Cave and Karst Science, Vol.42, No.1, (2015) 42–53
Transactions of the British Cave Research Association

ISSN 1356-191X

The caves of Giggleswick Scar – examples of deglacial speleogenesis?

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Abstract: The prominent Giggleswick Scar at the South Craven Fault extremity of the Carboniferous limestone of the Askrigg Block in North Yorkshire, UK, contains relict phreatic caves whose speleogenesis is enigmatic. This paper examines the local geomorphological evidence and proposes that some, but not necessarily all, karst features along and above the Scar formed after the Last Glacial Maximum. Building on a previous deglacial model for the Yorkshire Dales, it is hypothesized that inception fractures and bedding plane partings were created during isostatic uplift. These were then likely enlarged by dissolution in cold unsaturated meltwater beneath a local flowing deglacial ice-dammed lake that formed initially at an altitude of c.300m, with a catchment area of c. 2km^2. Rising cupolas outside Gully Cave were probably formed at c. 18ka BP by meltwater flowing up into a moulin within the ice, which continued to be cold-based farther south. As the ice-sheet slowly downwasted, the surface of the lake would have widened and lowered past the newly-formed cave entrances. Some of these were probably enlarged by freeze-thaw and lake-ice push and pull processes. Indeed, the heights of some enlarged entrances correspond to proposed stabilizing lake overflow levels. It is also assumed that the local ice-dammed lake coalesced with the main Settle glacial lake, until a jökulhlaup created a ravine above pre-existing glacial scoops in the limestone cliff. Thereafter, the lake split into two parts on each side of Buckhaw Brow, whilst still inundating the lower caves. If this hypothesis applies, it has wider implications for cave speleogenesis and sedimentation in the Yorkshire Dales.

Keywords: Cupola, deglaciation, dissolution, Giggleswick, ice-dammed lake, inception, jökulhlaup, tectonic

Received: 07 July 2014; Accepted: 12 February 2015

Giggleswick Scar is a prominent cliff line to the north of the B6480 road (the old A65) in North Yorkshire, seen when the valley of Ribblesdale has been crossed upon leaving Settle travelling in a westerly direction (Fig.1). The road skirts the cliff while climbing the steep hill of Buckhaw Brow at 250m AOD. After reaching this crest, the road descends and the cliff line becomes more spectacular. Where the road joins the A65 Settle Bypass (not shown on Figure 1), the cliff becomes more subdued, until the scarp is lost as the A65 takes a turn to the west towards Clapham (Fig.2). It is a classic example of a fault-line scarp along the line of the near-vertical South Craven Fault, which is included in the Geological Conservation Review (Huddart, 2002). This fault defines the southern edge of the Askrigg Block, a major structural unit characterized by platform carbonate deposition in early Carboniferous times. To the south lay the Craven Basin, where deeper water conditions prevailed and deposition of clastic sediment predominated. The South Craven Fault is the southern limit of a heavily faulted area. This is characterised by limestone grasslands whose northern limit is the North Craven Fault, a parallel feature 3km to the north. The area between the two faults is commonly referred to as the Craven Fault Zone. Fault throws are towards the south or southwest and equal about 700m at Giggleswick (Waters and Lowe, 2013, p.24). The local dip of the limestone averages c. 6° towards the southwest.

The scarp is formed of Dinantian (Mid Mississippian) carbonates of the Malham Formation, and the lower ground to the south, consisting of Namurian (Late Mississippian to Early Pennsylvanian) sandstones and siltstones of the Millstone Grit Group, is mainly drift covered (Arthurton *et al.,* 1988). The freshness of the topography (Fig.3) suggests that the fault was active during Tertiary times (Nicholson, 1990) and it is possibly still active (Versey, 1948). However, the fault scarp is so prominent because it was periodically eroded by ice throughout the Pleistocene. Evidence to show that ice flowed along the line of the fault scarp is provided by the scarp-parallel orientation of the roches moutonées, which form the rough of Giggleswick Golf Course below the scarp, and other streamlined depositional landforms in the area, including drumlins. These features indicate a late-stage ice flow direction southeastwards, towards the Ribble valley, although to the west of Austwick (which lies some 5km northwest of Buckhaw Brow) the ice flowed westwards (Raistrick, 1930: Fig.2). For convenience and clarity, rock climbers have divided the scarp into two sections. The crags to the northwest of Buckhaw Brow are termed Giggleswick Scar North (Musgrove, 2005; Fig.2); those to the southeast of Buckhaw Brow are termed Giggleswick Scar South (Fig.1). This pragmatic subdivision is retained in this paper, which proposes a preliminary assessment of local speleogenesis.

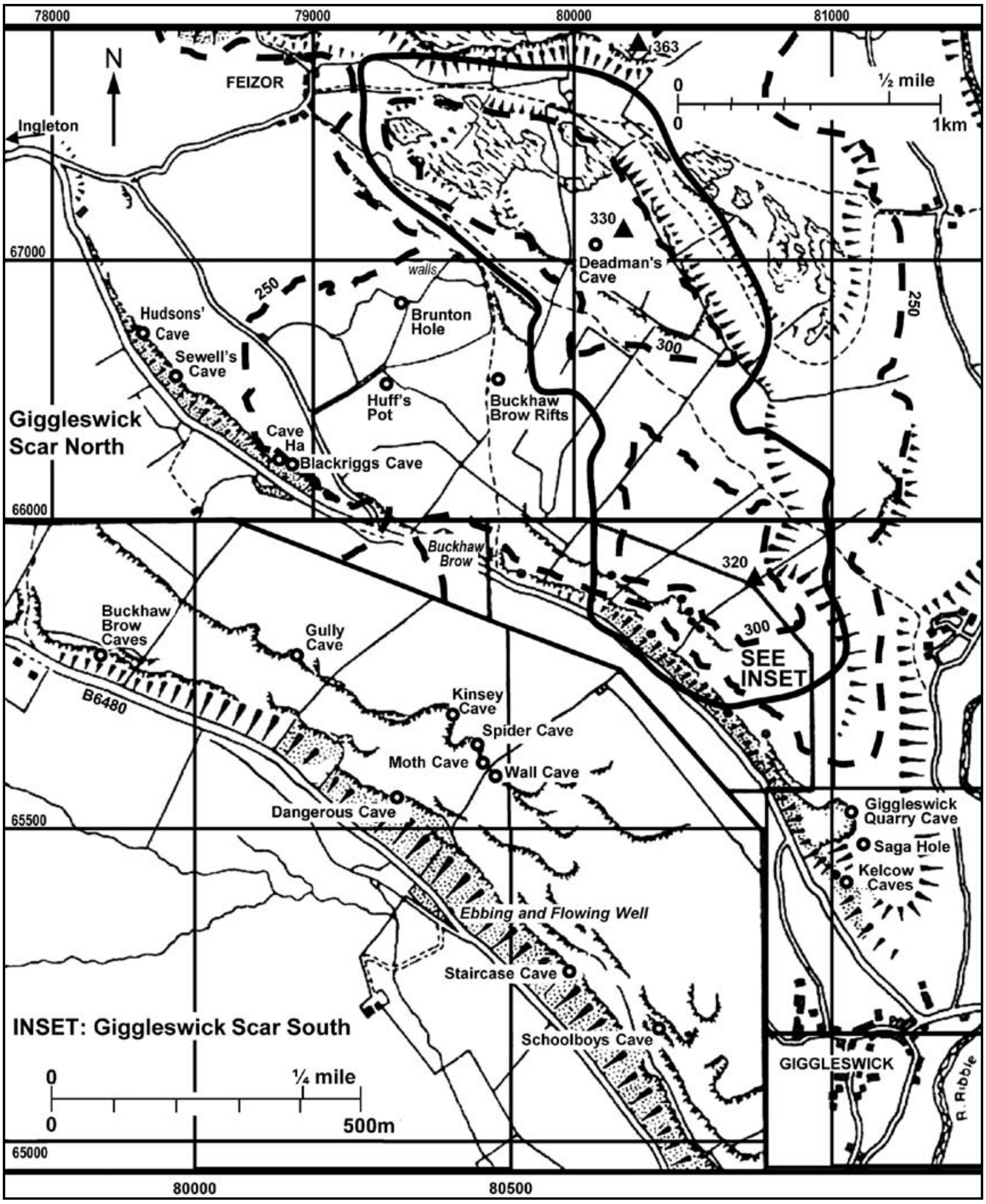


Figure 1: Map of the study area, showing the locations of the caves discussed in the text. 250m and 300m contours around two local summits of 320m and 330m are shown by heavy dashed lines. The heavy continuous line indicates the possible external periphery of a former glacial lake enclosed by ice. This would have surrounded the exposed nunataks, when local deglacial downwasting caused the surface of the ice-sheet to lower to c.300m.

