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The Chevin Landslides, Otley, West Yorkshire: An explanation of their distribution, morphology and development.

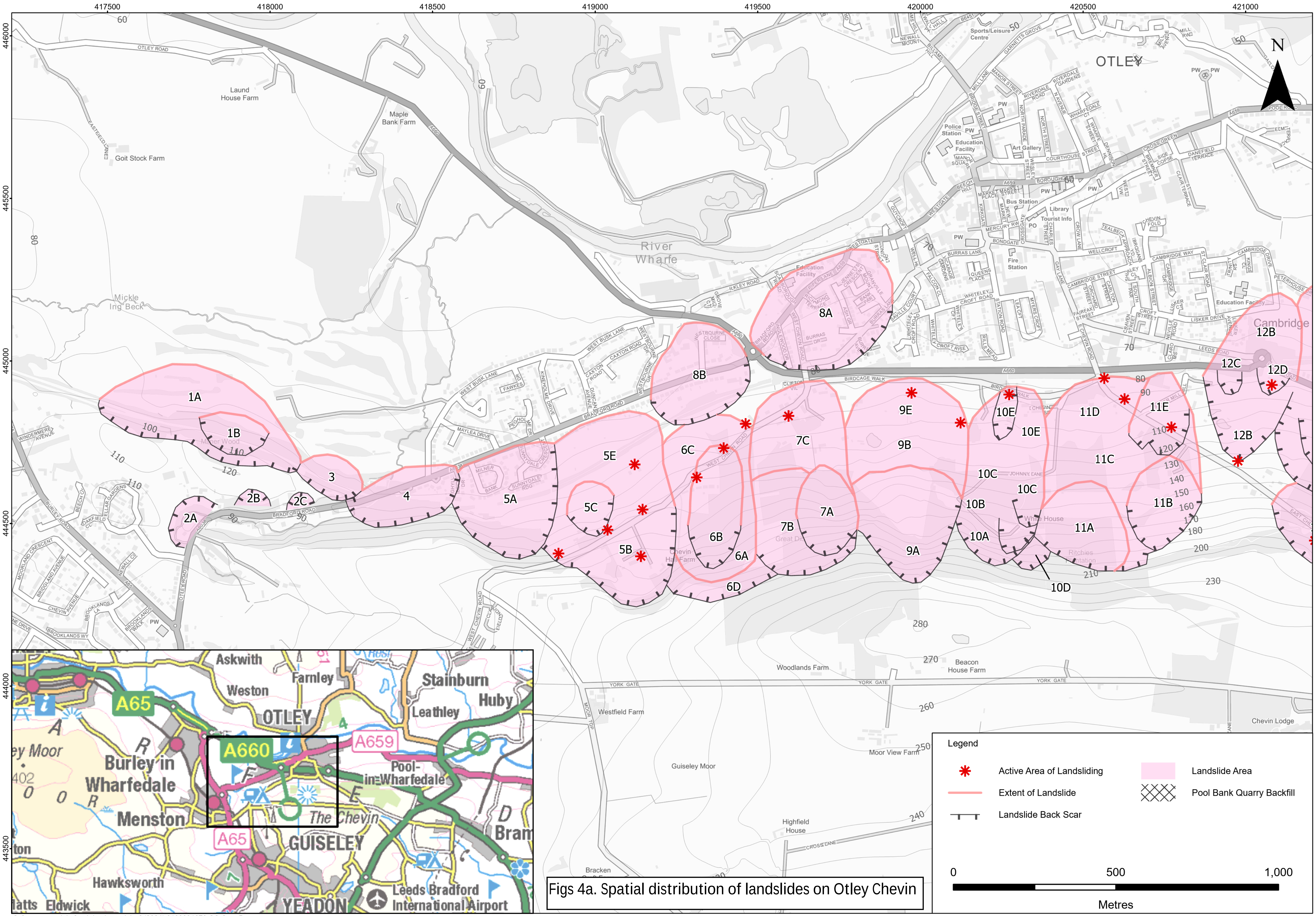
The Chevin Landslides, Otley, West Yorkshire: an explanation of their distribution, morphology and development

Dr Martin Cross, Leeds Geological Association, November 2024

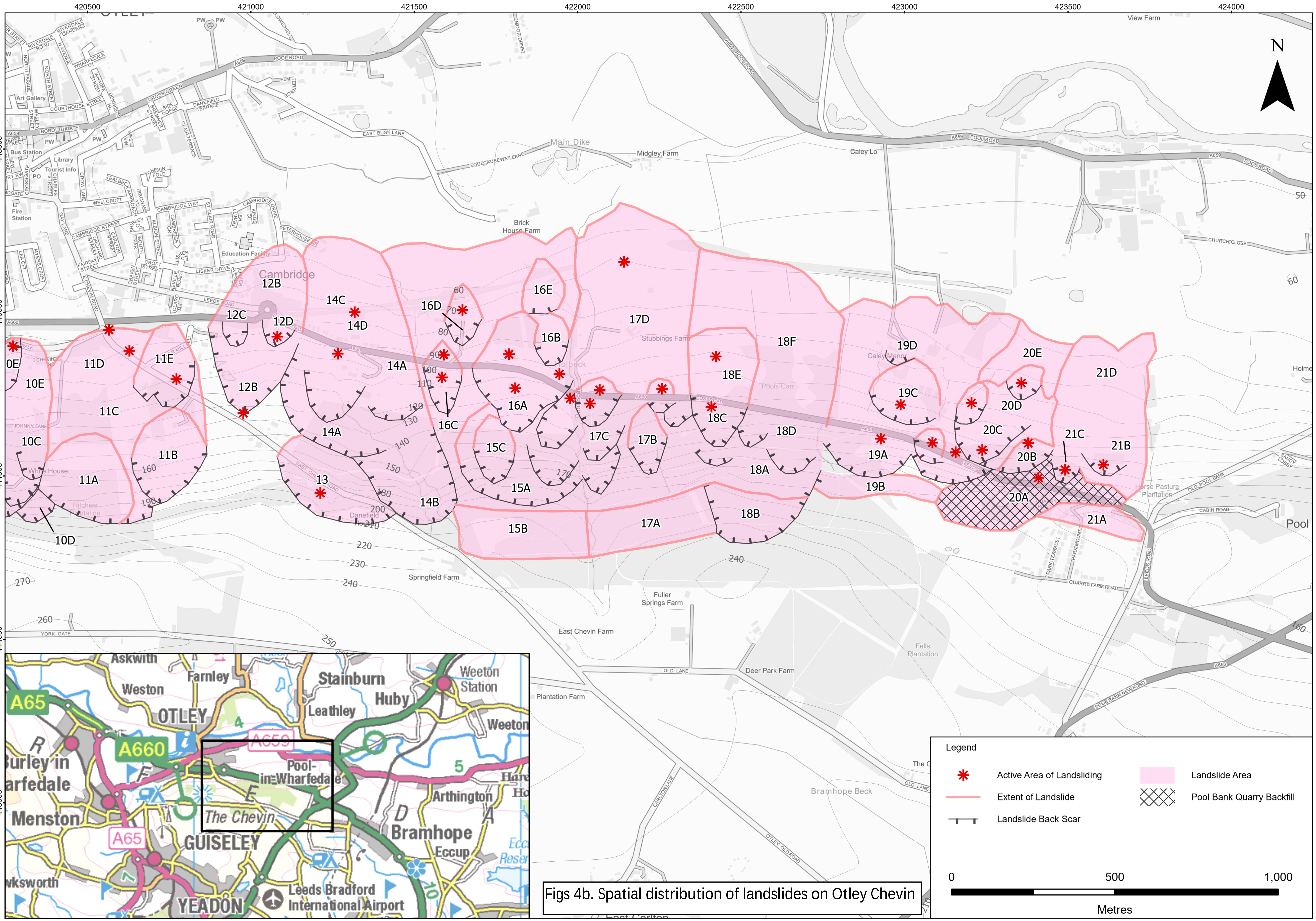
Abstract
The Chevin escarpment is noted for its exposed sandstone and gritstone edges along its crest, with large complex landslides affecting 4km^2 of the north-facing upper, middle and lower valley side slopes. The distribution and morphology of the Chevin landslide complexes are described in relation to their geology, geomorphology and hydrogeology. Large rotational landslides are prominent mass movement features on the upper and middle slopes, while mudslides and mudflows are present on the lower slopes. Factors responsible for landslide development include deglacial unloading, paraglacial stress-release processes and the reservoir principle of mass movement. The hydrogeological setting provides an ideal case for the application of the reservoir principle of mass movement. The north-facing scarp slopes have been affected by postglacial landsliding and are dominated by the geological structure of jointed gritstone/sandstone strata capping the escarpment crest (the reservoir), underlain by weaker and less permeable mudstones and shales where the large landslides have developed. The current state of stability of the deep-seated rotational landslides which reside on the upper and middle slopes and the mudslide/mudflow complexes located on the lower slopes is discussed in relation to climate change.

Otley Chevin
Otley Chevin (OS Grid Ref. SE204 442) is a steep escarpment located on the north-facing slopes of a 5km east-west trending ridge which overlooks the historic market town of Otley in mid Wharfedale, West Yorkshire, UK. The Chevin has attracted geological interest, notably for the exposed gritstone crags and fallen Millstone Grit blocks of Upper Carboniferous age at the summit and higher slopes and the large landslide complexes which affect approximately 4km^2 of the upper and middle north-facing valley side slopes between Otley and Pool Bank. There are many large deep-seated rotational landslides; one of the largest located above Otley is known as Great Dib. The Chevin landslide complexes are described in relation to their geology, geomorphology and hydrogeology. The factors responsible for their development are discussed, particularly, the deglacial unloading and paraglacial stress-release processes affecting the less competent mudstone strata forming the lower valley slopes and the reservoir principle of mass movement. The hydrogeological setting of Otley Chevin provides an ideal case for the application of the reservoir principle of mass movement, which was developed by the Institute of Geological Sciences now (British Geological Survey) following studies of landslides in sedimentary and glacially derived material. The north-facing scarp slopes of the Chevin which have been affected by postglacial landsliding are dominated by the geological structure of jointed sandstone strata capping the slope crest (the reservoir), underlain by weaker and less permeable mudstones and shales where the landslides develop. The current state of stability of the deep-seated rotational landslides which reside on the upper slopes and the mudslide/mudflow complexes located on the lower slopes of the Chevin under the present climate regime is also briefly discussed.

Study Area Location
Otley Chevin is a steep escarpment which faces north over the historic market town of Otley located in mid Wharfedale, West Yorkshire. Otley is located 16km north-west of Leeds and 16km north-east of Bradford (see location map inset on Fig. 4a. The Chevin escarpment is approximately 5km in length and approximately 250m in height. The study area is bounded by Old Pool Bank to the east (OS Grid Ref. SE234 443) and Gill Brow to the west of Otley (OS Grid Ref. SE181 444). The highest point of the Chevin, Surprise View, reaches 282m (Grid Ref. SE 2043 4419). According to the 1:50,000 Solid and Drift Geological map of Bradford, Sheet 69 (BGS 2000); the north facing escarpment of the Chevin comprises landslides extending approximately 4km^2. Several large deep-seated rotational landslides, together with associated mudslides and mudflows extend downslope to 0.75-1.0km at their maximum extent. The relevant topographical maps of the study area are Ordnance Survey 1:50,000 Landranger Sheet 104 Leeds and Bradford and 1: 25,000 Explorer Sheet 287 Lower Wharfedale and Washburn Valley.



Figs 4a. Spatial distribution of landslides on Otley Chevin



Figs 4b. Spatial distribution of landslides on Otley Chevin

Previous Studies

The name Chevin derives from the Brythonic *cefyn*, *cefn* or *cefu* meaning a ridge, or ridge of high land. The area extending from Otley Chevin to Caley Crags is a designated Local Geological Site (LGS). The West Yorkshire Geological Trust together with Leeds City Council and Friends of Chevin Forest have created The Chevin Park Geology Trail. The 3km geological trial includes several carved marker stones which provide information about the geological features of the Chevin. There is also an audio trail available on The Friends of Chevin Forest website, www.chevinforest.co.uk; the audio provides additional information on the geology and geomorphology of the Chevin and a description of the views that can be seen from the trail.

One of the earliest mentions of Otley Chevin was included in the prophecies of Old Mother Shipton (c. 1488-1561), where she predicted a catastrophic landslide would take place on the Chevin, with landslide debris covering parts of the town of Otley (Easton 1998; Araujo 2008). The Chevin was designated as one of the chain of beacons established as a nationwide communication system around the end of the eighteenth century when invasion by the French was threatened. A landscape history of Otley Chevin is provided by Laurence (2016).

The geology and landslides of Otley Chevin are described in the Memoir of the Geological Survey of Great Britain, Geology of the country between Bradford and Skipton, sheet 69 (England and Wales (Stephens *et al.* 1953). A more recent explanation of the 1:50 000 Series Sheet 69 Bradford, solid and drift edition, is provided by (Waters 1999). Detailed geological information about the 1:10 000 scale geological map of the Otley Chevin area is provided in the accompanying British Geological Survey Technical Report (Crofts 1995).

Previous investigations of Otley Chevin have been concerned with the stability of the Chevin escarpment in relation to the engineering design of the A660 and Otley Bypass. Several Ground investigation reports have been completed by West Yorkshire County Council and the West Yorkshire Highways Laboratory between 1930 and 1967; these mainly relate to specific areas of the A660 between Otley and Pool Bank which have been affected by landsliding (Robinson 1967). The Otley Bypass followed the alignment of a disused railway which was constructed through an area of complex mudslides located on the lower slopes of Otley Chevin (West Yorkshire Metropolitan Council Highway Engineering Technical Services (HETS) Laboratory 1972; Culshaw & Duncan 1975; Burgess 1976). Cooper (1984) described the geology, hydrology and stability of the landslides between the east side of Otley and Old Pool Bank. The British Geological Survey completed a geological desk study and aerial photograph interpretation together with fieldwork to map the landslides affecting A660 Leeds-Otley road (Cooper 1984). Cooper (1984) noted evidence of active landslide movement and recorded associated hydrological features such as surface water courses, springs and poorly drained areas.

The first UK reported anchored reinforced earth retaining wall (Silver Hill Mill Retaining Wall) was constructed in 1984 on the Otley Bypass (Snowdon *et al.* 1986). The retaining wall is 86m long with a maximum retained height of 6.0m. In addition to supporting a road embankment, the retaining wall also provided the southern abutment for a footbridge. The retaining wall was designed to protect several houses located approximately 20m behind the wall on Birdcage Walk. The retaining wall was constructed in an area comprising a complex series of mudslides and within an existing railway cutting constructed in 1860.

Geology

Geologically, the Chevin lies on the northern edge of the Yorkshire Coalfield. The geology of the area is shown on British Geological Survey 1: 50,000 Series, sheet 69, Bradford, solid and drift edition (BGS 2000). A brief explanation of the geology of the Bradford district is provided by Waters (1999). The geology of the northwest Leeds area is dominated by the solid geology of the Carboniferous Coal Measures and Lower Coal Measures, with the underlying Millstone Grit Group of the Namurian an important lithological unit within the site area (Waters *et al.* 1996). The Millstone Grit Group comprises about 1,800m of interbedded mudstone, siltstone and sandstone (see Fig. 1). During Namurian Times (c. 3.5 Ma) the rocks that crop out in the study area are interpreted as having been laid down in a deltaic environment, which eventually resulted in a sequence of generally coarse sandstones and fine-grained mudstones. The coals and seatearths resulted from the decay of the remains of forests that covered swamp areas of the delta system. Occasionally, marine incursions allowed deposition of muds including fossiliferous marine bands, which provide important isochronous marker horizons such as the Otley Shell Bed. Natural superficial deposits are also present on the Chevin escarpment slopes, which comprise glacial, periglacial and postglacial deposits. Peat is present on the upper part of Otley Chevin. Quaternary deposits, including Glacial Till and Head are present on the mid and lower valley-side slopes. Fluvioglacial sand and gravel deposits are present in the flood plain of the River Wharfe; river terrace deposits and colluvial deposits are also present on the basal slopes of the Wharfe valley (Dean & Lake 1991; Waters 1999). River Terrace Deposits, which represent fluvial depositional features isolated on the valley sides by entrenchment of the River Wharfe, are prominent on the north side of the river between Otley and Leathley. Deposits of Alluvium are widespread in the base of the Wharfe valley and typically comprise coarse granular gravel at the base with clays and silts present in the upper surface layers. The sand and gravel has been extensively worked historically, leaving several large, water-infilled sand and gravel pits and municipal landfills.

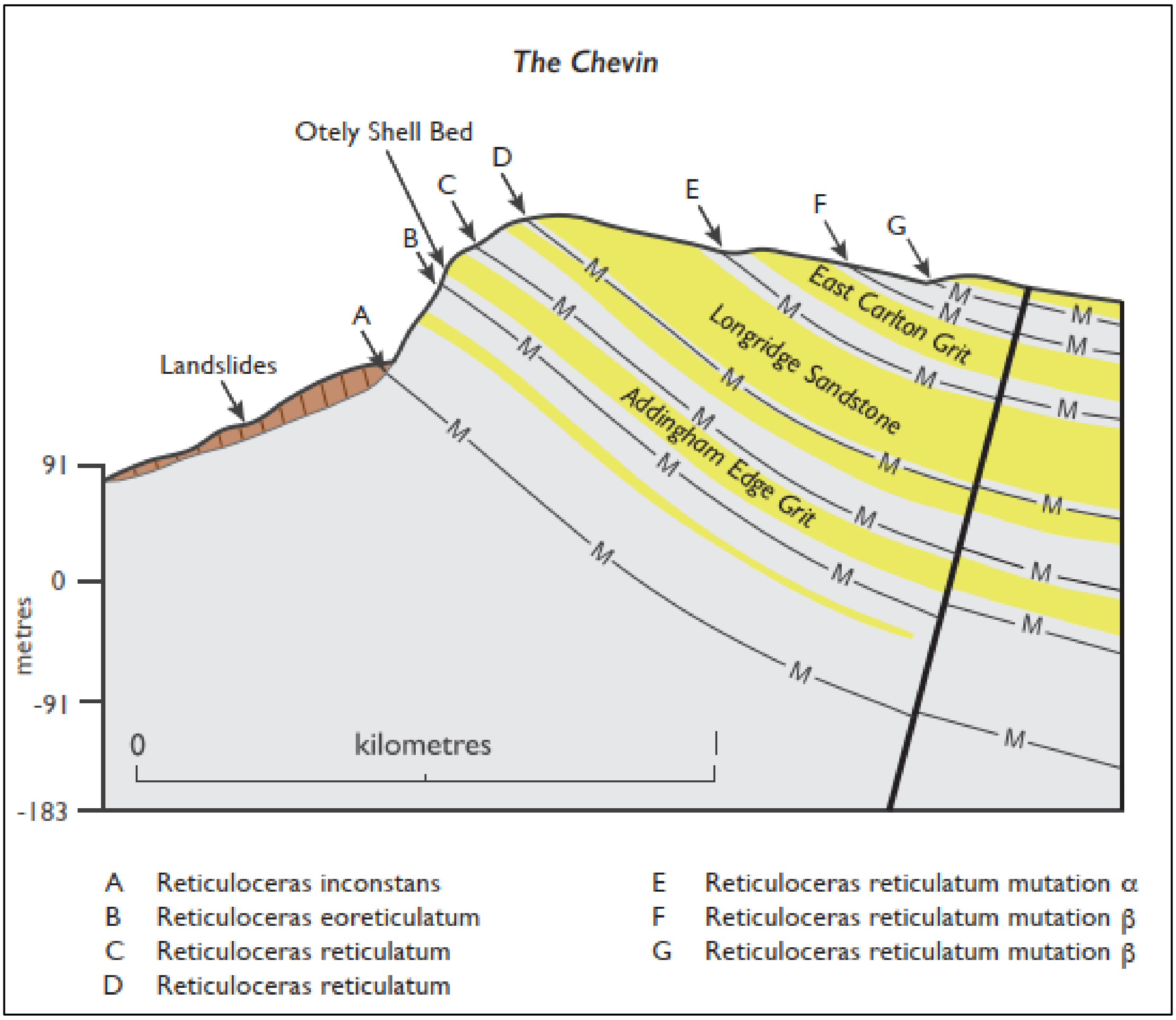


Fig. 1: Cross-section through Otley Chevin showing the higher beds (R1 and R2) stages of the Namurian Millstone Grit Series.

The rocks forming the base and mid slopes of the Chevin comprise undifferentiated mudstones and shales, with some thinly bedded siltstones and sandstones, of Namurian (Kinderscoutian R_1) age (see Fig. 2). These strata are capped by strata of the Namurian Millstone Grit Group, the Addingham Edge Grit (formerly known as the Caley Crags Grit), a coarse-grained sandstone giving rise to a conspicuous 'edge' from East Chevin Quarry (Grid Ref SE 2121 4448) to West Chevin Road. The Addingham Edge Grit consists of massive cross-bedded coarse grit, 25 to 30m in thickness, with a mudstone parting near the top. Most of the grits are felspathic and contain pebbles of quartz and felspar. The main bed of grit forms conspicuous rock faces along the top of the Chevin and is particularly well exposed at Caley Crags (SE2274 4436) and Great Dib landslide (SE 1998 4449), where it overlies the Otley Shell Bed. The Addingham Edge Grit is also clearly exposed in the disused East Chevin Quarry (SE 2121 4448), where it was once quarried and crushed for sand and gravel. About 25m of massive, coarse grit is exposed, with flaggy micaceous sandstone at the base, and a thin shale horizon.

The Addingham Edge Grit is in turn overlain in ascending order by the Long Ridge Sandstone and Doubler Stones Sandstone (formerly known as the Bramhope Grit), each separated by undifferentiated mudstone. The top of the Chevin is capped by the High Moor Sandstone (Bramhope Grit) or Upper Kinderscout Grit and the Doubler

Stones Sandstone (Bramhope Grit) or Lower Kinderscout Grit. The Doubler Stones Sandstone attains its maximum thickness at Beacon Hill (SE 199 442), just north of the former Yorkgate Quarry where it was quarried and crushed for sand. These units show a rapid variation in lithology, being dominated by gravel conglomerates and coarse pebbly sandstones, both of which have minor amounts of sand occurring interstitially. Clasts in these units are dominantly of white quartz, with <5% of other material. The coarser beds pass laterally into finer grained sandstones, and impressions of plant stems are common on bedding planes throughout the unit. Thin coals have been identified in beds close above the sandstone units. For example, the Morton Banks Coal seam overlies the Doubler Stones Sandstone (SE 1990 4412).

The dip slope to the south of the crest of the Chevin, which overlooks Airedale and the glacially eroded Guiseley Gap, comprises the East Carlton Grit through to the Guiseley Grit, separated by undifferentiated mudstone (see Fig. 1). The dip of the units along the Chevin is consistent, being about 15-18° to the south. The main sandstone units forming the Otley Chevin Escarpment are listed in Table 1.

Table 1. Sandstones of the Otley Chevin Escarpment

Sandstone (Former name)	Thickness (m)	Lithology
High Moor Sandstone (Bramhope Grit) or Upper Kinderscout Grit	0-21	Sandstone, very fine-grained to coarse-grained, upwards-fining, cross-bedded or massive, locally very micaceous, thinly bedded towards the top; sharp top typically marked by a ganister.
Doubler Stones Sandstone (Bramhope Grit) or Lower Kinderscout Grit	8-60	Sandstone, fine-grained to granular, micaceous, cross-bedded and laminated.
Long Ridge Sandstone (Bramhope Grit) or Lower Kinderscout Grit	5-58	Sandstone, fine to very coarse-grained, upwards-fining, massive and cross-bedded; sharp erosive base.
Addingham Edge Grit (Caley Crag Grit)	15-55	Feldspathic sandstone, medium to very coarse-grained, in parts pebbly, thickly cross-bedded.
Brocka Bank Grit	0-55	Sandstone, coarse-grained and massive or cross-bedded; lower leaf is fine-grained and massive.
Middleton Grit	0-30	Quartz-feldspathic sandstone, medium to coarse-grained, locally pebbly and cross-bedded.

