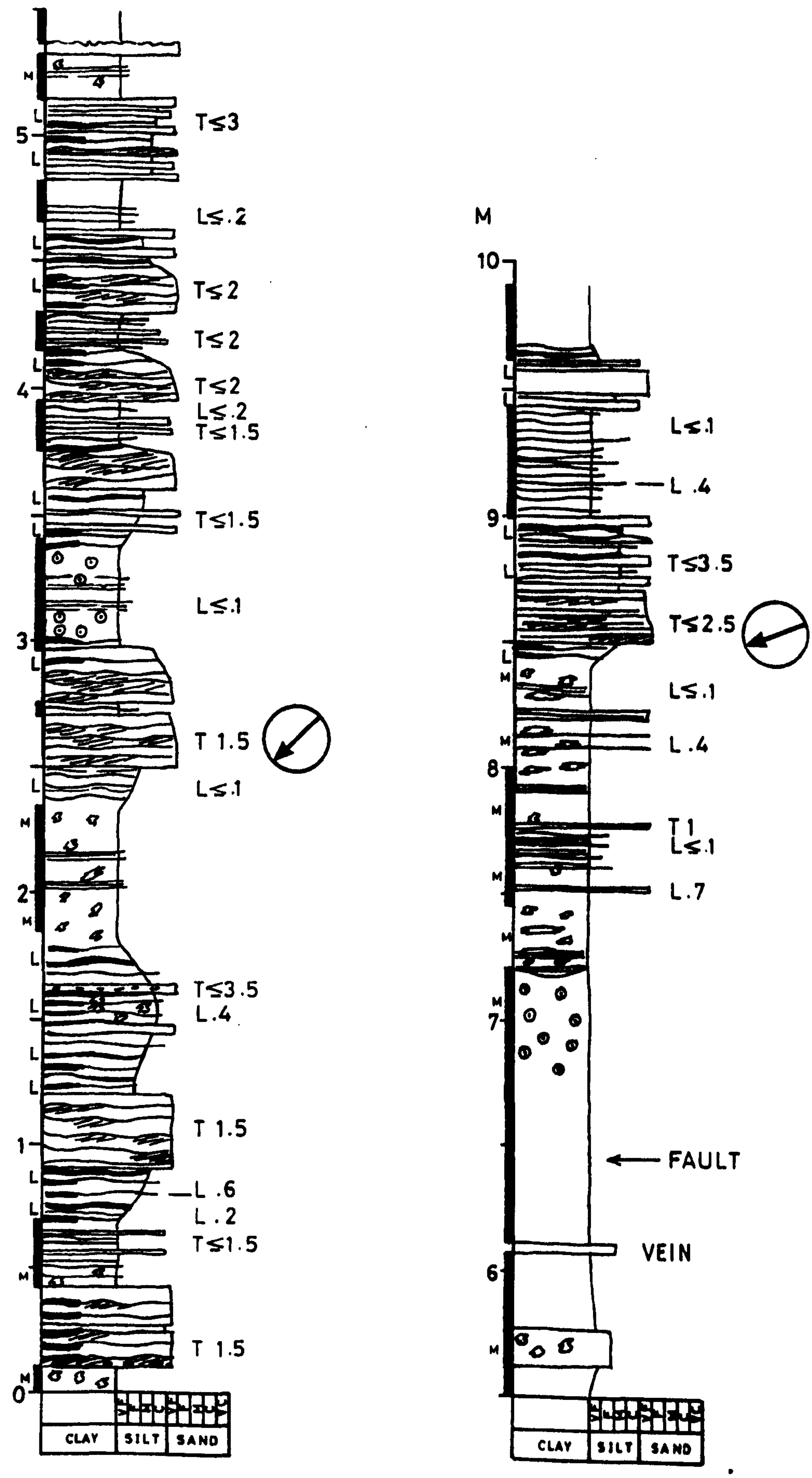



LOG 16 PIKE HOLLOW, SO 39988804
SYNALDS FORMATION
M
5
4
3
2
1
0
L .1
L .4 -.5
L .5 - 2
L .2 - .3
L .4 - 1
L .1 - .2
L≤0.05
L≤ .05
L≤.1
L .1 - .2
T 3
CLAY
SILT
SAND
M
9
8
7
6
L .1 - .2
L .1 - .2
L .3
T 3
L .4
T 2
L .5
CLAY
SILT
SAND

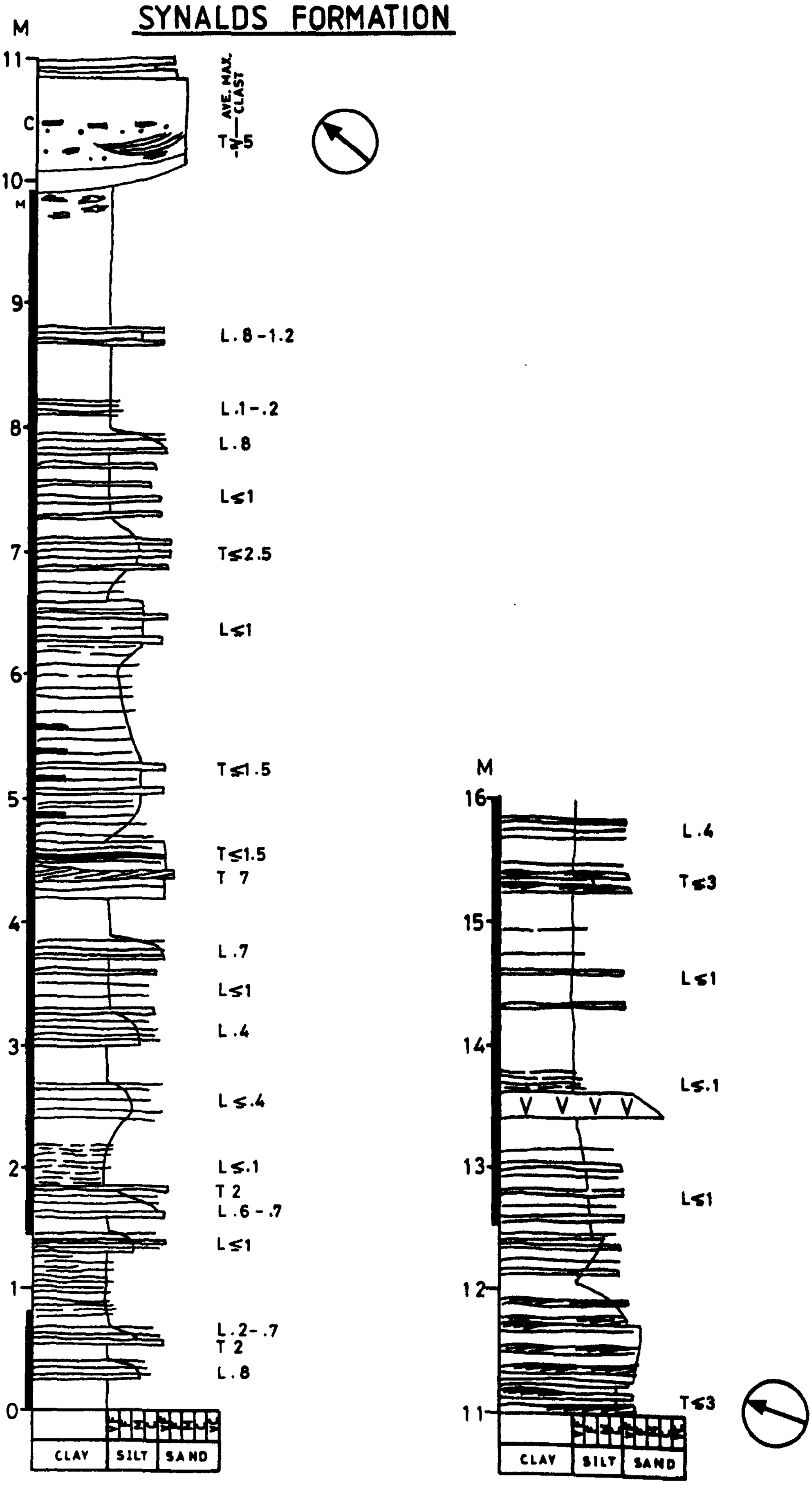
LOG 17 ASHES HOLLOW, GRID REF SO 43159321
SYNALDS FORMATION