The caves

The Giggleswick caves are commonly relict at present. Lateglacial and Holocene sediments encountered during various archaeological excavations show that hydrological processes have been more active in the past. Holocene relict tufa deposits also suggest that conditions have previously been more favourable for tufa production (Pentecost *et al*., 1990). Sediments in some caves are being disturbed by badger activity. Recent knowledge about the cave archaeology and palaeontology and the karst geomorphology of the Yorkshire Dales is provided by Lord *et al*. (2007), O'Regan *et al.* (2012), O'Connor and Lord (2013), Lord and Howard (2013) and Waltham and Lowe (2013). The caves and rock shelters are discussed below, generally ordered west to east for each sub-area. Grid references are within British National Grid square SD. "Axxx" gives the entrance floor altitude in metres AOD (above Ordnance Datum). Some information has been amended from the cave descriptions given by Brook *et al.* (1991), to accord with their location map, with later cave surveys by JAT, or with GPS readings taken on 13 June 2014. Greater Kelco Cave has been extended by ducking and diving (Hill, 2011) and Saga Hole is newly explored (Sellers, 2012). New Cave has been repositioned correctly and renamed as Christmas Cave: it was originally given incorrect coordinates to protect it (as used in Brook *et al.,* 1991). Human skeletal remains from the local caves have been studied by Leach (2015). The main caves are identified by name on the area map (Fig.1) and by number on the west to east section (Fig.4).

Figure 2: Giggleswick Scar North, looking east towards Buckhaw Brow along the line of the B6480 road (photo: P Murphy).

Figure 3: *Giggleswick Scar South, looking east towards Settle and the Ribble valley. The wooded cliff line of the fault scarp separates the rocky limestone high ground to the north from the rolling grasslands of the Millstone Grit Group to the south. The low ground in the middle distance is Giggleswick Golf Course (photo: P Murphy).*

Caves at Giggleswick Scar North

Situated near the base of a small upper scar, the westernmost feature is **Hudsons' Cave** (1: 78356673 A215), a small rock shelter that is 3m wide and 1m high, going back 2m to where loose clay and rocks meet the roof. **Sewell's Cave** (2: 78476658 A220) lies 200m southeast. This is an 11m-wide and 2m-high undercut in the cliff that goes back 4m. Prior to extensive archaeological excavations during the 1930s it was entirely concealed by limestone collapse material, scree and soil. The undercut resembles a phreatic half tube with shallow scallops (discussed below) 20–30cm long and some eroded flowstone on the rear wall. Two hidden smooth vertical holes in flowstone are >12cm long and 2–3cm wide.

The cave was used as a wolf den during the final part of the Late Glacial Interstadial c. 12.8ka BP (Lord *et al.,* 2012; O'Connor and Lord, 2013: Table 15.5; Lord and Howard, 2013). The two older AMS radiocarbon dated Late Glacial bear phalanges were recorded during the 1930s excavations from the scree outside the cave in a mixed cultural deposit containing Neolithic and Romano-British material. As no other brown bear remains were found during the excavations, it is possible the two phalanges originate from another cave and represent an instance of cultural re-deposition at a much later date. There is another undercut in the cliff 18m towards the northwest, at 78476659 A232, which is 7m long and ≤1m high above a rocky floor. Several rock shelters are given Cave Ha numbers in the archaeological literature.

Figure 5: *Cave Ha 1. The ascending cave passage entrance can be seen in the roof of the rock shelter (photo: P Murphy).*

Cave Ha 1 (3: 78856626 A238) is a large undercut, being 9m high and 24m long (Figs 5 and 6), which is called the Hollywood Bowl in the rock-climbing literature. There are possible paragenetic half-tubes in the roof and high up is a steeply ascending passage that decreases in size upwards. This is referred to by McKenny Hughes (1874), but was first entered in 1972 (Allanach and Champion, 1972) and later revisited in 1995 (Cordingley, 1996). It is described by McKenny Hughes (1874) and Brook *et al.* (1991) as an inlet passage. A prominent horizontal bed of limestone c. 6m above the floor appears to have moved since the undercut was formed, although the effects are partly obscured by tufa (Fig.6). Its lower bedding plane may mark the upper surface of the excavations.

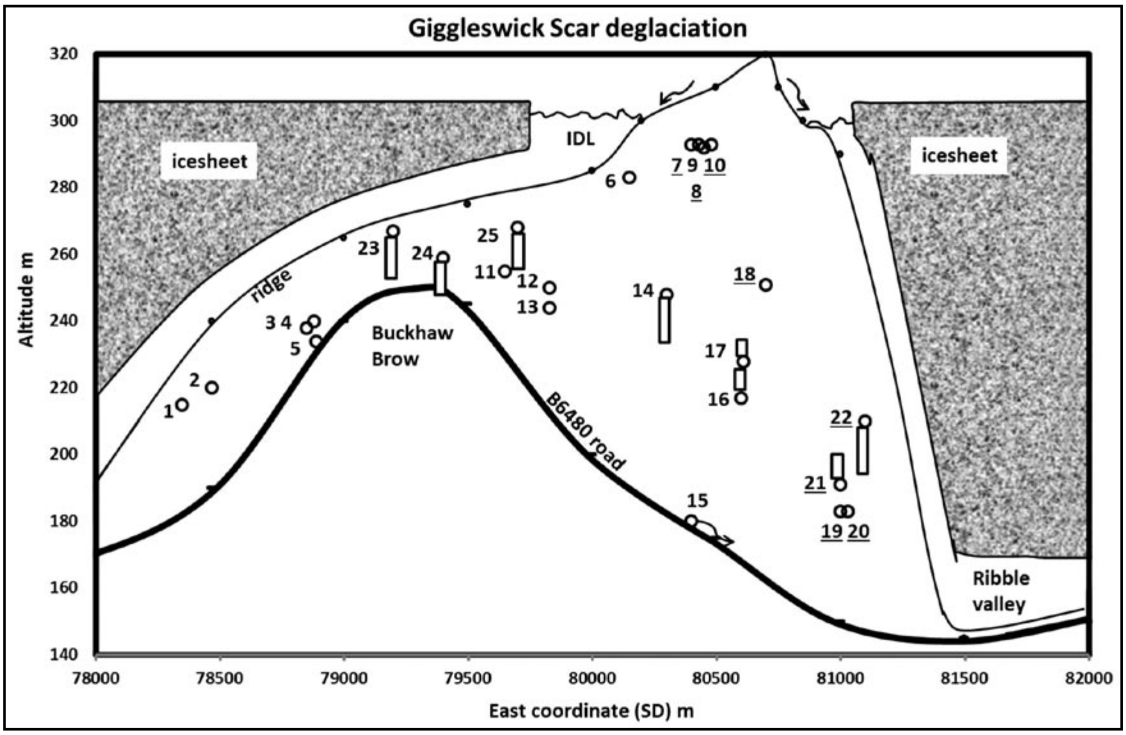


Figure 4: *Diagrammatic west to east section along Giggleswick Scar. This shows the altitudes of the caves discussed in the text and their vertical ranges, where appropriate. Caves with entrances that are significantly larger than their internal passages are indicated by underlined cave numbers.*
The potential deglacial situation is illustrated for when the ice-sheet surface downwasted to c.300m altitude. The local ice-dammed lake(s) (IDLs) could have fed recharge into subglacial waterways that flowed both westwards towards the Irish Sea and eastwards via Airedale towards the North Sea. The bedrock fractures would have been flooded by unsaturated flowing water that could cause cave inception, phreatic enlargement and subsequent sedimentation. As the surfaces both of the ice-sheet and the IDLs lowered past the cave entrances, each could have been enlarged in turn by annual freeze-thaw and lake-ice push-and-pull processes.

.***Figure 6:*** *The prominent horizontal bed of limestone c.6m above the floor of Cave Ha 1. This has sharp edges, indicating a likely neotectonic movement, although this is obscured by tufa on the right-hand side. This might also be the original floor level, before removal of sediments by antiquarians, from the evidence of lower dissolutional flutings (photo: D Checkley).*

Corrosion down this wall behind the previous sediment deposits could explain the observable fluted subcutaneous rock forms (Slabe, 1999), similar to those occurring in previously-sedimented cave passages. Large shallow scallops appear on the inner wall, which also supports a stalagmite boss.