The Namurian Millstone Grit Group lithologies described in Table 1 are of Marsdenian age; with an approximate age of 318 Ma. This chronostratigraphical sequence was originally established by Edwards *et al*. (1950) and later by Stephens *et al*. (1953). Borehole data, well sections and field correlations have been used to develop a summary of the key geological units in the area (Dean & Lake 1991) (see Fig. 2). To conform to current usage, the Millstone Grit Group has been mapped as a lithostratigraphical unit that includes all strata of Namurian age above the Upper Bowland Shales Formation, which is dominantly argillaceous and therefore, part of the Bowland Shales Group. Important modifications of the Kinderscoutian (R_1) succession have arisen from British Geological Survey boreholes on Rombalds Moor (Aitkenhead & Riley 1996). The exposed Millstone Grit Group Namurian strata comprises approximately 150m of alternating mudstone and sandstone ('rock' or 'grit') beds on the uppermost part of the northern escarpment of the Chevin (see Fig. 3).

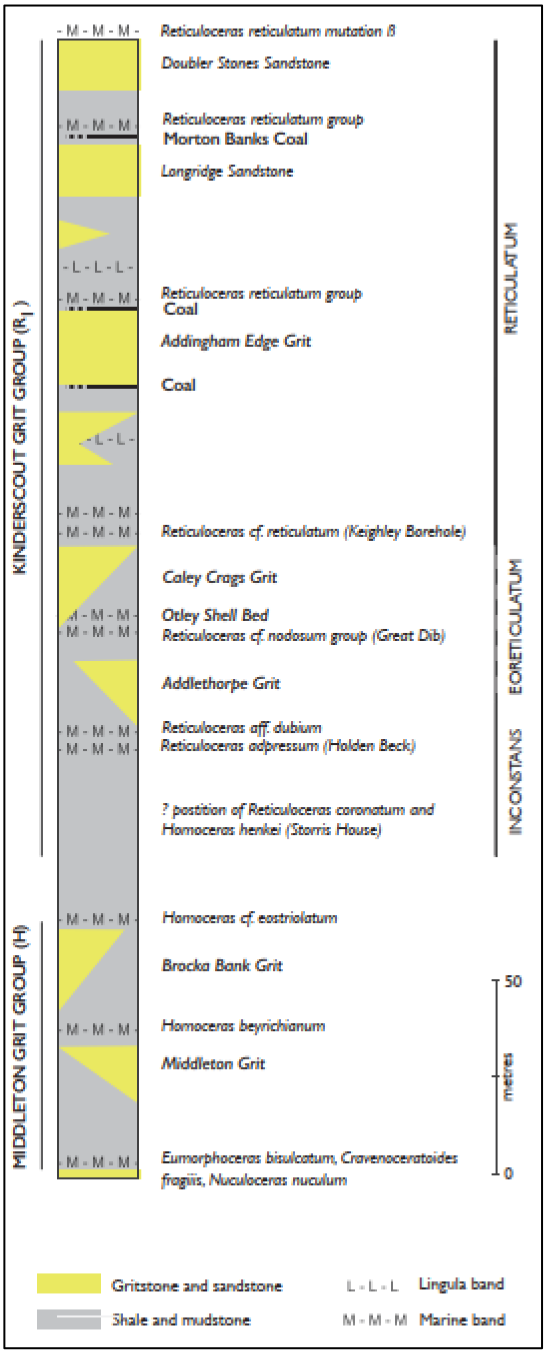


Fig. 2: Generalized geological section of the middle part of the Namurian Millstone Grit succession H and R1 stages.

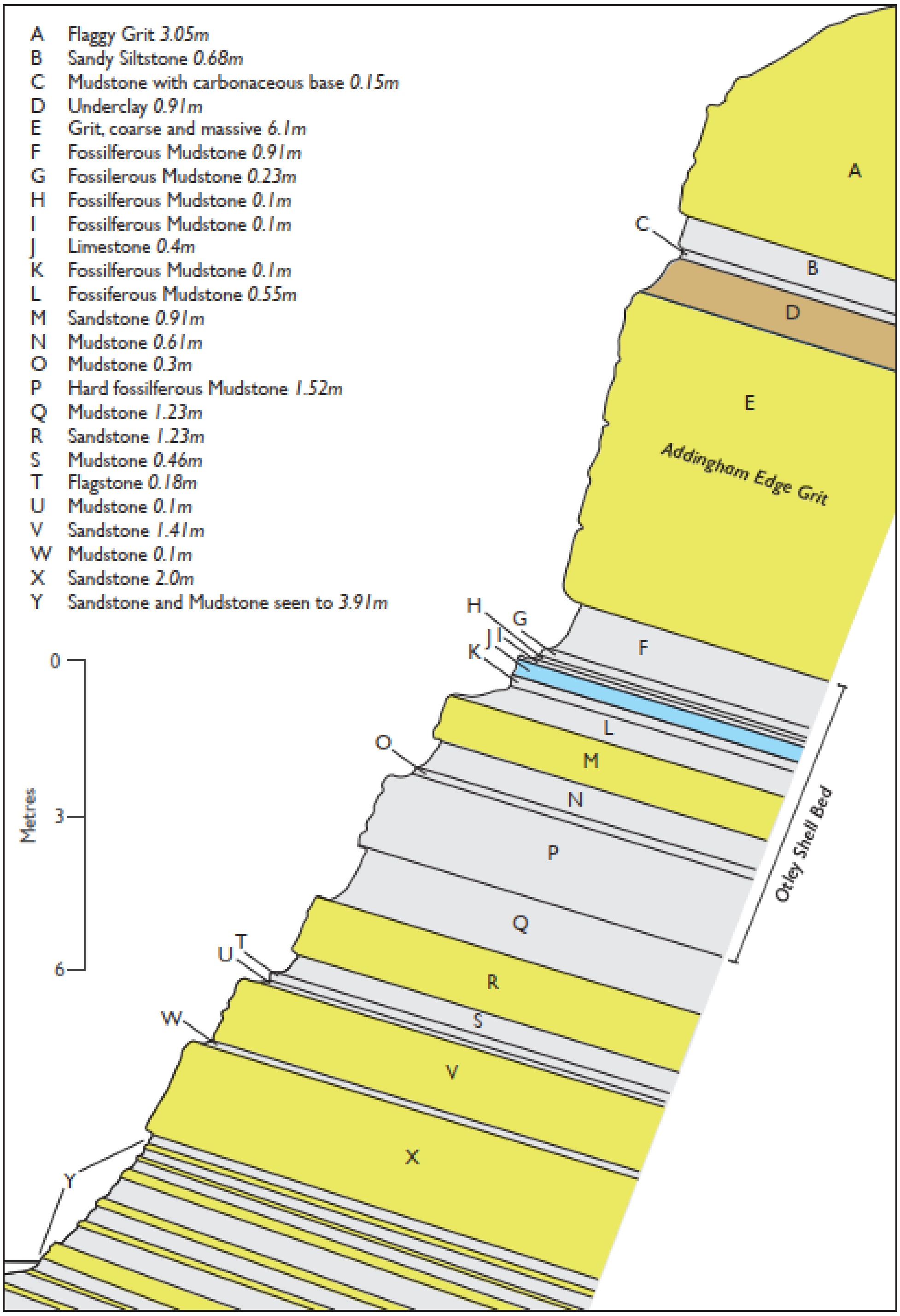


Fig.3. Cross-section through the back scarp of the Great Dib Landslide, Otley Chevin.

The rocks forming the lower slopes of the Chevin consist of mudstone and thinly bedded shales containing thin discontinuous sandstone layers and the Otley Shell Bed (see Fig. 3). The mudstone units of the Namurian are medium to dark grey and locally contain nodular sideritic ironstone. The mudstone units are not well exposed as they have been eroded and commonly form 'slacks' between outstanding sandstone or grit units. The best exposure of mudstone is within a small cutting to the north of West Chevin Road (SE 189 444), although it is doubtful that this material is *in-situ* as it is located in the toe of a large rotational landslide. Dark grey, micaceous shales also occur, and thin sandstone bands among the shales are typical. Two sandstone bands lie within 100m of the base of the Addingham Edge Grit, each ranging up to 15m in thickness. Another sandstone unit can be traced parallel to the slope for 150m westwards from Pig Farm (SE 200 445) where several small exposures of medium grained micaceous sandstone occur. True marine shales are confined to the marine bands, which commonly contain fossils and concretions of thin, impure limestones. These shales are typically dark grey or black and oily, and are usually calcareous, weathering to a yellow crumbling decalcified material termed 'gingerbread rock' (Stephens et al. 1953). One such unit is exposed just south of Chevin Hall (SE 191 442), and palaeontological evidence assigns these to *Reticuloceras eoreticulatum* zone (R_{1b}) (Dean et al. 2011).

The Otley Shell Bed is an important marine band and consists of a few metres of highly fossiliferous mudstone with sandstone beds and a limestone rib. A total of 16m of the Otley Shell Beds are located below the exposed Addingham Edge Grit within the Great Dib landslide (SE 200 443) (see Fig. 3). The Otley Shell Beds contain fossils of lamellibranchs and brachiopods with some cephalopods, trilobites and sponges.

Geological Structure of Wharfedale

During the Carboniferous, the Pennine Hills were part of a series of structurally controlled basins that strongly influenced sedimentation in the British Isles north of the Variscan Front. At the end of the Carboniferous period, a major collision between two tectonic plates culminated in the uplift of a high mountain range across southern Europe in a series of events called the Variscan Orogeny. From late Westphalian to early Permian times, during the Variscan Orogeny, basin inversion took place and the Carboniferous rocks of the Pennine region were uplifted. Considerable faulting is thought to have occurred with some major faults reflecting deep-seated structures. Subsequent substantial erosion and planation of the Carboniferous rocks also took place. This affected south-west England more than the rest of Britain. However, northern England was uplifted into the Pennine anticline, which trended north-south from the Midlands to southern Scotland. In addition, the Middle and Lower Wharfedale area was folded into an anticline which runs east-west along the Wharfe valley (Wharfedale). Otley Chevin is located on the south side of the fold so the rocks dip southwards towards Airedale. Reactivation of pre-existing faults, with the possible addition of minor fracturing and gentle folding of the Carboniferous rocks is likely to have occurred as a result of post-Variscan movements. It can be assumed that no significant tectonic activity has occurred since Tertiary times.

The regional dip of the Namurian Millstone Grit Series is gentle and mostly to the south-south-east. The strata are relatively undisturbed, with several major fractures and common instances of minor faulting (Dean & Lake 1991). The pattern of dominant faulting in the northern areas of the Leeds district is dominated by fractures which trend approximately NE-SE. Dean & Lake (1991), noted that subordinate sets of conjugate faults trending N-S and NW-SE link the major faults. Certain major faults are prominent, the most important of these to the Chevin being the Wharfe Valley Fault, which extends from Addingham Moor and enters Wharfedale south of Nesfield, where it has been traced beneath the alluvial deposits present within the valley floor to Otley, but no further.

Superficial Geology

Glacial Deposits

The Last Glacial stage (Devensian) was at a maximum about 17,000 B.P. and ice covered the high ground in the area of Otley Chevin. At the glacial maximum the Pennine hills at the head of Wharfedale were dominated by a large ice cap which covered Blea Moor, Ribble Head and Widdale Fell (Stone et al. 2010). The outer fringe of this ice cap also covered Dodd Fell, Penyghent and Fountains Fell, and the eastern edge fed a glacier which moved south-eastwards down Wharfedale (Raistrick 1931, 1933). This glacier had a profound effect on the topography and geomorphology of Wharfedale and left widespread deposits of Glacial Till and Glacial Sands and Gravels within the valley (Waltham 2007). As temperatures increased, ice melted, and ice sheets became thinner. The thickest ice was confined to valley glaciers in the final stages of the Devensian glacial period. It would have been possible to stand on the Chevin and look over glacial valleys, with ice moving down Wharfedale, Washburn Dale and Airedale and glacial meltwater streams filling the valley floors with glacial sands and gravels. By about 12,000 B.P., when the ice had melted completely, it left behind deposits of Glacial Till both within and on the valley sides of Wharfedale. During the Late Glacial and Post-Glacial periods, the Glacial Till present on the valley sides in places was either soliflucted, washed away or modified by colluvial processes to form Head Deposits. The Head Deposits are up to 2.0m thick on the valley slopes of the Chevin.

The Glacial Till deposits on the valley sides of the Chevin are very sandy and rich in gritstone debris that is difficult to differentiate from the weathered residual soils of the Millstone Grit Group strata. The Glacial Till typically weathers at the surface to a brown or yellow/brown, decalcified clay, and maximum thicknesses of up to 50m have been reported in the valley floor (Gross 1979). However, a thickness of 10 to 15m is probably more usual in the valley floor, and the deposit thins to a shallow depth higher up the slopes of the Chevin, not extending much above the 120m contour (Burgess 1976). Isolated pockets of Glacial Till and erratic clasts do occur on the higher slopes, but are rarely more than a few centimetres deep, indicating that more extensive glacial deposits existed over the slopes of the Chevin but have been removed by erosional slope transportational processes during paraglacial relaxation (i.e., the period of readjustment from glacial to non-glacial conditions (Benn and Evans, 1998). The crest of the Chevin was overridden by ice; relic patches of Glacial Till have been recorded on the Doubler Stones Sandstone (Hemingway 1957). Glacial striations were recorded in the same area, although these are no longer visible (Stephens *et al.* 1953) and agreed with the general eastward down-valley direction of the glacier movement postulated by Raistrick (1934). Raistrick (1927) mapped a series of terminal moraines within Wharfedale; he recognised six halt-stages in the Wharfedale glacier at Pool (1), Burley-in-Wharfedale (2), Middleton (3), Drebley (4), Kilnsey (5) and Skirfare Bridge (6). The halt-stages of the active ice fronts provide evidence of a periodicity in the climatic conditions towards the end of the Devensian; the terminal moraines representing periods when conditions were coldest, and the interval between moraines representing warmer periods when melting ice was rapid. The terminal moraines formed dams to produce a series of elongated proglacial lakes in Wharfedale upstream of the terminal moraine. The lakes have now been infilled with laminated clays and silts which overlie Fluvioglacial Sands and Gravels to form flat, frequently flooded, sections of the valley floor. One of the proglacial lakes was located below the Chevin between Burley-in-Wharfedale and Pool. Raistrick (1926) postulated that the moraine dammed lakes persisted well into prehistoric times based on archaeological evidence.

Several lateral moraines run parallel to the Wharfedale valley sides at various levels; good examples can be seen at Farnley, Leathley and Stainburn on the north side of Wharfedale. These are often associated with large glacial overflow channels which either trench the valley sides or cut across spurs and were eroded by melt water flowing along the edge of the Wharfedale glacier. With the retreat of the ice these channels were abandoned and now remain as pronounced dry valleys on the north side of the valley. Along the southern side of Wharfedale, few such channels can be recognised, because they are not easily distinguished from, and may even follow the outcrop terraces of Millstone Grit horizons. However, Raistrick (1931) postulated that at one time such a channel was cut at about 225m AOD across the eastern end of the Chevin by water flowing from the SE/NW trending Guiseley Gap and along the north-facing flank of the Chevin; however, this was only a small channel and was soon superseded. At a later stage of glacial retreat, drainage reverted to the main Wharfedale valley, and the water passed along the ice edge on the northern face of the Chevin from the Guiseley Gap (Stephens *et al.* 1953). This will have resulted in significant over-deepening of the valley floor and over-steepening of the basal slopes of the south side of the valley below the Chevin. Any deposits associated with erosion features which may have existed along the north facing slopes of the Chevin cannot now be distinguished because of the significant erosion which took place on the south side of the valley slopes below the Chevin and also because they have been masked by landslide deposits resulting from subsequent postglacial mass movement processes.

Periglacial Deposits

Periglacial climate processes controlled the geomorphological landscape adjacent to ice sheets and glaciers of Wharfedale. This climate resulted in a set of geological and geomorphological processes of which freeze-thaw was the most important. Mass transportation of soliflucted materials and wind action dominated the barren periglacial areas, where the ground was often perennially frozen (permafrost), and lead to the development of characteristic deposits such as Head and more deep-seated mass movement features.

Head Deposits consisting of rubbly solifluction deposits of local provenance occur on the lower slopes of the Chevin and extend to the valley floor. However, where the parent material involved in solifluction is sufficiently clayey, particularly on the lower mudstone and shale slopes, basal and internal shears can develop in the soliflucted material. The apparent gradation between Head Deposits and River Terrace Deposits in Wharfedale was noted by Stephens *et al.* (1953), who attributed the deposits to the process of Late Glacial solifluction. Evidence of periglacial activity on the crest of the Chevin is provided where the action of permafrost has disrupted bedding in the Doubler Stones Sandstone and Addingham Edge Grit leading to the development of involuted and cryoturbated structures (Hemingway 1957). The development of such structures in the locality was favoured by the abundance of percolating meltwater available during the summer and the emergence of the area as a nunatak as the ice sheets retreated around it. The top of the Chevin was therefore, subject to frost action from an early stage, whilst much of the surrounding area was still under ice.

Geomorphology

The present geomorphology of Mid Wharfedale is dominated geologically by the Millstone Grit Group, a thick succession of interbedded sandstones, siltstones and mudstones of Namurian age, the glacial and periglacial processes active within Wharfedale during the Devensian and the subsequent postglacial fluvial processes of the River Wharfe. As described, the Chevin escarpment is capped by sandstones and gritstones underlain by less competent interbedded mudstones and shales with thin sandstones. This bedrock series was partially covered by a layer of Glacial Till which extended over the scarp slope up to approximately the 120m contour. Landslides are a common feature in Wharfedale, occurring mainly on the northern facing slopes. In some areas the whole of the mid and lower scarp slopes of the Chevin are affected by landslide, mudslide and mudflow deposits. Waters et al. (1996) noted that landslides have developed within Glacial Till and other superficial deposits on natural slopes at inclinations of 11-18°; whereas in the case of landslides involving rock, the natural slopes are generally greater than 20°. The large rotational landslides on the Chevin are located on the upper parts of the valley side of the Chevin and have not been triggered by fluvial erosion but have been triggered by paraglacial processes. Large-scale rotational landsliding resulting from the paraglacial debutressing processes which took place during Late Glacial and Post-Glacial times would have been an important phenomenon affecting Pennine hillslopes including Otley Chevin (Shakesby & Matthews 1996; Ballantyne 2002).

The geomorphology of the Chevin has been greatly influenced during the Pleistocene by both ice sheets and valley glaciers and glacial processes. Wharfedale was affected by at least three glaciations, although evidence for the earlier two phases has been obliterated by the final Devensian phase (Waters 2000). There is growing evidence from the global sea-level record that the Last Glacial Maximum (LGM) occurred relatively early in the Late Devensian, from about 27ka BP and lasting about five thousand years (Lambeck & Purcell 2001). The limits of the glacier that occupied the southwest part of the southern Vale of York at the Last Glacial Maximum are defined by Friend *et al.* (2016) in relation to deposits at Lindholme (SE 7337 0643). These deposits had an erratic content of associated diamicts indicating sources from Stainmore and the Yorkshire Dales, including upper Wharfedale and along the Permo-Trassic outcrops on the west side of the Vale of York. Friend *et al*. (2016) suggest the advance is dated to an episode associated with a high level of pro-glacial Lake Humber within the Last Glacial Maximum. Devensian ice in the Vale of York, sourced from Northern England, reached destinations less far south than the east coast ice, and its southern limit was for a long time regarded as defined by the Escrick Moraine (SE 630 421). However, a group of deposits forming a distinct ridge called the Linton-Stutton gravels are now considered to mark the southern limit of an Early Main Dales Glaciation (Edwards 1937; Edwards *et al.* 1950). The Linton-Stutton gravels are located to the west of the strandline deposits and west of the diverted River Wharfe and Escrick Moraine and are thought to be a kame belt (Straw 2016).

Pennine valley glaciers would have occupied areas of low ground, such as the Aire, Wharfe, Nidd and Swale valleys, except during the maximum advance of the ice when they would have also extended over upland areas, resulting in the deposition of a range of glacial deposits. The maximum advance of the Devensian ice sheet in the study area is shown on the British Geological Survey 1: 50,000 Sheet 69, Bradford, to have reached the south of Bradford (Waters 2000). The most significant effects of the Devensian glaciation locally are the over-deepening of the Wharfe valley by over 40m, the deposition of Glacial Till, terminal and lateral moraines, glacial drainage channels and the formation, infilling and breaching of proglacial lakes, resulting in extensive lake flats. The final retreat of the glaciers left over-steepened valley sides in an unstable or metastable state with massive sandstone and gritstone strata overlying weak mudstones and shales.

The role of deglacial unloading and resulting paraglacial stress-release in conditioning or triggering slope failures in the weak mudstones and shale slopes was an important factor affecting Pennine valleys (Johnson & Walthall 1979; Donnelley 2008; Dowell & Hutchinson 2010; Cross 2011). Donnelly (2008) suggested that deep-seated landslide movements on Pennine slopes were possibly generated under conditions of periglacial erosion and weathering, during glacier retreat, deglaciation and associated processes of gravitational stress-relief of valley sides. This is most likely to have occurred during the Late Glacial and early Postglacial. This may have initiated the lateral spreading of upland moorland plateaux, subsequently resulting in fissuring, fault reactivation, tilting and subsidence. Research has shown that rock-slope failures tend to be concentrated on the middle and lower valley-side slopes within the area occupied by ice during the Last Glacial Maximum, and that their locations coincide with zones of inferred high glacial loading stress, consistent with interpretation of both bedrock disruption and large-scale rock-slope failures as paraglacial phenomena induced by stress-release following deglaciation (Ballantyne 2002; Ballantyne & Stone 2013; Ballantyne et al. 2014a; Cossart *et al.* 2008, 2017; Wilson 2009; McColl 2012).

The over-steepened southern side of the Wharfe Valley was therefore, particularly vulnerable to the development of large rotational and complex landslides on the middle valley slopes where massive sandstone strata overlay weaker mudstones and shales. Some research has shown that many landslides did not take place immediately following deglaciation but took place 1000-3000 years after ice-sheet deglaciation (Cruden & Hu 1993; Soldati et al. 2004; Dowell & Hutchinson 2010). The long delay probably reflects progressive rock-mass weakening initiated by deglacial stress-release and associated tensile rock mass damage. The Chevin landslides affect all bedrock strata up to, and including, the Addingham Edge Grit and were probably first initiated 12-15,000 years ago in the Late Glacial with landsliding continuing to take place during the Loch Lomond Stadial (nominally 11-10 ^{14}C ka BP) and Early Postglacial times during the late Boreal to mid-Atlantic (nominally 7.7 to 5.5 ^{14}C ka BP), (Dowell & Hutchinson 2010). Much of the lower slopes of the Chevin are covered by extensive mudslide and mudflow deposits; these are likely to be younger and took place throughout the postglacial period.

Landslide Morphology

Large deep-seated complex landslides are abundant throughout Wharfedale, particularly on the steep over-steepened scarp slopes which form the north facing side of the valley. They are especially well developed along the 5km face of Otley Chevin between Otley and Pool Bank. An overall geomorphological appraisal of the area reveals that the landslides can be generally split into two types and in two areas parallel to the slope. Firstly, above the 120m contour, large rotational deep-seated landslides are located on the steeper slopes, affecting solid bedrock; these also include associated smaller secondary rotational slides and mudslides on their surfaces. Secondly, downslope of these, secondary instability has led to the development of mudslides and mudflows which emerge from the debris apron toes of the higher deep-seated landslides to cover much of the lower slopes. These processes have been exacerbated by the emergence of redundant springs from below the debris apron toe of the deep-seated landslides. The deep-seated rotational landsliding was probably induced to a large extent by glacial over-steepening and debutressing of the north-facing slopes of the Wharfe valley, and movement probably commenced sometime after the Wharfedale glacial ice retreated after the Late Devensian glacial maximum and continued during early postglacial times.

