Syn-sedimentary deformation of the Ashover Grit (Pennsylvanian, Namurian, Marsdenian Substage) deltaic succession around Wirksworth, Derbyshire, UK

J. I. CHISHOLM[1*] and C. N. WATERS[2]

[1] *4 Park Street, Loughborough, LE11 2EG, UK*

[2] *British Geological Survey, Keyworth, Nottingham NG12 5GG, UK*

**Corresponding author* *(e-mail: jichisholm@googlemail.com)*

SUMMARY: Three sets of syn-sedimentary slides affecting delta deposits of the Ashover Grit have been identified in the area south and east of Wirksworth, Derbyshire, UK. The kilometre-scale slide masses, which are expressed clearly in the landscape, consist of arrays of fault blocks tilted back towards listric fault surfaces that sole out into underlying basinal mudstone. Relationships in one of the slide arrays, at Black Rock, suggest that the movements took place during the later stages of delta advance. There is a possibility that some minor compressional structures exposed in surface sections away from the fault blocks are also related to syndepositional movements.

This study results from detailed mapping of the area south and east of Wirksworth, Derbyshire (Fig. 1), carried out as part of the recent BGS resurvey of the Derby and Chesterfield sheets. It includes data from cored boreholes (Table 1) drilled in 1979 and 1983 for an aqueduct tunnel linked to Carsington Reservoir. These were logged by one of the authors (JIC) and BGS colleagues. The study has revealed that syndepositional structures are extensively developed within and beneath the Upper Carboniferous (Pennsylvanian, Namurian) Ashover Grit deltaic succession. It builds on the finding of similar structures in much the same stratigraphic interval around Birchover [SK 24 62], about 10 km to the north, where the structures were explored by boreholes (Chisholm 1977; see Fig. 1). There, sandstones of the Ashover Grit thicken on the downthrown sides of listric faults that sole out into the Bowland Shale Formation. Those sandstones are in turn overlain by sandstones that are not affected by the faulting, showing that the growth-fault movements occurred during deposition. This example provides a model for interpretation of the structures described here (see also Edwards 1976).

Comparable syn-sedimentary structures in Namurian deltaic deposits are described in detail from western Ireland (e.g. Gill 1979). Some slide faults there have been shown to develop in sets, each slide being followed by another one further upstream (Wignall & Best 2004). This cannot be demonstrated in the present case, because evidence for the direction of progression is lacking in the Wirksworth examples. In some of the western Irish exposures (e.g. Martinsen & Bakken 1990), extensional zones with listric faults are associated with down-dip compressional zones involving low-angle detachment surfaces. These examples raise the possibility that syndepositional compressional structures may exist in the present study area, and have caused us to examine possible evidence for these.

It has been necessary during the study to take account of the widespread post-Namurian structures that also affect these rocks. These include Variscan and post-Variscan tectonic folds and faults, and superficial Quaternary structures.

1. TECTONIC AND PALAEOGEOGRAPHIC FRAMEWORK

The study area lies within the Carboniferous Central Pennine Basin, which is located between the Craven Fault system to the north and the Wales-Brabant High to the south (Fig. 1). The basin initiated during the late Devonian and early Carboniferous as a series of grabens and half-grabens separated by horst blocks and tilt-block highs, produced in response to a phase of N–S extension and rifting (Leeder 1982; Fraser & Gawthorpe 2003).

The study area lies across the boundary between the Widmerpool Trough and the Derbyshire High (Fig. 1).

During the late Visean and early Namurian, rifting in the Pennine region was replaced by widespread thermal subsidence, and displacements along the block-bounding normal faults became minimal (Leeder 1982). The initial deposition of hemipelagic sediments (the Bowland Shale Formation) draped over the block and basin topography, but did not eliminate it. During the early to mid Namurian (Pendleian to Kinderscoutian), fluviodeltaic sediments of northerly provenance (Millstone Grit Group) entered the northern part of the basin and gradually encroached southwards, burying the palaeotopography. However, the southern part of the Central Pennine Basin remained unfilled until late Marsdenian times, when fluviodeltaic sandstones of northern provenance entered from its eastern side (Jones 1980, fig 13). The Ashover Grit represents the first such fluviodeltaic complex, which prograded from southeast to northwest along the axis of the Widmerpool Trough (Jones & Chisholm 1997). This was followed by the Chatworth Grit, with inflow from the east (Waters *et al.* 2008), and the Rough Rock, which entered this part of the basin from the east (Hallsworth & Chisholm 2008).

During the later Carboniferous, the south part of the Central Pennine Basin, including the study area, was subject to Variscan compression and partial inversion (Fraser & Gawthorpe 2003). This phase of deformation produced open folds along N-S axes; the Crich Anticline (Frost & Smart 1979, fig 59) is an example within the study area (Fig. 1). It also resulted in localized reversed movement on the Bonsall Fault, an early Carboniferous rift fault, with associated growth of the NW-SE trending Matlock Anticline (Smith *et al.* 2005). Extensional faulting followed, mainly on W–E and NNW–SSE lines. These Variscan and post-Variscan structures are distinguishable from the syn-sedimentary Namurian disturbances, in that the former also affect underlying and overlying Carboniferous strata. Minor structures affecting limited exposures cannot be categorized in this way, however.

