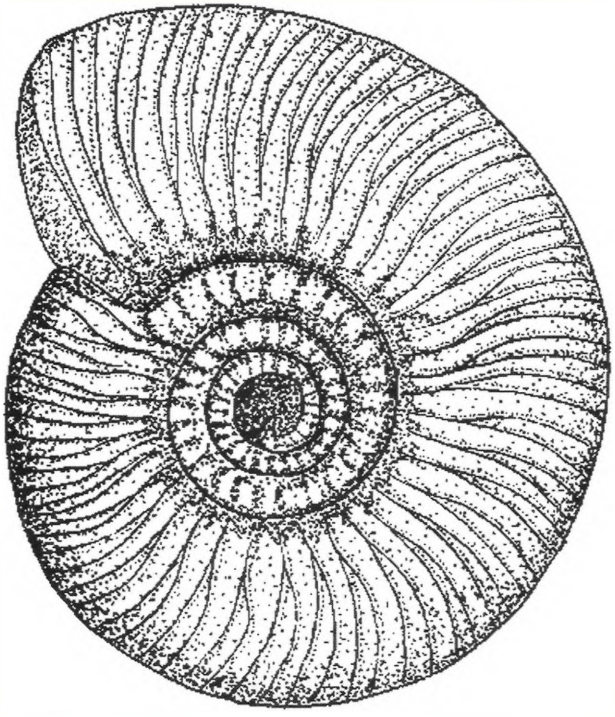
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EDITORIAL

This Edition of Proceedings is the third part of Volume 8. It covers the events of the Society from mid 2014 through to spring 2016 – a period when the Society has remained very active with our usual spread of winter lectures, summer field excursions, annual dinners and a number of special events.

I am particularly pleased to continue with our long standing tradition of publishing original articles on recent research in Cumbria. Additionally this time we have reports on our new Field Research Projects. A strength of the Society has always been a committed group of members willing to go out into the field and look at local rocks and landscapes. These projects aim to harness this combined strength of active people to produce data and information on our local geology. The two projects described here are hopefully just the start of what could become an annual feature of our programmes.

Cumbria has had a difficult winter in 2015/16 with severe flooding and adverse weather impacting on the Society. In spite of having to re-arrange some of our winter lectures and cancel our Member's Evening meeting in December, we have come through. The flooding has, however, highlighted the dynamic nature of our landscape. We are already compiling data on landscape changes (mainly erosional) resulting from the high winter rainfalls – hopefully future Proceedings will be a vehicle to put some of these changes on permanent record.

Can I thank all our contributors: the authors for their papers and the many members and friends who have written up the field excursions and provided other material.

Alan Smith (Editor)

[redacted]

TOWARDS AN INTEGRATED MODEL FOR HEMATITE MINERALIZATION AT THE MARGINS OF THE EAST IRISH SEA BASIN

Stephen F. Crowley

"A theorist goes astray in two ways: 1. The devil leads him by the nose with a false hypothesis. (For this he deserves pity.) 2. His arguments are erroneous and sloppy. (For this he deserves a beating.)"

Albert Einstein, 3rd February, 1915 (letter to Hendrick Lorentz).

Introduction

Economic iron mineralization in the form of hematite ore deposits occurs at a number of sites located on the margins of the post-Carboniferous East Irish Sea extensional sedimentary basin (Figure 1). Ore fields centred on Kirk Maughold (NE Isle of Man), Auchenleck (SW Scotland) and the Vale of Clywd (NE Wales) consist of a few relatively modest ore bodies with a total ore production from each ore field in the region of 10^4 to 10^7 tonnes. In contrast, the hematite ore deposits of the Cumbrian ore field (NW England), with a total production of high grade ore (45-55 weight % iron) in excess of 2 x10^8 tonnes, represented an iron resource of national importance during the mid-19th to mid-20th centuries (Smith 1924; Rose and Dunham 1977; Dodd 2010).

The origin of the Cumbrian ore deposits has long been the subject of discussion and controversy with a published record stretching back over 180 years (see reviews by Rose and Dunham 1977; Dunham 1984; Shepherd and Goldring 1993). In a description of the geology and hematite mineralization of south Cumbria, Rose and Dunham (1977) presented a speculative genetic model – here abbreviated to the R&D model – that attempted to account for the source, transport and precipitation of iron in the south Cumbria ore district and, by implication, the wider Cumbrian hematite ore field. Details of the model are described below, but it is sufficient to note at this stage that the model, as presented, failed to provide any substantive theoretical or empirical support for the apparent complex set of geological, geochemical and hydrogeological processes required to account for observed mineralization. This lack of scientific rigour places the model in the realm of conjecture and, while key components of the model possess an air of credibility, the model itself remains largely untested and subject to challenge (e.g. see Evans 1984; Schnellmann 1984; Shepherd and Goldring 1993; Rowe *et al* 1998; Chapman *et al* 2006a, b).

The objective of this paper is to examine the R&D genetic model by subjecting components of the model to critical scrutiny and, where possible, exposing them to quantitative tests using simple geological principles. The outcome of this assessment leads to some perhaps surprising conclusions regarding the origin of these ore deposits. Much of the following discussion is focussed on Cumbrian mineralization because of the magnitude of the deposits, availability of published information, and the fact that the R&D model was developed on the basis of experience gained from observation of the south Cumbrian ore field. However, it is an implicit assumption that the model should apply to the entire Cumbrian ore field, as originally proposed by Rose and Dunham (1977), and potentially be capable of accounting for all hematite mineralization on the margins of the East Irish Sea basin.