Along the upper slopes of the Chevin several large back scarps have developed at the rear part of large deep-seated rotational landslides e.g., Chevin Hall, Great Dib and Danefield Wood landslides. Below the back scarps the landslide morphology comprises a series of large, rotated blocks relating to a series of retrogressive rotational failures. Some of the landslides on the upper parts of the Chevin escarpment comprise a series of benches and terraces that run parallel to the contour lines. This morphology suggests multiple, successive rotational and translational landsliding. The deep-seated landslides therefore show evidence of various types of sliding mechanisms and can therefore be classed as complex landslides (Varnes 1978). In addition, many of the retrogressive and successive landslides have been denuded by shallow slumping, mudsliding and mudflow mass movement processes (Cruden & Varnes 1996). Below the rotated blocks which occur in the mid slope areas of the Chevin escarpment (i.e., elevations between 165-105m) are large spreads of hummocky landslide debris material. The morphology of the landslide debris area comprises of shallow slumps, shallow ridges and highly denuded mounds. A series of drainage channels and gullies have developed below springs and along seepage areas. In some areas within the landslide debris accumulations, poorly drained depressions have formed, sometimes containing small ponds.

Much of these landslide debris accumulations have been reworked by successive mudslides and in wetter places as discrete mudflows. The mudslides and mudflows extend over most of the lower slopes of the Chevin escarpment and are particularly well developed between slope elevations of 105m and 60m. The morphology of the mudslides comprises of a series of shallow spreads of material leaving discrete toe ridges and lobate features. The mudflows tend to be confined to more linear tracts within poorly drained shallow downslope drainage channels and gullies. The mudslides and mudflows have developed from the lower sections of the large deep-seated landslide debris aprons and are particularly well developed in the lower parts or the toe area of these landslides. The mudslides and mudflows comprise of reworked landslide debris, Glacial Till and solifluction Head Deposits. Some of the shallow lobate mudslides and mudflows on the lower valley side slopes comprise 1-2m depth of soliflucted Head Deposits below the ground surface. Although the general pattern of landsliding is as described, there are exceptions where discrete areas of mudsliding and mudflows occur on the upper and mid slopes and smaller shallow rotational failures also are present in some areas located on the lower slopes underlain by mudstones and shales. Burgess (1975) postulated that there may have been the potential for large deep-seated slip surfaces to develop and extend to the lower part of the Wharfedale valley. However, it should be noted that no deep-seated slip planes were identified in the main programme of site investigations for the

Otley Bypass in November 1972 (West Yorkshire Metropolitan Council Highway Engineering Technical Services (HETS) Laboratory 1972).

Landslide Distribution on Otley Chevin

The main areas of landsliding affecting the Otley Chevin escarpment are shown on Figs. 4a and 4b and are described below from west to east.

1. ***Milner Wood***

Milner Wood (SE 17862 44696) occupies higher ground overlooking Otley Golf Course to the north of the A6038 Bradford Road. The area is an interlocking spur at the western end of the Chevin escarpment which is separated from the main escarpment by Gill Brook. Further to the west of this interlocking spur is the Guiseley Gap which marks the end of the Chevin escarpment. On the north facing slope of Milner Wood are two landslides of different age. The older and larger landslide 1A (SE 17830 44693) comprises a large rotational failure with the scarp slope composed of mudstones and shales and thin sandstone bands located to the south of the disused railway line. The landslide has not rotated far, creating a pronounced bench and leaving a small area of landslide debris now covered by Milner Wood to encroach on the gentle slopes south of the disused railway which was constructed in cutting through the existing landslide near the base of the hillslope. A smaller rotational slide 1B (SE 17839 44712) has taken place within the older landslide affecting the back-scarp and rotated block of the older landslide; this also affects the southern side of the disused railway cutting.

2. ***Acrecliffe Farm***

A rotational landslide is located south of Acrecliffe Farm 2A (SE 17726 44540). The landslide has developed in mudstone on the western valley side of Mill Book, which curves round to the west of the A6038 Bradford Road. There are more recent smaller rotational landslides on the north side of the A6038 Bradford Road at 2B (SE 17916 44534) and 2C (SE 18093 44565). These have developed on the steep northern valley side of Gill Beck composed of mudstone and shale.

3. ***Dismantled Railway Bridge***

Immediately to the west of the dismantled railway bridge over the A6038 Bradford Road is a rotational landslide (SE 18232 44596) which has formed on the north-facing steep slope, south of the Golf Course. The back scarp comprises sandstone and the rotated block was cut through to construct the former railway cutting. Mudslides extend northwards onto the River Wharfe flood plain.

4. ***Gill Brow***

Further east of the dismantled railway bridge, below Gill Brow, south of the disused railway, are several rotational slides SE 18435 44528, with a complex of mudflows on the basal slopes of the north-facing escarpment. The mudflows extended across the area of the former disused railway line.

5. ***West Chevin / Chevin Hall***

Two large rotational landslides are located in an area known as West Chevin. Landslide 5A (SE19182 44374) is located to the northeast of West Chevin Delph. This is a complex rotational landslide as it also shows evidence of translational movement in the lower part of its accumulation zone. A very large deep-seated rotational landslide is located at Chevin Hall 5B (SE18994 44624). Both landslides have pronounced back scarps in the Addingham Edge Grit. Chevin Hall is located on one of the flatter upper rotated blocks. Landslide 5B has been affected by another successive failure; the back scarp of this failure is located below West Chevin Road (5C). From the toe of this landslide, a complex of mudflows extends north towards Chevin Grange (SE 192 447), and northeast towards West Chevin Road. Immediately north of the road at this point are a series of smaller rotational landslides in the mudstone sequence which extend westward from Chevin Grange to beyond Leeden (SE 189 444). Material from these landslides has moved north as a series of mudflows forming hummocky ground which extends to the Bradford Road (5E). Jones (1984) suggested that these mudflows were initiated by undrained loading at their crest by mudflows emerging from the Chevin Hall landslide (Hutchinson & Bhandari, 1971). West Chevin Road descends the Chevin through Landslide 5B and has been affected by instability over many decades. The road has been subject to various localised engineering remedial works including buttressing, slope drainage and constant pavement resurfacing. A significant landslide debris accumulation zone associated with Landslide 5B can be seen north of the second back scarp below West Chevin Road. The lower section of the debris accumulation zone coalesces with extensive mudslides and mudflows which extend northwards downslope to Bradford Road.

6. **Chevin Side**

A large area of complex landslides has taken place to the north of Chevin Side located to the east of Chevin Hall landslide (5B) (SE 193 444). It is postulated that this landslide took place following the Chevin Hall landslide as parts of the debris accumulation zone overlap those of the Chevin Hall landslide. The Chevin

Side landslide complex comprises of multiple rotational failures and mudslides which extend down slope north of West Chevin Road. In area 6A (SE19428 44587) a series of denuded multiple rotational landslides have taken place and these, in turn, have been affected by secondary slumping within the back scarps. Landslide 6B (SE19332 44490) comprises of successive rotational failures with significant translational movement taking place on the mid and lower parts of the accumulation zone. The landslide debris accumulation zone coalesces with mudslides and mudflows just to the south of West Chevin Road (Area 6C, SE19250 44769). These mudslides and mudflows extend down slope to Bradford Road. Most of the mudflow material appears to have originated from the area south of the cottages on West Chevin Road and adjacent to the Chevin Hall landslide (SE 193 444), where general collapse has occurred and many individual blocks of slipped material have been displaced downslope by mudflow. Burgess (1976) believed that this resulted from slumping along many closely-spaced slip planes with only small downslope movement. A similar mode of failure, not followed by general collapse, occurred just to the east of this area (SE 195 445), and may represent the development of retrogressive slumps. Excellent examples of mudslide terraces and mudflow lobate features can be seen to the north of Westville (SE 19289 44722). To the south of the Chevin Side landslide is an area of uneven ground underlain by large blocks of Addingham Edge Grit covered by a thin mantle of soil (6D). Several large boulders are exposed at the surface and were the result of rockfalls from the overlying Addingham Edge Grit. The size and frequency of boulders decrease downslope. This area of rockfall post-dates the Chevin Side landslide.

7. Ashwood House

The Ashwood House landslide can be split into three parts, 7A, 7B and 7C. Area 7A (SE19745 44618) comprises a multiple rotational landslide comprising two successive rotational slips. The back scarp of this landslide is 15-20m in height. The two successive rotational landslides are deep-seated possibly extending to 15-20m depth into the mudstones and shales. Area 7B (SE 19657 44600) is a complex landslide comprising a large rotational failure combined with a significant translational movement of debris material down slope. Area 7C (SE 19707 44706) comprises a large spread of mudslide and mudflow deposits which have formed down slope to the north of the landslide accumulation zones of landslides 7A and 7B. These mudslides and mudflows extend downslope north of Birdcage Walk. Mudslide terraces and mudflow lobate forms can be seen in the area around Ashwood House (SE 19694 44747). Gross (1979) recognised the effects of small-scale slumping along the mudslide toes.

8. West Chevin Bridge

The Glacial Till and Head Deposits to the north of West Chevin Bridge (SE 19671 44888) are up to 10m in thickness and appear to be highly disturbed (8A). These deposits are an extension of the mudslides associated with landslide complexes 6 and 7 described above. The deposits include Head Deposits in the upper part and mudslide deposits beneath the Head Deposits. Based on the data from the 1972 Otley Bypass site investigation, the fractured mudstones underlying the disturbed Glacial Till deposits in landslide are 8B exhibit dips of up to 30° down slope to the north (West Yorkshire Metropolitan Council, Highway Engineering Technical Services (HETS) Laboratory 1972). The dipping fractured bedrock may be the result of disturbance by deep-seated landsliding, or more probably by valley bulging which took place in the lower parts of Wharfedale valley. Valley bulging is thought to have taken place as a result of the relief of loading stresses during glacial retreat. The stress-release and resultant bulging developed because of the lengthening and steepening of slopes on less competent mudstone on removal of the ice, and where groundwater levels and pore-water pressures exceeded critical limits (Wilson 2014).

9. Great Dib

Great Dib is the name given to the steep back scarp of a very large deep-seated multiple rotational landslide (9A) (SE19983 44492). A schematic cross-section through Beacon Hill and Great Dib Landslide is shown in Fig. 5. The Great Dib landslide represents one of the largest rotational landslide on the Chevin. The back scarp comprises 20m of Addingham Edge Grit and the landslide includes a considerable thickness of underlying strata. Below the back scarp is a complex of large, rotated blocks which extend downslope to north of Spring Side (SE 19943 44843). The failure appears to have occurred along a non-circular slip plane, as the landslide has a marked translational element in its lower parts. The upper part of the Great Dib landslide, lying generally above 120m, consists of a series of large slump features which affect the higher beds of the argillaceous rock, and often involves the capping strata of the Addingham Edge Grit. Several secondary slumps have taken place on the steep edges of the rotated blocks. The large failures are separated by smaller multiple rotational landslides which have been modified by subsequent mudslide and mudflow mass movements. A small pond is present on a rotated block 9B (SE 19854 44756) to the south of Spring Side. This area is characterised by numerous small springs and boggy hollows. The lower part of the Great Dib landslide comprises a complex of frequently overlapping mudslides and mudflows which have developed from the higher rotational landslide debris aprons (9E). Pronounced mudslides and mudflows have developed down slope north of Oakdene, Spring Side and Woodlands to Birdcage Walk and extend north of Otley Bypass and probably are present further north in the now built up area (SE 19797 45091) which comprises of highly disturbed and reworked Glacial Till. The mudslides and mudflows are typically composed of numerous sub angular, poorly sorted fragments of bluish grey to brown, friable mudstone, 0.5 to 5cm in diameter, in a soft remoulded clayey matrix, but also incorporates large amounts of Glacial Till, which often contains rounded grit cobbles. This implies that the mudslides and mudflows must post-date glaciation in this area.

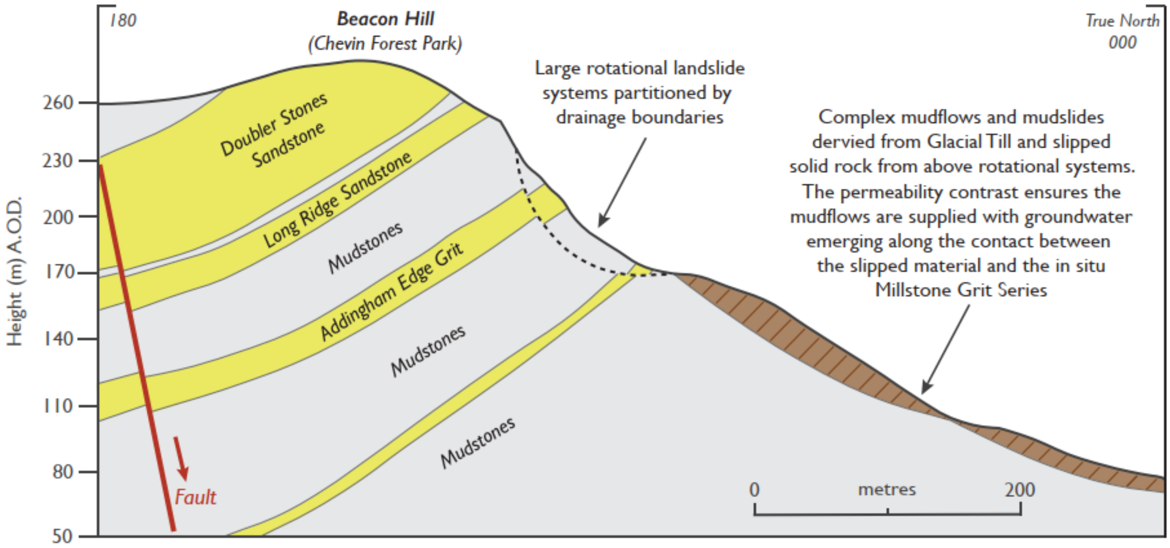


Fig. 5. Schematic cross-section through Beacon Hill, Otley Chevin, showing the geology and areas of rotational landslides, mudslides and mudflows.

10. White House

Several single deep-seated rotational landslides are located to the east of Great Dib Landslide and north of Ritchies Plantation (Landslides 10A (SE20076 44569), 10B (SE20156 44513), 10C (SE20304 44551) and 10D (SE20308 44442). The White house (SE 20309 44488) is located on the rotated block of landslide 10B. The ruins of another old house are also located on the same rotated block (10B) south of the White House. Several small slumps are located along the back scarp extending from the pig farm (SE 200 445) to the White House. The old building formerly located to the south of the White House was demolished due to severe structural damage. To the north of the pig farm is an active mudflow area extending towards Birdcage Walk (5E).

11. Cowburn Nursery

Ritchies Plantation (SE 204 444) contains two rotational failures, and the debris from these covers much of the wooded area (11A and 11B). Several small mudflows extend north towards Cowburn Nursery. Downslope of the main rotated blocks of landslides 11A and 11B is an extensive landslide accumulation area (11C) (area between Lawn Cottage (SE 20393 44600) and Cowburn Nursery (SE 20488 44696) comprising of disturbed and hummocky ground. Extensive mudslide and mudflow deposits can be observed north of the 120m contour of the Chevin escarpment. This area of mudslides and mudflows extends down slope affecting a large area (11D) (SE 20362 44796) to Birdcage Walk and further downslope to Otley Bypass. Particularly well-developed mudflow features can be seen around the Old Vicarage (SE 20322 44833), Beach Grove (SE 20526 44866) and Chevin House (SE 20595 44864). A borehole at Chevin House recorded evidence of at least three separate phases of mudslide and mudflow movement. There is another back scarp and deep-seated rotational landslide east of Chevin House at 11E (SE 20668 44622). The accumulation zone of this rotational landslide coalesces into mudslides and mudflows in the area around Chevin Villas (SE 20894 44929) and probably extends northwards beyond Otley Bypass and the A660 Leeds Road.

12. Northeast of Manby House

The area southeast of Manby House (SE20952 44645) shows little evidence of landslide disturbance, although solifluction Head Deposits are widespread in this area. Northeast of Manby House and downslope of East Chevin Road is a complex of small rotational landslides 12A (SE20825 44786). This area comprises hummocky ground and extends northwards into an area of mudflow extending to the A660 and further downslope to the north of the road (12B). A small artificial pond (Silver Mill Pond SE 20926 44808) is located on one rotated block. Silver Mills Reservoirs (SE 20954 44852) were also constructed and fed from the Silver Mills Spring east of the pond and Silver Mills Cottages (SE 209 448). More recent smaller rotational landslides 12C (SE 21028 44888) and 12D (Orchard House, SE 21061 44967) have developed within the older mudslide and mudflow complexes. Orchard House is located within a former clay pit located at the roundabout junction between Otley Bypass and the A660. The south facing slope of the former clay pit has exposed former mudslide deposits comprising reworked Glacial Till and shows evidence of continued instability.

13. East Chevin Quarry

East Chevin Quarry (SE 211 444) and the crags (see Photo 1) extending westwards to Ritchies Plantation show evidence of rockfalls). A schematic cross-section through East Chevin Quarry and the Birches Landslide complex below, is shown in Fig. 6. The slopes below the crags are covered with large grit boulders which decrease in size and number down slope. These are particularly evident in the field adjacent to Ritchies Plantation (SE 207 444) which has not been cleared for cultivation. The rockfalls have been caused by toppling failures.

Photo 1. Exposure of the Addingham Edge Grit to the west of East Chevin Quarry (west of Area 13).

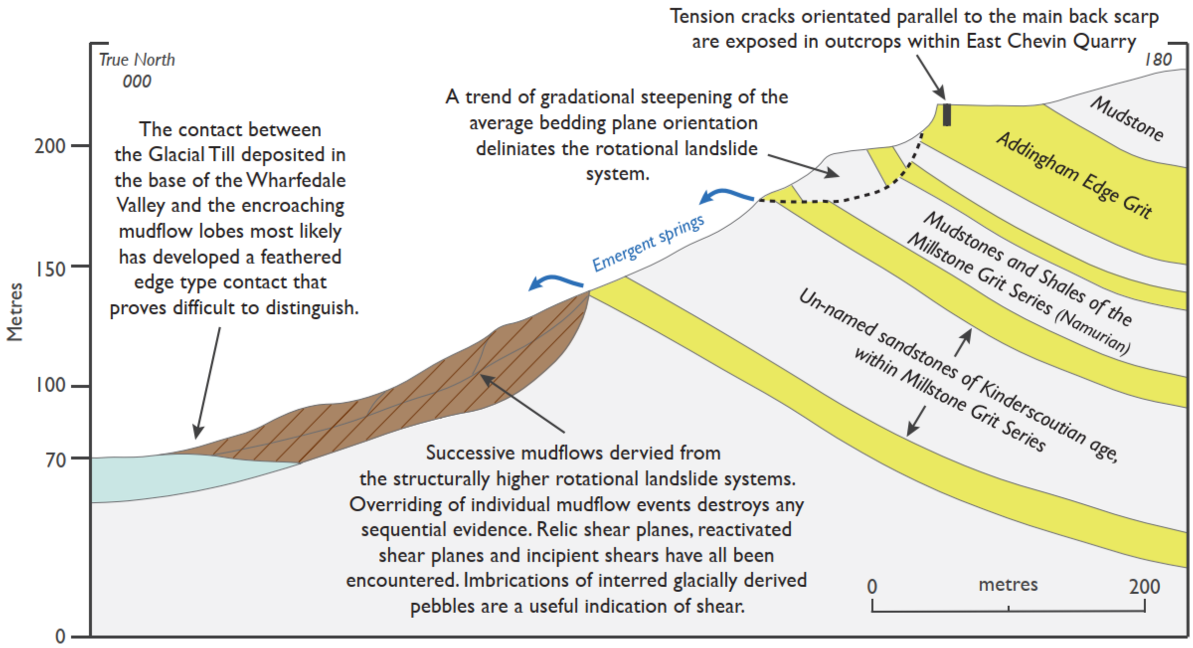


Fig. 6. Schematic cross-section through East Chevin Quarry and the Birches Landslide complex.

14. The Birches

North and east of The Birches (SE 210 446) extends a complex of deep-seated multiple rotational landslides (Landslides 14A (SE21143 44350) and 14B (SE21143 44762) (see Fig. 6). These landslides have well developed steep back scarps in the Addingham Edge Grit. North of the initial rotated blocks of landslides 14A and 14B, there are secondary rotational failures within an extensive area of mudslides and mudflows extending north of the A660 Leeds Road and Willow Bank 14C (SE 21177 45023). A series of well-developed mudflows can be seen north of the A660 Leeds Road around Willow Bank 14C (SE 21141 45061), Brunswick House 14D (SE 21230 44971) and Arnwood 14E (SE 21402 44954) (see Photo 2).

Photo 2: Mudflow toe area (Area 14D).

15. *Danefield Wood*

Danefield Wood (SE 212 445) is an area comprising several large deep-seated multiple rotational landslides (15A). The back scarps of these landslides have affected the Addingham Edge Grit as far south as Springfield Farm 15B (SE 21617 44225), (see Photo 3). Rockfalls have also affected the Addingham Edge Grit. Scattered boulders derived from outcrops of the Addingham Edge Grit are common on the upper slopes and increase in abundance further east. The large deep-seated rotational landslides have had the effect of lowering the crest of the Chevin escarpment along an east west alignment, 150m north of Springfield Farm (SE 21606 44376) in Danefield Wood (15B). The cambering has resulted in the beds of Addingham Edge Grit to tilt gently northwards. Small camber gulls (shallow linear depressions over tension cracks/joints) are present near the crest of the Addingham Edge Grit escarpment, where large blocks have tilted to the north. Several parts of the Addingham Edge Grit have become detached from the edge and have moved downslope. Northwards from this area, massive Addingham Edge Grit blocks litter the slope and mudslides and mudflows supporting the detached gritstone blocks are widespread over the lower slopes south of the A660. The debris aprons of the rotational landslides are marked by hummocky ground extending to the north and west. The northern extent of the hummocky ground of the main rotational landslides (15A) is marked by a line of springs (SE 21508 44687). Well-developed mudslides and mudflows (SE 21546 44922) extend northwards from the main accumulation areas of the rotational landslides (15C).

Photo 3. Steep back scarp of rotational landslide (Area 15A).

16. *Water Trough Bend*

Several small to medium-sized rotational landslides south of Water Trough Bend 16A (SE 21896 44515) have resulted in significant damage to the A660 over many decades. Numerous springs are located at the northern edge of the rotational landslides and widespread mudslides and mudflows (SE 21818 45074) have spread downslope northwards. A small rotational landslide scarp slope is located south of the A660 at Water Trough Bend (16A). The active area of landsliding at sub area (16B) has caused movement of the retaining wall constructed on the north side of the A660 (see Photo 4).

Active landslide movement at Water Trough Bend has previously affected the A660 Leeds Road over a length of 180m. Earliest reports of movement of the A660 date back to 1930. The A660 footpath and retaining walls have been damaged in several places (SE 21894 44812). Several springs emerge from the hillslope just to the south of Water Trough Bend (SE 21943 44635). Periodic remedial works have been required to stabilise the A660 Leeds Road at Water Trough Bend (SE 21986 44744). A series of shallow land drains were installed to the south, upslope of Water Trough bend in 1947. In 1964 twelve 40-58m long drains (30mm diameter) were drilled approximately at an angle of 5° above horizontal upslope north of the A660. These boreholes were installed with 115mm diameter porous concrete filter pipes, with an outfall constructed in a masonry retaining wall (Robinson 1967). The water from the drains is collected in a V-ditch and then into a carrier drain, which discharges in Holbeck, 100m to the east of the retaining wall outfall. After heavy rainfall the combined drain discharges were estimated to be 45,460 litres per day. The drainage helped to reduce movement of the shallow rotational landslides and mudslides north of the A660 until 1977, when the drains became blocked because of iron (III) hydroxide ($Fe(OH)3$) precipitation formed around the concrete filter pipes. Up to 23mm of horizontal displacement was recorded on the A660 carriageway together with subsidence of the northern retaining wall at Water Trough Bend. A further ground investigation was undertaken by West Yorkshire County Council in 1979. This confirmed that instability was confined to shallow depth and recommendations were given for the installation of counterfort drains above and below the A660; these drains were not installed.


Photo 4. Damaged retaining wall north of A660 (Area 16C).