Cave Ha 2 or **Blackriggs Cave** (4: 78886625 A240), is 18m west of, and 6m higher than, Cave Ha 3, above a small undercut and spring. It is 6m long to a minute sump, but might be merely a paraglacial slip rift below a roof of calcited fill. **Cave Ha 3** (5: 78896625 A234) was the biggest, but partial, archaeological excavation, which was beneath an obvious sediment level marked by discolouration and corrosion below a smoother limestone wall (Fig.7). Part of a phreatic arch can be seen, with weathered flowstone on the wall. AMS radiocarbon dated Early Neolithic human and animal bones within tufa deposits are present below more-recent limestone debris and scree (Leach, 2008; Lord and Howard, 2013, 243–244). **Cave Ha 4** (78956621: Fig.8) is a nearby ravine with vertical sides that is littered with jumbled blocks of limestone. It contains a possible opening below rocks in its west wall and a tiny phreatic rock arch, **Cave Ha 5**, at its eastern wall.

Figure 7: *Cave Ha 3, showing a partial phreatic roof and probable original sediment floor on the right-hand side at the junction of smooth limestone and the lower discoloured and corroded wall (photo: T Faulkner).*

Figure 8: *The Cave Ha 4 ravine, with steep sides and a floor of jumbled blocks of limestone (photo: T Faulkner).*

The ravine can be ascended to a dry wind gap through the ridge at c. 250m AOD, to emerge above a shallow valley (parallel to the Scar) that contains the lane to Feizor. **Cave Ha 6** is a narrow un-entered vertical slot on a joint high up in the cliff to the east of the wind gap. There are other lower dry gullies, but with rounded profiles (e.g. Fig.9), which descend through the ridge farther towards the east.

Caves at Giggleswick Scar South

The southern caves fall into two groups: five along the very top of the scarp and several caves that occur part way up the scar. Of those near the top of the scarp, **Gully Cave** (6, also called Antler Cave: 80156581 A283) has an obvious 2.4m-high entrance heading a wide gully above scree. It has parallel walls and parallel ceiling and floor, but closes after 4.5m, perhaps blocked by flowstone. Part of a rising chimney with cupolas appears above its entrance (Fig.10). A mound outside the entrance indicates previous excavation, which found red deer antlers, although O'Connor and Lord (2013) and Lord and Howard (2013) do not report any palaeontological or archaeological findings there. **Kinsey Cave** (7: 80406569 A293) is 36m long, with an excavated entrance 8m wide and 2.4m high, at the head of a larger dry valley with scree on both sides. Prior to archaeological excavations in the 1920s and 1930s, the entrance was blocked by clastic sediment; the cave was unknown until explored by Giggleswick schoolboys in about 1914. The phreatic roof (Fig.11) is undisturbed and displays many shallow scallops c. 20cm long with half tubes and rising tubes inside. Taking the radius of the main passage as 1.7m, the scallops indicate a peak mean flow speed and peak volumetric flow rate of 22cm s^{-1} and 2000 litres s^{-1}, if formed at 0°C (calculated from Faulkner, 2013, Eq.6). A prominent sharp bedding plane edge above the entrance probably indicates a neotectonic movement during deglacial unloading. Inside the cave, Lateglacial animal bones overlie waterlain cobbles, sands and basal clay. The oldest bone, an *Ursus arctos* partial skull, was dated to 14.7±0.4ka BP (O'Connor and Lord, 2013, p.234), demonstrating

Figure 9: *A dry valley with a rounded profile, east of the ravine at Cave Ha 4 (photo: T Faulkner).*

Figure 10: *Gully Cave entrance, showing part of a rising chimney with cupolas that carried an upward flow of water from its roof (photo: T Faulkner).*

the earliest possible bear colonization at the start of the Windermere Interstadial. Cones of pebbles and finer sediments have entered the cave at various points through roof chimneys now blocked at the surface. Pebbles of Lower Palaeozoic rock are present, probably transported from north of the North Craven Fault. It is possible that the Late Glacial faunal remains originally accumulated in a higher and now inaccessible part of the cave. Much sand and silt entered via roof chimneys after the cave had been used in Romano-British times. North of Kinsey Cave is an elongated shallow depression, with a low col at its western end, that leads to a dry valley and relict sink above the end of the cave. **Spider Cave** (8: 80456565 A292) is 30m long, at the head of a smaller dry valley with an obvious entrance below a 1m-long microfracture. Two short (11m- and 14m-long) caves nearby, **Moth Cave** (9: 80436563 A293) and **Wall Cave** (10: 80486560 A293, Figure 12), are at the scarp edge, with roofs c.2m below the ridge top. Wall Cave has two skylights to the surface above and vague shallow scallops 20–30cm-long that perhaps indicate outward flow. These upper caves mainly consist of near-horizontal phreatic passages. Three appear to have entrances that are c. 1m larger than internal continuations.

Figure 11: *The entrance to Kinsey Cave, showing phreatic roof and a sharp bedding plane above that likely indicates neotectonic movement during deglacial unloading (photo: T Faulkner).*

Figure 12: *Wall Cave. This is situated above scree, a few metres below the ridge top (photo: T Faulkner).*

In the lower group is **Superbobs Cave** or **Buckhaw Brow Cave 3** (11: 79656585 A255), whose entrance is a small hading rift in a cliff face. Beyond a squeeze over sticky clay and a small stream sink, a calcited rift ascends a series of small steps with pools until it becomes impassably narrow after 16m. 140m eastwards are **Buckhaw Brow Caves 1 and 2** (12 and 13: 79836580 A244 and A250) with a combined length of c. 45m, but which were at least partly mined. These previously-connected phreatic passages near a small fault have been exposed to daylight by surface erosion. The lowest passage displays shallow scallops. 50m to the east along the cliff is a tiny resurgence and 5m farther is an unexcavated possible phreatic cave entrance with a dry valley below it. **Dangerous Cave** (14: 80306558 A248) has a steep entrance shaft from a small ledge beside a forested cliff. It is 14m deep and 40m long to a 9m-high aven with half-tubes in its roof. Both the dry entrance shaft (Fig.13) and the aven contain massive flowstones that might appear too large to have formed during the Holocene. The behaviour of the **Ebbing and Flowing Well** (15: 804654 A180) beside the road is described by Gunn and Bottrell (2013, p.164).

Figure 13: *The entrance shaft in Dangerous Cave, against a massive relict deposit of flowstone (photo: D Checkley).*

Figure 14: *Cave 1: a karstic remnant ascending chimney with cupolas on the west side of the large Giggleswick Quarry (photo: T Faulkner).*

Staircase Cave (16: 806653 A217) has a 14m-long ascending passage to a small aven. **Christmas Cave** (17) is in the cliff, c. 10m east, and 11m above it, and is 10m long to a single chamber with speleothems. **Schoolboys Cave** (18: 807652 A251) has 40m of passages, with a roomy entrance chamber and two looping oxbows.

The longest Giggleswick cave is **Greater Kelco Cave** (20: 810646 A183). This is 121m long, phreatic, and descends gently along crawls over sediments to the north, from the far southeast end of the scarp. Two medium-sized scallops on the wall of the 1m-wide and 2m-high passage were measured as being c.7 and 8cm long. These indicate a peak mean flow speed and peak volumetric flow rate in the ranges 44–61cm s^{-1} and 880–1220 litres s^{-1}, if formed between 10 – 0°C. Nearby are large speleothems at head height. The entrance floor and that to **Lesser Kelco Cave** (19: 9m long and 37m to the northwest) were lowered by c. 4m and c. 2m, when clay and silt sediments were removed during archaeological excavations. After heavy rain, these caves function as resurgences. AMS dating of bones from both the Kelco Caves has so far identified only one specimen older than the Early Neolithic: a single brown bear canine from Greater Kelco Cave, AMS dated to 12.2±0.2 ka BP (Edwards *et al.* 2014), i.e. within the latter part of the Younger Dryas stadial.

Giggleswick Quarry Cave (21: 810649 A191) was encountered by the now inactive large limestone quarry towards the eastern end of the scarp. It was then the longest local cave at 43m, with a large entrance to a 24m-long, 8m-wide and 9m-high chamber, but it has since been destroyed (Anon, 1974). Caves were also entered and explored in the quarry in 1931 and 1932 (Craven, 2007). **Cave 1** is a karstic remnant at 810650 that appears to be part of an ascending chimney with cupolas (Fig.14). These features lie within the mass of the limestone, c. 200m north of the South Craven Fault. **Saga Shaft** is the remaining part of a quarried cave on the upper tier. It descends 8m to a small rift chamber and a short passage that exits above a vertical quarry face. **Saga Hole** (22: 81106484 A210m, 48m long) has a low, but wide, entrance chamber with sticky deposits reported to be loess, but which might contain quarry dust. The smaller continuing passage leads to descending crawls that reach 16m below the entrance. **Cave 2** is a small opening in the south face of the quarry, at the same altitude, that is almost completely filled with quarry waste. **Cave 3** is the probable location of a cave mentioned by Mitchell (1937), dye tested to a small resurgence below Kelco Caves.

Several of these caves in the lower group have more vertical development than in the upper group, but they are also phreatic fragments. Some entrances appear to be 1–2m larger than internal passages.