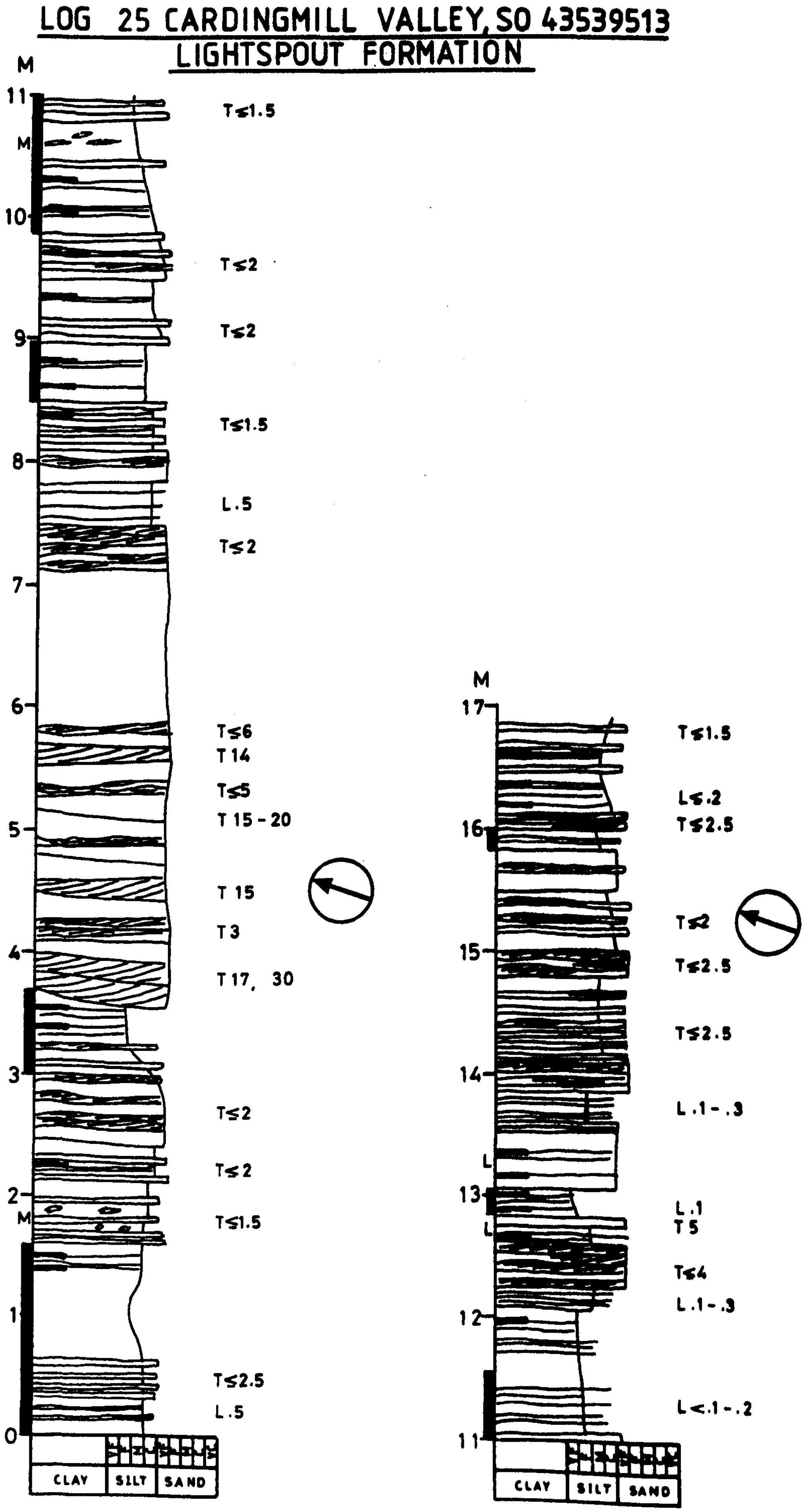


LOG 19 CARDINGMILL VALLEY, SO 44309491
SYNALDS FORMATION

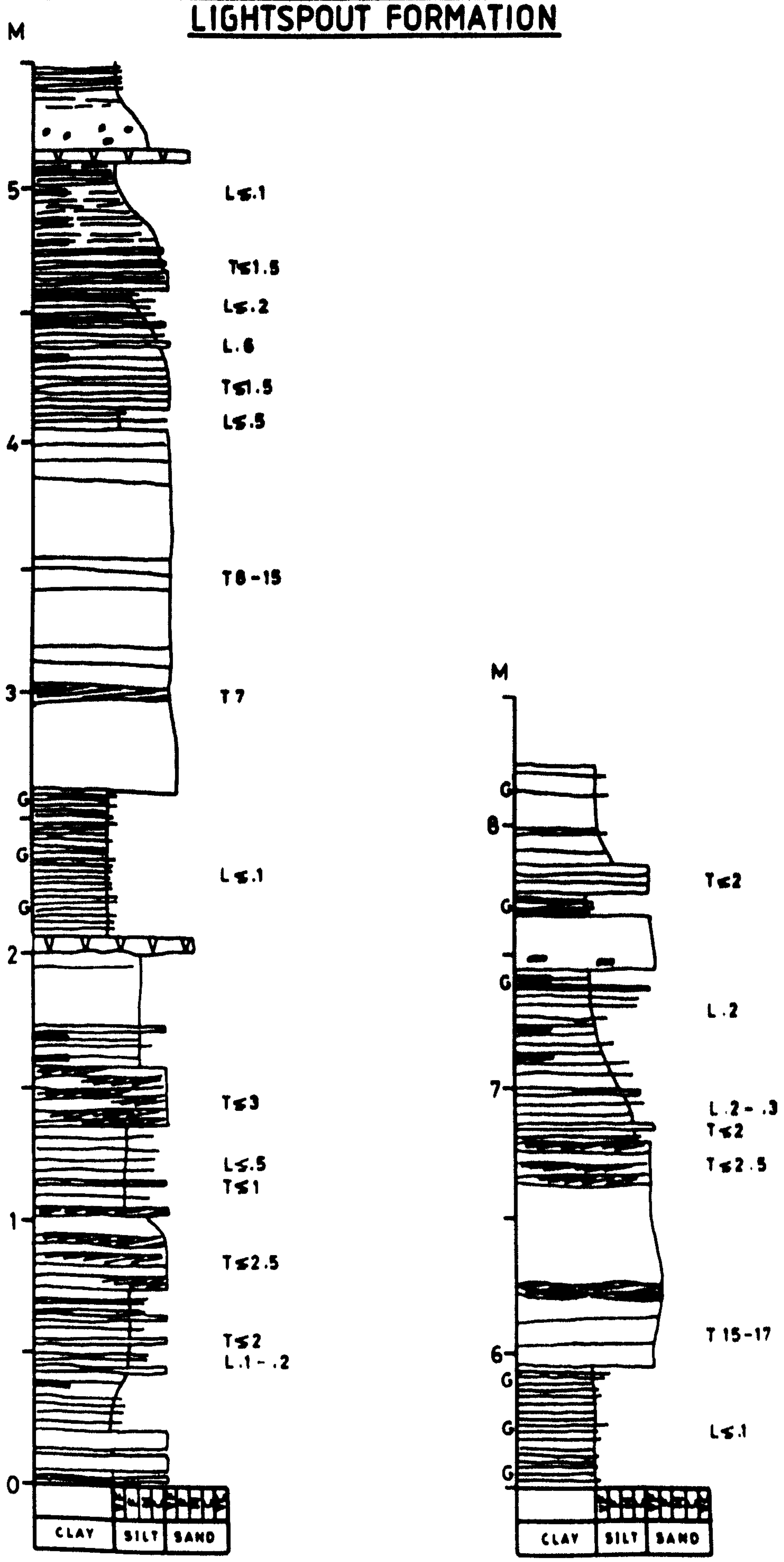


LOG 20 ASHES HOLLOW, SO 42639308
SYNALDS FORMATION

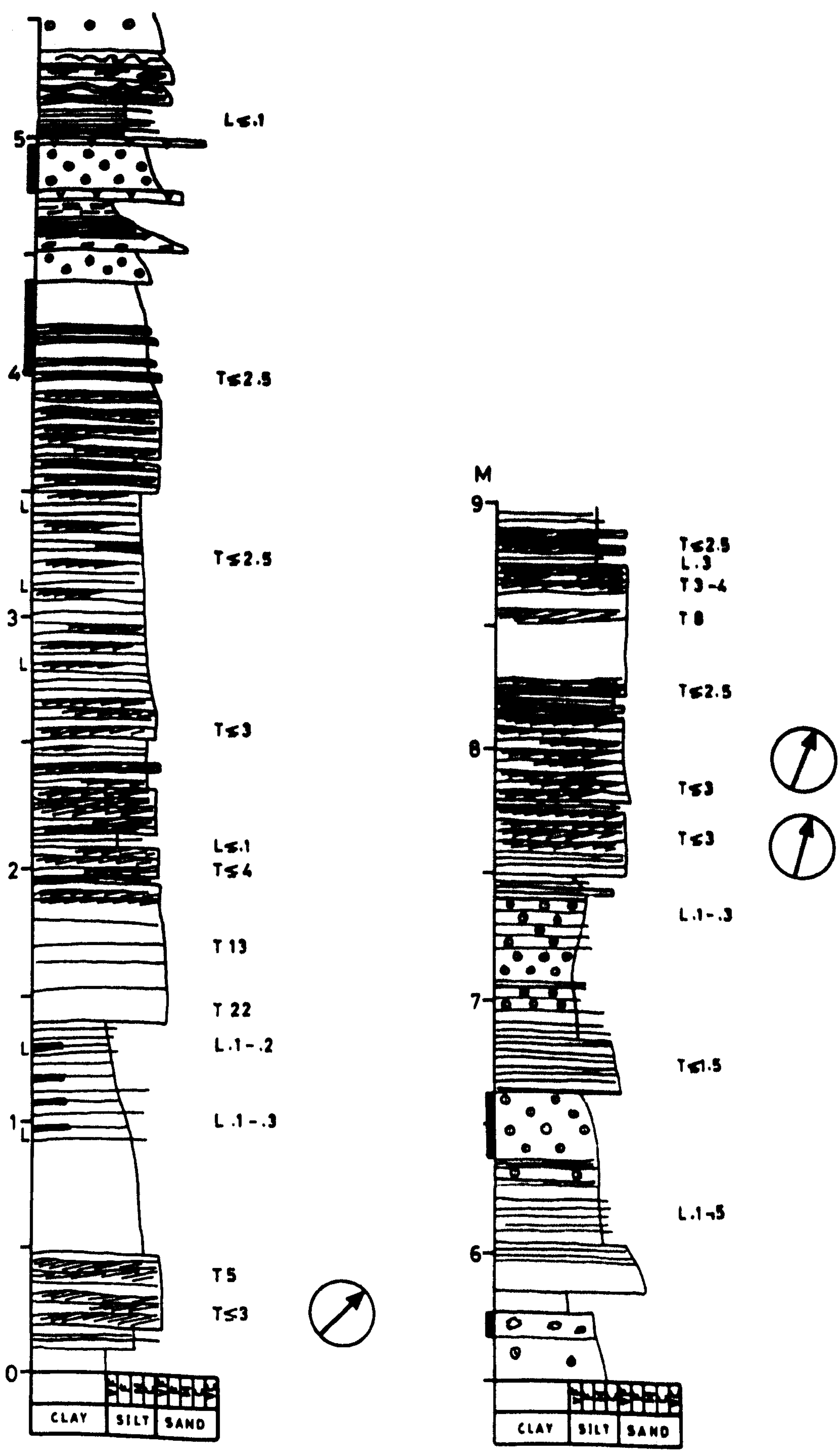



LOG 25 CARDINGMILL VALLEY, SO 43539513
LIGHTSPOUT FORMATION