North of the A660 at Water Trough Bend are active mudslides and mudflows which have spread out over an area of 150m^2. The mudslide terraces include fresh tension cracks and the mudflow toe lobes show evidence of active pressure bulging. The mudslides on the north side of the A660 Leeds Road have also been affected by small rotational landslides (16B, C, D and E). They also show signs of recent movement through the appearance of tension cracks. An area of active movement is located at SE 21530 44799, where reactivation of a mudflow lobe has taken place just to the north of Hawes House 16C (SE 21538 44794). The mudslides and mudflows extend over a widespread area north of Russell Farm (16D) (SE 21635 44983). The slopes in this area have been cut and the area has been redeveloped for a house and garden.

Leach (1984) carried out a geophysical investigation of the mudflows north of the A660 at Water Trough Bend and correlated the findings with previous ground investigations carried out in November 1979 and June 1984 by West Yorkshire County Council. Leach (1984) confirmed a series of overlapping mudflows, identified by material colour changes and discrete planar discontinuities. The thickness of the mudflows ranged up to 6.0m in thickness. The mudflows comprised silty sandy clay, with discrete sandy lenses; sandstone fragments were present throughout the mudflow deposits. Beneath the mudflow material was dark grey stiff clay, occasionally sandy with discrete gravel horizons. Leach (1984) described this material as disturbed landslide material. Below the landslide material, completely weathered mudstone was present containing shale fragments and this transitioned into more competent/intact mudstone and shale bedrock. The previous boreholes and geophysical investigation revealed a clearly recognisable broken zone at 10 to 15m depth bgl for approximately 100m from the A660 downslope within the mudstone/shale bedrock, relating to a deeper zone of rotational movement and probably emanating from the arcuate back scarps observed south of the A660 (Leach 1984).

17. Poolscar Wood

There are several large deep-seated rotational landslides in Poolscar Wood (17B) (SE 22111 44436). These rotational landslides have destabilised the main Addingham Edge Grit and resulted in significant cambering of the gritstone strata 17A (SE 22199 44314). Several springs are present within the landslide accumulation zone within area 17c (SE 22216 44495); a number of these are captured within new culverts which have been constructed beneath the A660. North of the A660 is a large area 17D (SE 22202 45094) of mudslides and mudflows. Several small rotational landslides and slumps have developed on the mudslide and mudflow deposits. Several springs also emerge from the toe areas of the mudslides and mudflows.

18. Pools Carr

South of the A660 at Pools Carr (SE 22592 44405), the valley side comprises of a series of large deep-seated rotational landslides which extend from the Addingham Edge Grit escarpment (18A) (see Photo 4). The back scarp (5m in height) of one of the landslides 80m south of the A660 has resulted in a large block failure in the Addingham Edge Grit (18B) (see Photo 5). Below the back scarp of the rotational landslide are a series of slumps, each with a throw of several metres, each of these is separated by a small back scarp. The back scarps have associated marshy areas. North of the A660 are numerous mudslides and mudflows and abundant springs. West of Pools Carr is an active landslide and mudflow 18C (SE 22493 44777). Northwards movement of this active area has displaced the northern retaining wall of the A660 by approximately 1m over a length of 120m; the wall has also subsided by 0.3m. Several springs emerge beneath the retaining wall on the north side of the A660. The presence of several small back scarps and associated mudflow features indicate that the landslide is a complex of rotational landslides with associated mudflows. South of the A660 the slope is hummocky and littered with many large, detached gritstone blocks 18D (SE 22581 44602).

Photo 5. Gritstone Edge comprising Addingham Edge Grit showing large open joint behind the face (Area 18B).

Extensive mudslides and mudflows are present north of the A660 in area 18F. Leach (1984) carried out a geophysical survey of the mudslides and mudflows north of the retaining wall on the A660 (18E). Leach (1984) determined two and sometimes three sets of overlapping mudflows on the northern slopes below the A660. These comprised a 1.1m layer of soft brown sandy clay with boulders and fragments of sandstone (mudflow deposit). This material overlay soft to firm sandy silty clay containing organic horizons between 1.5 and 4.0m thick (mudflow deposit). This material, in turn, overlay stiff sandy clay containing gravels (remoulded Glacial Till disturbed by mudslides). This material overlay completely weathered mudstone and shale (5.0-6.3m bgl) which transitioned into more competent mudstone bedrock. However, Leach (1984) identified a discrete zone of broken mudstone and shales at a depth of 10.15m bgl 40m north of the A660 carriageway further downslope. This broken ground was interpreted as a slip surface associated with the rotational slides indicated by conspicuous back scarps to the south of the A660.

19. Caley Crags

To the north of Caley Crags, the Chevin escarpment falls away steeply; this area coincides with the back scarps of several large rotational landslides 19A (SE 22876 44293). The Addingham Edge Grit escarpment is cambered 19B (SE 22739 44360) and several large gritstone blocks have become detached and moved down slope onto the rotational landslide accumulation zone (see Photo 6). Numerous springs emerge near the northern edge of the main accumulation zone of the rotational landslides. Below this zone, extensive mudslide and mudflows are present and extend north of the A660 19C (SE 23166 44780). An active landslide located in area 19D (SE 23257 45130) has developed where excavations have taken place for the extension of farm buildings.

Photo 6. Rafted boulders which have become detached from gritstone edge comprising Addingham Edge Grit (Area 19B).

20. Pool Bank Quarry

The Addingham Edge Grit was worked in Pool Bank Quarry (20A) between 1800 and 1930; when it closed it had a maximum depth of 25m. The quarry waste materials were tipped on the valley side slopes to the north of the quarry (20B). The quarry waste was tipped on an area previously affected by extensive rotational landsliding 20C (SE 23383 44598). The quarry waste was tipped over an area of $100m^2$ north of the A660 as a large steep-sided fan. Leach (1984) noted from exposures that the quarry waste had been placed on a discrete layer of sandy clay containing sandstone fragments. Leach (1984) attributed this layer to clearance of material from the crest of the escarpment before quarrying commenced. Some movement has taken place above this sandy layer in places. However, the undrained loading caused by this quarry waste material south of the A660 has mainly been responsible for reactivation and movement of former shallow mudflows, mudslides and slumps on the surface of the former hummocky landslide debris material present below the tipped material 20D (SE 23453 44801). Numerous springs are also present in the landslide accumulation zone of landslide (20D). Further downslope springs and seepages can be seen at the toe of the former landslide debris zone and also at the base of mudflow lobes (20E). Cooper (1984) estimated that over 20m of movement of the former landslide had taken place as a result of the quarry waste being placed on the pre-existing landslide. There is a large back scarp which coincides with the north side of the A660 above the area of tipped quarry material. Over 40m of the A660 carriageway has been affected by landslide movement, with tension cracks forming in the carriageway and northern pavement on a regular basis during the 1970s. The northern edge of reactivated mudflow features has formed toe bulges of 3-4m in height (20E). A small active landslide is located near 20C (SE 23299 44577) which has formed within the back scarp of a pre-existing landslide at SE 23425 44689.

Pool Bank Quarry has now been infilled with municipal landfill and landscaped; the quarry was filled above the original valley slope levels. The quarry restoration will have significantly changed the hydrogeological and surface water regime of the Chevin escarpment at Pool Bank Quarry. Cooper (1984) postulated that the effects resulting from the landfill (i.e., change in loading near an existing landslide caused by the additional weight of fill and possible resultant changes in water table level and groundwater flow) may provide conditions favourable for triggering a significant failure of the slope. Cooper (1984) noted that landslides 20C (SE 23299 44577) and 20F (SE 23292 44458) extend back into the Addingham Edge Grit, almost as far back as Pool Bank Quarry 21B (SE 22935 44395). It is possible that the excavation of the Addingham Edge Grit from the quarry has produced an area of weakness in the remaining sandstone escarpment which might, under certain conditions, form a failure plane. It is also possible that the lack of support to the north of the escarpment due to Landslide 20B, the weight of the fill in Pool Bank Quarry, and possible increases in the water table level and groundwater flow may produce destabilising conditions of the escarpment slope.

Ground investigations were carried out across the Pool Bank Landslide areas in July 1984. Eight boreholes were completed as part of this investigation and inclinometers and piezometers were installed within the boreholes. The boreholes confirmed that the mudstone and shale bedrock north of the A660 is overlain by landslide debris up to 4.5m thick. The quarry waste has been tipped over the landslide debris. The bedrock in this area also comprises thin beds of sandstone. The bedrock comprised zones of broken ground interpreted to be the result of former rotational slip planes within the bedrock. These areas of broken ground recorded elevated piezometric surfaces.

21. Low Bank Plantation

The Addingham Edge Grit exposed in the A660 road cutting is cambered 21A (SE 23635 44381), with the dip changing from gently southwards to gently northwards near the junction with Old Pool Bank Road. The jointing in the rock becomes more open northwards towards the edge of the escarpment. Below the escarpment there are several small rotational landslides 21B (SE 23782 44598), some of these are partly concealed by quarry waste 21C (SE 23684 44633) from Pool Bank Quarry. The lower part of the slope 21D (SE 23992 44913) comprises of numerous well-developed mudflow features.

Geotechnical Properties of Namurian Millstone Grit Series

Millstone Grit Sandstone

The Millstone Grit Series sandstones have been characterised by Robinson & Williams (2005) and assigned a general description as a medium or coarse grained mainly feldspathic unit, with feldspar contents reaching 27.5%, (Aitkenhead & Riley 1996), making them sub-arkosic or in extreme cases arkosic. The large-scale crossbedding is also noted. The broad geotechnical aspects are outlined as marked lateral variations in thickness, and high lateral variability with abrupt passages from sandstone members to shale/mudstone members. Robinson & Williams (2005) noted that the Millstone Grit Series sandstones exhibit a relatively high strength and low porosity. These physical properties provide the rock with a high resistance to weathering. This is essentially the primary factor for the development of the prominent Millstone Grit (Addingham Edge Grit) outcrops along the length of the Chevin escarpment. Robinson & Williams (2005), define the outcrops as angular to sub-angular, with rounded crags and boulders also present.

The Namurian Millstone Grit sandstone which forms the Addingham Edge Grit escarpment of the Chevin exhibits distinct strength characteristics, which influences its weathering and general geotechnical behaviour. Xidakis (1976) carried out triaxial compression tests on dry cylindrical rock specimens (height 50mm x width 25mm) of sandstone and Gritstone taken from various Gritstone edge exposures on the Chevin. The results of multiple triaxial compression tests using the same series of confining pressures (0, 5, 10, 15, 20, 25 Mn/m^2) were combined to calculate the envelope parameters of angle of friction and apparent cohesion (see Table 2). A mean unconfined compressive strength of 28.2MPa, a friction angle of 66.8° and an apparent cohesion of 6MPa were determined for dry specimens.

Table 2. Summary of triaxial compression testing of sandstone rock specimens (Xidakis 1976)

Specimen Description	Experimental Mean Triaxial Test Results		Combined Mohr Envelope Results		
	Unconfined compressive strength (UCS) (MPa	Angle of friction (Ø) (degrees)	Friction angle (Ø) (degrees)	Cohesion (c) (MPa)	Unconfined compressive strength (UCS) (MPa)
Dry coarse sandstone (Chevin Top)	23.22	66.2	64.5	7	34.2
Dry Addingham Edge Grit (Caley Crags)	10.25	68.2	69.5	7	36.2
Dry medium grained sandstone (East Chevin Quarry)	13.4	59.3	66.5	4	14.2
Mean	15.62	64.6	66.8	6	28.2
Saturated medium grained sandstone (East Chevin Quarry)	10.6	60.0	68.0	2	

Xidakis (1976) noted that the values for angle of friction obtained from the Mohr envelope tend to be higher than the experimental results and that the unconfined compressive strength (UCS) values derived from the Mohr envelope are significantly higher in some cases than the experimental results. Xidakis (1976) also found a 21% reduction in UCS between dry and saturated samples of medium grained sandstone from East Chevin Quarry. Dyke & Dobereiner (1991) found the reduction in UCS in dry and saturated sandstones varied between 24 and 36%. Hawkins & McConnell (1992) found that the relationship between water content and the UCS of sandstones could be described by the exponential equation:

$$\sigma'_{c(w)} = ae^{-bw} + c$$

Where:

$\sigma'_{c(w)}$ is the UCS (MPa)

w is the water content (%)

a, b and c are constants (b is a dimensionless constant defining the rate of strength loss with increasing moisture content).

Hawkins & McConnell (1992) determined the UCS of dry Millstone Grit sandstone to be a maximum of 127.5 MPa. The Millstone Grit sample tested was described as medium-coarse grained with characteristic massive crossbedding. A sample of the mudstone end member of the Millstone Grit series was also tested and gave a maximum unconfined compressive strength for a dry sample of 59.3MPa (Hawkins & McConnell 1992). This UCS value is significantly higher than those reported by Xidakis (1979). Hawkins & McConnell (1992) also reported the tangential Youngs Modulus (Etan) for a dry Millstone Grit sandstone specimen and a dry mudstone end member of the Millstone Grit series specimen as 23.1MPa and 12.8MPa respectively.

Hawkins & McConnell (1992) concluded that the degree of strength sensitivity to water saturation is primarily controlled by the mineral composition and to a lesser extent by the rock micro-fabric. On further analysis of the Hawkins & McConnell (1992) data, Vasarhelyi (2003) calculated the mean loss in UCS after complete saturation was approximately 24.1% and that the relationship between saturated UCS satisfied a linear function. Vasarhelyi (2003) stated that for any given Millstone Grit Series sandstone unit, the compressive strength decreases by approximately 20% on saturation. There is a similar relationship applicable to the deformability moduli (Vasarhelyi, 2003). The UCS reduction on saturation calculated by Xiadakis and Vasarhelyi are consistent and a value of 20% can be used to provide an indication of the order of UCS reduction on saturation of the Millstone Grit Series sandstones. The exposed Millstone Grit Strata on edges of the Chevin will have become saturated during thaw periods during periglacial conditions which followed the last glacial maximum.

Xidakis (1976) also carried out axial Point Load tests on air dried core samples, semi prepared specimens and irregular lumps of Addingham Edge Grit from East Chevin Quarry. The uniaxial compressive strength was calculated using the methodology of Broch & Franklin (1972). The test results are shown in Table 3. The results are similar to those obtained by the experimental triaxial compression tests which are also included in Table 3. The Mohr envelope used for the derivation of UCS, Ø and c was obtained using the Bieniawski's equation (Bieniawski 1976).

Table 3. Summary of Point load testing and Triaxial testing of sandstone rock specimens (Xidakis 1976).

Rock Test	Specimen Type	I500 (MPa)	Uniaxial Compressive Strength (UCS) (MPa)	Mean friction angle (Ø) degrees.	Mean Apparent cohesion (c) (MPa)
Point Load Test	Cores	1.86	23.25	63	5
Point Load Test	Semi prepared specimens	2.26	28.50	65	7
Point Load Test	Irregular Lumps	2.06	25.75	64	6
Triaxial Compression Test	Mean Mohr Envelope	n/a	29.00	66.04	6
Triaxial Compression Test	Mean Experiment Result	n/a	16.00	65.03	n/a

Jones (1984) carried out Schmidt Hammer Tests on the Addingham Edge Grit in East Chevin Quarry. A mean UCS of 42MPa was determined based on 20 no. tests carried out on the quarry face. Leach (1984) carried out axial point load tests on standard samples of medium grained sandstone and coarse grained gritstone from the Addingham Edge Grit in East Chevin quarry, the results are shown in Table 4.

Table 4. Unconfined Compressive Strength of Addingham edge Grit derived from axial Point Load Tests (Leach 1984)

Description of Addingham Edge Grit from East Chevin Quarry	Point Load Test – Unconfined Compressive Strength (UCS) (MPa
Medium grained fresh quarzitic sandstone with 5mm laminations.	58.5
Medium grained micaceous quarzitic sandstone, unlaminated.	38.3
Very coarse grained fresh quarzitic gritstone.	28.1
Mean	41.6

The Addingham Edge Grit varied in strength depending on the lithology but had a mean UCS of 41.6MPa. This is significantly higher than the UCS determined by Point Load tests by Xidakis (1976). However, the Addingham Edge Grit can be classified as moderately strong (25-50MPa) according to BS 5930 (2015). Jones (1984) investigated the effect of the degree of weathering in reducing the UCS of the Addingham Edge Grit. Jones (1984) found that the weathered medium grained sandstone exhibited a lower strength (20.75-35MPa) mean UCS of 27MPa compared to unweathered sandstone; a reduction of 35%. Spivey (1982) carried out Schmidt Hammer (Type N) tests on the Addingham Edge Grit within Hanging Stones Quarry on Ilkley Moor. Tests were undertaken on medium grained sandstone and gritstone. The gritstone provided UCS tests ranging between 16 to 25Mpa and the medium grained sandstone 40 to 48MPa.

Spivey (1982) carried out axial Point Load tests on standard and non-standard samples of Addingham Edge Grit sandstone and gritstone on Ilkley Moor. The results of the Point Load testing are shown in Table 5.

Table 5. Axial Point Load tests on standard and non-standard samples of Addingham Edge Grit sandstone and Gritstone (Spivey 1982).

Addingham Edge Grit Description	Standard/Non Standard Specimen	Point Load Test – UCS (MPa)
Grey, medium-grained, slightly weathered sandstone.	Standard	89.7
Grey, coarse-grained, slightly weathered sandstone.	Standard	73.9
Light brown, thinly bedded, medium-grained, moderately weathered sandstone.	Non-standard	47.0
Yellow-orange, coarse-grained, moderately weathered sandstone.	Standard	31.8
Purple, very coarse-grained, strongly weathered sandstone.	Standard	12.8
Grey, moderately weathered gritstone.	Standard	41.5
Orange, moderately weathered gritstone.	Standard	31.9
Light grey, moderately weathered gritstone.	Standard	25.7
Orange, very coarse, strongly weathered gritstone.	Non-standard	6.12

The axial point load test results demonstrate the importance of lithology, with medium grained sandstone providing higher URS compared to coarse grained sandstone of gritstone. The results also show how the weathering grade is also an important factor affecting the strength of sandstones and gritstone.

The nature of the varied sandstone and gritstone outcrops along the Chevin escarpment is attributed to the action of severe periglacial conditions during the Quaternary (Robinson & Williams 2005). The Millstone Grit Series sandstone units outcrop on either side of the last Devensian ice limit, although all outcrops were once encompassed by the earlier Quaternary ice sheets. The many block fields and clitter slopes along the Chevin escarpment can be linked to the intense freeze-thaw processes operating under a periglacial climate. Robinson & Williams (2005), highlight the important fact that the Millstone Grit Series sandstones were immediately peripheral to the ice limits for much of the Devensian. The angularity of outcrops do not reflect the effects of periglacial processes but rather the rock hardness. The absence of any substantial surface rind or crust on the Millstone Grit Series sandstones has prohibited the development of polygonal cracking. The second surficial property of the Millstone Grit Series outcrops is their relative impervious nature. This property has allowed the development of weather pits and runnels which can be seen on the crags at Beacon Hill (SE 1989 4425), which are indicative of past storage and flow of water on and over the surface of the Millstone Grit Series sandstone (Robinson & Williams 2005).

From the pre-existing information available, the Millstone Grit Series sandstones are characterised by a rock mass strength controlled primarily by discontinuities and strength anisotropy. These features are the result of the rocks formational processes and from past regional structural deformation. Fracturing and faulting have been induced on both a large and small scale. The more recent periglacial activity that dominated the Quaternary has allowed the nucleation of further jointing and fracturing on the outcrop scale, leading to disintegration of the rock mass and production of block fields. The action of water is known to have produced surficial discontinuities in the form of linear erosion features.

Mudstones and Shales

The intervening mudstones and shales within the Millstone Grit Series, shown on the 1: 50,000 solid geological map, Sheet 69, Bradford, are shown as Millstone Grit (undifferentiated). The mudstones are grey to black, weathering to orange-brown, mottled pale grey, planar laminated and micaceous, or massive. Commonly, they contain non marine bivalves. Generally, ironstone nodules are common in mudstones, ranging in size from a few millimetres to tens of centimetres in diameter. The mudstones are commonly overlain gradationally by siltstones, which are typically medium grey with flaser and lenticular bedding, ripple cross-laminations and parallel lamination and commonly contain plant debris. The siltstones grade both vertically and laterally into sandstones.

Ground investigations for the anchored retaining wall on the Otley Bypass confirmed 3.5m of solifluction/mudslide materials overlying completely weathered to weak mudstone (Snowden et al. 1986). Undrained and consolidated undrained triaxial tests with pore-water pressure measurements were carried out on the weathered mudstone. Undrained shear strengths were extremely variable, ranging between 21 and 210 kPa, with an average of 70 kPa. Effective stress parameters were found to lie in the following ranges: Effective angle of shearing resistance (Ø') = 22° to 34° and Effective cohesion (c') = 10 kPa to 30 kPa.

Xidakis (1976) carried out triaxial compression tests on dry cylindrical rock specimens (height 50mm x width 25mm) of a thinly bedded sandy-micaceous shale and mudstone derived from cores using rotary coring methods. Under zero confining pressure the specimens failed giving a friction angle of 65° and a uniaxial compressive strength (UCS) of 8-10 MPa. Xidakis (1976) also carried out axial point load tests on saturated cores and air-dry semi-prepared pieces of shale and mudstone. The results of the point load testing are shown in Table 6.

Table 6. Results of point load test on saturated and dry specimens of mudstone and shale (Xidakis 1976).

Type of Specimen	Point Load Test	Mohr Envelope Results	
	Unconfined Compressive Strength (UCS) (MPa)	Cohesion (c) (MPa)	Friction angle (∅) (degrees)
Dry shale/mudstone (Semi-prepared specimens)	19.13	5	61.7
Saturated shale/mudstone (cores)	15.94	4	61.1

The UCS of the shale and mudstone specimens determined by axial point load tests appears very high and is 50% higher than the UCS strength of 8-10MPa determined by the triaxial compression tests. The UCS results from the point load tests on shale and mudstone are also higher than the experimental UCS results for dry sandstone at 16.0 MPa. It is possible that the point load tests were undertaken on samples of a strong siltstone rather than weak mudstone and shale; these can be found as thin layers within the mudstone. Xidakis (1976) determined a reduction of 17% in the UCS between dry and saturated mudstone and shale specimens.

Snowdon et al. (1986) carried out undrained and consolidated undrained triaxial tests with pore-water pressure measurement on weathered mudstone for the design of the Silver Mill Hill Retaining Wall. Undrained shear strengths were extremely variable, ranging between 21 kPa (0.021 MPa) and 210 kPa (0.21 MPa), with a mean of 70 kPa (0.070 MPa). Effective stress parameters were found to lie within the following ranges: Effective angle of shearing resistance (Ø') = 22° to 34° and Effective cohesion (c') = 10 kPa (0.01 MPa) to 30 kPa (0.03 MPa).

As part of the ground investigations for the Otley Bypass undrained triaxial tests were undertaken on samples of weathered mudstone from boreholes at the location of the Chevin Bottom Bridge at the bottom of West Chevin Road. The mudstone was described as shaley with occasional silt partings. The mudstone was clayey in part and water-bearing. The bedding planes dip between 0° and 10° from the horizontal. The weathered mudstone was approximately 4.0m in thickness and then becomes less weathered and more competent with depth. The undrained triaxial tests carried out on the weathered mudstone gave effective angles of internal friction (Ø') of 26° to 30° and an effective cohesion c' = 50 kPa.

The mudstone and shale units on the Chevin are poorly exposed. An exposure below a property called Leeden on West Chevin Road comprised dark grey, fine grained, finely laminated, fissured, friable mudstone. The mudstone and shale weathers rapidly to a brown colour, and iron hydroxides are deposited along fissures. Occasional nodular concretions of siderite, up to 5cm in length, also occur, and these are highly weathered. Jones (1984) completed slake durability tests on two samples of the mudstone and shale exposed in the slope north of Leeden on West Chevin Road. The results confirmed that the two samples tested to have a very low to low slake durability. On the Chevin where mudstone and shale units become exposed in cuttings, they weather rapidly and rarely form outcrops, suggesting the rocks are very weak (1-5Mpa) to weak (5-25MPa).