2. STRATIGRAPHY AND SEDIMENTOLOGY

The Namurian succession throughout the Central Pennine Basin shows a diachronous upward transition from dark grey mudstone, deposited in basinal environments, to interbedded mudstone, siltstone and sandstone deposited in fluviodeltaic environments. Superimposed on this general trend is a cyclical arrangement of lithologies produced by changes of sea level. The mudstone succession, formerly referred to in the Midlands as the Edale Shale or Edale Mudstone Group, is now classified as the Bowland Shale Formation of the Craven Group (Waters *et al*. 2009). In the south part of the basin, the formation extends from the *Cravenoceras leion* Marine Band, the base of which defines the base of the Pendleian Regional Substage (early Serpukhovian Stage), up to strata of the early Marsdenian (R_2) Regional Substage (mid Bashkirian Stage). The fluviodeltaic deposits are classified as the Millstone Grit Group (Fig. 2). For descriptive purposes, we follow the practice of dividing the succession into sedimentary cycles (Holdsworth & Collinson 1988).

The deposits concerned in this study are of mid-Marsdenian (R_{2b}) age, belonging to the $R_{2b}5$ sedimentary cycle, with a likely representative of the underlying $R_{2b}4$ cycle (Fig. 2). The $R_{2b}4$ cycle extends from the base of the *Bilinguites eometabilinguis* Marine Band to the base of the *B. metabilinguis* Marine Band; the $R_{2b}5$ cycle extends from the *B. metabilinguis* Marine Band to the base of the *B. superbilinguis* Marine Band. However, faunas in the R_{2b} bands are very similar, and faunal determinations currently available for the study area do not distinguish $R_{2b}4$ from $R_{2b}5$ bands with certainty. For present purposes (Fig. 2) we assume that the highest band proved in boreholes around Kirk Ireton and at outcrop in Lumb Brook [SK 3313 4676] is the $R_{2b}5$ band, and that the next lower is the $R_{2b}4$ band. The higher band appears to be impersistent, being so far undetected in the Ashleyhay area. Sandstones in the $R_{2b}5$ cycle are named Ashover Grit, but those in the $R_{2b}4$ cycle are currently unnamed: to simplify the text we include the latter in the Ashover Grit. They are the earliest fluviodeltaic sediments of northerly provenance to enter this southeastern part of the Central Pennine Basin, and define the local base of the Millstone Grit Group.

2.1. Lithofacies description

Regional development of the $R_{2b}5$ cycle, with a full account of the lithologies and their interpretation, was described by Jones (1980) and Jones & Chisholm (1997, Units 1-5). The succession in the present study area comprises five broad lithological categories, shown in Figure 2 and described in summary below. Transitions between facies are normally gradational. The descriptions are based partly on the boreholes and partly on exposures.

Facies 1. Basinal mudstones (equivalent to Unit 1 of Jones & Chisholm 1997). These consist of up to 25 m of dark grey, carbonaceous, variably fissile mudstone and siltstone, with marine faunas concentrated in thin beds ("marine bands"). They were deposited on the basin floor in areas distant from encroaching deltas. The marine bands represent deposition during periods of high sea level, whereas intervening unfossiliferous mudstones represent periods of lower sea level and/or lowered salinity.

Facies 2. Delta slope siltstones (equivalent to part of Unit 3 of Jones & Chisholm 1997). These consist of up to 230 m of laminated, micaceous, carbonaceous siltstone, with mudstone and fine-grained sandstone interbeds. The sandstones, up to 1 m thick, show a variety of sedimentary structures: normal grading and disturbed lamination with randomly oriented micas are both common. Thicker sandstones are described separately (*Facies 3*, below). All lithologies show zones of closely spaced minor penecontemporaneous faults and minor slump structures. These sediments are interpreted as the deposits of a variety of gravity flows emanating from delta distributaries into water of varying salinity above delta slopes.

Facies 3. Delta slope sandstones (equivalent to part of Unit 3 of Jones & Chisholm 1997). These form mappable features in the landscape, and are shown on Figures 3, 7, and 10. Most consist of fine- to medium-grained (rarely coarse-grained), pale grey, well sorted sandstone in poorly laminated or massive units up to about 15 m thick, grouped together in sets up to about 30 m thick, interbedded with siltstone of the type described above. Features commonly present are well laminated bed tops, ragged mudstone/siltstone intraclasts, and sharp erosive bases. Sole marks on some beds include small flutes, tool marks and load structures. Also present, but more rarely, are beds of current ripple laminated sandstone up to 25 m thick. The poorly laminated sandstones are interpreted as delta-slope (including outer mouthbar) deposits from confined and unconfined density currents that formed offshore from active fluvial distributaries during flood events; the ripple laminated sandstones are interpreted as deposits of traction currents emanating from the same distributaries.

Facies 4. Delta front and delta top sandstones (equivalent to Unit 4 of Jones & Chisholm 1997). These are in units up to 70m thick, consisting of fine- to very coarse-grained sandstone, with subordinate siltstone. Some sandstone beds contain scattered quartz pebbles up to 3 cm in size. Cross-bedded, parallel-laminated, current ripple-laminated and weakly laminated to massive lithologies are all present. The sandstones are interpreted as the deposits of traction currents in mouthbars and channels of fluvial distributaries. Unusually thick beds of weakly laminated sandstone are a distinctive component in some of the slide blocks.