M
T≤1.5
T≤2
T≤2
T≤1.5
L.5
T≤2
T≤6
T 14
T≤5
T 15 - 20
T 15
T 3
T 17, 30
T≤2
T≤2
T≤1.5
T≤2.5
L.5
CLAY
SILT
SAND
T≤1.5
L≤.2
T≤2.5
T≤2
T≤2.5
T≤2.5
L.1 - .3
L.1
T 5
T≤4
L.1 - .3
L<.1 - .2

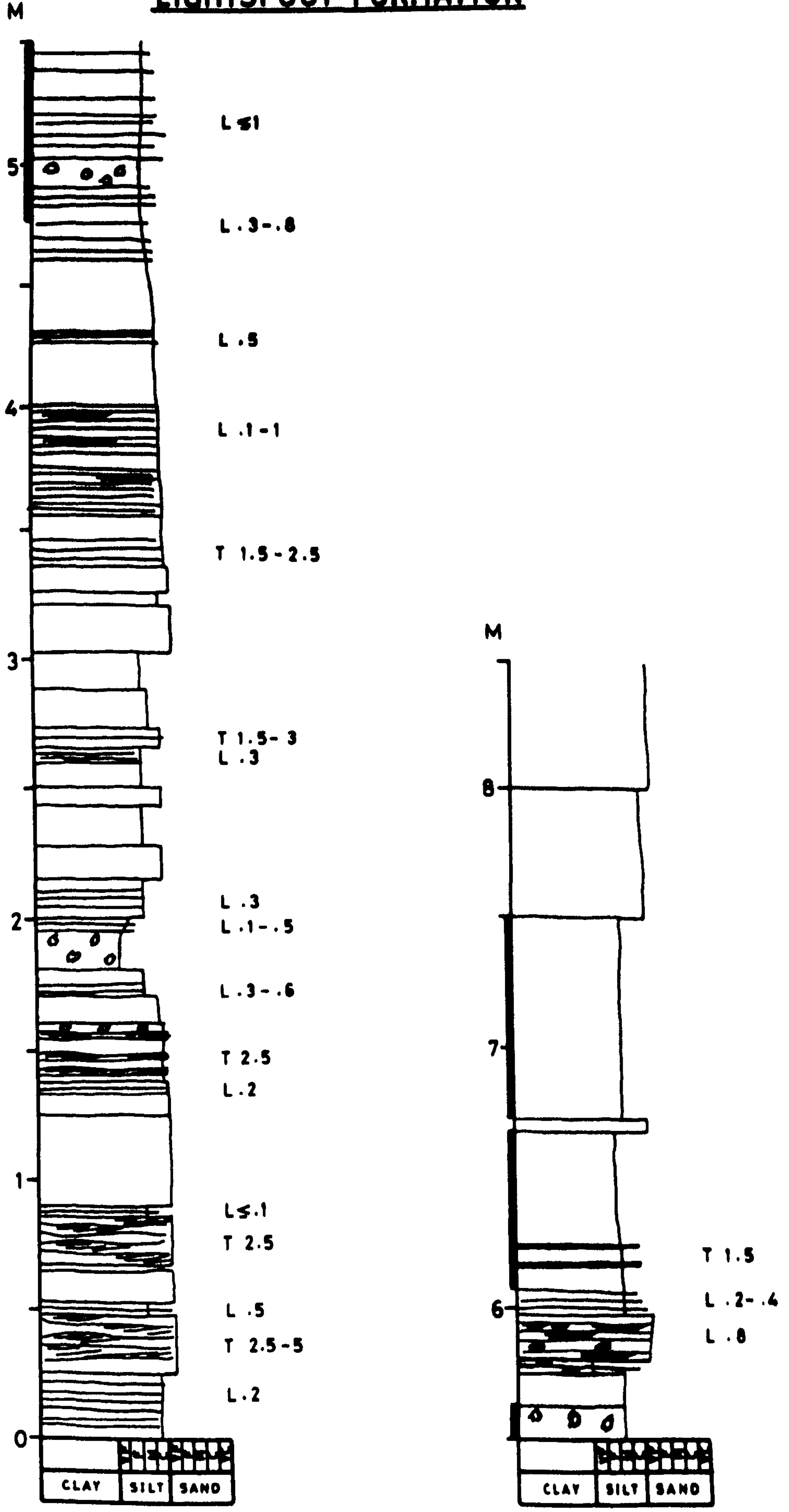
LOG 26 LIGHTSPOUT HOLLOW, SO 43429508
LIGHTSPOUT FORMATION