As part of supplementary ground investigations for Otley Bypass, two boreholes A and B were undertaken within mudstone strata at the eastern end of the Bypass near the Old Orchard (SE 211 449) (Culshaw & Duncan 1975). The samples recovered from borehole A (Ground level 73.21m AOD) consisted entirely of grey mudstone. Down to approximately 2.0m bgl the mudstone was weathered, from 0.5 to 1.1m bgl being highly weathered but the degree of weathering decreases with depth. The weathering is characterised by iron along the shaley partings and the more broken nature of the rock. The groundwater table was encountered at 1.7m bgl. Below 2.0m depth bgl the mudstone is unweathered blue-grey, weak, shaley and slightly silty. Jointing is generally high or medium angled (between 0° and 45° relative to the axis of the cores). Throughout the succession to 10.15m bgl lithological banding occurs, the mudstone varying between almost a siltstone and a true mudstone. No regularity was noticed in this variation in lithology. In Borehole B the mudstone was encountered beneath 12.2m of superficial deposits comprising mudslide and mudflow material (reworked Glacial Till). The mudstone was unweathered between 12.2 and 24.0m bgl. Moisture content and Atterberg test results are provided for weathered and unweathered samples of mudstone from boreholes A and B (see Table 7).

Table 7. Atterberg test results on weathered/unweathered mudstone from Boreholes A and B (Culshaw and Duncan 1975).

Laboratory Test	Weathered Mudstone BHA	Unweathered Mudstone BHA	Unweathered Mudstone BHB
Moisture Content	15 to 16%	8 to 13%	6 to 12%
Plastic Limit	26%	19 to 23%	15 to 21%
Liquid Limit	46%	32 to 38%	25 to 35%
Plasticity Index	20%	12 to 17%	9 to 16%
Liquidity Index	-0.50	-0.47 to -1.25	-0.50 to -1.11

Borehole A shows high moisture content recorded within the weathered mudstone. Moisture content varied between 8 and 13% within the unweathered mudstone, gradually decreasing with depth as fissures and pores close up with increasing overburden pressure. Plastic limit varies little in the unweathered mudstone ranging from 32-38%. Consequently, the plasticity index remains virtually constant, between 11 and 14%, except for values of 17% at 7.10m and 7.75m bgl. This may reflect a slight decrease in the proportion of silty material in these samples. The liquidity index is low throughout the mudstone succession; one of the highest values being in the wetter near surface mudstone. However, this had a value of -0.5, otherwise the liquidity index varied between -0.47 and -1.25. All the liquidity indices are negative; no remoulded mudstone was encountered in Borehole A. The values obtained indicate that the mudstone is in a brittle state.

Discussion on rock testing on the Millstone Grit Series Rocks

Xidakis (1976) provided a summary of triaxial and point load testing conducted on samples or air-dried and saturated samples of sandstone and mudstone and shale collected from Otley Chevin (see Table 8). Based on the values shown in Table 8 the strength of the sandstone is relatively high and is capable of permitting the sandstone to stand at a near vertical slope, even in saturated conditions when the rock exhibits a lower strength. Xidakis (1976) found that on saturation of the sandstone specimens this resulted in a reduction in the UCS by 33% and a reduction in cohesion by 67%. Broch & Franklin (1972) found a reduction in the UCS of saturated sandstone samples of 30% and Colback & Wild (1965) of 48%. Therefore, there is reasonable correlation with Xidakis (1976) test results. However, it should be noted that the compressive strength tests relate to intact rock and not the in-situ rock mass which is faulted, jointed and distorted by cambering. The strength of rock parallel to its discontinuities is significantly lower than the intact rock. The cohesion intercept, for instance, parallel to the rock discontinuities approach almost zero, or to that of the joint filling (gouge material) (Hendron 1968). Therefore, in estimating the sandstone strength, it is important that shear tests have to be made along the joint surfaces or the strength parameters of the gauge material must be taken into account.

Table 8. Results of triaxial compression and point load tests on air-dried and saturated samples of sandstone and mudstones and shales (Xidakis 1976).

Rock Type	Dry/Saturated Sample	Uniaxial Compressive Strength (UCS) (MPa)	Cohesion (c) (MPa)	Friction angle (Ø) (degrees)
Sandstone	Dry	16.0	6	42
Sandstone	Saturated	10.6	2	38
Sandstone fault gouge	Dry	n/a	0.0075	12
Mudstone/Shale	Dry	19.0	4	33
Mudstone/Shale	Saturated	16.0	5	32

The gritstone edge failures on the Chevin mainly take the form of toppling, supplemented by sliding in some cases. This type of failure mechanism demonstrates that the tensile strength and the friction resistance are the most important parameters of the sandstone strata. The tensile strength is relatively low, UCS/10 = 1 KPa in saturated conditions as shown in Table 8. However, the stability of the sandstone slopes does not depend upon the strength of the sandstone, but mainly upon the strength parameters and stability of the underlying mudstone and shale strata. Any yield in the supporting mudstone and shale strata, results in the development and opening of joints in the overlying sandstone, reduction in strength, creation of large tension cracks, detachment of the individual blocks, augmented by frost action and failure by toppling or cambering. Erosion of the underlying mudstones and shales beneath the sandstone layers, which dip inward to the slope leaves the rock mass unsupported, which fails by bending and shearing (e.g., failed blocks of sandstone below Caley Crags). Erosion of the sandstone itself helps to widen its discontinuities and accelerate failure.

Characteristic slope angles and threshold or limiting slopes

A number of authors have interpreted the occurrence of characteristic slope angles as limiting or threshold angles for various types of mass movement and have recognised that many upland areas of Britain may be interpreted as relic features of post glacial and periglacial mass movements (Hutchinson 1967; Hutchinson & Bhandari 1971; Carson & Petley 1970; Weeks 1969; Chandler 1970a, 1972; Chandler et al. 1973; Early & Skempton 1972; Rouse & Farhan 1976; Anderson et al. 1980). Threshold or limiting slopes may be regarded as the boundaries above, or below which, particular geomorphological processes cannot operate (Chandler 1970b; Carson & Petley 1970; Anderson et al. 1980; Brooks et al. 1993; Montgomery 2001; Cross 2011). Various mass movement processes, including deep and shallow landslides, mudslides and mudflows, may set limits to the angle at which a particular slope type may be maintained.

Prediction of the angle of ultimate slope stability (β_L) is site-specific as it depends on a detailed understanding of slope properties and piezometric conditions at shallow depth at the time of failure (Anderson et al. 1980; Cross 2019). However, it is possible to use more simplistic methods to assess limiting slope stability angles and the prevailing geomorphological processes associated with them. Taking into consideration the triaxial compression and point load testing results obtained by Xidakis (1976) it is possible to explain the limiting angle (β_L) for characteristic slopes on the mid and lower parts of the Chevin. Xiadakis (1976) noted that triaxial compression and diametral point load tests taken parallel with the mudstone and shale bedding planes are near zero and specimens could easily be sheared by hand. If a cohesion of zero (c = 0) is taken for a weak fractured mudstone, and a friction angle of 32°, then the mechanical behaviour of a weak mudstone can be regarded to be similar to a hard fissured clay. Assuming these parameters for slopes developed on completely weathered and very weak mudstones it is possible to determine the limiting slope stability (β_L) of these slopes by applying the Infinite Slope Stability Assessment (ISSA) method initially proposed by Skempton & DeLory (1957). Versions of the ISSA model have been used by engineering geomorphologists to determine the limiting slope stability (β_L) and to help explain hillslope morphology, development of 'straight' slopes and natural hillslope evolution processes (Anderson et al. 1980; Rouse 1989; Cross 2010, 2011, 2019).

Based on the Skempton & Delory (1957) equation below and the test results of Xidakis (1976), an assessment of the limiting slope stability of completely weathered and very weak mudstone slopes can be made as follows:

$$F = \frac{c' + (\gamma_s - \gamma_w) \times Z \times cos^2\beta \times \tan\varphi'}{\gamma_s \times Z \times \sin\beta \times \cos\beta}$$

Where:

F = Factor of Safety

c' = Effective cohesion of mudstone

γ_s = Saturated density of mudstone (Xiadkis 1976)

γ_w = Density of water

Z = Depth

φ'= Effective friction angle of mudstone (Xidakis 1976)

ψ_f = Slope angle (to the horizontal)

For c' = 0 and water seepage parallel to the ground surface, the water pressure is $\mu = \gamma_w Z cos^2\beta$ and the equation becomes:

$$F = \frac{\gamma_s - \gamma_w \times \tan\varphi'}{\gamma_s \tan\beta}$$

The factor $\gamma_s - {}^{\gamma_w}/_{\gamma_s} = {}^1/_2$

If at limited equilibrium F = 1; $\tan\beta$ = $\tan\varphi'$ or $\beta_L = {}^{\varphi'}/_2 = 16^{\text{o}}$

The limiting slope stability (β_L) angle of 16° was identified by Xidakis (1976) to be a typical slope angle observed on the mid and lower slopes of the Chevin and within areas affected by shallow landslides, mudslides, mudflows and regolith developed over mudstone. The author also recognised this characteristic limiting slope category to be important in research carried out in the Southern Pennines (Cross, 1987, 1998, 2010, 2011, 2019). This limiting slope category coincides with the category 15-18° - Convex creep slopes. This category corresponded to the limiting slope angle for Head Deposits and regolith exhibiting shallow planar movement and in shallow or deep-seated translational landslide debris. The author recognised the characteristic slope categories on south Pennine Slopes as shown in Table 9 (Cross 2011).

Table 9. Characteristic slope angle categories in the South Pennines (Cross 2011).

2° - 6° Structurally controlled plateaux or valley floors
Upland plateaux with peat or regolith cover, localised creep mass movements and gulley erosion; lowland valley floors with alluvial floodplains or mudslide and earth flow debris from higher landslides.
12° - 14° Lower seepage slopes
Block debris from creep and solifluction, with gullies below seepage areas; hill-wash and soil creep are active.
15° - 18° Convex creep slopes
Threshold angle for head and regolith, with shallow planar movement on shallow or deep-seated translational landslides.
19° - 23° Mid-level transportational slopes
Slope unit most affected by mass movements, forming the steeper valley side slopes, covered with landslide debris; threshold angle for fossil scree on mudstone and shale strata, and for sandy regolith.
24° - 28° Talluvial slopes
On talus at the threshold angle for talluvium; forming the steeper scarp slopes and back scars of rotational landslides.
29° - 32° Fossil scree slopes
On fossil scree of sandstone and gritstone, affected by deep-seated rotational slides and translational slides; slides may have lower debris flow components.
33° - 36° Dry scree slopes at the foot of free-faces
Drained scree at the foot of sandstone and gritstone free-faces; at the threshold angle for dry scree.
37° - 45° Slopes on fractured or jointed rock
On fractured and drained rock; common below free-faces with rockfalls and toppling failures.
>45° Gritstone edges and free-faces
Gritstone edges and sandstone free-face.

The author identified characteristic limiting slopes at 17.7° to be important in the southern Pennines (Cross 2011), which is similar to the predicted mean limiting angle for solifluction materials and correlates with the slope angles associated with shallow planar failures in South Wales identified by Rouse & Farhan (1976). Most of the characteristic slope categories recognised by the author are also applicable to the Chevin slopes, although the category 33-36° Dry scree slopes at the foot of free-faces are not present on the Chevin slopes. This type of material appears to have been either covered by solifluction material or has been incorporated within Head Deposits and moved further downslope. The author also identified characteristic limiting slope angles(β_L) of 21.0-21.8° to be important in the southern Pennines; this represents the threshold angle for sandy soil regolith (19.6-21.8°), this slope category also appears to be an important characteristic limiting slope angle on the Chevin according to geomorphological mapping of the Chevin slopes undertaken by Xidakis (1976), Jones (1984), Cooper (1984) and Germaine (2007).

Discontinuity Survey of Addingham Edge Grit

Jones (1984) completed a discontinuity survey on the Addingham Edge Grit at East Chevin Quarry and on the exposed escarpment to the west of the quarry. Three major and two minor discontinuity sets were observed and display the following characteristics:

i) Bedding, orientation 18/148°, spacing 4m, tight, rough.
ii) Major joint, orientation 78/060°, spacing 3m, open, clean, smooth.
iii) Major joint, orientation 78/133°, spacing 3m, open, clean, smooth.
iv) Minor joint, orientation 78/044°, spacing 0.5m but irregular, tight, smooth.
v) Minor joint, orientation 82/338°, spacing 0.5 but irregular, tight, smooth.

Bedding was observed by Jones (1984) to be poorly developed, and usually only in evidence where softer beds have weathered out to leave small overhangs at the foot of the exposed gritstone face. The two major sets are very persistent and these form the rock face. They are concentrated in groups, so that in some areas spacing is very wide (up to 5m) and in others the joints are closely spaced (about 15cm). This has the effect of dividing the rock mass into very large blocks which are separated by areas of more intensive jointing. In areas of intensive jointing these are subject to failure by ravelling. Masonry dentition works have been constructed in several places on the base of East Chevin Quarry to prevent failure of the quarry face and rocks falling onto a footpath located below the quarry. The majority of the joints are open and apertures vary from narrow to moderately wide (6mm to 150mm). Surfaces are usually smooth, although surface waviness of mean wavelength 1m and maximum amplitude 5cm is usually present. Infilling is usually absent, but where observed comprised of a very soft, coarse grained non-cohesive sand (completely weathered grit). All joints were dry at the time of the survey.

Jones (1984) used the discontinuity survey data to provide an indication of the slope stability of the Addingham Edge Grit and the likely failure mechanism. Jones (1984) assumed a mean strike of the rock face of 080°. The orientation of the major discontinuity sets were plotted on a stereonet and the slope stability was analysed using the methodology of Hoek & Bray (1981). Jones (1984) concluded that the rock face is stable against plane failure, but that wedge failure is possible by sliding along the intersection of the major joint sets. However, failure is only possible when the following condition is satisfied:

$$\psi_f > \psi_i > \phi$$

Where:

ψ_i= is the plunge of the discontinuity intersection;

ψ_f = is the apparent dip of the slope face measured in the direction of sliding;

ϕ = is the angle of internal friction of the rock mass.

For these data, ψ_i is 78°, and if a value of ϕ = 45° is assumed for the rock mass, it is evident that failure can only occur where ψ_f is > 78°. This corresponds to a true dip of the face of >87° (i.e., nearly vertical). Such steep faces were not observed in the Addingham Edge Grit at Caley Crags. The most likely failure mechanism in the Addingham Edge Grit is that of toppling. This involves the rotation of columns or blocks of rock about a fixed base. However, the orientation of the bedding discontinuity upon which they sit (defined by the two major joint sets) is not favourable since it dips into the slope (i.e., south), and the face is therefore, stable under present day conditions. Failure by secondary toppling (Hoek & Bray 1981) has been active in the area in the past, when instability of the lower slopes of the Chevin caused undercutting of the gritstone. This mechanism has also operated in the East Chevin Quarry, where extraction of rock has undercut the slopes.

Spivey (1982) carried out a discontinuity survey and stereographic projections of the edges formed in Addingham Edge Grit on Ilkley Moor. He concluded that the orientation and geometry of the discontinuities of the Addingham Edge Grit are not the dominant factor in causing the failure of the rock slopes. The main cause of toppling and wedge failures is the higher rate of weathering of weakly cemented sandstone, which occurs in beds of approximately 1.0m in thickness within the Addingham Edge Grit. When the centre of gravity of a grit block acts eccentrically due to cambering or weathering of the weakly cemented sandstone on the exposed face, failure will occur. The weathering out of the mudstone at the base of the grit also contributes to failure at the mudstone/gritstone interface where this is exposed on the scarp slopes.

Geotechnical Properties of Superficial Deposits

The geotechnical properties of the superficial deposits on the lower slopes of the Chevin have been investigated by West Yorkshire County Council Highways in 1970 for the Otley Bypass. This ground investigation comprised 48 no. exploratory holes (i.e., 26 no. cable percussive boreholes, 16 no. rotary boreholes and 6 no. trial pits). Further site investigations were also undertaken in 1975 and 1976 at the eastern end of the Otley Bypass (Old Orchard) designed by The Institute of Geological Sciences (Culshaw and Duncan 1975; Burgess 1976). A series of ground investigations were undertaken by West Yorkshire Highways and Technical Services (HETS) in 1977, 1979 and 1984 to investigate areas of landsliding affecting the A660. Ground investigations including laboratory testing of Glacial Till, Head Deposits and Mudstone were undertaken for the design of the reinforced earth retaining structure on the Otley Bypass adjacent to Birdcage Walk (Snowden et al. 1986. MSc students at Leeds University have also undertaken laboratory testing on the rock and superficial deposits present on the Chevin (Spivey 1982; Jones 1984; Leach 1984; Xiadakis 1985 and Germaine 2007). This test data set has been used by the author to compile characteristic geotechnical properties for Glacial Till (Table 10), Head Deposits (Table 11), Landslide Deposits Tables 12 and 13) and Mudslide and Mudflow Deposits (Table 14) located on Otley Chevin.

Glacial Till

The geotechnical properties of the Glacial Till on Otley Chevin are presented in Table 10. The unweathered Glacial Till generally comprises, firm becoming stiff dark grey, sandy clay, with frequent rock fragments and many sub-angular and sub-rounded cobbles and boulders of coarse sandstone. The grading envelope of the Glacial Till ranges from coarse gravel to fine sand, with between 15% and 35% of silts and clays. The gravel content consists mainly of sandstone and limestone, with traces of mudstone and shale. The gravel content is sub-angular to rounded. The weathered Glacial Till is brown to orange/brown with grey mottles and is a soft to firm consistency.

Table 10. Geotechnical properties of Glacial Till

Property	Range	Mean
Bulk unit weight (γ_b)	20-22kN/m^3	21kN/m^3
Water content (w)	16-26 %	18%
Liquid Limit (w_l)	23-38%	32%
Plastic Limit (w_p)	13-23%	19%
Plasticity Index (I_p)	8-18%	14%
Undrained shear strength (C_u)	20-105kN/m^2	44kN/m^2
Peak Effective cohesion (c')	6-19kN/m^2	8kN/m^2
Peak Effective angle of internal friction (φ')	20-36°	26°
Residual effective cohesion (c'_r)	0-15kN/m^2	6kN/m^2
Residual effective angle of shearing (φ'_r)	16-28°	22°

The plasticity data indicates the finer fraction of the Glacial Till to be of low plasticity. The water content/plastic limit relationship confirms the consistency to be stiff. The mean peak and residual angles of internal friction shows a small reduction in the angle of shearing resistance which is a characteristic of low-plasticity materials.

Head Deposits

The geotechnical properties of the Head Deposits on Otley Chevin are presented in Table 11. The Head Deposits comprise thinly bedded sandstone fragments in a clayey silt matrix. The Head Deposits comprise of reworked Glacial Till which has been translocated downslope by solifluction processes. Solifluction sheets up to 3.5m in thickness were identified at the location of the Silver Mill Hill retaining wall (Snowdon et al. 1986).

Table 11. Geotechnical properties of Head Deposits

Property	Range	Mean
Bulk unit weight (γ_b)	19-20.5kN/m^2	20kN/m^2
Water content (w)	11-32 %	22%
Liquid Limit (w_l)	24-52%	35%
Plastic Limit (w_p)	15-33%	21%
Plasticity Index (I_p)	5-24%	17.5%
Peak Effective cohesion (c')	6-16kN/m^2	8kN/m^2
Peak Effective angle of internal friction (φ')	21-34°	26°
Residual effective cohesion (c'_r)	0kN/m^2	0kN/m^2
Residual effective angle of shearing (φ'_r)	16-26°	20.5°

Landslide Deposits

The landslide deposits can be split into two types depending on their location and predominant rock type content. Type 1 comprises coarse sandstone cobbles and boulders in a matrix of yellow/grey clayey sandy silt. This type of landslide deposit was mainly present on the upper Chevin slopes and formed part of the large debris aprons on the larger rotational landslides. The geotechnical properties for Type 1 Landslide Deposits are presented in Table 12. Type 2 comprises dark grey clay, with frequent cobbles and boulders of mudstone and shale. These deposits were present on the mid to lower slopes where the underlying mudstone and shale strata have been incorporated within rotational and translational landslides. The geotechnical properties for Type 2 Landslide Deposits are presented in Table 13.

Table 12. Geotechnical properties of Type 1 Landslide Deposits present on the upper Chevin slopes

Property	Range	Mean
Bulk unit weight (γ_b)	21-22kN/m^2	21kN/m^2
Water content (w)	8-35 %	22%
Liquid Limit (w_l)	24-38%	34%
Plastic Limit (w_p)	18-28%	21%
Plasticity Index (I_p)	6-17%	11%
Undrained shear strength (C_u)	20-65kN/m^2	33kN/m^2
Peak Effective cohesion (c')	0-15N/m^2	8kN/m^2
Peak Effective angle of internal friction (φ')	22-38°	30.3°
Residual effective cohesion (c'_r)	0kN/m^2	0kN/m^2
Residual effective angle of shearing (φ'_r)	19-25°	22°

Table 13. Geotechnical properties of Type 2 Landslide Deposits present on the mid to lower Chevin slopes

Property	Range	Mean
Bulk unit weight (γ_b)	20-22kN/m^2	21kN/m^2
Water content (w)	13-35 %	24.3%
Liquid Limit (w_l)	25-53%	41%
Plastic Limit (w_p)	17-27%	23.5%
Plasticity Index (I_p)	8-19.5%	10%
Undrained shear strength (C_u)	5-105kN/m^2	42.3kN/m^2

Mudslide/Mudflow Deposits

The geotechnical properties of the Mudslide/Mudflow Deposits on Otley Chevin are presented in Table 14. The mudflow deposits generally comprised of brown, clayey sand and silt with some coarse gravel rich bands. These deposits were mainly on 12-15° slopes on the mid to lower slopes of Otley Chevin. The mudflows sometimes overlap and have prominent toe lobes of 1-2m in height. The mudflows are a prominent feature at the foot of the debris aprons of larger rotational and translational landslides. At Chevin Grange (SE 193 447) mudflow deposits were noted to be between 3.4 and 4.0m in thickness.

Table 14. Geotechnical properties of Mudflow Deposits

Property	Range	Mean
Bulk unit weight (γ_b)	19-21kN/m^2	19.%kN/m^2
Water content (w)	17-43 %	31%
Liquid Limit (w_l)	34-56%	46%
Plastic Limit (w_p)	20-32%	26.5%
Plasticity Index (I_p)	14-31%	23%
Undrained shear strength (C_u)	11-22.5kN/m^2	17kn/m^2
Peak Effective cohesion (c')	6-20kN/m^2	9kN/m^2
Peak Effective angle of internal friction (φ')	20-26°	22°
Residual effective cohesion (c'_r)	0kN/m^2	0kN/m^2
Residual effective angle of shearing (φ'_r)	17-23°	18°

Jones (1984) investigated the active mudflow area at Pig Farm on the mid slopes of Otley Chevin (SE 201 446). The depth of the active mudflow was investigated at various locations along the mudflow track and varied between 0.72m to 1.4m depth.

A supplementary ground investigation was completed in 1975 for the Otley Chevin Bypass; this investigation included a borehole within an area affected by mudsliding. A cable percussion borehole with follow-on rotary coring was completed to identify the number of phases of mudflow/landslide activity in a typical mudslide/mudflow lobe and to examine the mudstone succession beneath, to the level of the proposed bypass, for slip-planes. The borehole was completed south of the Old Orchard at (SE 211 449). The solid-superficial geological boundary was at 12.2m bgl. The superficial deposits comprised several different types of reworked Glacial Till. The solid geology comprised of a fractured shaley mudstone. The top 2.8m of the borehole comprised a relatively homogeneous layer of weathered Glacial Till. The rounded boulders and cobbles consisted of iron-stained sandstone and exhibited a wide range of weathering states from slightly weathered to completely weathered. The matrix material comprised silty clay, such as might have been derived from the underlying mudstone. The appearance at 1.5m bgl of gravel-size particles and the progressive increase in abundance of these with depth possibly indicates some sort of gravity sorting which is more likely to be a result of mudflow movement than of glacial movement.

The increase in wetness of the material at 2.8m coincides with the change in lithology, where the material below 2.8m is firmer, less sandy and less permeable than the material above it. Between 2.8 and 5m depth the material showed the same progressive increase in gravel sized particles with depth, indicating overall disturbance and sorting with depth. Between 5.15 and 7.35m the superficial material changes; the overall lithology is similar to the material above but the alignment of the mudstone particles in the matrix indicates that the material is intact. At 8.1m depth bgl a disturbed structure is observed within the deposits, but the disorder gradually decreases from 10m to 12.15m bgl where the mudstone is described as intact.