Comment [SGM1]: These consist of beds of ... sandstone up to 70m thick, with...?

Facies 5. Fine-grained delta top deposits (equivalent to Unit 5 of Jones & Chisholm 1997). These were not encountered in the boreholes, but sporadic exposures show that they form units up to about 15m thick, consisting of siltstone and mudstone with thin beds of fine-grained sandstone and local palaeosols. Deposition is inferred to have been in delta-top environments such as levees and delta-plain lakes, away from the distributary channels.

Comment [SGM2]: siltstone and mudstone packages up to about 15m thick?

2.2. Large-scale facies organization

Jones & Chisholm (1997) described the regional development of the $R_{2b}5$ cycle in the south part of the Central Pennine Basin. It consists of an upward-coarsening deltaic succession bounded by marine maximum flooding surfaces, the *Bilinguites metabilinguis* Marine Band below and the *Bilinguites superbilinguis* Marine Band above. Their account includes deltaic deposits now considered to lie in the $R_{2b}4$ cycle, as discussed above. Palaeocurrent flows were towards the northwest. In the Staffordshire outcrops, some 30 km to the west of the present study area, a tubidite fan is described at the base of the delta slope succession, and an incised valley fill up to 80 m thick represents the minimum sea-level between two marine band maxima. In the Derbyshire outcrops, including the present study area, the period of minimum sea level is represented by palaeosol developments and small locally incised channels in the fine-grained delta top deposits (their Unit 5). The main incised valley fill is located in the extreme south, and does not enter the present study area.

In the present study area, the presumed $R_{2b}4$ cycle (see above) is well represented at Kirk Ireton (Fig. 2) and in the Lumb Brook area. It shows an upward transition from basinal mudstones, with a maximum flooding surface taken to be the *Bilinguites eometabilinguis* Marine Band, into a delta-slope environment where siltstone interdigitates with fine- to medium-grained (rarely coarse-grained) sandstones deposited mainly as turbidite lenses. A marine band, assumed to be the *Bilinguites metabilinguis* Marine Band, ends the cycle. At Kirk Ireton, the presumed R_{2b}5 cycle consists of thin basinal mudstones with a marine band (*Facies 1*) overlain by delta slope siltstones and sandstones (*Facies 2 and 3*). No delta top deposits are preserved (Fig. 2).

In the Ashleyhay outcrop (Fig. 2), the higher marine band has not been found, and the R_{2b}4-5 cycles cannot be separated with certainty. The combined cycle shows an upward transition from basinal mudstones (*Facies 1*) into delta slope siltstones and sandstones (*Facies 2 and 3*). The sandstones are lenticular, some probably channelized; most are poorly laminated or massive, but a few are current-ripple laminated. The sandstones appear randomly distributed through the siltstones; there is no turbidite fan at the base of the delta slope. These delta-slope sandstones are common in some areas but rare, even absent, in others. They are, however, particularly concentrated in the area of listric faulting at Ashleyhay (see below). The delta-slope siltstones and sandstones pass up into the sandstone-dominated part of the cycle, where sandstones (*Facies 4*) are coarser-grained and were deposited in distributary channels and mouthbars. An upper sandstone unit deposited after an erosional episode caused by a fall in sea level was recognized regionally by Jones & Chisholm (1997), but cannot be distinguished with certainty in the present study area due to lack of suitable exposures. A variable thickness of finer-grained beds, with local palaeosols *(Facies 5)*, completes the cycle. The overlying *Bilinguites superbilinguis* Marine Band has been recorded at several places in the present study area (Frost & Smart 1979, p.135).

3. SLIDE STRUCTURES

Slide structures have been recognized in three separate areas (Fig. 1). In each of these they form groups of two or three fault blocks, where the faults downthrow in the same direction and where strata in the hangingwall are tilted back towards the fault planes. The widest fault block measures 1650 m along the fault plane, and the largest vertical displacement is about 110 m. In these respects (Table 2) the slide structures are broadly similar to those described at Birchover, where listric geometry of one of the faults was confirmed by boreholes. The direction of downthrow, broadly towards the northwest, is parallel to the regional palaeoflow direction of the Ashover Grit (Jones & Chisholm 1997). The maps showing the outcrop disposition of the fault blocks (Figs 3, 7, 10) are based on landscape features and bedrock exposures, with some borehole evidence. Evidence for possible slide movements outside the fault blocks is examined in a later part of the paper (Section 4).

Comment [SGM3]: = Jones & Chisholm 1997, or Chisholm 1977?

3.1. Black Rock, Cromford

On the hillside south of Cromford, a set of faults affects the Ashover Grit (Figs 3-6). The width of the faulted zone is about 1650 m. Sandstones in individual fault blocks are tilted back towards the fault planes at angles up to 25° (corrected for tectonic dip), and vertical displacements across the faults are estimated at up to c.80 m (Fig. 4). The succession in the fault blocks differs from that in the unfaulted area above the fault blocks in several respects, partly shown in Figure 4. These features are comparable with those described at Birchover (Chisholm 1977; Table 2), suggesting that the faults are syn-sedimentary listric faults linked to one or more low-angle detachment surfaces in the underlying mudstones, rather than tectonic faults linked to the mineral vein system in the limestone beneath (as was implied by earlier maps: British Geological Survey 1978; Ford 2005) or Quaternary landslide detachment planes.