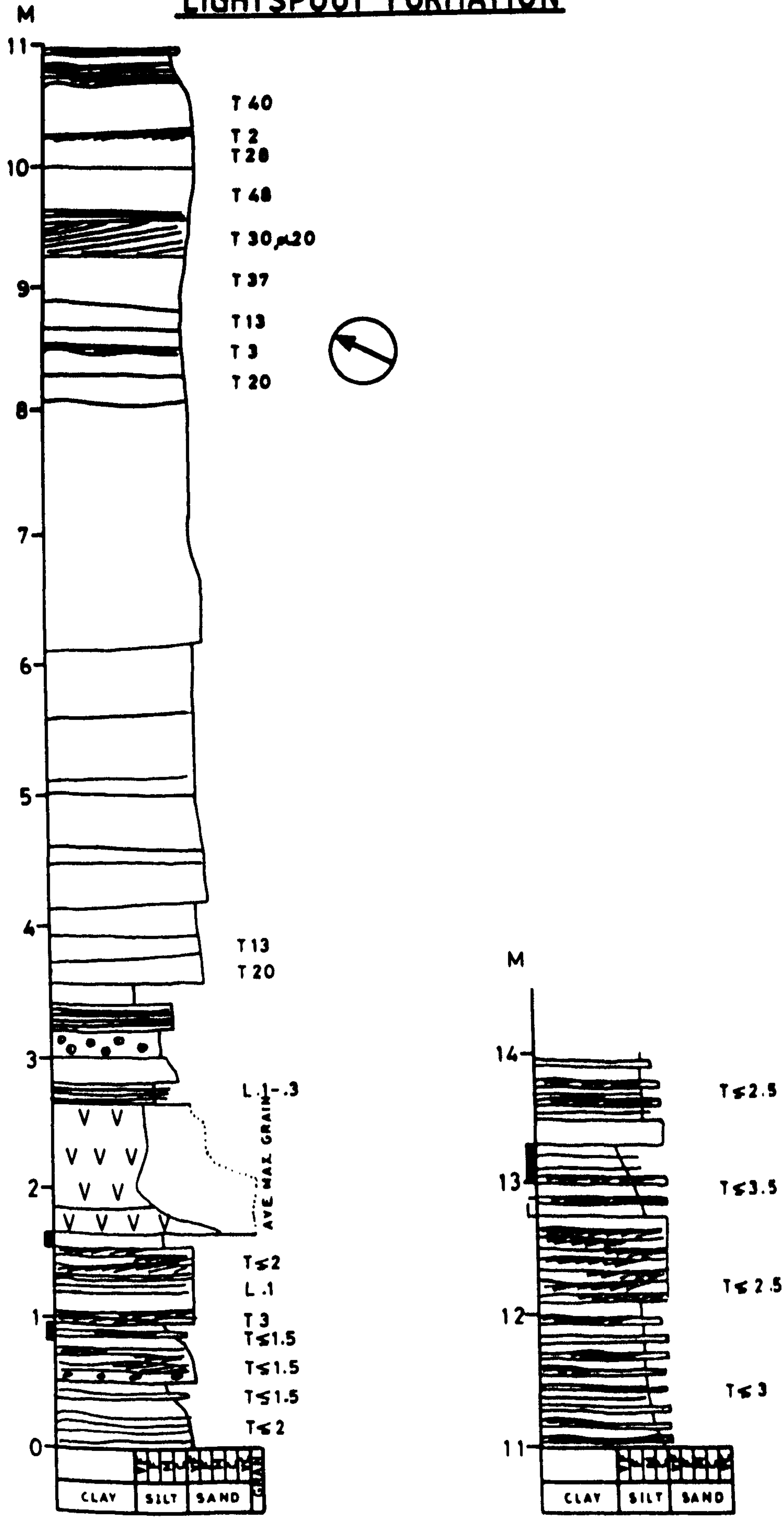
LOG 27 LIGHTSPOUT HOLLOW, SO 43069507
LIGHTSPOUT FORMATION



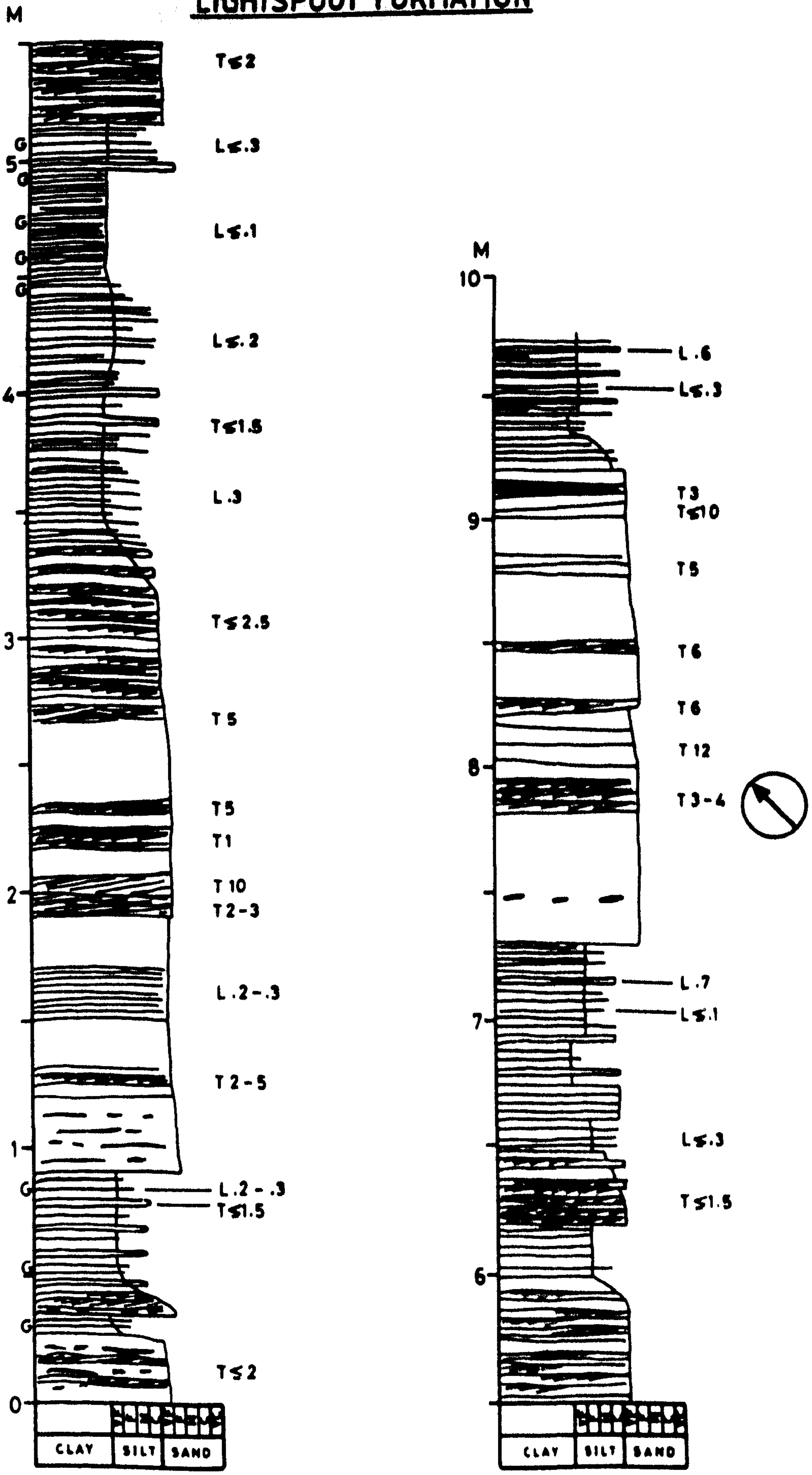
LOG 28 LIGHTSPOUT HOLLOW, SO 43049506
LIGHTSPOUT FORMATION



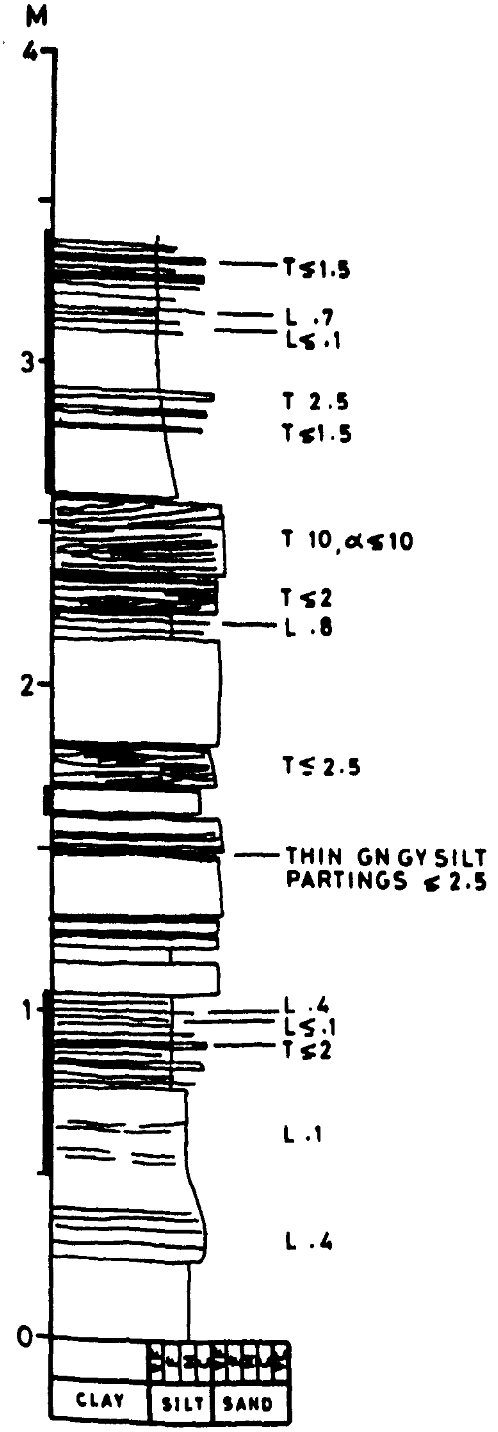
LOG 29 LIGHTSPOUT HOLLOW, SO 43029505
LIGHTSPOUT FORMATION