This sequence shows three divisions of superficial material exist over the solid mudstone bedrock. An upper disturbed zone extending to 5m depth bgl and containing a variety of lithologies is followed by a less-disturbed zone extending to 8m bgl and then by a more disturbed zone which merges into bedrock at 12.0m bgl. Culshaw & Duncan (1975) suggested that this sequence demonstrated successive mudflow/landslip activity and that the more disturbed zones at the top and bottom of the sequence represent mudflow activity and the middle zone is related to a mudslide rather than a mudflow, although the two types of mass movement may have occurred contemporaneously. No evidence of slickensides or discontinuities was observed in the U4's and core of solid mudstone to 28.10m depth bgl and the test results did not indicate any remoulding of the mudstone. It was concluded that that the toe of any slip in the upper part of the valley side above, probably surfaced at a higher elevation on the slopes of the Chevin and that the mudstone on which the bypass has been located is founded on mudstone which is in situ.

Geotechnical laboratory testing including the determination of moisture content (mc), plastic limit PL and liquid limit (LL) was undertaken on samples of the superficial materials within each of the three divisions described above. From these, plasticity index and liquidity index were calculated. The former index is the difference between the liquid and plastic limits and the latter is defined as LI = (mc – PL) / (LL – PL). The liquidity index, therefore, can assume any value, positive or negative, depending on the value of moisture content. When the moisture content is equal to the plastic limit, the liquidity index has a value of zero, and when the moisture content equals the liquid limit, the index has a value of unity. Therefore, if the liquidity index of the soil is negative it is in a fairly brittle condition, if it is between zero and unity the soil is plastic, and if greater than unity the material can flow as soon as its microstructure is disturbed. The critical liquidity index dividing disturbed from undisturbed soil lies between -0.03 and +0.15. It is likely that soils exhibiting liquidity indices within this range have experienced some remoulding. The liquidity index values reported were mainly within the critical range. Between 0 and 3.5m the liquidity index is mostly positive improving at 4m. Between 4m and 7m the liquidity index is negative but close to the zero abscissa, improving at 8m and between 8 and 10m the liquidity index is close to zero and improves at 11.5m. Therefore, there is clear evidence of three zones of instability above the solid bedrock in the borehole, each being delimited by a more stable desiccated top and a wetter base. The most plastic zone is the uppermost, whilst the lowest is closest to the 'stable' profile shown for the solid mudstone bedrock. The depths of the zones delineated compare reasonably well with those previously interpreted from the lithological evidence described.

Paraglacial Geomorphological Processes

Landsliding on the Chevin is a result of numerous factors, including both static and dynamic factors. Paraglacial geomorphology is the study of the ways in which glaciated landscapes adjust to non-glacial conditions (Church & Ryder 1972). This definition was redefined by Ballantyne (2002) as, 'non-glacial earth surface process, sediment accumulation, landforms, landsystems and landscapes that are directly conditioned by glaciation and deglaciation.' Paraglacial processes consist of preparatory factors that reduce stability overtime without initiating movement (McColl 2012). Since deglaciation, valley slopes have experienced paraglacial stress-release in response to differential deglacial loading. Glacial over-steepening as a characteristic of glaciated valleys, results in formation of steepened valley sides and deep valleys that increase the shear and self-weight stresses within the rocks forming the valley slopes. These effects may generate tensile conditions within the lower slope strata, promoting rock mass weakening (Augustinus 1995). The response of glacially-steepened rock to deglaciation is conditioned by bedrock lithology and structure, glacial history, degree of glacial modification and in particular by joint density and the orientation and inclination of joints and other planes of weakness relative to that of the valley side slope. These slopes are then susceptible to failure as glacial retreat removes support at the base of the valley (Sellier & Lawson 1998; Jarman 2006 2009; Ballantyne 2006; Sellier 2008; McColl 2012).

Failures of paraglacial origin also result from stress-release (Cossart *et al.* 2008). Glaciers can create new non-tectonic joint systems, distinguishable by various features, such as an increase in joint density towards the surface. Part of the resulting ice-load deformation is stored within the rock mass as strain energy. Decompression of the rock masses during deglaciation results in release of strain energy, resulting in major changes in the orientation of the principal stress field and development of a zone of tension within the slope. Relaxation of tensile stresses during and after glacial ice down-wastage causes rebound or stress-releases within the rock. This results in the propagation of an internal joint (fracture) network, with associated loss of cohesion along joint plains and reduction of internal locking stresses. These types of joints tend to have little or no filling material, and sometimes cross-cut or terminate at tectonic fractures (McColl 2012). In the case of counter-dip slopes such as the Chevin escarpment, the main mechanism associated with weakening of basal slope strata is the development of neo-joints about the former glacier trim-line. These vertical neo-joints become deeper in relation to debutressing and vacuum due to glacier disappearance.

Fig. 7 shows a schematic diagram of the development of rockfalls and rotational landslides as a result of paraglacial debuttressing and over-deepening of the base of a hillslope with a counter dip geological structure as in the case of the Chevin. Fig. 7 shows that landsliding processes are driven by a combination of debuttressing (higher at the base of the hillslope due to former ice-thickness) and over-steepening of the lower part of the hillslope (due to glacial erosion, whatever the geological structure). In the case of the counter dip slope, the failure mechanism involves the development of neo-joints just above the former trim line (Stage 1). These neo-joints are near vertical and can become deeper in relation to debuttressing and vacuum due to glacier down-wasting (Stage 2). This rock fracturing may initially generate rockfalls or may evolve into a rotational landslide (Stage 3).

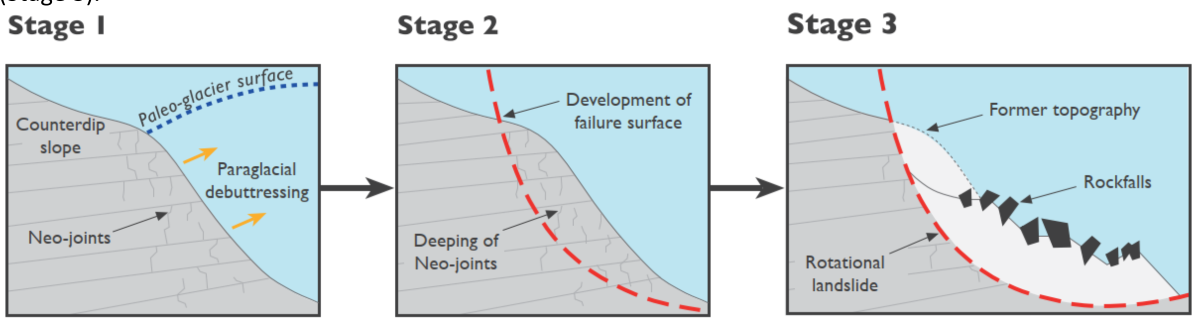


Fig. 7. Development of rockfalls and rotational landslides as a result of paraglacial debutressing and over-deepening at the base of the hillslope with a counter dip geological structure.

McColl (2012) suggested that sheeting joints are likely to be generated by high surface-parallel stresses when a glacial mass is present over steep valley side-slopes. Glacial erosion within steep glaciated valleys will remove lateral confinement providing a locale for stress-release and subsequent formation of sheeting joints. This results in rock shattering generating rockfalls, which may evolve into rotational landslides (Sellier 2008). The resulting stress-release fractures may occur very soon after deglaciation or even during ice-thinning or retreat; however, the timescale for their development following glacial stress-redistribution remains unclear (McColl, 2012).

Non-tectonic joint systems significantly alter the stability and slope-failure pattern of steep sided glacial valleys. Paraglacial jointing can also be coupled with slow movements leading to sackung features causing deep-seated gravitation deformation of the entire hillslope, in relation to the development of normal faults (Cossart et al. 2013). Other paraglacial processes following deglaciation include retrogressive movement of the zone of potential rock-slope failures. This includes jointing (at local scale) and faulting (at slope-scale) this weakens the internal cohesion of the bedrock and encourages water seepage, this in turn, makes the displacement of material easier through large-scale landsliding (Cossart et al. 2013). Various triggering factors can change the slope from a ‘marginally stable’ to ‘actively unstable’ state (McColl 2012). Undercutting from glacial debutressing is an important factor influencing long-term slope failure (Holm et al. 2004). Slope failures on the Chevin were probably triggered after sufficient glacial retreat or ice-thinning had taken place at the base of the valley slope and when the buttressing ice was no longer supporting the weaker mudstones and shales at the base of the Chevin.

Dadson & Church (2005) produced a landscape evolution model of an idealised glaciated valley during the period following retreat of glacier ice. The model included landsliding and fluvial sediment transport processes and a detailed tectonic displacement field with a set of geomorphological rules (including paraglacial processes i.e., rock shattering, bedrock sliding). The paraglacial exhaustion conceptual model indicates that these processes of destabilisation have a temporal window in which to operate. These factors exert a maximum influence on landscape response shortly after deglaciation, followed by a progressive decline in frequency of failures. However, this model does not consider changing climatic influences (Ballantyne 2002). Paraglacial processes have proven to dominate areas in the absence of periglacial processes; however, processes of freeze-thaw and permafrost activity are also important to consider when these are active in the landscape (Kellerer-Pirklbauer et al. 2012). Moragues et al. (2019) recognised three groups of paraglacial processes associated with slope instability processes in Patagonia which have some similarities with the Chevin during postglacial periglacial conditions.

i) Terrain morphometric parameters: Terrain elevations >700m AOD, slope gradients between 25-45°, east-northwest slope aspects with greater insolation, concave curvature of terrain and slight to moderate roughness (0.40-0.65).
ii) Conditioning factors: Glacial deposits, weathered rock outcrops and vegetation cover.
iii) Triggering factors: Groundwater infiltration, proglacial lakes, surface infiltration by rainfall, thaw and runoff, variation of air and soil temperatures and variation in proglacial lake levels.

The dominant cause of instability in periglacial landscapes is freeze-thaw activity. Where locations of weakness, such as joints or bedding planes exist, mechanical weathering further reduces rock-face stability through repetitive cycles of freeze-thaw activity. During rainfall or snow-melt events, water penetrates into joints and induces frost-heaving (vertical force) and frost-thrusting (horizontal force) as the water freezes due to compressional forces from the water expanding by about 10% (Taber 1930; Rempel et al. 2004). Over time, these compressional forces exert stress on the jointed rock face causing incremental movement away from the slope face. Climatic variability influences periglacial processes, providing alternative means, in addition to glacial morphodynamics, to affect landscape response.

The causes and timing of large Pennine landslides are poorly understood. There are still major gaps in the understanding of Holocene geomorphological activity in the Pennines. The nature, causes, and timing of major landslide failures are still incompletely understood.

Paraglacial rock-slope evolution models indicate that rock-slope failure frequency is highest following deglaciation, eventually slowing to a lesser frequency over time (Ballantyne 2002). Research has been undertaken using ^{14}C dating and cosmogenic isotope analysis to assess the frequency of rock-slope failures after deglaciation (Cossart et al. 2008; Wilson 2009; Ballantyne & Stone 2013). It should be noted that this research focussed on rock-slope failures rather than deep-seated landslides. However, the findings of this research provide some interesting conclusions in relation to the timing of failure after deglaciation. Ballantyne et al. (2014b) used Beryllium 10 (^{10}Be) exposure dating of rock slope failures on the Isle of Jura in Scotland. They concluded that rock-slope failures occurred at least 720-2,240 years after deglaciation, with the probability of failures peaking circa 2,000 years after deglaciation, consistent with a millennial-scale delay model. The time-span of the rock-slope failures also coincided with the period of maximum glacio-isostatic crustal uplift, suggesting that failure may have been triggered by uplift-driven seismic events acting on fractured rock masses.

Ballantyne *et al.* (2013) also dated rock-slope failures using cosmogenic ^{10}Be in northwestern Ireland. All rock-slope failures took place within 5,000 years following deglaciation at circa 17.4 ka. Most occurred within 2,000 years after deglaciation. They concluded that paraglacial stress-release and associated fracture propagation were important in reducing rock masses to critical stability, although earthquakes caused by Late Glacial glacio-isostatic rebound and/or release of stored tectonic stresses may have triggered failure in some or all cases. This supports the view that paraglacial rock-slope failure activity in tectonically stable intraplate terrains was concentrated within a few millennia following deglaciation.

Hydrogeology and Mass Movement Processes

The hydrogeological conditions which prevailed during the development of landslides following deglaciation on Otley Chevin provided an ideal setting for the application of the 'Reservoir Principle of mass movement' (Denness 1972). The Reservoir Principle was developed following studies undertaken by the Institute of Geological Sciences of UK landslides which have developed in sedimentary and glacially derived material. The Reservoir Principle of mass movement is a general principle applicable to many areas of landslides in the UK and describes the effect of groundwater draining from an overlying permeable stratum onto an underlying impermeable stratum; the permeable layer provides the reservoir of groundwater which is continuously released. The principle may be summarised as the overall mechanism by which a landslide complex degenerates more rapidly from a 'solid' to a 'liquid' state than would be the case if the groundwater flow were not present. The north-facing scarp slopes of Otley Chevin which have been significantly affected by landsliding are dominated by the geological succession of jointed gritstone and sandstone strata capping the crest of the Chevin (the reservoir), underlain by less permeable mudstone and shale lithology where the landslides develop. The underlying mudstone and shale sequence, although demonstrating some secondary permeability properties, is still less permeable than the overlying gritstones and sandstones. Groundwater percolates through the jointed gritstones and sandstones until it encounters the underlying relatively impermeable mudstones. The mudstone strata intercept the groundwater flow, changing the flow path from a mainly vertical to near horizontal (parallel to the interface). An effective head is maintained over the underlying mudstone and shale strata, keeping them permanently saturated. This leads to the softening of the mudstone and shale strata and reduces the effective stress in the saturated material that produces a weak zone, within which failure surfaces develop over time leading to slope failure. Failure is most likely to occur where pore-water pressure is enhanced by impaired drainage (at impermeable layers in the succession) or in areas of greater water flow through, joints, uncemented bedding plains or along fault zones. The acceptance of surface runoff into the reservoir is rapid but its discharge may continue at a decreasing rate until the reservoir is discharged. In the sequence described the majority of the groundwater discharges at a spring line at the top of the impermeable layer where this intersects the valley slope face.

The result of this process on the Chevin is that the surface zone of the mudstone layer softens due to the intake of groundwater, while at the same time deforming due to processes of stress-relief due to debutressing following deglaciation and loading forces exerted by the overlying heavy gritstone edges. In the event of slope failure, usually in the form of deep-seated rotational failure, the landslide debris which accumulates below the rotated block(s), permits easier access of the groundwater. As pointed out by Denness (1972) the abundance of groundwater within the debris apron which develops after the initial rotational mass movement, therefore, tends to generate a more or less viscous flow of landslide debris. This process, therefore, creates extensive debris aprons, often including discrete debris flows within the landslide complex. The reservoir principle of mass movement therefore, resulted in the development of a complex landslide system on the Chevin which is demonstrated by the morphology of the landslide complex. The morphology of the Chevin landslides corroborates that the landslides degenerated rapidly after initial stages of rotational failure from a solid state to a more saturated viscous flow of debris, thereby creating widespread debris aprons extending several hundreds of metres downslope. The effect of the available water from the reservoir, therefore, facilitated rapid mass movement of landslide debris as multiple or successive failures or mudflows, so steepening parts of the landslide system and thereby, generating further landslides to renew mass movement of already failed material (Denness 1972).

Two pronounced zones of springs and seepages are located on the north-facing slope of the Chevin, one zone is located at the upper part of the escarpment just below the gritstone capping at approximately 165m AOD and a lower zone of springs at approximately 105m AOD which coincides with the toe of many of the deep-seated rotational landslides. An area of intermittent springs, seepages and marshy ground is also present on the lower slopes (typically between 60-90m AOD) of Otley Chevin which have been affected by widespread mudflows (Cooper 1984).

Although the overall dip of strata forming Otley Chevin is 15°-18° to the south, towards Airedale, there is enough percolating groundwater to supply springs and seepage areas on the north-facing escarpment of the Chevin. The movement of water towards the north-facing escarpment slope of the Chevin is assisted by the following factors:

1) The intense jointing of the gritstone and sandstone beds, which are crossed by two sets of long and open joints. This jointing pattern extends into the mudstones and shales below the gritstones/sandstones.

2) The intense faulting of the Chevin area also supports the establishment of continuous zones of percolation along zones of fault-disturbed strata.
3) The existence of a long established proglacial lake following deglaciation within Wharfedale between Burley-in-Wharfedale and Pool will have changed baseflow conditions at the base of the Chevin, resulting in high groundwater levels and the rapid recharge of aquifers in the watershed.
4) The disruption of the gritstone/sandstone bedding by periglacial processes following deglaciation, including freeze/thaw and cambering processes, together with paraglacial stress-release processes within the underlying mudstones and shales, will have resulted in highly fractured rock masses, allowing greater transmissivity of groundwater through strata on the north-facing escarpment slope.
5) Large volumes of meltwater would have been generated at times of thaw, following deglaciation of the Wharfedale glacier.

All the above factors would have resulted in the high secondary permeability characteristics of the strata forming the north-facing escarpment of the Chevin. The hydrogeological conditions of the Chevin therefore, satisfy the conditions required for applying the Reservoir Principle.

The continuous supply of water from a sandstone aquifer overlying a relatively impermeable rock, results in setting up and maintaining high pore-water pressure attributing to rapid failure of the slope. Many of the larger landslides on the Chevin involved deep-seated rotational movements with accompanying cambering of the competent sandstone strata at the crest of the slope. After the initial deep-seated rotational slide had developed, further movement was initiated by the following factors:

1) Deglacial unloading and paraglacial stress-release processes affecting the less competent mudstones and shales on the lower valley slopes.
2) The reduction of the shear strength of materials along the slip planes of deep-seated landslides to a residual value.
3) The disruption of the former drainage channels caused by the large deep-seated rotational failure.
4) The increased softening of disturbed and fractured mudstone through unchannelled groundwater flow.
5) Remoulding of mudstone material during failure, which resulted in increasing the water-holding capacity of the material and reducing their shear strength to residual values.

Favourable conditions for the Reservoir Principle would have occurred in Wharfedale after the last glacial maximum when excess meltwater penetrated the aquifers on the crest and north-facing slopes of the Chevin as a result of melting ice. The water released during thaw, could not penetrate beyond the permafrost barrier on the lower slopes, hence superficial and residual weathered materials would have become saturated promoting solifluction and widespread mudflow mass movement processes. This sequence of events helps explain why the large deep-seated rotational slides and hummocky debris aprons are located on the upper slopes of the Chevin and the mudslides, mudflows and solifluction materials are located below the debris aprons on the mid and lower slopes.

Age of the Chevin Landslides

The age of the large rotational Chevin landslides has not been determined by radiocarbon dating. However, there is evidence that the large deep-seated rotational landslides on the upper and mid slope areas of the Chevin took place sometime between the last glacial maximum (Devensian Glaciation, circa 12ka BP) and Early Post-glacial period (10-6ka BP). The evidence for this includes:

1. The inclusion of Glacial Till deposits within the large deep-seated Chevin landslides (Culshaw & Duncan 1975; Burgess 1976; Cooper 1984, Waters et al. 1996).
2. Dating of other large deep-seated Pennine landslides by pollen and radiocarbon analysis from >10,000 years BP to 6,200 years BP. (Scarth's Fields – 11,135 years BP, Dowell & Hutchinson 2010; Alport Castles - 10,000 Years BP, Johnson & Vaughan, 1983; Lawrence Edge – 7,500 years BP, Johnson & Walthall 1979; Skempton et al. 1989; Buckstones Moss, Muller 1979; Didsbury Intake A - 7,400 years BP, Tallis & Johnson 1980; Cown Edge, Far Cowms - 7,200 year BP, Franks & Johnson 1964; Bradwell Sitch A - 7,100 years BP, Tallis & Johnson 1980; Holme House Wood – 6,480, Dowell & Hutchinson 2010; Millstone Rocks – 6,200 years BP, Johnson & Walthall 1979; Coldside – 5,860 years BP, Redda & Hansom 1989).

3. Many studies of Late Glacial geological sequences demonstrate high incidence of large landslide failures in the Pennines. Waters et al. (1996) identified over 200 landslides in the region which involved Late Glacial sequences.

The trigger mechanisms causing first-time failure have been discussed previously. Following the first-time failures which took place between the Devensian and end of the Early Post-Glacial period, landslide activity has continued to take place on the Chevin during Post-glacial and historical times; this is supported by the geomorphology of the landslides as follows.

i) The geomorphology of the landslides demonstrates distinct stages in landslide development on the Chevin. For example, the Chevin Hall deep-seated rotational landslide (1A) is older than the mudflow (1B) north of it; also, it appears to be older than landslide (2B) east of it, as the landslide accumulation zone of landslide 2B covers that of 1A. Such relationships can be derived in many instances on the Otley Chevin escarpment. However, without the use of radiocarbon dating or pollen analysis it is difficult to infer the time of failure of individual slides.
ii) The mudflow lobes north of Chevin House, suggest three phases of mass movement (Culshaw & Duncan 1975). An upper mudflow zone up to 5m depth. A less disturbed zone possibly produced by sliding rather than flow between 5-8m; and a third zone of more disturbed deposits belonging to another mudflow between 8-12m. These movements appear to have taken place in short time intervals.
iii) The age of deposits present on the valley sides varies over the Chevin escarpment.

Dowell & Hutchinson (2010) recognised landslide events in neighbouring Airedale to be concentrated during the following time periods.

A) During or just before the Loch Lomond Stadial, nominally 11-10 ^{14}C ka BP.
B) In the late Boreal to about the mid-Atlantic, circa 7.5 to 5.5 ^{14}C ka BP.
C) In the late Sub-boreal to the early Sub-Atlantic, circa 3-2 ^{14}C ka BP.
D) Cold late Sub-Atlantic (Pollen zone late VIII, Little Ice Age or Neoglacial.

These periods all correspond to cold wet periods; Dowell & Hutchinson (2010) suggested the following reasons for the higher occurrence of landslides during these periods as follows:

1) Toe erosion; e.g., from melt-water channels and melt-water augmented rivers, and the sudden discharge of pro-glacial lakes;
2) Raising of groundwater pressures resulting from increased rainfall, reduced evaporation, and freezing of groundwater outlets; and
3) Alternations of freeze-thaw, particularly at and near the ground surface, causing ground heave (mainly by the formation of segregated ice, with greater increase in ice content and accompanying loosening of structure), followed by thaw-consolidation, generating high water contents, collapse of structure and large excess pore-water pressures.

It is demonstrated by the Chevin landslide geomorphology that many of the first-time deep-seated rotational failures have subsequently retrogressed and that later adverse climate conditions probably played a part in their further destabilisation and subsequent denudation morphological features. For example, many of the larger landslides have complex solifluction sheets which were formed as a result of freeze-thaw processes at fairly shallow depth during the Early Post-glacial period. The heave on freezing during cold periods will have helped to produce a loose, ice-rich structure, which on thawing would have given rise to high water contents and excess pore-water pressures. These processes would have destabilized the active layer and resulted in the periodic translational downslope movement of the solifluction mantle and in some cases may have resulted in the reactivation of more rapid mass movement features such as mudflows and mudslides. More recent mass movement has taken place on the lower slopes of Otley Chevin particularly on pre-existing slip surfaces in the upper parts of solifluction sheets, mudflows and mudslides. There are many examples of recent instability activity relating to the Otley Chevin landslides; these areas are described below.

Slope Stability Analyses

Germaine (2007) carried out slope stability modelling for the upper and middle slope sections of the Chevin at Beacon Hill (SE 199 443) and East Chevin Quarry (SE 209 444) using Slope/W limit equilibrium software (Geo Studio, Version 6.21, 2007). The geotechnical parameters used for the slope stability analyses are shown in Table 15.

Table 15. Geological Sequence and geotechnical parameters used for the stability analyses (Germaine 2007).