3.1.1. Succession in the unfaulted footwall area south of the faults (Figs 3, 4)

The Peak Limestone Group (Visean age) is overlain by a condensed succession of basinal mudstone (late Visean to mid Namurian age), which passes up through delta-slope siltstones (*Facies 2*) to delta-top and fluvial sandstones (*Facies 4*) of the Ashover Grit. Delta-slope sandstones (*Facies 3*) are few in this area. The delta-top and fluvial sandstones, exposed in old quarry sections along the escarpment, are fine- to very coarse-grained, with scattered quartz pebbles. They form two separate units (Fig. 4).

The *older unit*, at Barrel Edge, consists of well-bedded lithologies, with some faintly laminated and massive beds. Scattered quartz pebbles are up to 20 mm in size. Cross bedding, in sets up to at least 3.5 m thick, shows consistent palaeoflow to the northeast (Figure 3). At Barrel Edge Quarry (SK 2947 5565), a 20-m thick section shows faintly laminated to massive sandstones dipping NE at angles up to 19° (corrected for tectonic dip), with a few thin beds of sandy siltstone. Some beds are erosive-based, with concentrations of mudstone and siltstone intraclasts. The depositional dips suggest that these sandstones are the toesets of large (>20m thick) low-angle cross-beds, part of the same unit as that exposed at Barrel Edge, 300 m to the south (Figure 4).

The *younger unit*, at Cromford Moor, consists mainly of cross bedded sandstone in sets up to at least 2.5 m thick, showing palaeoflow to the northwest. Quartz pebbles are up to 25 mm in size. Field relations (Figure 3) suggest that it cuts down into the top of the highest fault block, though exposures of the contact are lacking.

3.1.2. Succession in the fault blocks

Exposures in the fault blocks show well bedded sandstone overlain by faintly laminated sandstone; unexposed siltstone is inferred beneath (Figures 3 & 4). The well bedded, partly cross bedded, sandstone ranges up to very coarse-grained, with scattered quartz pebbles up to 15 mm in size, and extends across the full width of each fault block. Palaeoflow was consistently to the northeast. The faintly laminated sandstone, up to very coarse-grained and locally granular, with rare quartz pebbles up to 30 mm in size, rests erosively on the well-bedded sandstone (Fig. 6A, C). It ranges up to 33 m in thickness (at Black Rock itself: SK 293 557), but is laterally impersistent and dies out within some of the fault blocks (Fig. 3). Both lithologies are tilted back towards the fault planes at comparable angles (Fig. 4). The differences in succession between the fault blocks and the unfaulted footwall area are enough to show that the fault blocks are not Quaternary landslides derived from the hillside above, so must be explained in terms of slide processes active during Ashover Grit deposition.

The most proximal slide, at Black Rock, provides the best exposures of structural and sedimentological relationships within a slide block, and between it and the footwall (Fig. 6). The unusually thick development of faintly laminated sandstone in the slide block suggests a close link between that lithology and the slide movement, and the strongly erosive base (Fig. 6) suggests that deposition was preceded (or accompanied) by incision. The general similarity of dip between it and the underlying well-bedded sandstone can be explained in two ways: either that both had formed before the slide moved, or that the faintly laminated sandstone was deposited in the hangingwall hollow as the slide developed, with the lamination conforming to the shape of the hollow. Both mechanisms may have operated in the Black Rock slide.

In the footwall, at Barrel Edge (Fig. 4), there is no equivalent of the thick, faintly laminated sandstone of the fault block. The footwall succession (see above) consists of cross-bedded sandstone, including large, inclined, faintly laminated beds with intraclasts and siltstone interbeds exposed at Barrel Edge Quarry. Again, two explanations are possible: either the thick, faintly laminated sandstone formed only in the fault blocks, or it formed in both areas and was eroded and replaced by a later sandstone in the footwall area.

In view of the uncertain interpretation of relationships outlined above, it is not possible to provide a detailed succession of events in the Black Rock slide, but it seems likely that the faintly laminated lithology was first deposited in the hangingwall hollow during the slide movements, due to an acceleration of the river flow across the developing fault scarp. Any submerged nickpoint formed there would probably have progressed upstream into unconsolidated sand, leading to incision of part of the river channel, and deposition of faintly laminated sand in the footwall area also. However, the footwall succession contains large inclined beds unlike those in the slide block, suggesting that deposits there were eroded and redeposited after slide movement was complete.

Within the slide blocks at Birchover (Chisholm 1977), similar relationships between erosive-based, faintly laminated sandstone and pre-existing well-bedded sandstone suggest that events may have followed a similar course in some of the fault blocks there also.