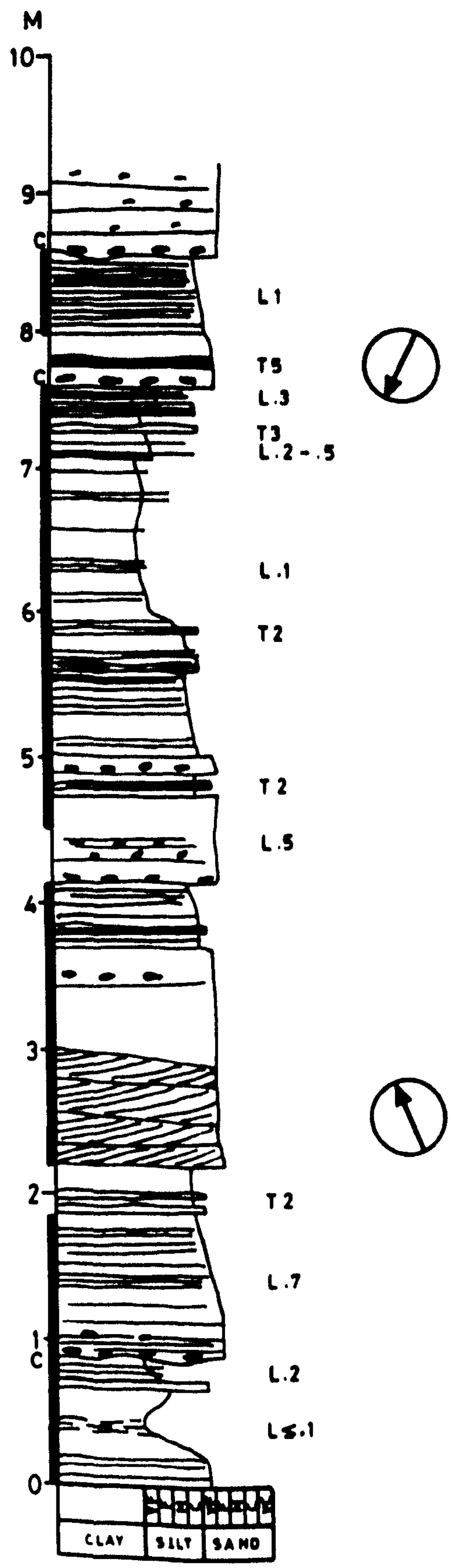
LOG 30 LIGHTSPOUT HOLLOW, SO 42869514
LIGHTSPOUT FORMATION

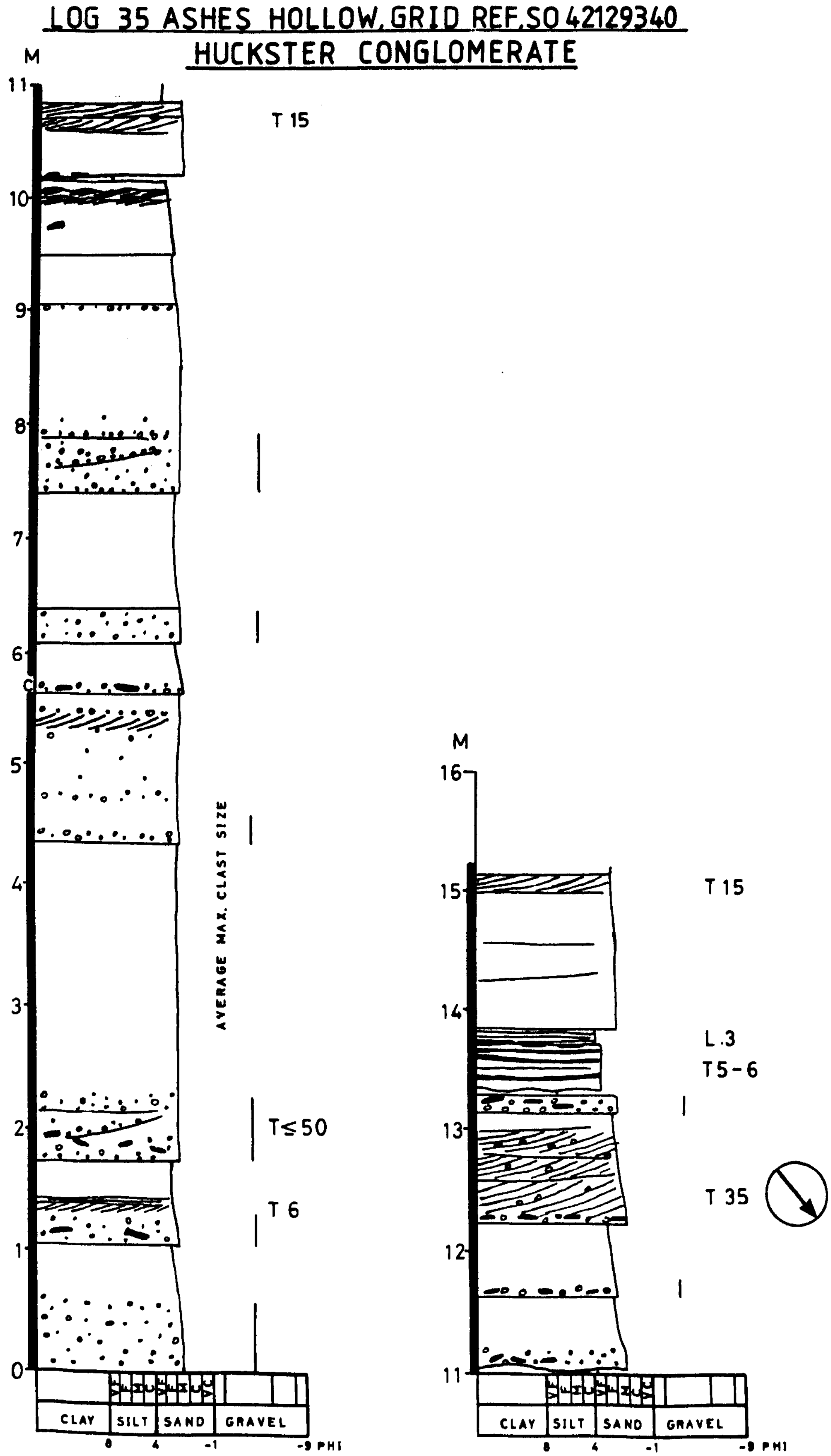


LOG 31 CARDINGMILL VALLEY, SO 43429548
LIGHTSPOUT FORMATION

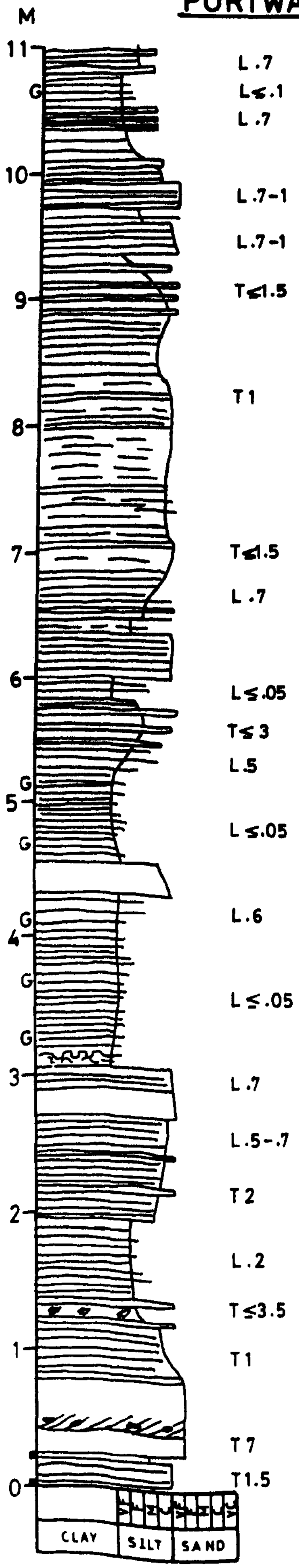


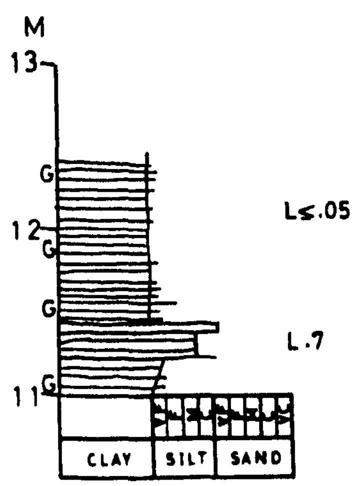
LOG 33 ASHES HOLLOW, SO 42139335
LIGHTSPOUT FORMATION