Caves north of Giggleswick Scar

Brook *et al.* (1991) describe four karst features in the plateau area up to 600m north of the Scar, which are now partially filled or blocked. **Huff's Pot** (23: 792665 A267) is a narrow rift, 15m deep, where a stream sinks. **Brunton Hole** (24: 794668 A259) is a choked rift, 11m deep. **Buckhaw Brow Rifts** (25: 797665 A268) are two choked shafts, 11m and 4.5m deep. **Dead Man's Cave 1 and 2** (80056704 A325) are 1km north of Giggleswick Scar, above scars on the north side of a footpath in a shallow valley leading to Feizor. They comprise a single west-to-east-aligned relict phreatic through-loop, 41m long, with a vocal connection between the two caves. The lowest point underlies a low hillock 7m above. The floor along the whole conduit consists of various sediments with human and animal bones, without exposed bedrock. The western entrance roof at Dead Man's 1 has been raised by >1m compared with the internal passage and this entrance chamber is about three times wider.

Geomorphological and glaciological observations

There is a distinct contrast between the caves at Giggleswick Scar North and at the other two areas and initially the geomorphological evidence needs to be considered separately.

Geomorphology at Giggleswick Scar North

The North caves are large scoops in the cliff line, with no significant development into the rock mass, other than the roof tube of Cave Ha 1. These 'rock shelters' are perched above the general ground-level to the south of the fault scarp. Their presence may arise partly from differential erosion at the many fluctuations of the ice-sheet thickness early in MIS2, at the start of the last phase of the Devensian glaciation, when ice probably flowed towards the southeast. It is unlikely that these erosional forms survive from a prior glaciation, because inherited scoops would have been erased subsequently by ice flowing along a different trajectory. Nevertheless, their present morphology does not resemble a P-form produced entirely by wet- or till-based mechanical glacial erosion (Benn and Evans, 2010). Neither do the scoops resemble long phreatic conduits in limestone truncated by glacial removal of a southern wall. Such fortuitous removal, just coincident with a scarp-aligned cave passage, would likely leave complete phreatic arches in places, but none are known. However, the scoops do show features of dissolutional phreatic development such as pocketing and scalloping. Scallops are cuspate, oyster-shell-shaped, dissolution depressions in cave walls that are asymmetrical in the direction of flow and are formed at 'scallop dominant discharge' during floods. They have a smooth slope on their downstream side and a steep cusp on the upstream side (Curl, 1974; Murphy, 2001). The presence of the scallops confirms that these scoops were enlarged by dissolution in aggressive, turbulent, water that was not saturated with calcite (Faulkner, 2004). This presumably occurred whilst ice formed the south wall and part of the floor and roof of an englacial conduit at the contact with the scarp edge. Half-tubes in the roof of Cave Ha 1 and shallow indeterminate scallops at its rear seem to indicate palaeo-water flow in an uphill direction towards the north. This would demonstrate artesian flow contrary to, and independent of, the present topography, presumably to resurge on the scar top. This would make the chimney passage an outlet, not an inlet.

Cave Ha 1 was clearly associated with a hydrological regime that likely operated during glaciation and / or deglaciation and was very different from the present dry environment. Subglacial waters are characterized by low temperature and a low partial pressure of CO_2 that arises from a lack of vegetation. Above limestone, they can also exhibit a high saturation ratio of calcium carbonate, from intimate contact with rock flour (Tranter *et al.* 1993). These conditions could apply at present to slow-moving or static water in subglacial lakes in Antarctica and in the thin films of water below warm-based ice-sheets that lead to proglacial flow at glacier snouts. The unknown local thickness of the ice-sheet at the Devensian Last Glacial Maximum (LGM) has been modelled as reaching c. 1000m AOD (Evans *et al.,* 2009). Such ice was probably thick enough to be warm-based from time to time. If so, geothermal heating via the bedrock was then sufficient to overcome heat loss from the surface in the cold climate, so that it could melt the lower layers of ice. Such melting would occur below 0°C, because water freezes at negative °C temperatures at higher pressures, a process known as glaciohydraulic supercooling. However, the topographic high of Buckhaw Brow would have impeded the ice movement to the southeast, forcing the base of the ice to flow upwards. This could cause any basal water to experience a lower pressure, making it freeze again and block off rising outlets at the Brow (e.g. Dow *et al*., 2014), whilst perhaps allowing water to rise slowly via Cave Ha 1, which is at a lower altitude. However, small volumes of subglacial water that were not supplied from the surface would have flowed only at the speed of the ice and become saturated with calcite from the contact with limestone. They would remain non-aggressive and therefore not be capable of enlarging passages, scoops or scallops by dissolution in the time available. Thus, it is unlikely that dissolution along the scoops and the scallop pattern at Cave Ha 1 occurred as part of a sinuous subglacial drainage system at the LGM, when ice moved along the scar and any water flow was directed both by the surface slope of the ice-sheet and the underlying topography.

Geomorphology at Giggleswick Scar South and the area north of Giggleswick Scar

The South caves are relict phreatic passages situated in topographically hanging positions that are also unrelated to present hydrology. They commonly lead northwards, away from the scarp edge, and appear to have acted as palaeo-resurgences along the scarp line. When explaining isolated relict phreatic fragments in topographically 'impossible' situations, the loss of the continuing cave to erosion is commonly invoked. This might not be the explanation at Giggleswick, because the fault scarp is the southern limit of the Askrigg Block carbonate facies, and Pleistocene ice to the south of the fault was underlain by clastic strata. The limestone here was first exposed at the surface at c. 1.3Ma BP (Waltham and Lowe, 2013, Fig. 4.13). At that time, it could have continued only c. 200m farther southwest than the line of the present scarp top. There would then have been insufficient hydraulic head to drive significant cave development, if the overlying topography was fairly flat, although embryonic flows via a fault-zone aquifer might have previously initiated proto-conduits along inception horizons (Lowe and Gunn, 1997).

Gully Cave, Kinsey Cave and Spider Cave are positioned along the highest scarp, at the heads of dry limestone valleys up to 60m long that peter out below c. 280m AOD. Such valleys are commonly interpreted as formed by fluvial erosion of permafrosted bedrock, which seals any open fractures in the epikarst and forces streams to flow on the surface. However, they have small present catchment areas and their existence requires another explanation. Their orientations (Fig.1) and U-shaped profiles (similar to Fig.9) suggest at least partial formation by glacial erosion when ice was locally flowing towards the ***southwest***. This could have occurred during part of the time of the last glaciation in the area from c. 27–19ka BP (Telfer *et al*., 2009), additionally during earlier glacial maxima, or if this local flow was soon diverted eastwards by the main flow. During deglaciation, if and whilst the limestone was still permafrosted, the dry valleys could have acted as a focus for a subglacial channel flow, as interpreted by Huddart (2002). Their widths greatly exceed the diameters of the caves and some do not have a cave at their head. Thus, these valleys were not created as pocket valleys by the active functioning of the caves as resurgences. However, they could have captured cave discharges, to modify the external appearance in a way that is typical of retreat features associated with karst springs. The alignments of some scree deposits in this area might indicate a scarp-parallel flow direction for water or ice, rather than just exhibiting the effects of shattered limestone falling down in periglacial or paraglacial conditions. However, the upper parts of these dry valleys were not smoothed away by the ice flow towards the southeast during the late deglacial phase of the Devensian, as discussed below.

After initiation along suitable palaeo or neotectonically-opened inception horizons, the size and position of these upper group caves also cannot be accounted for by a corresponding interglacial catchment area behind the scarp. This, their isolation from present hydrologically active conduits and the presence of the oldest dated archaeological sediments confirms that these features commonly enlarged prior to both the Holocene and the warm Windermere interstadial from 14.7–12.9ka BP. The common absence of vadose entrenchments and the existence of the sediments also confirm that the caves remained relict throughout most of these periods. As at Giggleswick Scar North, it is unlikely that these phreatic conduits formed during the LGM. Hence, the functioning of the resurgences and their flows into the dry valleys are probably constrained to the time of the main Devensian, and/or an earlier, deglaciation. Above these caves is a series of depressions that might have fed waters into the conduits at that time. Indeed, Spider Cave and Wall Cave are directly connected to the surface through grykes in the limestone pavement, and Kinsey Cave is assumed to contain blocked roof chimneys. Similar timing might also apply to Deadman's Cave and the various rifts north of Giggleswick Scar (Fig.1); these rifts could connect to unknown palaeo-resurgences near Buckhaw Brow.