3.2. Ashleyhay

A group of six faulted masses around Ashleyhay is interpreted as two sets of listric fault blocks, each linked to a low-angle detachment surface (Figs 7-9). This interpretation is based on similarities to the examples at Birchover (Chisholm 1977), particularly the size and shape of the fault blocks and the back-tilting of strata towards the faults. The width of the fault blocks ranges up to 1300 m; the tilt angle varies up to 12°, based on dip slope angles at surface (Fig. 7), but up to 15° in borehole M1 (Fig. 8). Vertical displacements up to about 140 m can be inferred from the constructed sections (Fig. 8). A feature not evident at Birchover is a slight doming of the strata in the fault blocks (Fig. 7). The slide movements were directed towards the northwest and west, subparallel to the regional palaeocurrent direction, and affected both delta slope and delta top deposits. Delta slope sandstones *(Facies 3)* are thicker and more numerous in the fault blocks and adjacent footwall areas than they are in areas east of the outcrop. Delta top deposits (*Facies 4*) are preserved in two slides only here, and are poorly exposed (unlike at Black Rock and Birchover), so firm evidence for the timing of the slides is lacking. The constructed sections (Fig. 8) give an impression of sudden thickness changes of the delta slope sandstone at the most proximal fault, with an implication of fault movement during deposition of the slope deposits. This is misleading: the sandstones are markedly lenticular, and thickness changes between the outcrop (Fig. 7) and the section line can explain the different thicknesses, so there is no necessary implication of a fault movement at that time. This conclusion is favoured also by the similarity of grainsize and facies between footwall and hangingwall sandstones. The simplest interpretation is that the timing of the slides was the same as at Black Rock.

3.3. Kirk Ireton

A few kilometres west of Ashleyhay, an outlier of Ashover Grit is centred on Kirk Ireton (Fig. 1). The succession comprises an alternation of delta-slope siltstone and sandstone *(Facies 2 and 3)* above basinal mudstone *(Facies 1)*; no delta top sandstone is preserved (Fig. 2). An area of about 0.5 km^2 at the north end of the outlier near Callow incorporates southward dipping sandstones in three fault blocks (Fig. 10A), in a deformation style comparable with that at Black Rock and Ashleyhay. The faults are interpreted as a set of listric faults with downthrow to the NNW (Fig. 10B). The slide blocks collectively are about 950 m wide, and the angle of southward tilt ranges up to 13° (corrected for tectonic dip). Vertical displacements of 20 m and 40 m can be estimated for the two slide blocks shown in the constructed section (Fig. 10B). The timing of the movement cannot be closely constrained in this example, because no delta top beds are preserved in the outlier. The slides probably lie in the $R_{2b}4$ cycle (see above), so could conceivably have been formed during that cycle of deltaic influx, but the similarity of the Callow slide blocks to those at Ashleyhay and Black Rock suggests that a common origin is more likely.

4. MINOR STRUCTURES OUTSIDE THE LISTRIC FAULT BLOCKS

Many published accounts of gravitational slide structures (e.g. Gawthorpe & Clemmey 1985; Martinsen 1989; Martinsen & Bakken 1990) show an extensional region of listric faulting linked downslope to a region of compressional structures (Fig. 11). Areas distal to the listric fault arrays in the Wirksworth area now form a gentle anticline, the Derbyshire Dome, where erosion has removed evidence of any downslope region of compressional structures that may once have existed. However, a range of small-scale structures of extensional and compressional origin affects the delta slope deposits outside the listric slide blocks. In surface sections (Fig. 1, A-G) these lie beneath the main concentrations of delta slope (Ashover Grit) sandstones, in which regional dips are very low, so that the minor structures are apparently discordant to the broader structure. These sections were examined in order to test whether any of the minor structures relate to the compressional component of another slide system that may have been located to the south of the present outcrops. The aqueduct boreholes also contained minor structures: dip values were measured but dip directions could not be determined, so the existence and trends of possible folds could not be assessed.

4.1. Description and discussion of minor structures

The discontinuous nature of the surface exposures makes a full description of the minor structures impossible, but some relationships between the various types of structure can be illustrated by reference to the larger exposures (Figs 12, 13) described below.

4.1.1. Soft-sediment structures. The difference between soft-sediment structures and consolidated structures is not clear-cut, as noted below. In the surface sections, true soft-sediment structures show convolution and disruption of lamination limited to small thicknesses of strata. They are clearly related to gravitational down-slope movements of small thicknesses of sediment newly deposited on the delta slope. In the aqueduct boreholes, such structures were recorded at all levels in the delta slope succession.

4.1.2. Structures in semi-consolidated sediment. Figure 13A shows a low-angle fault exposed in Franker Brook (Fig. 1, D). It steepens upwards and may be a small listric slide plane. The hangingwall contains a sandstone roll that has been deformed while soft, and the footwall is affected by steep joints that appear to end upwards at the fault. Figure 13B, also in Franker Brook, shows a steep fault associated with a small fold, and with a soft-sediment sandstone roll in the fault zone. Both exposures appear to involve the juxtaposition of sediment showing different degrees of consolidation, suggesting that these structures formed during sedimentation, perhaps in small slides. The further implication, that the Namurian delta slope comprised sediment in all states of compaction, is consistent with evidence from the large listric slides, which took place while sediment was still being deposited on the delta top, but which are notably coherent structures.