Geological Stratum	Geotechnical Parameters
Medium Coarse Grained Sandstone	
Unsaturated unit weight (γ)	23.23 kN/m^3
Peak angle of friction (φ_{peak})	38°
Cohesion (c)	6650 kPa
Fine Grained Mudstone	
Unsaturated unit weight (γ)	26.75 kN/m^3
Peak angle of friction (φ_{peak})	28°
Cohesion (c)	2120 kPa
Disturbed Landslide Deposits	
Unsaturated unit weight (γ)	19.9 kN/m^3
Peak angle of friction ($\varphi_{residual}$)	18°
Drained Cohesion (c)	21 kPa

For the Beacon Hill section area, the defined critical slip surfaces gave factors of safety ranging from 1.31-9.4 (see Fig. 5). The analysis showed the potential for future landsliding to be within the disturbed landslide deposits. The analysis assumed a piezometric surface close to the existing ground surface. For the East Chevin Quarry section area (see Fig. 6), the defined critical slip surfaces gave factors of safety from 3.78-10.68. The analysis again showed the potential for future landsliding to be within the disturbed landslide deposits. The analysis assumed a piezometric surface close to the existing ground surface. It should be noted that the slope stability models used to analyse the two different section areas of the Chevin by Germaine (2007) were based on simplified geological ground models and used assumed geotechnical parameters as shown in Table 15. The results of the analyses suggested the possibility of a failure taking place along a deep-seated failure surface within bedrock is negligible in both the landslide section areas analysed. However, the analyses showed that potential future failures can take place within the existing disturbed landslide, mudslide and mudflow deposits present in both of the landslide section areas analysed.

Recent Mass Movement Activity

Figures 4a and 4b showing the spatial distribution of landslides on the Chevin also highlight areas of active landsliding with the red asterisk. The main areas of more recent landslide activity are described below.

West Chevin Road

West Chevin Road which traverses the western side of the Chevin towards Guiseley has been affected by active landslides in many areas. Several sections of the road have been affected by landslide movement and these areas have been stabilised by the use of masonry buttressed retaining walls to support the road. West Chevin Road was closed between 1947 and 1951 due to severe damage by landsliding. Walls along the north side of West Chevin Road have collapsed in several areas and have been repaired over many decades. Dry stone and cemented masonry walls along the south side of West Chevin Road display many large cracks and bulging in many areas. A number of large houses in the area to the north of Birdcage Walk and West Chevin road have also been affected by extensive cracking (Chevin Grange SE 193 446, Chevin House SE 207 448, Woodlands SE 202 448 and Chevin Hall SE 189 443). The pavement of West Chevin Road has suffered from many large transverse tension cracks, bulging and differential settlement and has required continuous repairs by reinstatement and patching in many areas. There is an active mudflow south of Bird Cage Walk (SE 200 446) known as the Pig Farm Mudflow. The mudflow extends for 80m to the north from a farm track and is 25m in width. The mudflow is composed of grey/brown, plastic sandy clay, with numerous weathered sandstone and mudstone fragments.

East Chevin Road

East Chevin Road which traverses the eastern side of the Chevin towards Danefield and Yeadon has also been affected by recent landsliding activity. The walls and footpath on the northern side of the road have moved downslope in several areas with the mortared masonry walls showing evidence of cracking and bulging. A large house formerly located east of Beech Grove on the south side of East Chevin Road was so badly damaged by landsliding that it had to be demolished in the late 1960s. There have been continuous drainage problems on the south side of the road; these drains have often not coped with excess rainfall and surface runoff causing rivers of water to run down East Chevin Road spreading debris across the road. There are many small springs which emerge on the northern side of the road and these have reactivated shallow landsliding and mudslides in the fields immediately north of East Chevin Road. Recent excavations for the enlargement of a depot east of the Cattle Market have reactivated movement within a former mudslide complex, this has, in turn, destabilised the walls above the depot on the northern side of East Chevin Road. Recent toppling failures have occurred in the Addingham Edge Grit within East Chevin Quarry (SE 209 443).

Old Pool Bank to East Otley

Between Old Pool Bank and the eastern outskirts of Otley, the A660 Leeds-Otley Road crosses several areas affected by landslides, mudslides and mudflows. The A660 has had a long history of instability problems and numerous small ground investigations followed by remedial works have been carried out to rectify local failures along the road. The British Geological Survey (BGS) carried out a geological assessment of the mass movement features affecting the A660 between Old Pool Bank and the roundabout for Otley Bypass for West Yorkshire Metropolitan County Council Directorate of Traffic and Engineering between December 1983 and January 1984 (Cooper 1984). This assessment included a review of aerial photography to identify landslides, mudslides and mudflows together with hydrological information associated with the A660 alignment. This was followed by site fieldwork to check and map the identified features, to look for evidence of recent mass movement activity and to accurately plot the location of springs and other drainage features.

British Geological Survey recognised the following areas of active mass movement adjacent to the A660:

1) Old Orchard - The Old Orchard house is located east of the junction between Otley Bypass and the A660 (SE 211 449). The house occupies a former clay pit which has been excavated into the reworked Glacial Till and former mudslide deposits. The north-facing cut slope (approx. 1:2 gradient) of the old clay pit has failed behind the house. During 1982-83 a shallow landslide and mudflow took place on the eastern cut slope affecting a boundary fence and trees. The shallow landslide developed when a land drain in the field 70m to the south of the house was breached and water flowing down slope caused a small slump to develop at the crest of the former clay pit cutting slope. Movement was confined to the top 1.5m of Glacial Till.
2) Hawes House – a small back scarp and mudflow has developed to the north of Hawes House. This mudflow poses a potential risk to both the stability of the house and the A660.
3) Water Trough Bend - There are several active areas of landslide and mudflow in the area known as 'Water Trough Bend'. This area has been investigated several times by Leeds City Council including a large investigation following landslide movements affecting the A660 between 1964 and 1966. The landsliding caused damage to the road pavement and displacement of the footpath and walls along the north side of the A660. Movement of walls and footpath continue to occur at Water Trough Bend.
4) Pools Carr – West of Pools Carr is an active complex of mudflow which has displaced the A660 northern retaining wall over a length of 120m by up to 1.0m. This complex area of mass movement comprises rotational landslides and associated mudslides.
5) Caley Crags – South of the A660 below Caley Crags are several rotational landslides and mudflows. The A660 has been constructed over these features. The reactivation of a shallow failure within a mudflow has taken place to the north of the A660 displacing road fencing and causing tilting of woodland. Also, further north, a 5m deep cutting for the construction of new farm buildings at Deals Farm has failed. The cutting was excavated through an area affected by mudflow.

British Geological Survey (BGS) recommended that particular attention should be given to the alteration or disturbance of surface and underground drainage particularly where these coincide with identified former mass movement features. Where water emerges at the ground surface, it should be prevented from infiltrating back into the landslide, mudslide and mudflow areas and causing problems further down slopes.

The most significant area of active and potentially unstable landsliding was identified in the area affected by the Pool Bank Quarry Landslide. BGS identified several areas of active landsliding taking place within the Pool Bank Quarry landslides below Pool Bank Quarry Landfill site and to the north of the A660. These active areas of landsliding were taking place in the former quarry waste materials which had been tipped over pre-existing landslides between 1800 and 1950 when the sandstone quarry was in operation. The Pool Bank Quarry was infilled with domestic refuse between 1966 and 1984. BGS noted that surface water runoff from the restored landfill had caused a small mudflow to develop on the south side of the A660 (Cooper 1984). BGS recommended that further investigations of the slopes to the south and north of the A660 be undertaken for the following reasons:

a) To investigate the effects resulting from the restored landfill (i.e., changes in loading near to an existing landslide by the additional weight of fill and possible changes in the groundwater table and groundwater flows).
b) BGS suggested it was possible that the Pool Bank Quarry which worked the Addingham Edge Grit may have produced a line of weakness, which might, under certain conditions, form a future deep-seated failure plane.
c) BGS postulated that it was possible that the lack of support to the north of the A660 due to active landsliding taking place in the former quarry waste materials, the weight of the landfill in Pool Bank Quarry, and possible increase in the groundwater table and groundwater flow in the landfill area south of the A660, could combine to produce conditions favourable to triggering a sizeable landslide.

BGS therefore, recommended that the Pool Bank Quarry Landfill site should be investigated, that the groundwater regime and groundwater flow in and around the landfill should be monitored, and that a detailed stability analysis be undertaken of the former Pool Bank Quarry and the existing Pool Bank Quarry Landslide area to the north of the A660 (Cooper 1984). Based on the BGS recommendations, Leeds City Council Geotechnical Department carried out further ground investigations at Pool Bank Quarry Landfill and the Pool Bank Quarry Landslides north of the A660. They also undertook slope stability analyses for a series of potential slip surfaces relating to the former Pool Bank Quarry profile, Pool Bank Quarry Landfill and the Pool Bank Quarry Landslides north of the A660. As a result of these ground investigations and stability analyses, major engineering remedial works were designed in the late 1980s and constructed during the mid-1990s. The remedial works included the construction of an anchored retaining wall to support the A660 below the Pool Bank Landfill area. In addition, a series of inclined drainage boreholes were drilled beneath the A660 and inclined upslope. These boreholes were drilled to intercept specific areas of groundwater seepage identified in the ground investigations on the south side of the A660. The inclined boreholes were fitted with perforated pipes with a geotextile filter surround to form a series of inclined drains. The drainage water from the inclined drains was then transferred into carrier drains to be conveyed downslope. The carrier drains were designed to prevent any drainage water to seep into unstable landslide areas identified within the former quarry waste, tipped on the slopes north of the A660.

Discussion

The continued growth of Otley has meant that construction activity is now taking place on areas of potentially unstable ground with inherent geotechnical problems. Consequently, there is a greater requirement for a better understanding of the geotechnical behaviour of these unstable areas for future planning and development. At present the large deep-seated rotational landslides on the Chevin appear to be relatively stable under the prevailing climatic conditions and without significant human interference on natural slopes. However, the shallow landslides, mudslides and mudflows located on the middle and lower slopes are vulnerable to changes in their geometry and groundwater conditions. The majority of shallow landslides that occur today on the Chevin are not first-time failures but result from the reactivation of old failure surfaces. Landslides (whether active or dormant) are common features on the north-facing slopes of the Chevin. There are probably many shallow landslides on the lower parts of the Chevin that have lost their surface morphological evidence of mass movement processes through natural degradation processes or farming practices or have been obscured through urban development. However, shear surfaces relating to these mass movement processes remain below ground. These relic shear surfaces are now increasingly being recognised as important features in ground investigations being carried out for larger development projects, such as the proposed access road for the East of Otley Development.

Figs 4a and 4b show the location of active areas of landsliding on the Chevin escarpment (Red asterisks). These areas of active landsliding are essentially located on the lower slopes of the Chevin and coincide with existing areas of shallow landslides, mudslides and mudflows. Many of the shallow landslides, mudslides and mudflows on the lower slopes of the Chevin are only marginally stable and can be easily reactivated through excavations to create steeper slopes, interference with slope drainage to elevate piezometric levels and by undrained loading by placing fill over former landslides. The reactivation of shallow landslides, mudslides and mudflows, therefore, presents a significant risk to development of the mid and lower slopes of the Chevin. The risk of reactivating former shallow landslides, mudslides and mudflows is also increasing due to the effects of climate change. There is now growing evidence of climate change with longer wetter periods in autumn, winter and spring and longer drier periods in the summer. At the same time there is increasing evidence of man-made interference with drainage on the upper and mid slopes of the Chevin relating to forestry operations, farming practices and also both residential and commercial development on the lower slopes within the suburbs of Otley. The identification of areas which have been affected by mass movement processes on the Chevin is therefore, of importance to the local authority, planners and developers. The construction of roads, buildings and placement of fill on former landslides, mudslides and mudflows can potentially reactivate movement since the failure criteria are significantly less than those on a stable slope.

A number of measures are suggested to help maintain the existing conditions of the Chevin and prevent potential reactivation of slope failures.

i) Completion of a series of geological hazard maps of the Chevin area which delineate, the geology, geological structure, geomorphology, hydrogeology, landslide types and areas of active instability.
ii) Construction of an Engineering Geological landslide hazard map of the Chevin area.
iii) Long-term surveillance landslide hazard maps recording changes in landform morphology through periodic LIDAR mapping, aerial photo and drone surveys.
iv) Installation of groundwater monitoring installations and maintenance of long-term monitoring of groundwater within key locations within the Chevin landslide complex.
v) Detailed Engineering Geological and Geotechnical Guidance to be provided to the Local Planning Authority in relation to any planning applications for development within mapped landslide areas on Otley Chevin.
vi) Guidance for site investigation requirements and geotechnical requirements for foundation design of any developments taking place in mapped landslide areas on Otley Chevin.
vii) Better drainage measures within the areas affected by landslides using the following:
    a) Installing shallow surface drains, approximately 1m deep, at the crest of landslipped areas, which will serve to collect the surface runoff and divert it away into the existing surface drainage channels.
    b) Installing a system of deep land drains within areas affected by landsliding to help reduce piezometric levels and prevent surface runoff and ponding.
    c) Installing horizontal drains at the more susceptible locations, such as along West Chevin Road and Pool Scar, to augment the deep land drains and prevent reactivation of shallow landslides, mudslides and mudflows.
viii) Excavations in areas affected by landsliding may reactivate failures on pre-existing slip surfaces, steep cuttings may fail due to high pore-water pressure and the presence of slope materials with residual shear strength. The geometry of such movements cannot be predicted without detailed ground investigation to provide site-specific data on the geotechnical engineering properties of the ground. The possible kinds of failure to be encountered are small rotational slides in the upper part of the landslide area, translational slides along the interface between slipped material and the mudstone/shale bedrock, and large slides due to reactivation of the old slip surfaces.

    As the groundwater table is near the surface in the mid and lower slopes of the Chevin and slope materials are disturbed and at residual strength, instability of cuttings is highly probable. The stability of slopes excavated to 1:2 in landslipped material and Glacial Till will be problematic without groundwater lowering during excavation, long-term drainage, or the use of retaining walls. Cutting slopes steeper than 1:2 or deep cuttings should be avoided on the mid to lower slopes of the Chevin as demonstrated in the steep cutting slopes for the depot located off East Chevin Road, south of the Cattle Market (SE 206 449).

ix) The mid and lower slopes which have been affected by shallow landslides, mudslides and mudflows are marginally stable. Careful foundation design is required to prevent reactivation of relic slip surfaces in such areas. Massive foundations and heavily loaded structures should be avoided. Construction of several structures simultaneously in landslide prone ground should also be avoided. Lightly loaded structures may be possible in some areas. Piled foundations are recommended, however, the design will require bespoke ground investigation. Careful and detailed ground investigation is required to derive geotechnical parameters for detailed slope stability analysis. The allowable bearing capacity must be determined for all structure foundations as this can vary greatly depending upon the degree of disturbance of the ground.

x) Main roads constructed within areas affected by landsliding should be avoided or designed very carefully and only after detailed ground investigation and a programme of piezometric and ground movement monitoring. The engineering design for the proposed access road off the Otley Bypass roundabout to the proposed East Otley Development is posing significant problems due to the presence of unstable ground conditions north of the A660 Leeds Road/Otley Bypass roundabout.

Similar issues to those described above are also faced by other towns along the Wharfedale valley including, Burley in Wharfedale, Ben Rhydding and Ilkley.

Conclusions

The north-facing escarpment of the Chevin comprises landslides covering an area of 4km^2. There are over twenty-one large deep-seated landslide complexes with associated mudslides and mudflows, some of which extend 0.75-1km at their maximum extent. The large landslides have developed in the mudstone and shale strata which are overlain by massive, jointed gritstones and sandstones of the Millstone Grit Series. The rocks dip gently to the south out of the valley.

Above the 120m contour, large deep-seated rotational landslides are located on the upper and mid slopes, affecting solid bedrock of the Addingham Edge Grit, mudstone and shale strata; these also include associated secondary rotational slides and mudslides on their surfaces. Downslope of these deep-seated landslides, secondary instability has led to the development of mudslides and mudflows which emerge from the debris apron toes of the higher deep-seated landslides.

The morphology of the landslides on the upper and mid slopes of the Chevin comprises prominent back scarps and a series of large, rotated blocks and terraces relating to a series of retrogressive rotational failures. This morphology suggests multiple, successive rotational and translational movement. The deep-seated landslides show evidence of various types of landsliding mechanisms and can, therefore, be classed as complex landslides. Many of the retrogressive and successive landslides have been denuded on the upper and mid slopes by shallow slumping, mudslides and mudflow mass movement processes. Much of the landslide debris accumulations have been reworked by successive mudslides and in wetter areas as discrete mudflows. The mudslides and mudflows extend over most of the lower slope elevations between 105-60m.

The morphology of the mudslides comprises a series of shallow spreads of material leaving discrete toe ridges and lobate features. The mudflows tend to be confined to more linear tracts within poorly drained shallow drainage channels and gullies.

The sandstone and gritstone strata show three dominant joint sets, two vertical and one horizontal. The prevalent mode of failure in the sandstone and gritstone exposed edges is toppling accompanied by some sliding. The main cause of toppling is the cambering of the sandstone and gritstone edges caused by plastic deformation of the underlying mudstone and shale strata.

The most significant effect of the Devensian glaciation was the over-deepening of the Wharfe valley by 40m. The final retreat of the glaciers left over-steepened valley sides in an unstable or metastable state with massive sandstone and gritstone strata overlying weak mudstones and shales.

The role of deglacial unloading and resulting paraglacial stress-release in conditioning or triggering slope failure in the weak mudstone and shale slopes was an important factor affecting the Chevin escarpment

Laboratory UCS and point load testing on sandstone, gritstone, mudstone and shale demonstrate a significant reduction in strength when saturated. The strength of mudstone and shale across the bedding planes is relatively high but parallel to the bedding planes the strength is almost zero. The residual strength of mudstone and shale in the saturated condition is significantly lower than their unsaturated peak strength. The hillslope angles within the landslide/mudslide disturbed areas are near to their angle of repose.

The main factors contributing to instability of the hillslopes are geological structure, paraglacial over-steepening and stress-release processes, lithological variations of the strata augmented by the availability of excess water. Although the overall dip of strata forming Otley Chevin is 15°-18° to the south, towards Airedale, there was enough percolating groundwater to supply springs and seepage areas on the north-facing escarpment of the Chevin. The movement of water towards the north-facing escarpment slope of the Chevin would have been assisted by the following:

- Intense jointing of the sandstone/gritstone beds, which are crossed by two sets of long and open joints. This jointing pattern extends into the mudstone/shale strata;
- Faulting of the Chevin supports the establishment of continuous zones of percolation along zones of fault-disturbed strata;
- The existence of an established proglacial lake following deglaciation between Burley-in-Wharfedale and Pool will have affected base-flow conditions, resulting in high groundwater levels and the rapid recharge of aquifers in the watershed;
- Disruption of the gritstone/sandstone bedding by periglacial processes following deglaciation, including freeze/thaw and cambering processes, together with paraglacial stress-release processes within the underlying mudstones and shales, will have resulted in highly fractured rock masses, allowing greater transmissivity of groundwater through strata on the north-facing escarpment; and
- Large volumes of meltwater would have been generated at times of thaw following deglaciation of the Wharfedale glacier.

All the above factors would have resulted in the high secondary permeability characteristics of the strata forming the north-facing escarpment of the Chevin. The hydrogeological conditions of the Chevin therefore, satisfy the conditions required for applying the Reservoir Principle. The Reservoir Principle of mass movement is applicable to the Chevin landslides, where groundwater was supplied continuously by the overlying beds of jointed sandstone and gritstone, developing and maintaining high pore-water pressure within the underlying mudstones and shales. As a result, the surface zone of the mudstone layer softens due to the intake of groundwater, while at the same time deforming due to processes of stress relief due to debutressing following deglaciation and loading forces exerted by the overlying heavy gritstone edges. In the event of slope failure, usually in the form of deep-seated rotational failure, the landslide debris which accumulates below the rotated block(s), permits easier access of the groundwater. The abundance of groundwater within the debris apron which develops after the initial rotational mass movement, therefore, tends to generate a more or less viscous flow of landslide debris. This process, therefore, creates extensive debris aprons, often including discrete debris flows within the landslide complex. The Reservoir Principle of mass movement therefore, resulted in the development of a complex landslide system on the Chevin which is demonstrated by the morphology of the landslide complex. The morphology of the Chevin landslides corroborates that the landslides degenerated rapidly after initial stages of rotational failure from a solid state to a more saturated viscous flow of debris, thereby creating widespread debris aprons extending several hundreds of metres downslope. The effect of the available water, therefore, is to facilitate rapid mass movement of landslide debris as multiple or successive failures or mudflows, so steepening parts of the landslide system and thereby, generating further landslides to renew mass movement of already failed material.

The continuous supply of water from the sandstone aquifer overlying a relatively impermeable rock resulted in setting up and maintaining high pore-water pressure attributing to rapid failure of the slope. Many of the larger landslides on the Chevin involved deep-seated rotational movements with accompanying cambering of the competent sandstone strata at the crest of the slope. After the initial deep-seated rotational slide had developed, further movement was initiated by the following factors:

- Deglacial unloading and paraglacial stress-release processes affecting the weak mudstones and shales on the lower valley slopes;
- The reduction of the shear strength of materials along the slip planes of deep-seated landslides to a residual value;

- The disruption of the former drainage channels caused by the large deep-seated rotational failure;
- The increased softening of disturbed and fractured mudstone through unchannelled groundwater flow; and
- Remoulding of mudstone material during failure, which resulted in increasing the water-holding capacity of the material and reducing their shear strength to residual values.

Favourable conditions for the Reservoir Principle would have occurred in Wharfedale after the last glacial maximum when excess meltwater penetrated the aquifers on the crest and north-facing slopes of the Chevin as a result of melting ice. The water released during thaw, could not penetrate beyond the permafrost barrier on the lower slopes, hence superficial and residual weathered materials would have become saturated promoting solifluction and widespread mudflow mass movement processes. This sequence of events helps explain why the large deep-seated rotational slides and hummocky debris aprons are located on the upper slopes of the Chevin and the mudslides, mudflows and solifluction materials are located below the debris aprons on the mid and lower slopes.

The causes and timing of large Pennine landslides are poorly understood. There are still major gaps in the understanding of Holocene geomorphological activity in the Pennines. However, dating of Pennine landslides suggests that larger deep-seated landslides did not take place immediately following deglaciation, but took place 1000-3000 years after deglaciation. The long delay reflects the importance of progressive rock-mass weathering initiated by deglacial stress-release processes and associated tensile rock mass damage. The mudslides and mudflows affecting the lower slopes of the Chevin are younger than the deep-seated landslides on the upper and mid slopes and took place throughout the postglacial period.

The present hydrological conditions are different from those that initially caused the landslides. However, there is still enough water entering the landslide disturbed areas. The existing drainage is inadequate and further drainage development is required.

The deep-seated landslide areas on the upper and mid slope areas of the Chevin appear to be in equilibrium and relatively stable under the present hydrological conditions without interference of the natural slopes. Climate change involving long wet autumn, winter and spring months has resulted in raising groundwater levels to near ground surface within areas affected by previous shallow landslides, mudslides and mudflows on the lower slopes of the Chevin. This has caused reactivation of mudflows and mudslides on parts of the lower slopes of the Chevin. Changes to slopes affected by shallow landsliding, mudslides and mudflows such as by undrained loading or over-steepening may cause reactivation of former slip surfaces and the initiation of slope failure.

A greater emphasis is required in relation to landslide risk management and the resilience of the engineering measures adopted as part of the landslide risk management process for future development taking place on the Chevin. These assessments should include the compilation of engineering geological / geotechnical hazard plans for the Chevin area; these plans should be used to support planning decisions for any future development on the Chevin. Regular inspection and special guidance for planning and construction activities are required for any proposed development within mapped landslide areas on the Chevin. Special guidance for ground investigation, foundation design and engineering climate resilience is also required for proposed development taking place within mapped areas of landsliding on the Chevin. Periodic LIDAR surveys should be completed to determine any areas indicating continued movement on the Chevin. Geotechnical instrumentation including surface movement monitoring and groundwater monitoring should be considered within key areas to monitor areas of potential high risk of landsliding which may affect existing infrastructure.