The possibility that the upper group caves and Deadman's Cave enlarged to their present sizes along long flow paths at some, perhaps pre-Quaternary, time, prior to the exposure of the limestone at c. 1.3Ma BP, needs to be considered. However, in that case, some could be expected to be much larger than their rather consistent metre-scale size. Additionally, their closeness to the present surface would then be coincidental, rather than a natural epikarstic occurrence during deglaciation. Compare also the arguments of Cooper and Mylroie (2014) for a postglacial enlargement of shallow caves in Joralemon Park, New York State. The enlarged sizes of some of the South entrances hint that two different processes or stages were involved in their speleogenesis. Enlarged relict entrances commonly cannot be accounted for by simple periglacial or Holocene frost shattering or by paraglacial stress release. Such collapsed material would remain *in situ* and not be carried or dissolved away by a stream, if none existed. Human removal during quarrying is possible here, but unlikely, because much limestone scree has not been removed.

Deductions from previous studies

The relationships among caves, glaciation and deglaciation have been studied in other northern karst regions and the evidence left by Devensian deglaciation has been analysed for parts of the Yorkshire Dales.

Northern glaciated karsts

The present interglacial use of a karst conduit by subglacial waters has been well documented at Castleguard Cave, below the Columbia Icefield in the Rocky Mountains of Canada (Ford *et al.,* 2000). However, this water resurges below the cave entrance, which itself is below and outside the ice field. Thus, its present hydrology does not represent an LGM environment, when the resurgence would have been blocked by ice, nor a deglacial situation, because its upstream ends are blocked by cold ice. Ice contact speleogenesis has been proposed at a number of sites in Norway, all in metamorphic limestone (Lauritzen, 1982; Lauritzen, 1984). Those caves were commonly interpreted as being pre-existing karst conduits that have been utilized by subglacial waters, as evidenced in some caves by a second phreatic episode in the conduits' history. This might result in the development of paragenetic half tubes, rock pendants and possibly a reversal of flow direction in the system (which could occur during deglaciation). However, these interpretations were based on arguments that preceded clear understandings of dissolutional cave-wall retreat rates before and after chemical breakthrough (Palmer, 1991) and under meltwater conditions (Faulkner, 2006a). Kvithola in northern Norway, a vertical system close to the wall of a glaciated valley, was thought to be a direct result of an uncertain ice marginal speleogenesis. Instead, it appears to be a good example of deglacial speleogenesis, with recharge from an ice-dammed lake (IDL; Lauritzen, 1986: Fig.5). That figure can be generalized by enlarging the size of the IDL so that it completely submerges the underlying cave. The full flow path was probably to an outlet back on to the surface of the ice-sheet, somewhere lower than the top of the cave, so that its speleogenesis complies with the later Faulkner (2007) model. Another hypothesis is that some Norwegian maze caves were formed under warm-based ice-sheets (e.g. Skoglund *et al*., 2010; Skoglund and Lauritzen, 2011). However, it is difficult to see how such a process could operate at glacial maxima, because such subglacial water would have become saturated with calcite. These papers did not consider speleogenesis during deglaciation, and no pre-Quaternary passages are known in Scandinavia.

From extensive studies of caves in the marble stripe karst of the central Scandinavian Caledonides, Faulkner (2007, 2009, 2010) proposed a model of speleogenesis invoking inundation by deglacial IDLs. This accounts for phreatic enlargement by dissolution for the many short relict caves there, which are also not related to the present landscape. When large ice-sheets deglaciate, strong fluctuations in diurnal and seasonal ablation rates can create IDLs around emerging nunataks and ridges that may submerge karst fractures and cave conduits. With down-wasting of the ice sheet, water from these lakes may later flow 'backwards', i.e. uphill, over a lowering succession of cols that provide periodic stability in lake levels between collapses. When the ice sheet lowers below the level of the lowest col, IDLs lower into the heads of valleys. The latest deglacial hydrology of ice sheets is complex, with several possible flow route types, which can vary in space and time. Thus, depending on local environmental conditions and stage, the melting ice plus rainfall on the exposed catchment area can flow via the IDLs 'forwards' along supraglacial channels or cut surface meltwater spillways around valley sides. The water can also flow downhill into englacial conduits within the enclosing and continuing ice-sheet or glacier, or into subglacial reservoirs and waterways along Nye channels (Nye, 1973) in the bedrock beneath, to discharge beyond the glacial snout. Thus, all parts of the landscape can become temporarily submerged in turn by meltwater. Some conduits may be submerged during successive deglaciations. Faulkner (2006b) briefly surmised that a similar deglacial model could have applied in the Yorkshire Dales.

Deglaciation in the Yorkshire Dales

In several papers from 1926–1934, Arthur Raistrick provided evidence from laminated lake-mud up to 10m thick of large and up to 150m-deep deglacial IDLs and lower moraine-dammed lakes of long duration in Wensleydale, Wharfedale, Ribblesdale and along the Aire Valley towards Bradford, as summarized by Faulkner (2012). These were as high as 420m AOD at the Widdale col and included a large lake at Settle at 158m AOD (Raistrick, 1930). Also from the evidence of laminated lake clays, Earp *et al*. (1961, 233–239) described a series of IDLs that lowered from 1120–225 feet (337–68m) AOD along the Lancashire River Calder catchment, as ice retreated southwards down the Ribble valley towards the Irish Sea. More recently, Delaney *et al.* (2010) proposed the previous presence of IDLs in the Manchester embayment, 45km to the south of the study area. Bridgland *et al.* (2011) reported deglacial IDLs in the lower parts of the Swale and Ure valleys, an area that they called the 'Yorkshire Lake District'. One of the meltwater lakes there survived long enough to deposit glacio-lacustrine clay more than 18m thick.

Because early, high-level, IDLs had small catchment areas, they would have been almost static and deposited only fine sediments. Indeed, the widespread, but discontinuous, clay and silt deposits in the Yorkshire Dales, previously described as loess, might partly consist of such lacustrine sediments, as evidence of their prior existence. If their upper water levels matched an annual fall of the ice surface, they would also leave no observable terraces. Thus, direct diagnostic evidence of high-level IDLs cannot always be expected. Instead, their earlier presence can be inferred from the evidence of coarser-grained sediments and overflow channels, as the IDLs migrated to lower levels with higher recharges from larger catchment areas. Such water that flowed rapidly into subglacial waterways could form 'tunnel valleys' that can be observed on land or detected under the sea, after deglaciation is complete. Evidence that this occurred in the Yorkshire Dales is provided by the re-use of a karst conduit at 245m AOD in Illusion Pot in the Kingsdale/Chapel-le-Dale region, 15km to the northwest of the study area (Murphy *et al.,* 2001; Murphy, 2001). Studies of the sedimentary fill revealed underground dune forms that would be classed as eskers if deposited by deglacial melt waters in tunnel valleys directly under the ice. The caves discussed herein lie between 325m and 183m AOD, and the evidence above confirms the presence of deglacial IDLs from 420m to 158m AOD in the Yorkshire Dales. Thus, it is reasonable to assume that the limestone in the Giggleswick Scar area was inundated by water from a lowering IDL during the Devensian deglaciation. Indeed, Raistrick (1930: Fig.2) identified overflow channels at Settle, 3km southeast of Giggleswick Scar, at 1200 feet, 975 feet and 930 feet AOD, i.e. at 360m, 292m and 280m AOD, plus two lower ones. These drained water into Airedale along the eastern side of the Ribblesdale glacier, from an IDL at its head. This occurred whilst the lower ground to the south of the South and Middle Craven faults was still covered by the lowering ice sheet, whose surface probably also dipped gently towards the south.

Deglacial dissolution

The water in IDLs and subglacial waterways would initially be almost completely unsaturated with calcite and therefore aggressive, despite the low temperature and low P_{CO2} conditions. At 0°C and a P_{CO2} of 0.02%, the maximum dissolutional cave wall retreat rate remains at c. 0.35mm a^{-1} (Faulkner, 2006a). This is still one third of the maximum 1mm a^{-1} rate at 10°C and a P_{CO2} of 1% in a typical temperate interglacial climate (Palmer, 1991), and could be supplemented by mechanical erosion. This almost certainly means that the latest phreatic enlargement of the scoops at Giggleswick Scar North occurred when they were still covered by ice and meltwater flowed between the ice and the scarp wall during Devensian deglaciation. Such a flow could have persisted for hundreds of years. A similar mechanism could explain the speleogenesis of the various relict phreatic passages in both the upper and lower groups at Giggleswick Scar South and at Deadman's Cave. The various bed displacements (Figs 5, 6 and 11) provide evidence of cm-scale neotectonic movements along the South Craven Fault. These were probably caused seismically and/or aseismically by isostatic rebound during deglaciation. Together with deglacial water, they provided the conditions for tectonic inception (Faulkner, 2006c). The fractures and bedding plane partings could enlarge to achieve widths and diameters greater than one metre in the time available.