4.1.3. Structures in apparently consolidated sediment. Low-angle fault zones in apparently consolidated sediment are exposed at three places. In the first exposure (Fig. 12A), an excavation in basinal mudstone just below one of the main listric faults near Ashleyhay (Fig. 1, F), low-angle faults dipping to the northeast are associated with small semi-recumbent folds. The inferred direction of stress alignment, NE-SW, is at right angles to that of the listric fault block above (Fig. 7). In the second exposure (Fig. 1, A), several kilometres distant from any of the listric faults, low-angle faults in basinal mudstone are again associated with recumbent folds, but the dip of the faults suggests NW-SE compression, consistent with the direction of movement in the slide blocks (Fig. 12B). In the third exposure (Fig. 1, G), also several kilometres distant from any of the listric faults, a sandstone interbedded with delta slope siltstone and mudstone is cut off by a low-angle fault dipping to the southeast at 35°. Small folds close below (Fig. 13C) also suggest NW-SE compression. Two of the three compressional movements recorded in these exposures are consistent with the direction of Namurian slide movements, but one is not.

Minor folds not demonstrably connected to faults are common in the surface sections. They are mainly short periclinal structures, symmetrical and asymmetrical, with flanking dips up to vertical (Fig. 13D). A stereoplot (Fig. 14) of dip measurements combined from 4 km of stream section (located on Fig. 1, A-E) shows that the steepest dips are generally towards the east, and are perhaps more easily understood in relation to E-W Variscan compression than to the NW-SE compression that would be expected from Namurian sliding. Minor folds could not be identified in the borehole cores, for the reason given above.

Steeply dipping minor faults were present in some of the boreholes. In borehole M13 (Table 1), for example, a 35 m-thick interval with dips up to vertical is terminated downwards by a fault plane mineralized with dolomite and pyrite. This could have been interpreted as a low-angle thrust fault with a substantial zone of disturbed strata above it, but the field mapping around the site demonstrates that the fault is a steep dip-slip structure with a narrow zone of normal drag close above it. The presence of mineralization suggests a post-Variscan date.

4.2. Causes of the minor structures

Three possible causes of the minor structures are considered: Quaternary valley bulging processes (suggested for nearby sections by Aitkenhead & Chisholm 2003); Namurian syn-sedimentary movements; and Variscan and post-Variscan tectonic movements. The minor structures show several contradictory features, as noted above, and it has proved impossible to assign a definite age and origin to many of them.

4.2.1. Quaternary effects. Parallelism between valley trend and bedding strike might indicate a measure of control by valley bulging, but in only two of five lengths of logged stream section (Figure 1: A, Kirk Ireton: E, Shipley Brook) is a sub-parallel arrangement evident. In the other three there is considerable divergence. We conclude that some Quaternary superficial movements may have taken place, but that if so they would have modified already existing structures.

4.2.2. Namurian slide movements. Soft-sediment structures formed by small slip movements on the delta slope are clearly of Namurian age. Structures affecting semi-consolidated sediment must also be of Namurian age, but they include folds and faults similar to those found in apparently consolidated sediments, raising the possibility that some of the latter structures are also Namurian. The low-angle faults (Fig. 12), in particular, may fall into this category.

4.2.3. Variscan and post-Variscan tectonic movements. Compressive stresses during late Carboniferous inversion of this part of the Pennine Basin produced open folds trending roughly N-S (such as the nearby Crich Anticline) and folds trending roughly NW-SE (such as the hangingwall fold on the Bonsall Fault near Matlock) (Fig. 1). Post-Variscan tensional movements gave rise to normal faults in many orientations. These Variscan and post-Variscan stresses must have affected the area, and are a possible cause of some of the minor structures. An impression of E-W compression given by the dip data (Fig. 14) is a possible indication of Variscan influence.

Some of the steep faults (such as the fault in borehole M13, above) are probably of post-Variscan date, but others, such as that shown in Figure 13B, are associated with complex deformations more easily explained by Namurian movements.

5. CONCLUSIONS

1. The Ashover delta system prograded from the southeast producing delta slopes dominated by silt and sand deposits, the sand being introduced mainly by density currents. There is no evidence of large-scale failure of the slope at this stage. A high proportion of delta slope sandstones in and around the Ashleyhay slides contrasts with a low proportion around Black Rock. There is, therefore, no exact correlation between location of delta distributaries and location of slides.

2. Continued northwestward progradation of the deltaic system allowed the delta front and delta top fluvial sands of the Ashover Grit to build out over earlier slope deposits. Loading of the delta top, in a succession already partly compacted and lithified, resulted in rotational failures of the delta front along listric slide surfaces. The failures produced sets of rotational blocks, with evidence of continued sedimentation during and after the slide movements. Progradation of the fluviodeltaic system over the delta slope succession cannot be demonstrated in the Kirk Ireton area, where no delta top sandstones are preserved.

3. The slide-block arrays described here are apparently limited to a zone close to the south side of the Derbyshire limestone shelf, suggesting that their distribution might have been controlled by sea-bed palaeotopography as well as by the direction of delta advance. The slides might have been triggered by accelerating delta advance as the water depth decreased towards the upstanding limestone shelf. The slide blocks at Birchover are situated within the limestone shelf area, but perhaps in a location where the palaeoslope continued upwards, so that delta advance would have been into shallowing water.