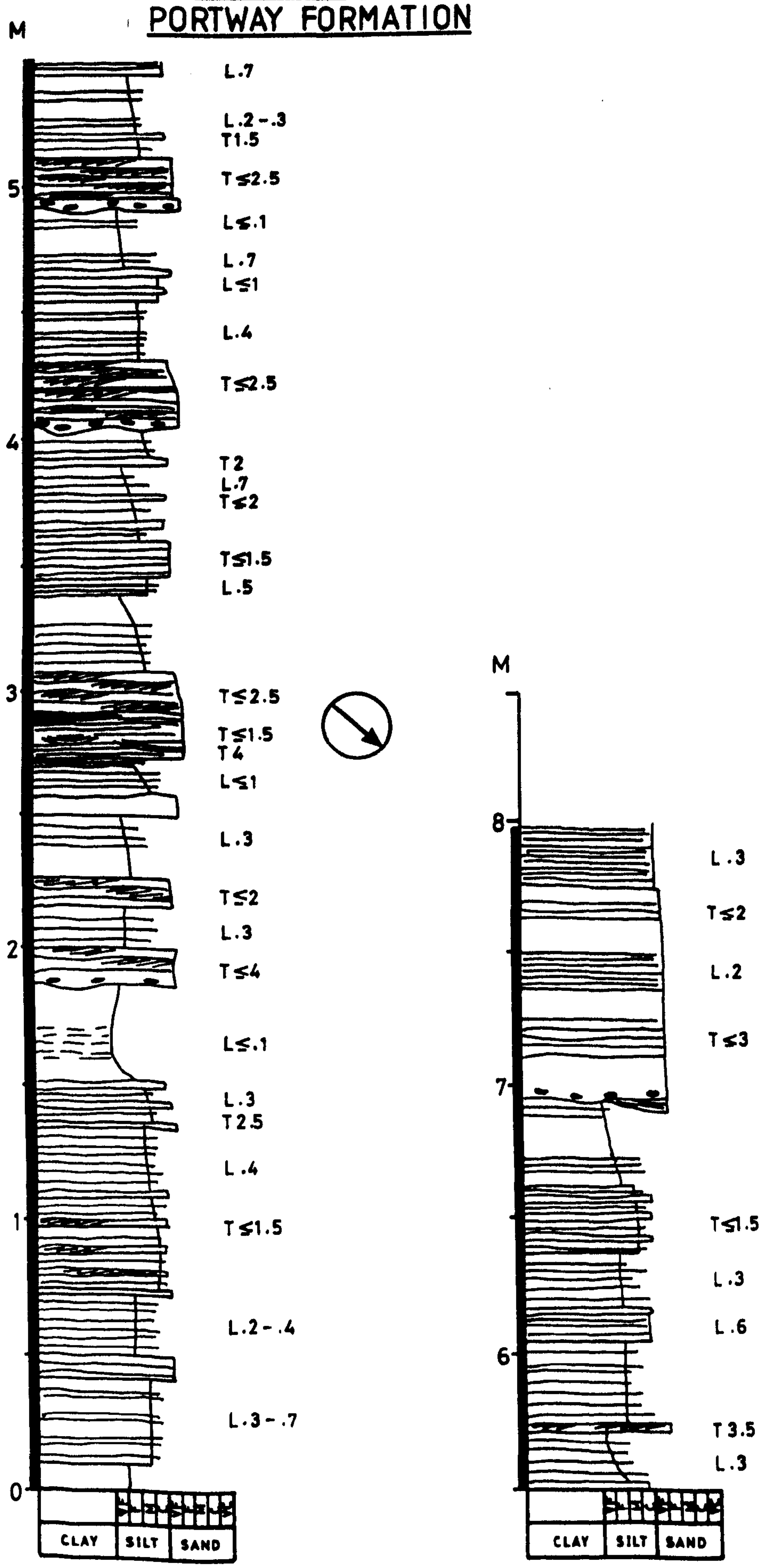

LOG 35 ASHES HOLLOW, GRID REF. SO 42129340
HUCKSTER CONGLOMERATE
M
T 15
AVERAGE MAX. CLAST SIZE
T≤50
T 6
CLAY
SILT
SAND
GRAVEL
8
4
-1
-9 PHI
M
T 15
L.3
T5-6
T 35
CLAY
SILT
SAND
GRAVEL
8
4
-1
-9 PHI