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References

Aitkenhead, N. and Riley, N.J. 1996. Kinderscoutian and Marsdenian successions in the Bradup and Hag Farm boreholes, near Ilkley, West Yorkshire. Proceedings of the Yorkshire Geological Society, 51, 115-125. https://doi.org/10.1144/pygs.51.2.115

Anderson, M.G, Richards, K.G. and Kneale, P.E. 1980. The role of stability analysis in the interpretation of the evolution of threshold slopes. Transactions of the Institute of British Geographers, 5, 100-114. https://doi.org/10.2307/622101

Araujo, F. 2008. Mother Shipton: Secrets, Lies Prophecies. Scotts Valley, CA. ISBN 13: 9788562022005.

Augustinus, P.C. 1995. Glacial valley cross-profile development: the influence of in situ rock stress and rock mass strength, with examples from the Southern Alps, New Zealand. Geomorphology, 14, 87-97. https://doi.org/10.1016/0169-555X(95)00050-X

Ballantyne, C.K. 2002. Paraglacial geomorphology. Quaternary Science Reviews, 21, 1935-2017. https://doi.org/10.1016/S0277-3791(02)00005-7

Ballantyne, C. K. 2006. The Loch Lomond Readvance on north Arran, Scotland: glacier reconstruction and paleoclimatic implications. Journal of Quaternary Science, 22, 343-359. https://doi.org/10.1002/jqs.1059

Ballantyne, C.K. and Stone, J.O. 2013. Timing and periodicity of paraglacial rock-slope failures in the Scottish Highlands. Geomorphology, 186, 150-161. https://doi.org/10.1016/j.geomorph.2012.12.030

Ballantyne, C.K., Wilson, P. Schnabel, C. and Sheng X. 2013. Late glacial rock slope failures in north-west Ireland: age, causes and implications. Journal of Quaternary Science, 28, 789-802. https://doi.org/10.1002/jqs.2675

Ballantyne, C.K., Sanderman, G.F., Stone, J.O. and Wilson, P. 2014a. Rock-slope failure following Late Pleistocene deglaciation on tectonically stable mountainous terrain. Quaternary Science Reviews, 86, 144-157. https://doi.org/10.1016/j.quascirev.2013.12.021

Ballantyne, C.K., Wilson, P., Gheoghiu, D. and Rhodes, A. 2014b. Enhanced rock-slope failure following ice-sheet deglaciation: timing and causes. Earth Surface Processes and Landforms, 39 (7), 900-913. https://doi.org/10.1002/esp.3495

Benn, D.I. and Evans D.J.A. 1998. Glaciers & Glaciation. Arnold, London, 734pp. ISBN 0-340-58431-0.

Bieniawski, Z.T. 1976. Rock Mass Classification in Rock Engineering. In: Bieniawski, Z.T., (Ed), Proceedings of the Symposium on Exploration for Rock Engineering, Johannesburg, Balkema, Cape Town, 1, 97-106. ISBN: 0-471-60172-1.

British Geological Survey, 2000. Bradford, England and Wales Sheet 69, Solid and Drift Geology 1:50,000, Ref. No. DFD069, British Geological Survey, Nottingham.

Broch, E. and Franklin, J. 1972. The Point-Load Strength Test. International Journal of Rock Mechanics and Mining Sciences, 9, 669-697. https://doi.org/10.1016/0148-9062(72)90030-7

Brooks, S.M., Richards, K.S. and Anderson, M.G. 1993. Approaches to the study of hillslope development due to mass movement. Progress in Physical Geography, 17, 32-49. https://doi.org/10.1177/030913339301700103

Burgess, I.C. 1976. Preliminary Report on the Otley Chevin landslips. Institute of Geological Sciences, Yorkshire and East Midlands Unit, Institute of Geological Sciences, WN/EG/84/021, 11.02.76, Nottingham. https://nora.nerc.ac.uk/id/eprint/513966

Carson, M.A. and Petley, D.J. 1970. The existence of threshold hillslopes in the denudation of the landscape. Transactions of the Institute of British Geographers, 49, 71-95. https://doi.org/10.2307/621642

Chandler, R.J. 1970a. The degradation of Lias clay slopes in an area of the East Midlands. The Quarterly Journal of Engineering Geology, 2, 161-181. https://doi.org/10.1144/GSL.QJEG.1970.002.03.01

Chandler, R.J. 1970b. Solifluction on low angled slopes in Northamptonshire. The Quarterly Journal of Engineering Geology, 3, 65-69. https://doi.org/10.1144/GSL.QJEG.1970.003.01.05

Chandler, R.J. 1972. Periglacial mudslides in Vest Spitsbergen and their bearing on the origin of fossil solifluction shears in low angled clay slopes. The Quarterly Journal of Engineering Geology, 5, 223-241. https://doi.org/10.1144/GSL.QJEG.1972.005.03.02

Chandler, R.J., Pachakis, M. and Wrightman, J. 1973. Four long-term failures of embankments founded on areas of landslip. The Quarterly Journal of Engineering Geology, 6, 405-422. https://doi.org/10.1144/GSL.QJEG.1973.006.03.17

Church, M. and Ryder, J.M. 1972. Paraglacial sedimentation: a consideration of fluvial processes conditioned by glaciation. Geological Society of America Bulletin, 83, 3059-3072. https://doi.org/10.1130/0016-7606(1972)83[3059:PSACOF]2.0.CO;2

Colback, P.S.B and Wild, B.L. 1965. The influence of moisture content on the compressive strength of rock. National Mechanical Engineering Research Institute, 19pp.

Cooper, A.H. 1984. The geology, hydrology and stability of the landslips between Otley and Old Pool Bank, West Yorkshire. Technical Report WN/EG/84/21C, January 1984, British Geological Survey, Nottingham. https://nora.nerc.ac.uk/id/eprint/513966

Crofts, R.G. 1995. Sheet explanation for the 1:10,000 scale geological map Bradford SE24NW. Technical Report SE24NW, WA/97/6. British Geological Survey, Nottingham.

Cossart, E., Braucher, R. Fort, M. and Bourlés, D.L. 2008. Slope instability in relation to glacial debutressing in alpine areas (Upper Durance catchment, southeastern France): Evidence from field data and ^{10}Be cosmic ray exposure ages. Geomorphology, 95 (1), 3-26. https://doi.org/10.1016/j.geomorph.2006.12.022

Cossart, E., Mercier, D., Decaulne, A. and Feuillet, T. 2013. An overview of the consequences of paraglacial landsliding on deglaciated mountain slopes: typology, timing and contribution to cascading fluxes. International Journal of the French Quaternary Association, 24 (1), 13-24. https://doi.org/10.4000/quaternaire.6444

Cossart, E., Mercier, D., Coquin, J., Decaulne, A., Feuillet, T., Jóhnsson, H.P. and Saemundsson, D. 2017. Denudation rates during a postglacial sequence in Northern Iceland: example of Laxárdalur valley in Skagafjörour area. Geografiska Annaler: Series A, Physical Geography, ISSN: 0435-3676. https://doi.org/10.1080/04353676.2017.1327320

Cross, M. 1987. An engineering geomorphological investigation of hillslope stability in the Peak District of Derbyshire. PhD Thesis (unpublished), University of Nottingham, 466pp. https://eprints.nottingham.ac.uk/id/eprint/11324

Cross, M. 1998. Landslide susceptibility mapping using the Matrix Assessment Approach: a Derbyshire case study. In: Maund, J.C. & Eddleston, M. (eds) Geohazards in Engineering Geology. Geological Society, London, Engineering Geology Special Publications, 15, 247-261. https://doi.org/10.1144/GSL.ENG.1998.015.01.26

Cross, M. 2010. The use of an open-sided direct shear box for the determination of shear strength of shallow residual and colluvial soils on hillslopes in the south Pennines, Derbyshire. North West Geography, 10 (2), 8-18, ISSN 1476-1580.

Cross, M. (2011). Slope geomorphology and threshold slopes at Callow Bank, Hathersage, Derbyshire. Mercian Geologist, 17 (4), 243-248.

Cross, M. 2019. Sensitivity analysis of shallow planar landslides in residual soils on south Pennine hillslopes, Derbyshire, UK. Bulletin of Engineering Geology and the Environment, 78, 1855-1872. https://doi.org/10.1007/s10064-017-1195-0

Cruden, D.M. and Hu, X.Q. 1993. Exhaustion and steady state models for predicting landslide hazards in the Canadian Rocky Mountains. Geomorphology, 8 (4), 279-285. https://doi.org/10.1016/0169-555X(93)90024-V

Cruden, D.M. and Varnes, D.J. 1996. Landslide types and processes. In: Turner, A.K., Schuster, R.L. (eds) Landslides investigation and mitigation. Transportation Research Board, US National Research Council Special Report 247, Washington D.C., Ch. 3, 36-75. http://onlinepubs.trb.org/Onlinepubs/sr/sr247/sr247.pdf

Culshaw, M.G. and Duncan, S.V. 1975. A preliminary assessment of stability conditions at the eastern end of the proposed bypass at Otley, West Yorkshire. Report No. 75/16, December 1975, Institute of Geological Sciences, British Geological Survey, Nottingham.

Dadson, S. and Church, M. 2005. Postglacial topographic evolution of glaciated valleys: a stochastic evolution model. Earth Surface Processes and Landforms, 30 (11), 1387-1403. https://doi.org/10.1002/esp.1199

Dean, M.T. and Lake, R.D. 1991. Geology of the Northwest Leeds district. Technical Report WA/91/41, British Geological Survey, Nottingham, 1 Jan, 1991. ASIN: B0018PBA1E.

Dean, M.T., Browne, M.A.E., Waters, C.N. and Powell, J.H. 2011. A lithostratigraphical framework for the Carboniferous succession in northern Great Britain (Onshore). British Geological Survey Research Report, RR/10/07, 174pp. http://www.bgs.ac.uk/downloads/browse.cfm?sec=1

Denness, B. 1972. The Reservoir Principle of Mass Movement. Institute of Geological Science Report 72/7, British Geological Survey, Nottingham. ISBN 10: 0118805924.

Donnelly, L.J. 2008. Subsidence and associated ground movements on The Pennines, northern England. Quarterly Journal of Engineering Geology and Hydrogeology, 41 (3), 315-332. https://doi.org/10.1144/1470-9236/07-216

Dowell, R.W.R. and Hutchinson, J.N. 2010. Some landslides in Airedale, Yorkshire, and their incidence in relation to paleoclimate compared with that indicated generally for southern Britain and NW Europe. Quarterly Journal of Engineering Geology and Hydrogeology, 43, 333-344. https://doi.org/10.1144/1470-9236/08-031

Dyke, C.G. and Dobereiner, L. 1991. Evaluating the strength and deformability of sandstones. Quarterly Journal of Engineering Geology, 24 (1), 123-134. https://doi.org/10.1144/GSL.QJEG.1991.024.01.13

Early, K.R. and Skempton, A.W. 1972. Investigation of the landslide of Walton's Wood, Staffordshire. The Quarterly Journal of Engineering Geology, 5, 19-41. https://doi.org/10.1144/GSL.QJEG.1972.005.01.04

Easton, J. 1998. Mother Shipton: Prophecies of Ursula Sontheil. Fenris Press. ASIN: B01HC9KT0C.

Edwards, W. 1937. A Pleistocene strandline in the Vale of York. Proceedings of the Yorkshire Geological Society, 23, 103-118. https://doi.org/10.1144/pygs.23.3.103

Edwards, W., Mitchell, G.H. and Whitehead, T.H. 1950. Geology of the District North and East of Leeds. Memoir of the Geological Survey of Great Britain, Ref. No. DF070, British Geological Survey, Nottingham.

Franks, J.W. and Johnson, R.H. 1964. Pollen analytical dating of a Derbyshire landslip: the Cown Edge landslides, Charlesworth. New Phytologist, 63, 209-216. https://doi.org/10.1111/j.1469-8137.1964.tb07373.x

Friend, R., Buckland, P., Bateman, M.D. and Panagiotakopulo, E. 2016. The 'Lindholme Advance' and the extent of the Glacial Maximum in the Vale of York. Mercian Geologist, 19 (1), 18-25. ISSN 0025-990X.

Germaine, G.T. 2007. Geomorphological and geotechnical investigation of the landslides of the Otley – Ilkley escarpment for purposes of risk assessment and landslide stabilisation. Unpublished MSc Dissertation, University of Leeds.

Gross, P. 1979. Problems of slope stability in central Wharfedale. Unpublished MSc Dissertation, Leeds University.

Hawkins, A. and McConnell, B. 1992. Sensitivity of sandstone strength and deformability to changes in moisture content. Quarterly Journal of Engineering Geology, 25, 115-130. https://doi.org/10.1144/GSL.QJEG.1992.025.02.05

Hemingway, J.E. 1957. The Bramhope Grit and its structures on Otley Chevin. Transactions of the Leeds Geological Association, 7, 1, 43-52.

Hendron, A.J. 1968. Mechanical properties of rocks. In: Rock Mechanics in Engineering Practice Stagg, K.G. and Zienkiewicz, O.C. (eds.), Ch. 2, John Wiley and Sons, NY.

Hoek, E. and Bray, J. 1981. Rock slope engineering, 3rd edition, Institution of Mining and Metallurgy, London. https://doi.org/10.1201/9781482267099

Holm, K., Bovis, M. and Jakob, M. 2004. The landslide response of alpine basins to post-Little Ice Age glacial thinning and retreat in southwestern British Columbia. Geomorphology, 57 (3), 201-216. https://doi.org/10.1016/S0169-555X(03)00103-X

Hutchinson, J.N. (1967). The free degradation of London Clay cliffs. Proceedings of the Geotechnical Conference on Shear Strength of Natural Soils and Rocks 1, 113–118. Oslo, Norwegian Geotechnical Institute.

Hutchinson, J.N. and Bhandari, R.K. (1971). Undrained loading, a fundamental mechanism of mudflows and other mass movements. Géotechnique 21 (4), 353–358. https://doi.org/10.1680/geot.1971.21.4.353

Jarman, D. 2006. Large rock slope failures in the Highlands of Scotland: characterisation, causes and spatial distribution. Engineering Geology, 83, 161-182. https://doi.org/10.1016/j.enggeo.2005.06.030

Jarman, D. 2009. Paraglacial rock slope failure as an agent of glacial trough widening. Geological Society, London, Special Publications, 320, 103-131. https://doi.org/10.1144/SP320.8

Jones, T.N. 1984. The engineering geology and hydrogeology of the slopes above the Otley Bypass. Unpublished Dissertation, University of Leeds.

Johnson, R.H. and Vaughan, R.D. 1989. The Cowms Rocks landslide. Geological Journal, 24, 359-370. https://doi.org/10.1002/gj.3350240408

Johnson, R.H. and Walthall, S. 1979. The Longdendale landslides. Geological Journal, 14 (2), 135-158. https://doi.org/10.1002/gj.3350140211

Kellerer-Pirklbauer, A. Lieb, G.K., Avian, M. and Carrivick, J.L. 2012. Climate change and rock fall events in high mountain areas: numerous and extensive rock falls in 2007 at Mittlerer Burgstall, central Austria. Geografiska Annaler: Series A, Physical Geography, 94 (1), 59-78. https://doi.org/10.1111/j.1468-0459.2011.00449.x

Lambeck, K. and Purcell, A.P. 2001. Sea-level change in the Irish Sea since the Last Glacial Maximum: constraints from isostatic modelling. Journal of Quaternary Science, 16 (5), 497-505. https://doi.org/10.1002/jqs.638

Laurence, A. 2016. Otley Chevin: A landscape history. Pioneer Press Limited, Skipton. ASIN: B007SGZN10.

Leach, A.R. 1984. A guide to the landslipped slopes traversed by the A660(T) between Otley and Pool Bank, West Yorkshire. Unpublished MSc Dissertation, University of Leeds.

McColl, S.T. 2012. Paraglacial rock-slope stability. Geomorphology, 153-154, 1-16. https://doi.org/10.1016/j.geomorph.2012.02.015

Montgomery, D.R. 2001. Slope distributions, threshold hillslopes, and steady-state topography. American Journal of Science, 301 (4-5), 432-454. https://doi.org/10.2475/AJS.301.4-5.432.

Moragues, S, Lenzano, M.G. Moreiras, S. and Lenzano, I. 2019. Paraglacial geomorphology associated with slope instability in the North Branch of the Argentino Lake, Argeinean Patagonia. Geographical Research Letters, 45 (1). https://doi.org/10.18172/cig.3786

Muller, R. 1979. Investigating the age of a Pennine landslip. Mercian Geologist, 7, 211-218.

Raistrick, A. 1926. The Glaciation of Wensleydale, Swaledale and adjoining parts of the Pennines. Proceedings of the Yorkshire Geological Society, 20, 366-410. https://doi.org/10.1144/pygs.20.3.366

Raistrick, A. 1927. Periodicity in the glacial retreat in West Yorkshire. Proceedings of the Yorkshire Geological Society, 21, 24-28. https://doi.org/10.1144/pygs.21.1.24

Raistrick, A. 1931. The Glaciation of Wharfedale, Yorkshire. Proceedings of the Yorkshire Geological Society, 22, 9-30. https://doi.org/10.1144/pygs.22.1.9

Raistrick, A. 1933. IV. The Glacial and post-glacial periods in West Yorkshire. Proceedings of the Geologists' Association, 44, (3), 263-269. https://doi.org/10.1016/S0016-7878(33)80004-6

Raistrick, A. 1934. The correlation of glacial retreat stages across the Pennines. Proceedings of the Yorkshire Geological Society, 22, 199-222. https://doi.org/10.1144/PYGS.22.3.199

Redda, A. and Hansom, J.D. 1989. Mid-Flandrian (Atlantic) landslide activity in the South Pennines. Proceedings of the Yorkshire Geological Society, 47, 207-213. https://doi.org/10.1144/pygs.47.3.207

Rempel, A.W., Wettlaufer, J.S. and Grae Worster, M. 2004. Premelting dynamics in a continuum model of frost heave. Journal of Fluid Mechanics, 498, 227-244. https://doi.org/10.1017/S0022112003006761

Robinson, B. 1967. Landslip stabilisation by horizontally bored drains. West Riding County Council, Highways Department.

Robinson, D.A. and Williams, R.B.G. 2005. Comparative morphology and weathering characteristics of sandstone outcrops in England, UK. Sussex Research Online, Oai:sro.sussex.ac.uk:11528. Corpus ID: 131556569.

Rouse, W.C. 1989. The frequency of landslides in the South Wales Coalfield. Cambria: a Welsh geographical review 15 (2), 167-179. Corpus ID: 128567551.

Rouse, W.C. and Farhan, Y.I. 1976. Threshold slopes in South Wales. The Quarterly Journal of Engineering Geology, 9, 327-338. https://doi.org/10.1144/GSL.QJEG.1976.009.04.05

Sellier, D. 2008. Les glissements translationnels paraglaciares et l'evolution des grands reliefs monoclinaux du Sutherland occidental (Highlands d'Ecosse). Bulletin de l'Association des Geographes Francais, 2, 141-152.

Sellier, D. and Lawson, T.J. 1998. A complex slope failure on Beinn nan Cnaimhseag, Assynt, Sutherland. Scottish Geographical Magazine, 114 (2), 85-93. https://doi.org/10.1080/00369229818737036

Shakesby, R.A. and Matthews, J.A. 1996. Glacial activity and paraglacial landsliding in the Devensian Lateglacial: evidence from Craig Cerrig-gleisiad and Fan Drinarth, Fforest Fawr (Brecon Beacons), South Wales. Geological Journal, 31 (2), 143-157. https://doi.org/10.1002/(SICI)1099-1034(199606)31:2%3C143::AID-GJ704%3E3.0.CO;2-K

Skempton, A.W. and Delory, F.A. 1957. Stability of natural slopes in London Clay. Proceedings of the 4th International Conference on Soil Mechanics and Foundation Engineering, London, 2, 378-381.

Skempton, A.W. Leadbeater, A.D. and Chandler, R.J. 1989. The Mam Tor landslide, North Derbyshire. Philosophical Transactions of the Royal Society of London, A, 283, 493-526. https://doi.org/10.1098/rsta.1989.0088

Snowdon, R.A., Darley, P. and Barratt, D.A. 1986. An anchored earth retaining wall on the Otley Bypass: Construction and early performance. Research Report 62, Transport and Road Research Laboratory, Crowthorne, Berkshire.

Soldati, M., Corsini, A. and Pasuto, A. 2004. Landslides and climate change in the Italian Dolomites since the Late glacial. Catena, 55, 141-161. https://doi.org/10.1016/S0341-8162(03)00113-9

Spivey, A.Q.M. 1982. An initial investigation of the engineering geology of the Wharfe valley, south of Ilkley, West Yorkshire. Unpublished MSc Dissertation, Leeds University.

Stephens, J.V., Mitchell, G.H. and Edwards, W. 1953. Geology of the country between Bradford and Skipton. Memoir of the Geological Survey of Great Britain, Sheet 69 (England and Wales. British Geological Survey, Nottingham.

Stone, P., Millward, D., Young, B. Merritt, J. W., Clarke, S.M., McCormac, M. and Lawrence, D.J.D. 2010. British Regional Geology: Northern England (Fifth edition). British Geological Survey, Keyworth.

Straw, A. Devensian glaciers and proglacial lakes in Lincolnshire and southern Yorkshire. Mercian Geologist, 19 (1), 39-46.

Taber, S. 1930. The mechanics of frost heaving. Journal of Geology, 38, 303-317. https://doi.org/10.1086/623720

Tallis, J.H. and Johnson, R.H. 1980. The dating of landslides in Longdendale, north Derbyshire, using pollen analytical techniques. In: Cullingford, R.A., Davidson D.A. & Lewin, J. (eds) Timescales in Geomorphology. Wiley, Chichester, 189-205. https://doi.org/10.1002/esp.3290060509

Varnes, D.J. 1978. Slope movement types and processes. In: Schuster R.I. Krizek, R.J. (eds) Landslides, analysis and control, Special Report 176: Transportation Research Board. National Academy of Sciences, Washington D.C., 11-33.

Vasarhelyi, B. 2003. Some observations regarding the strength and deformability of sandstones in dry and saturated conditions. Bulletin of Engineering Geology and the Environment, 62, 245-249.

Waltham, T. 2007. The Yorkshire dales: Landscape and Geology. Crowood Press, Marlborough. ISBN 9781 86126 972 0.

Waters, C.N. 1999. Geology of the Bradford district: A brief explanation of the geological map Sheet 69 Bradford. British Geological Survey, Nottingham. https://nora.nerc.ac.uk/id/eprint/509911

Waters, C.N., Northmore, K.N. Prince, G. and Marker, B.R. 1996. A geological background for planning and development in the City of Bradford Metropolitan District, Vol. 2: A technical guide to ground conditions. Technical Report WA/96/1, British Geological Survey, Nottingham.

Weeks, A.G. 1969. The stability of natural slopes in south-east England as affected by periglacial activity. Quarterly Journal of Engineering Geology, 2, 49-61. https://doi.org/10.1144/GSL.QJEG.1969.002.01.04

West Yorkshire Metropolitan Council, Highway Engineering Technical Services (HETS) Laboratory, 1972. Otley Urban District Liverpool-Preston-Leeds (Directly Maintained) Trunk Road A.660 Proposed Otley Bypass and Future Improvements at Leeds Road, Otley. Laboratory Report Lab/1/660/RA, Volume 1. West Yorkshire Metropolitan Council Highway Engineering Technical Services, Osset, West Yorkshire 23.11.1972.

Wilson, P. 2009. Rockfall talus slopes and associated talus-foot features in the glaciated uplands of Great Britain and Ireland: periglacial, paraglacial or composite landforms? In: Knight, J. & Harrison, S. (eds), Periglacial and Paraglacial processes and environments. The Geological Society, London, Special Publication, 320, 133-144. https://doi.org/10.1144/SP320.9

Wilson, P. 2014. Valley bulging. In: Encyclopaedia of Environmental Change, (pp. 1138), SAGE Publications. ISBN: 9781446247112.

Xidakis, G.S. 1985. Landslips in Wharfedale, West Yorkshire and their engineering geological significance. Unpublished MSc Dissertation, University of Leeds.