4. Small-scale soft-sediment disturbances are common in the delta slope deposits, but most other minor structures cannot be assigned unambiguously to any one of the possible periods of stress. There is some evidence for small Namurian slide movements outside the listric

fault blocks, but the extent of such movements cannot be assessed. Some of the minor structures probably relate to Variscan and post-Variscan stresses, and there may have been some overprint by Quaternary valley-bulging movements in present-day incised valleys.

Acknowledgements

Severn Trent are thanked for permission to publish information on the aqueduct boreholes. These were logged by Ian Chisholm, Howard Johnson, David Lowe, Jack Pattison and Neil Aitkenhead. Copies of the logs are held in the BGS National Geoscience Data Centre. Identification of marine band faunas was carried out by Dr W.H.C. Ramsbottom and Dr N.J. Riley. We thank Dr C.M. Jones for guidance on the stratigraphy of the R_{2b} Substage, and Drs J.N. Carney, P.R. Wilby and two anonymous rewiewers for their helpful comments on the manuscript. Colin Waters publishes with the permission of the Executive Director, British Geological Survey, Natural Environment Research Council.

REFERENCES

AITKENHEAD, N. & CHISHOLM, J. I. 2003. Excursion 6. The Edale Shales and Ashover Grit of the Ecclesbourne valley and Alport Hill. *In*: HORTON, A. & GUTTERIDGE, P. (editors). *The Geology of the East Midlands*. Geologists' Association Guide No. 63, 49-57.

British Geological Survey. 1978. *Sheet SK 25 NE, Solid and Drift geology, 1:10,560 scale.* Keyworth, Nottingham: British Geological Survey.

CHISHOLM, J. I. 1977. Growth faulting and sandstone deposition in the Namurian of the Stanton Syncline, Derbyshire. *Proceedings of the Yorkshire Geological Society*, **41,** 305-323.

EDWARDS, M.B. 1976. Growth faults in Upper Triassic sediments, Svalbard. *Bulletin of the American Association of Petroleum Geologists,* **60,** 341-345.

FORD, T. D. 2005. The Geology of the Wirksworth Mines: A Review. *The Bulletin of the Peak District Mines Historical Society*, **16,** No.2, 1-42.

FRASER, A.J. & GAWTHORPE, R.L. 2003. *An Atlas of Carboniferous Basin Evolution in Northern England.* Geological Society, London, Memoir 28.

FROST, D. V. & SMART, J. G. O. 1979. *Geology of the country north of Derby*. Memoir of the British Geological Survey, Sheet 125 (England & Wales).

GAWTHORPE, R. L. & CLEMMEY, H. 1985. Geometry of submarine slides in the Bowland Basin (Dinantian) and their relation to debris flows. *Journal of the Geological Society, London*, **142,** 555-565.

GILL, W. D. 1979. *Syndepositional sliding and slumping in the West Clare Namurian Basin, Ireland.* Geological Survey of Ireland Special Paper No.4.

HALLSWORTH, C.R. and CHISHOLM, J.I. 2008. Provenance of late Carboniferous sandstones in the Pennine Basin (UK) from combined heavy mineral, garnet geochemistry and palaeocurrent studies. *Sedimentary Geology,* **203,** 196-212.

HOLDSWORTH, B.K. & COLLINSON, J.D. 1988. Millstone Grit cyclicity revisited. *In:* BESLY, B.M. & KELLING, G. (editors). *Sedimentation in a synorogenic basin complex.* Blackie, Glasgow and London, 132-152.

JONES, C.M. 1980. Deltaic sedimentation in the Roaches Grit and associated sediments (Namurian R_2b) in the south-west Pennines. *Proceedings of the Yorkshire Geological Society,* **43**, 39-67.

JONES, C. M. & CHISHOLM, J. I. 1997. The Roaches and Ashover Grits: sequence stratigraphic interpretation of a 'turbidite fronted delta' system. *Geological Journal*, **32,** 45-68.

LEEDER, M. R. 1982. Upper Palaeozoic basins of the British Isles: Caledonide inheritance versus Hercynian plate margin processes. *Journal of the Geological Society, London*, **139,** 481-494.

MARTINSEN, O.J. 1989. Styles of soft-sediment deformation on a Namurian (Carboniferous) delta slope, Western Irish Namurian Basin, Ireland. *In:* WHATELEY, M.K.G. and PICKERING, K.T. (editors). *Deltas: sites and traps for fossil fuels.* Geological Society, London, Special Publications **41,** 167-177.

MARTINSEN, O.J. & BAKKEN, B. 1990. Extensional and compressional zones in slumps and slides in the Namurian of County Clare, Ireland. *Journal of the Geological Society, London,* **147,** 153-164.

SMITH, N.J.P., KIRBY, G.A. & PHARAOH, T.C. 2005. Structure and evolution of the south-west Pennine Basin and adjacent area. *Subsurface memoir of the British Geological Survey.*

WATERS, C.N., CHISHOLM, J.I., BENFIELD, A.C. and O'BEIRNE, A.M. 2008. Regional evolution of a fluviodeltaic succession in the Marsdenian (late Namurian Stage, Pennsylvanian) of the Central Pennine Basin, UK. *Proceedings of the Yorkshire Geological Society*, **57,** 1-28.