LOG 37 ASHES HOLLOW, SO 41939367
PORTWAY FORMATION

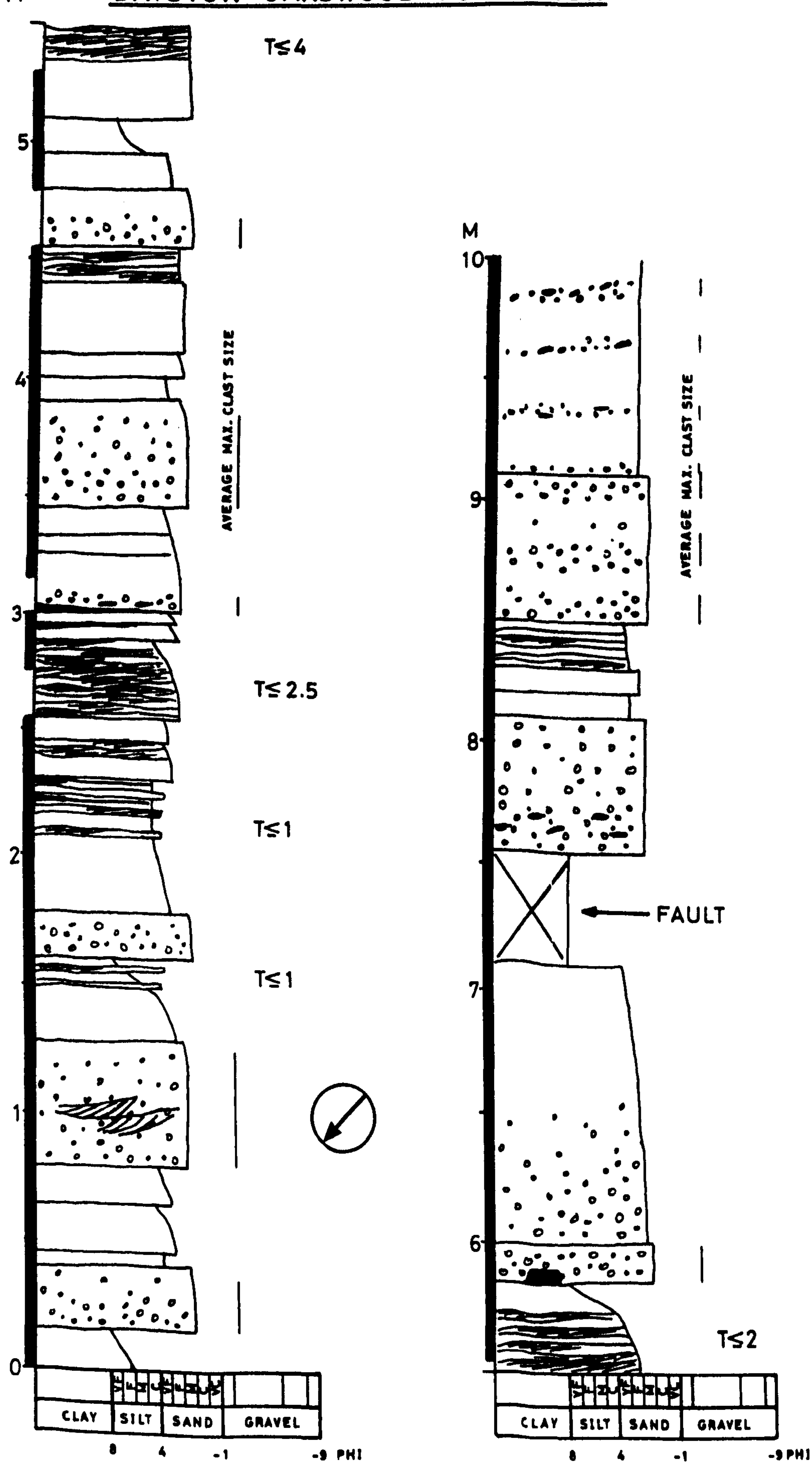




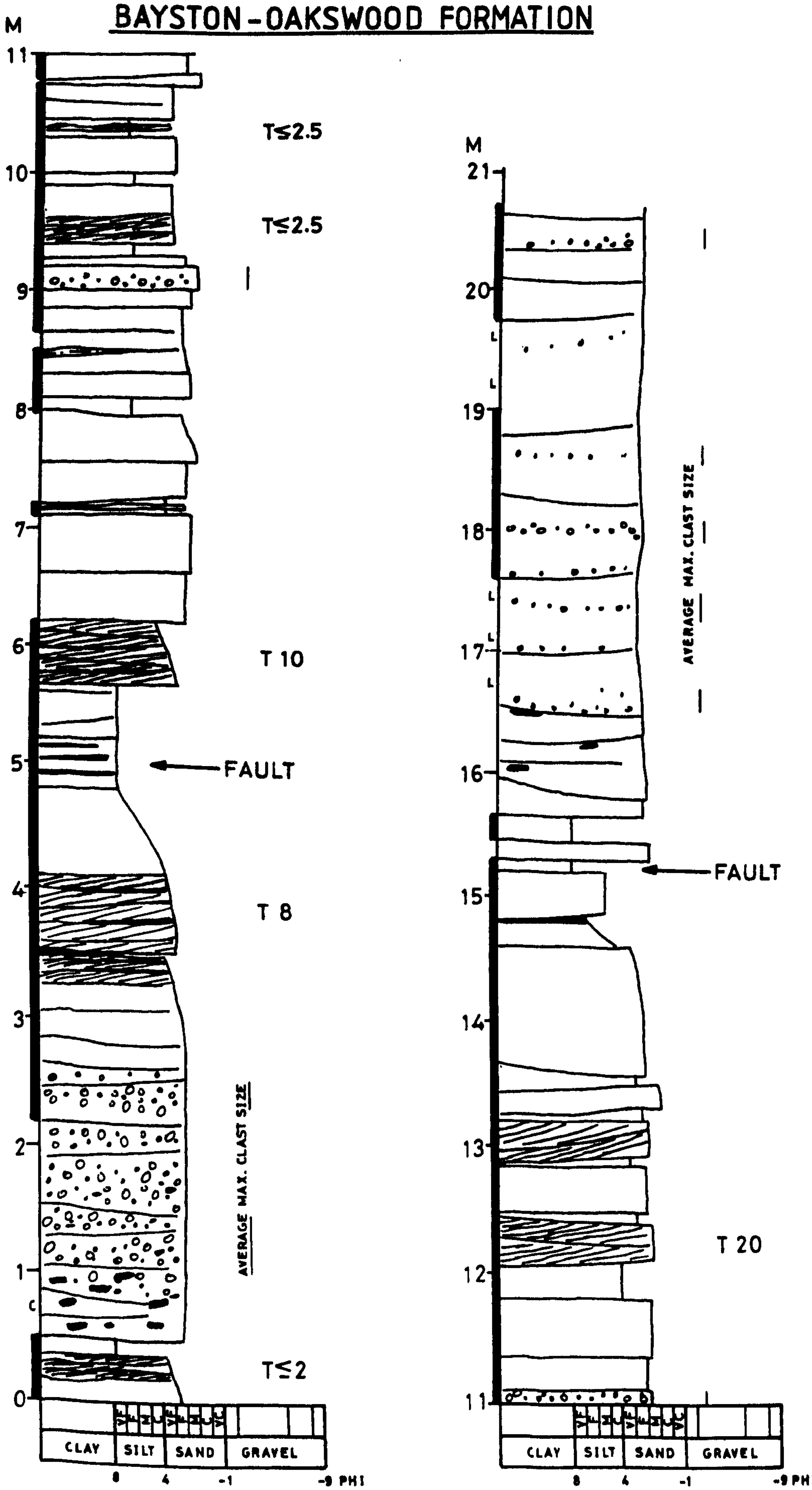
LOG 39 HAWKHAM HOLLOW, SO 43199752
PORTWAY FORMATION



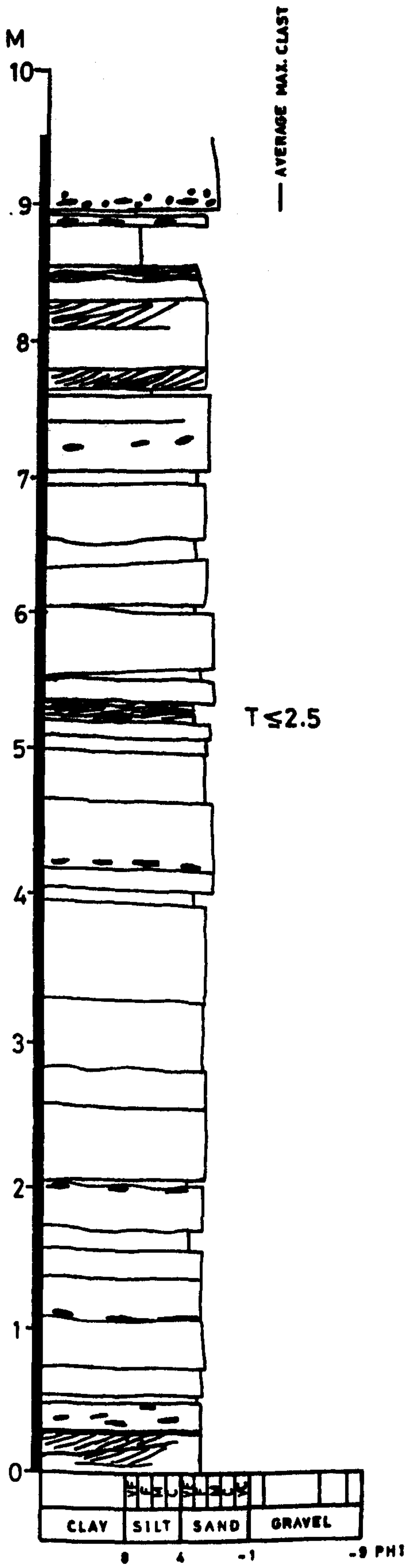
LOG 41 HAUGHMOND QUARRY, SJ 54391485
BAYSTON-OAKSWOOD FORMATION



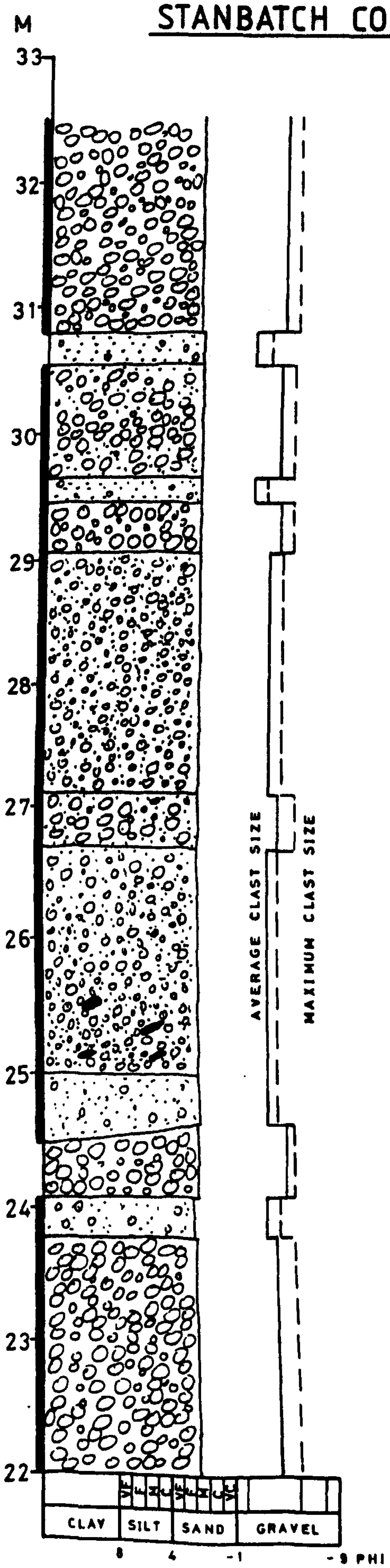
LOG 42 HAUGHMOND QUARRY, SJ 54351474
BAYSTON-OAKSWOOD FORMATION