Entrance enlargement

The enlarged sizes of some entrances to caves in both groups at Giggleswick Scar South seem significant in their development. Faulkner (2005a) discussed the likely processes involved in the modification of cave entrances in Norway by marine action. However, such entrance widths and heights tend to taper inwards, whereas those at Giggleswick commonly do not. Additionally, the Giggleswick caves are all at too high an altitude to have been below sea level during the Devensian glaciation. Hence, their enlargement by marine action can be disregarded. The probability that more than 24 entrances with a significantly phreatic morphology in central Scandinavia were enlarged as the levels of ice-dammed lakes lowered past them during deglaciation was proposed by Faulkner (2005b). Those caves have sub-horizontal entrances that have been widened somewhat compared to continuing passages, although it is not a universal phenomenon. Their roofs and floors tend to remain parallel to each other, which distinguishes them from those enlarged by marine action. None of those enlarged entrances is into a mainly-vadose cave, all of which, on other evidence, entrenched only during the Holocene. This supports the concept that entrance enlargements of phreatic caves that were formed before deglaciation was complete are associated with IDLs. Such IDLs all drained away during late deglaciation, before they could influence the younger vadose caves. Although IDLs do not generate the erosional power associated with marine waves and tidal movement, they experience melting and freezing cycles each year, which could cause considerable frost damage, particularly when the surface of the IDL was coincident with a cave opening. In central Scandinavia, Faulkner (2005b) showed that enlargement by IDLs was favoured early during deglaciation at caves on western slopes, assisted by increased temperature cycling, rather than on eastern slopes that remained more in shadow each afternoon. The southwest-facing aspect of Giggleswick Scar would also favour such a process. The enlargement of phreatic cave entrances by the freezing and melting of the surface water in a lowering IDL might be the last act after the flow from the IDL formed the internal passages.

It is assumed that IDLs lower at a similar rate to that of the constraining ice-sheet surface, which in the Faulkner (2005b) study area was calculated to be at a fast mean rate of c. 0.5m a^{-1} during the rapid Holocene warming. Matthews *et al.* (1986) calculated that an IDL that formed in southern Norway during the Little Ice Age froze annually to a depth of 1.4m. This means that a 2.5m-high entrance could be in contact with winter ice during the period that such an IDL lowered by about 4m, i.e. ≥8 years during deglaciation. It would also experience temperature cycling above and below freezing in wet conditions for a period longer than that. Matthews *et al.* (1986) also reported that the rock-cut platforms in metamorphic bedrock beside the IDL were eroded by freeze-thaw mechanisms at the fast rate of 1.4–7.1cm a^{-1}. The frost-shattered debris was removed from the lake walls by lake-ice push and pull. According to Worsley (1975), who studied lake-ice push features at Grasvatn in Norway, the stress at a lake shore caused by the thermal expansion of ice prior to melting is effective up to lake diameters of 3–4km. This suggests that the walls of any cave entrance packed with solid ice would then be highly stressed. If these processes also applied inside 2.5m-high cave entrances, then each existing entrance could widen >1m during deglaciation. In fact, the entrance enlargements listed by Faulkner (2005b) vary from 2–10m, suggesting an even faster mechanism within the narrow confines of a cave entrance, and / or deeper freezing and / or effective action over a period >8years. Such a longer timescale could apply if the cave entrance coincided roughly with an IDL level stabilized for several years by overflow at a col.

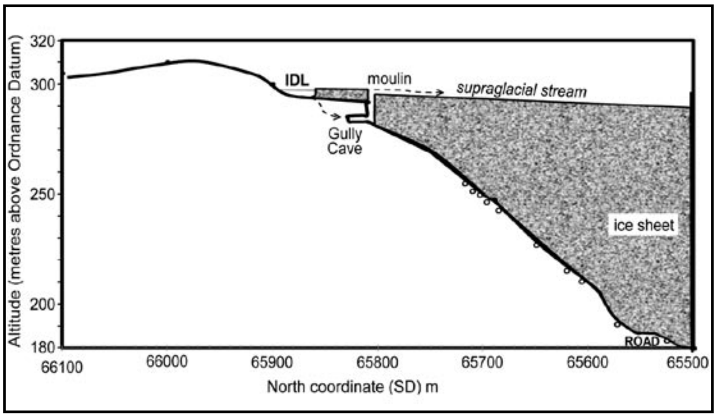
A hypothetical formation model for Giggleswick Scar

Most of the glacial erosion of the scoops at Giggleswick Scar North likely occurred when the ice flowed over Buckhaw Brow towards the southeast, during the Devensian LGM.

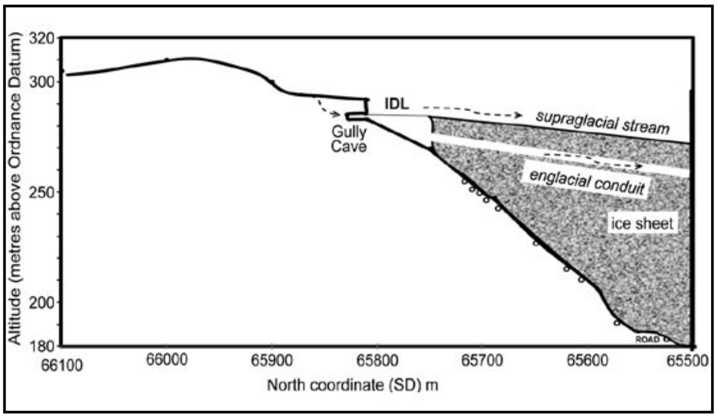
From the previous observations and discussion, it seems probable that the higher group of karst features at Giggleswick Scar South developed after the LGM, rather than either during an earlier deglaciation or much sooner, within a pre-Quaternary buried karst. From the c. 18ka BP dating of the deglaciation of the boulders at Norber Brow, some 5.5km northwest and at a similar mean altitude of 287m AOD (Wilson *et al*., 2012), the lowering of the ice sheet past these caves occurred at a similar time.

The period following deglaciation, until the start of the Windermere Interstadial at 14.7ka BP, is commonly assumed to have been periglacial and windy, with low precipitation (mainly snow). There was a mean

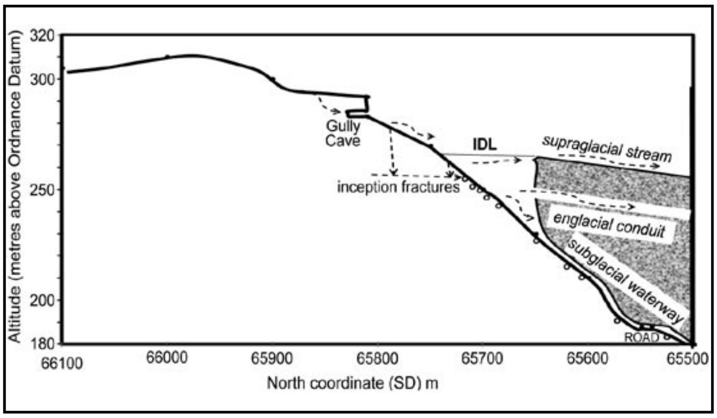
Figure 15a, b and c (below): *Conjectural stages of the deglaciation at Gully Cave at Giggleswick Scar South: North–South sections. Dashed arrows show the direction of meteoric and melt water flow. The projected elevations of lower caves on the Scar are shown by open circles (at their correct altitudes but at north coordinates that are incorrect due to the projection).*



(a) *An ice-dammed lake, with a total catchment area of approximately 2km², was established against the continuing cold-based ice-sheet when the ice margin approached the cliff at Gully Cave. Some water penetrated rock fractures inside the existing cave and flowed outwards and upwards to create a roof-level conduit and the ascending chimney and cupolas by dissolution (see Figure 10), whilst flowing upwards into a moulin within the ice-sheet, and then away down the surface of the ice-sheet. At this stage the ice-sheet must have remained cold-based without a local englacial conduit. This flow regime continued until the ice margin passed the cliff, or the lake level fell below the level of the cave.*



(b) *An englacial conduit might then have captured the water from both the cave and the now more extensive IDL en route to a lower exit on the ice-sheet surface. The cave passage could have enlarged phreatically by dissolution below the initial roof-level conduit, and perhaps later by vadose entrenchment. Other cave entrances on the Scar probably widened by annual freeze-thaw cycles as the lake lowered past them, with limestone blocks removed by ice push-pull and by the effects of short vadose flows.*



(c) *Eventually the IDL lowered below the level of Gully Cave. If the ice-sheet then became warm-based locally, some water could also have flowed into either a Nye channel or a subglacial waterway. If cold-based ice was reached, the outflow would then have followed another englacial conduit. The diagram also gives an indication of how all of the fractures within the adjacent limestone would have been inundated as part of the local deglacial flow regime, which would promote dissolutional enlargement along cave conduits as the ice-sheet lowered. Thus, the caves at lower levels along the Scar could have developed by similar mechanisms.*

January temperature between −20°C and −25°C and a mean July temperature of ≤10°C (Wilson *et al*., 2012). The non-regeneration of the ice sheet after 18ka BP implies that most winter snowfall melted during each subsequent summer. Permafrost expansion continued during early deglaciation, which was later characterized by enhanced river flows farther east (Bridgland *et al*., 2011, pp244–245). Evapotranspiration during deglaciation must have remained low, because of the low temperature and absence of vegetation.