Waters, C .N., Waters, R. A., Barclay, W. J. & Davies, J. R. 2009. *Stratigraphical framework for Carboniferous successions of Southern Great Britain (Onshore).* British Geological Survey Research Report, RR/08/07.

Wignall, P.B. and Best, J.L. 2004. Sedimentology and kinematics of a large, retrogressive growth-fault system in Upper Carboniferous deltaic sediments, western Ireland. *Sedimentology*, **51**, 1343-1358.

<u>TABLES</u>

Table 1. Boreholes mentioned in text and figures.

Borehole name	Borehole number in BGS archive	Grid Reference
M 1	SK 25 SE/59	SK 2947 5191
M 6	SK 25 SE/60	SK 2969 5197
M 13	SK 25 SE/67	SK 2672 5073
WA 16	SK 25 SE/44	SK 2908 5182
WA 18	SK 25 SE/45	SK 2914 5183
WA 18a	SK 25 SE/46	SK 2923 5185
WA 19	SK 25 SE/47	SK 2963 5195
WA 20	SK 25 SE/48	SK 2997 5201

Table 2. Comparison of features of listric fault blocks. Birchover data from Chisholm 1977, figures 1 and 4.

Locality	*width of widest fault block*	*backward tilt in fault blocks (corrected for tectonic dip)*	*direction of displacement*	*vertical displacement*
Black Rock	1650m	up to 25°	NNW to NNE	up to c.80m
Ashleyhay	1300m	up to 15°	NW & WSW	up to c.110m
Kirk Ireton	950m	up to 13°	NNW	up to c.40m
Birchover	2300m	up to c.40°	W to WNW	up to c.70m

FIGURES

Figure 1. Bedrock geology of the study area, based on BGS mapping (igneous and post-Carboniferous rocks not shown). Slide structures are present at Black Rock, Ashleyhay and Callow, and shown in Figures 3, 7 and 10. Minor structures described in text are exposed at Localities A-G. Inset map: basement structure of surrounding area, based on Waters *et al.* (2009, figure 1).

Figure 2. Generalized stratigraphic sections for the Ashleyhay and Kirk Ireton areas, based on boreholes and surface information. Stratigraphical context of the Marsdenian Substage also shown.

Figure 3. Outcrop geology of the Black Rock area. Dips in slide blocks are corrected for tectonic dip of 6° to the east.

Figure 4. Two scale sections around Black Rock, superimposed to show relationship of stratigraphy in slide blocks (section 2) to that of footwall (section 1). Lines of section are about 300 m apart, as shown in Figure 3. View is towards SE, to match orientation of Figure 5.

Figure 5. Panoramic view of the Black Rock area, from viewpoint (SK 282 566) above Dene Quarry. Vertical scale exaggerated by about 30%.

Figure 6. Details of the most proximal of the Black Rock slides.
A. Sketch section along the scarp NE from Black Rock, showing erosive relationship between faintly laminated sandstone and well bedded sandstone, and location of C & D. Scales are approximate.

B. Map showing location of section A, simplified from Figure 3.

C. Erosive contact of faintly laminated sandstone on well bedded sandstone (SK 2949 5582). View to southwest.

D. General view of Black Rock (SK 293 557) looking east. Faintly laminated sandstone dips south towards slide fault (not exposed).

Figure 7. Outcrop geology of the Ashleyhay area. Tectonic dip is almost flat. See Figure 2 for generalized succession.

Figure 8. Scale sections through the Ashleyhay area. Location of sections shown in Figure 9.

,

Figure 9. Panoramic view of Ashleyhay slides seen from viewpoint at SK 275 506 on west side of Ecclesbourne Valley. Not all sandstones are shown. Vertical scale exaggerated by about 50%, to emphasise the sandstone dip slopes. The Black Rock slides are hidden behind the scarp at Barrel Edge.

Figure 10. Simplified map (A) and section (B) showing geology of the north part of the Kirk Ireton outlier. For explanation of stratigraphy see text and Figure 2.

Figure 11. Block diagram to illustrate compressional and extensional parts of a slide system, based partly on an example in western Ireland (Martinsen 1989, figure 6).

Figure 12. Sketches of surface sections showing compressional structures in basinal mudstone below the Ashover Grit.

A. South side of an excavation (SK 2904 5182) near Ashleyhay (Figures 1,F & 8).

B. South side of a ravine (SK 2720 5020) near Kirk Ireton (Figure 1,B).

Figure 13. Minor structures in stream sections below the slope sandstones. Scale bars are approximate. Long dashed lines emphasise bedding, short dashed lines are faults. See text for discussion.

A. Franker Brook (SK 3060 4853), view to SE: juxtaposition of different states of compaction.

B. Franker Brook (SK 3056 4843), view to SE: juxtaposition of different states of compaction.

C. Lumb Brook (SK 330 466), view to SE: asymmetrical folds

D. Stream near Kirk Ireton (SK 2754 5007), view to SW: asymmetrical fold with vergence to ENE .

Figure 14. Stereoplot of poles to bedding dips from stream sections in the lower part of the delta slope at Kirk Ireton (Figure 1, A-C), Franker Brook (Figure 1, D) and Shipley Brook (Figure 1, E). Dips between 0° and 30° are almost randomly distributed but dips between 60° and 90° tend in a generally eastward direction.