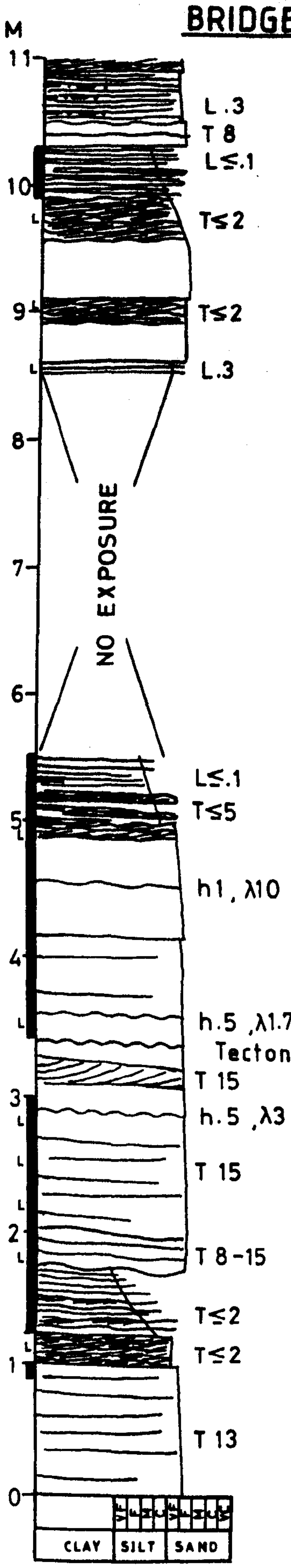
LOG 44 WESTCOTT, SJ 40370128
BAYSTON-OAKSWOOD FORMATION

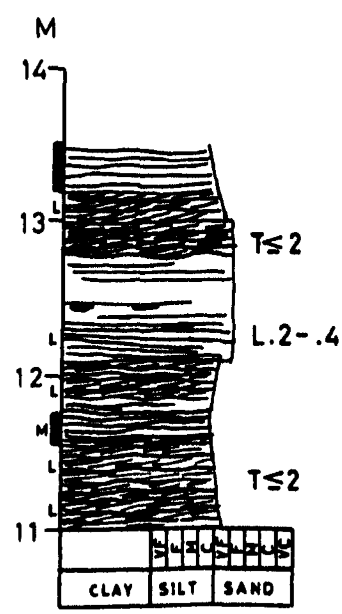


LOG 45 (cont'd) BILBATCH, GRID REF, SO 41549559
STANBATCH CONGLOMERATE

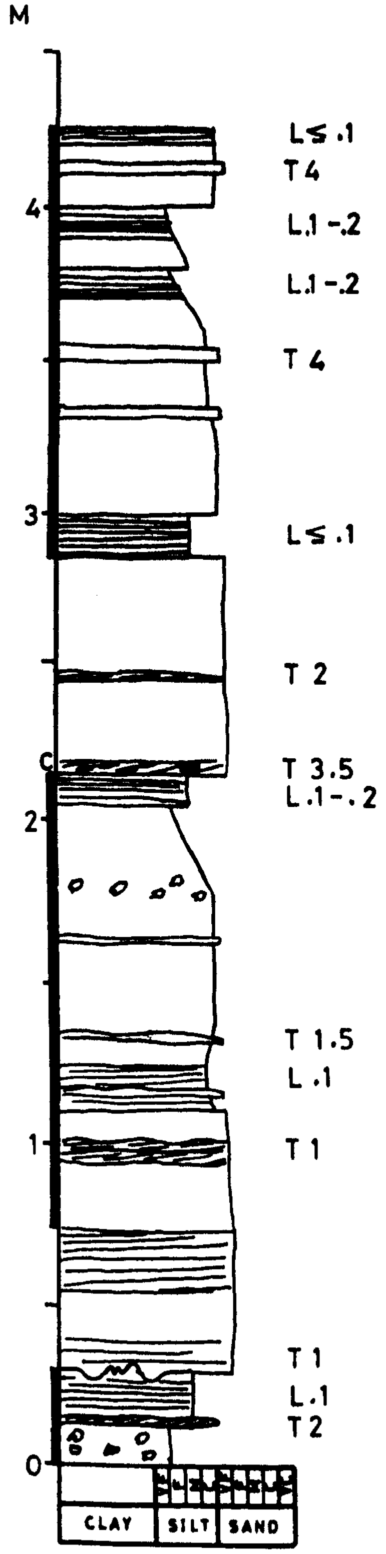


LOG 46 COTHERCOTT, GRID REF, SJ 40703031
BRIDGES FORMATION

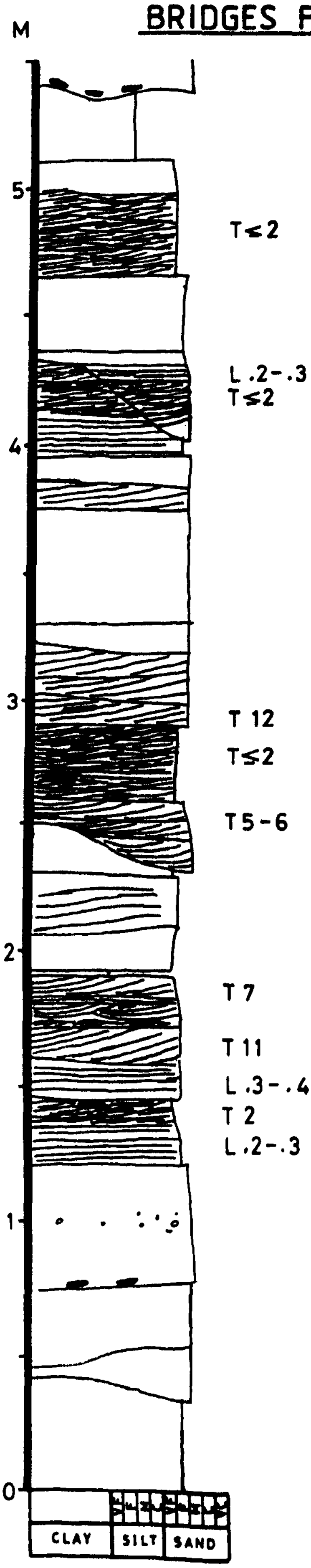


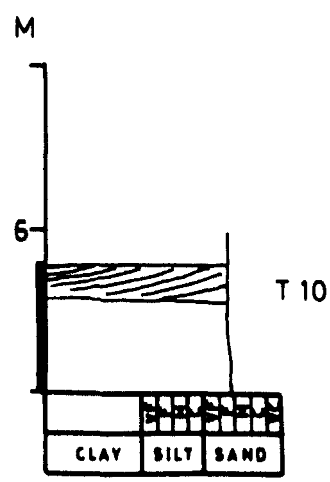


LOG 49 WENTNOR, SO 38929280
BRIDGES FORMATION

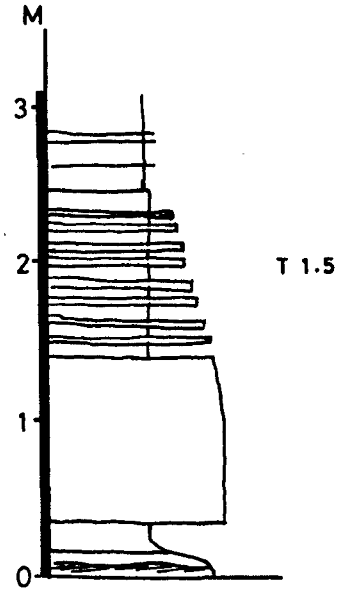


LOG 50 RIVER EAST ONNY, SO 38949613
BRIDGES FORMATION

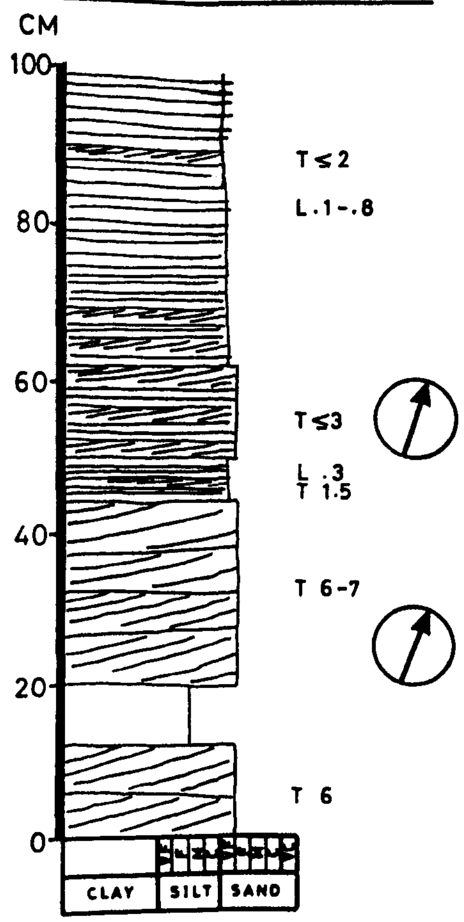




LOG 51 COATES FARM
SO 39199565
BRIDGES FORMATION



LOG 52 RIVER EAST ONNY
SO 38989615
BRIDGES FORMATION



LOG 53 RAGLETH HILL
SO 45409178
RAGLETH TUFF FORMATION