When the ice-sheet surface lowered towards c. 300m AOD, the two local peaks north of the Scar at 330m and 320m AOD were exposed as nunataks. IDLs would necessarily have formed around them in summer as the rock warmed, to create an initial 'Figure of Eight' type moat around them both, with a catchment area of c. 2km² (Fig.1). It is assumed here that this IDL formed separately from the IDL at 360m AOD in upper Ribblesdale. Importantly, the aggressive meltwater would also have inundated the fractures and bedding plane partings in the limestone, created or re-activated by deglacial isostatic rebound, and enlarged them into phreatic cave conduits as it flowed from the northern plateau area out to the scarp edge. Such a sequence is illustrated conjecturally in Figure 15. This explains how the roof-level conduit and the outside ascending cupolas at Gully Cave could have formed by dissolutional flow upwards, into a moulin in the ice that was perhaps associated with a small crevasse above the cliff. For this to occur, the moulin would have been the only initial exit route from Gully Cave. The local ice south of it would then have remained cold-based, without englacial conduits. Hydrogeological calculations for the possible early peak mean flow speed of 1.6cm s^{-1} and peak volumetric flow rate of 12 litres s^{-1} at Gully Cave are given in the Appendix. Ascending cupolas are also known to be formed during hypogenic processes. However, in this case, their location in the epikarstic scarp wall suggests that dissolution before and during ice-contact offers a simpler explanation for their formation.

The dissolutional depth of the cupolas and the half-height of the upper, phreatic, part of the Gully Cave entrance is c. 10cm (Fig.10). At a deglacial chemical wall retreat rate of 0.35mm a^{-1}, this suggests that this early part of the flow regime persisted for c. 300 years, until the ice margin retreated south beyond the cupolas, or until the lake surface dropped to a lower level. The other four upper group caves are situated c. 10m higher, near 293m AOD, at about the height of the highest Gully Cave cupola. This almost coincides with the major Raistrick (1930) IDL level of 292m AOD at Settle. Entrance widening by ice push-and-pull effects at this level appear to have occurred at Kinsey, Spider and Wall caves. This at least suggests that this local IDL had coalesced with the Settle and Ribblesdale IDL by this time, causing it to remain stable for several years. The absence of widening at Gully Cave might be because of its depth below this stable lake surface, before it dropped to the next Raistrick (1930) level of 280m AOD, which is beneath it. To form these upper group entrances up to 2.4m high in cold phreatic conditions could take c. 2200 years, if mechanical erosion brought the mean wall retreat rate up to 0.55mm a^{-1} (as deduced at Pool Sink, which is 20km to the northwest, by Checkley and Faulkner, 2014). The peak flow rate through Kinsey Cave of 2000 litres s^{-1} was 20 times greater than the mean annual flow rate from a local 2km² catchment area (Appendix). This compares well with the initial peak Holocene rate at Pool Sink, which was 14 times greater than the present mean flow rate, but that was after the LGM deglaciation was complete (derived from Checkley and Faulkner, 2014).

The morphology of the lower part of the Gully Cave entrance (Fig.10) is unclear. It possibly indicates continuing phreatic flow into an englacial conduit (Fig.15b) and/or a subglacial waterway, followed by vadose entrenchment by a partly meteoric stream above permafrost that continued to flow at a similar rate into the same channels (Fig.15c), or down-valley after the ice melted. If this height of c. 2m above a sedimented floor was accounted for by vadose chemical and mechanical entrenchment at rate of 0.55mm a^{-1}, that would take c. 3600 years. This suggests that significant flows continued after 18ka BP, whilst presenting a puzzle, because of the known low precipitation until 14.7ka BP. However, it is also possible that these flows could have come from the degradation of the permafrost in unconsolidated sediments, by the creation of so-called 'thermokarstic' thaw lakes (Summerfield, 1991). These can form above any lithological surface, as reported farther east by Bridgland *et al.* (2011). This process perhaps also operated during the latter part of the Younger Dryas stadial, with possible enhanced flows during the Little Ice Age as well. Such flows could explain the entrenchment at Gully Cave and the apparent sterility of its deposits (apart from the antlers), and could also explain some of the sporadic high peak flow evidence at Pool Sink, as reported by Checkley and Faulkner (2014: Fig.8).

The above reasoning suggests that local down-wasting and the retreat of the ice sheet margin were slow processes before and after 18ka BP, in persistent permafrost conditions. Thus, nunataks and glacial lakes were probably prevalent throughout the final, deglaciating, part of the LGM in the Yorkshire Dales, which is consistent with the Raistrick reports of large and long-lasting IDLs. The formation of the local IDL might also explain the survival of the dry valleys below the upper caves. Their petering out below 280m AOD could then have been caused by erosion by the active glacier that continued to flow to the southeast, beside a long arm of the Settle IDL that was up to 60m wide. Assuming that the dissolutional creation of the cupolas outside Gully Cave lasted for 300 years during the stable IDL level at 292m AOD, until it suddenly dropped to 280m AOD, then the mean ice lowering rate was c. 4cm a^{-1} at that time. If this rate is representative for this period, then, compared with an ice height at 287m at c. 18ka BP, the Raistrick (1930) overflow channel at 360m AOD occurred roughly at 19.8, that at 292m at 18.1 and that at 280m at 17.8ka BP.

Deadman's Cave is situated only a few metres below a subdued peak at 330m AOD, and is significantly higher than the other caves discussed here. During deglaciation as previously described, it could only have been covered by a small annular IDL around this peak for a few years, before the IDL would have lowered below it. Thus, the water from that IDL would have had insufficient time to enlarge its passage to its maximum internal size of 1m x 1m. Two alternatives can be considered. Firstly, this and all the other caves discussed here could have been inundated earlier, beneath a larger and deeper IDL that initially surrounded Smearsett Scar at 363m AOD, 1km farther north (Fig.1). However, from the discussion about the formation of the ascending chimney at Gully Cave, and perhaps at Cave Ha 1, an initial 'wide IDL' melting process is unlikely. Otherwise, the whole of Gully Cave would have been submerged by the IDL and upward moulin flow via its entrance would be impossible The more likely second alternative is that Deadman's Cave is much older, perhaps being formed beneath an earlier Devensian stadial deglacial IDL, or during the deglaciation from MIS6–5e, at the start of the Eemian interglacial, when the local peak was higher. The passage could then have been slightly enlarged and sedimented, and its main entrance further enlarged, during the final Devensian deglaciation. The elevations of ten of the lower caves at Giggleswick Scar South are also shown in Figure 15. These could have also enlarged by dissolution when meltwater flowed through them as the ice-sheet and the inundating IDL lowered, within this hypothetical model of Devensian deglaciation. Five of these caves also appear to have widened entrances, at altitudes of 251, 210, 191 and 183m AOD. Some of these might correspond to the unspecified heights of the two lowest overflow channels shown by Raistrick (1930: Fig.2). However, the presence of substantial, but as yet undated, flowstones in several of these caves at least hints that some survive from an earlier Devensian interstadial or from the Eemian Interglacial. In that case, they could have been partly shortened or de-roofed during the Devensian glaciation and then enlarged internally and at their entrances during the final deglaciation.

By the time the local ice-sheet had down-wasted to an altitude of 250m, the narrow ring of meltwater would have had a perimeter roughly 17km long (Fig.1). However, it seems likely that this annular IDL had already coalesced with the long IDL that extended 13km up Ribblesdale from Settle, as far north as Selside. This combined IDL would have had a total surface area of c. $10km^2$. The underlying fall of the ground surface south of Giggleswick Scar would have caused a sympathetic local southern downward slope in the surface of the ice-sheet, allowing the ice to collapse at the wind gap above the Cave Ha 4 ravine. The consequential jökulhlaup flood would have caused meltwater to flow down and create the ravine and partly flow westwards under the ice beside the scoops.

Globally, many deglacial IDLs experienced catastrophic jökulhlaups, which perhaps lasted for several hours whilst lake levels suddenly fell several metres. Such superfloods were studied by Rudoy (2002) in the mountains of southern Siberia, and various papers about these processes in northern America were reviewed by Fisher *et al*. (2002). The characteristic relief forms of water-forced hollows, niches, "drillpots", outburst gorges with box-like profiles (as seems to be the case at Cave Ha 4), dry canyons, and dry, possibly stepped, waterfalls are formed by ***evorsion***, the destruction of rock caused by the rotation of water that falls sub-vertically. Rudoy (2002) suggested that maximum jökulhlaup discharge, $Q_{max} = 0.0075V^{0.667}m^3s^{-1}$, where Vm^3 is the IDL volume, and gave examples of velocities up to $32.5ms^{-1}$. Walder and Costa (1996) gave mean relationships of $Q_{max} = 0.0046V^{0.66}m^3s^{-1}$ for tunnel-drainages from 26 lakes and $Q_{max} = 0.11V^{0.44}m^3s^{-1}$ for non-tunnel collapses.

If the Settle IDL level fell 5m whilst creating the Cave Ha 4 ravine, the discharged volume was c. $5x10^7m^3$. Using the Rudoy (2002) formula, the maximum flow rate was c. $1000m^3s^{-1}$. The flow velocity would create dominant mechanical erosion (Faulkner, 2013), preventing the formation of scallops during the jökulhlaup itself. However, the IDL probably stabilised at the new level, as perhaps supported by the enlarged entrance to Schoolboys Cave at 251m AOD. Such scallops could then form during continuing recharge along the Cave Ha flow route between the ice and the scarp wall. Initially, this flow could also have created the ascending chimney at Cave Ha 1 (which perhaps reached the surface at 247m AOD), in a manner somewhat similar to the creation of the ascending chimney outside Gully Cave. The col at Buckhaw Brow is also at an altitude of 250m AOD, making it another candidate as an overflow channel. However, it lies 300m south of Cave Ha 4, which the ice margin would have reached first. Nevertheless, after its summit became exposed above the ice, it would necessarily have spit the large IDL into separate western and eastern lakes. It seems likely that the Cave Ha 4 ravine, which is unique along Giggleswick Scar, provided the only overflow route for the Settle IDL that was on the western side. Those reported by Raistrick (1930) were all on the eastern side.

The main Settle IDL, with its arm along the lower slopes of Giggleswick Scar South, would then have continued to flow eastwards, as discussed by Raistrick (1930). The flow would have passed through all the lower caves (Figs 1, 4 and 15), as discussed above. Josberger *et al*. (2006) reported that the summer temperature structure of the ice-dammed, 90m-deep, Berg Lake in Alaska consists of three layers. The upper layer extends from the surface at up to 14°C to c. 8m depth at 4°C. The middle layer spans the depth range 8–50m, with variable temperatures from 4°C to nearly 0°C. The lower layer extends below 50m depth, where the temperature remains constant at nearly 0°C. Apparently, sediment discharge stabilises the situation whereby the densest water at 4°C overlies the colder, less-dense water. It is tempting to speculate that Berg Lake, whose recent and earlier surface altitude has varied from 135–195m and up to 275m, below an ice cliff up to 60m higher, may provide an analogue for a deglacial IDL on the southern side of Giggleswick Scar.

Conclusions

Two contrasting groups of caves exist along the line of the Giggleswick Scar fault scarp. Those north of Buckhaw Brow are mainly isolated phreatic pockets within scoops along the Scar that were probably eroded during the Last Glacial Maximum. The additional phreatic dissolution can be explained as occurring where the route of a subglacial channel parallel to the fault scarp impinged on the carbonate bedrock on the north side of the South Craven Fault, during an episode of the Devensian deglaciation. The caves south of Buckhaw Brow are mainly relict phreatic conduits that probably enlarged from inception horizons when submerged by a lowering ice-dammed lake that recharged subglacial and/or englacial melt waters towards the Settle glacial lake, with which it later coalesced. This is supported by the enlargement of several entrances that could have occurred as the lake level lowered past them, especially if stabilized at that level by an overflow channel or col. However, incomplete archaeological excavation hinders this interpretation. It would not have been possible for older and longer passages to have been truncated significantly by late Pleistocene glaciations, because the limestone did not then continue much farther south of the existing cliff line. The different characteristics of the features north and south of Buckhaw Brow may arise from less tectonic fracturing on the northwestern side and perhaps stronger and more-continuous deglacial hydrological flow regimes towards the east rather than the west, despite the local dip being towards the west. The details of the evolution of ice flow westwards and eastwards between Austwick and Buckhaw Brow, as part of the whole British–Irish Ice Sheet that waxed and waned in thickness and extent during the period 27–19ka BP, are unknown. However, it is likely that late deglacial ice and meltwater flowed southwestwards on the western side of Buckhaw Brow, whilst they flowed southeastwards on the eastern side. The timing of the deglacial speleogenesis has been partially clarified by the fortuitous finding of the ascending cupolas outside Gully Cave. This study suggests that at least some of the caves at Giggleswick Scar South did not enlarge until the last few thousand years of the Devensian glaciation, which reached its final local slow demise after 18ka BP in continuing permafrost conditions. Additionally, Deadman's Cave and some of the larger, lower group, caves might survive from a deglaciation prior to an earlier Devensian interstadial or from the Eemian Interglacial. The deglacial process might not apply to the rising chimney seen in Giggleswick Quarry. Being well-within the mass of limestone, a pre-Quaternary hypogenic origin could be surmised for its formation.

This postulated speleogenetic scenario is in concert with the more general model of deglaciation and cave evolution in the Yorkshire Dales that was suggested by Faulkner (2006b). However, additional study is required to test the hypothesis of local deglacial speleogenesis.

Sedimentological studies of superficial deposits presently described as 'loess' are needed to help establish whether they might be at least partly lacustrine in origin rather than directly aeolian. The spatial distribution of these deposits should also be examined, to determine whether they coincide with places where there might have been stabilized ice-dammed lakes.

Additionally, sedimentological studies of underground clastic deposits are needed to help establish whether their depositional environments can be correlated with the hypothesized stages of deglacial hydrology that probably varied from the static to those with high flow rates.

As more of the Giggleswick Scar cave systems and their sediment deposits are studied in detail, it might then become possible to generate a common and more rigorous model of speleogenesis and sedimentation for this part of the region, which potentially contains a large and possibly unique archive of information about the history of the British Quaternary.

Acknowledgements

We thank David Checkley for his company during the fieldwork associated with this study and, in particular, for helping to measure scallops during visits to many of the discussed caves on 13 June 2014. David Lowe, Tony Waltham and another, anonymous, referee provided complementary, comprehensive and thoughtful sets of geological and geomorphological review comments, based upon their knowledge of the Giggleswick Scar area and the Dales region in general. These comments enabled the hypothesis to be strengthened by comparison with other possible modes of speleogenesis and by more rigorous use of earlier literature.

Appendix:

A conjectural flow regime at Gully Cave during deglaciation.

It is assumed that the rising cupolas above Gully Cave can be treated rather like scallops, and that the peak mean water velocity (V) analysis of Curl (1974) applies. From Figure 10, the 'scallop' length $\lambda \sim 100$cm. An effective flow diameter ~ 100cm (radius r = 50cm) also seems applicable, for a tubular conduit between the limestone and the ice sheet. The applicable formula given by Curl (1974) was re-arranged by Checkley and Faulkner (2014) as:

$$V = (2200\mu/\rho\lambda)[2.5\mathrm{x}2.3\log_{10}(r/\lambda) + 5.65]\ \mathrm{cm\ s^{-1}}$$

where μ is the dynamic viscosity of water at 0°C = 0.01793 gramme $\mathrm{cm^{-1}s^{-1}}$ (Ford and Williams, 2007, p.112) and ρ is the density of water at 0°C = 1 gramme $\mathrm{cm^{-3}}$.

Substituting gives:

$$V = 1.6\mathrm{cm\ s^{-1}}$$

and thus the peak volumetric flow rate:

$$Q = \pi r^2 V/1000\ \mathrm{litres\ s^{-1}} = 12\ \mathrm{litres\ s^{-1}}.$$

An effective precipitation rate of 1m a^{-1} for a 1km^2 catchment gives a mean discharge rate of 32 litres s^{-1}. The present annual precipitation rate given by the Meteorological Office for Malham Tarn (some 10km to the east of Giggleswick Scar) is 1550mm a^{-1}. The precipitation rate during deglaciation is not known, but is assumed to be less than the current rate, as discussed above. However, taking the sum of the effective precipitation rate and the ice melting rate as 1.5m a^{-1}, the mean discharge rate from a 2km^2 catchment area is 96 litres s^{-1}.

Thus, for the rising chimney at Gully Cave to have functioned as a dissolutional conduit to form the cupolas, and to melt the ice above as a moulin, it needed to take only c. 12% of the mean flow from the ice-dammed lake, even during scallop dominant discharge. This result seems reasonable, and indeed, when the IDL surface lowered to a level close to that of the cliff above Gully Cave, both the cave conduit and the ice-marginal flow route might have taken much more of the flow.

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