It is important to note that no new information is presented here. Fortunately, this does not prevent us from re-evaluating available observational data and examining critically existing hypotheses in the light of modern geological thinking, modifying current paradigms based on new insights into the ore-forming process, and indicating the possible direction of future research.

What do we know?

Although the history and field geology of individual mines within the Cumbrian ore field is well documented (Smith 1924; Rose and Dunham 1977; Young 1985; Dodd 2010), surprisingly few modern geological and geochemical data are available to test and refine potential ore deposit models. Even when published information is reported the data are presented either in the form of a review or conference abstract with little detailed description or critical interpretation. The following summary is confined to those aspects of the mineralization that impact directly on the development and examination of ore deposit models. Readers are referred to the work of Smith (1924), Rose and Dunham (1977) and Shepherd and Goldring (1993) for additional details.

Geographical distribution: The broader Cumbrian hematite ore field (Figure 1) may be divided into three distinct districts with apparently little or no mineralization occurring between each district: (1) a Northern District (west Cumbria) forming an arcuate belt of deposits stretching from Lamplugh in the northeast to Cleator Moor and Bigrigg on the north-western margin of the Lake District massif, and the Egremont and Beckermet group of mines to the south, (2) a Central District (western Lake District) centred on Boot, and (3) a Southern District (south Cumbria) including Millom, Ulverston and the Furness peninsular.

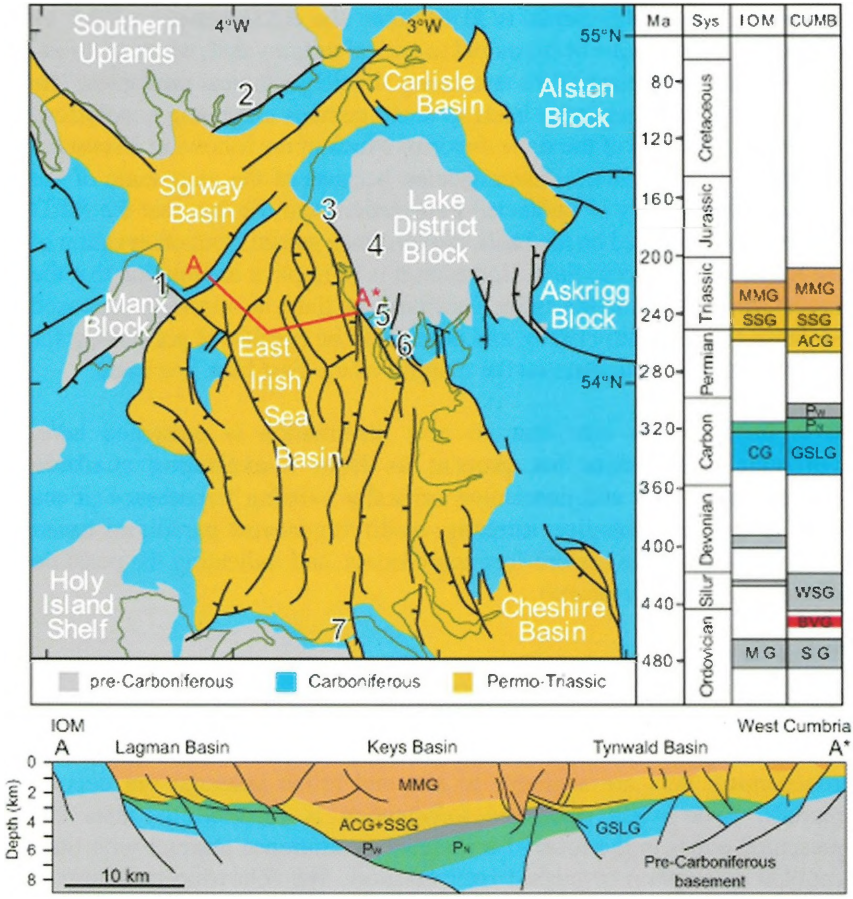


Figure 1. Geological setting of hematite mineralization on the margins of the East Irish Sea basin showing the location of ore fields (modified from Akhurst *et al* 1998). West-east cross-section of the northern sector of the basin reconstructed from seismic and exploration borehole data (after Jackson *et al* 1995). Ore fields: 1 – Maughold Head, Isle of Man. 2 – Auchenleck, SW Scotland. 3 – West Cumbria. 4 – Central Cumbria. 5 – Millom (South Cumbria). 6 – Furness peninsular/Ulverston (South Cumbria). 7 – Vale of Clwyd, NE Wales. Stratigraphic columns for the Isle of Man (IOM) and the East Irish Sea basin/western Cumbria (CUMB). Abbreviations: ACG- Appleby Group and Cumbria Coast Group. BVG – Borrowdale Volcanic Group. CG – Craven Group. GSLG – Great Scar Limestone Group. MG – Manx Group. MMG – Mercia Mudstone Group. P_N – Pennsylvanian (Namurian). P_W – Pennsylvanian (Westphalian). SG – Skiddaw Group. SSG – Sherwood Sandstone Group. WSG – Windermere Supergroup.

The distribution of ore bodies is primarily related to the prevailing geology with all major deposits of the Northern and Southern districts hosted by Carboniferous (Mississippian) marine limestones of the Great Scar Limestone Group. Only minor or sub-economic mineralization occurs within Early Ordovician clastic metasediments of the Skiddaw Group (Northern District), Late Ordovician Eskdale granite (Central District), and the Late Permian St Bees Evaporite (Magnesian Limestone) Formation (Northern District) and Early Triassic St Bees Sandstone Formation (Northern District).

An important second order control on ore deposit distribution, specifically relevant to the location of limestone-hosted deposits of west and south Cumbria, is linked to the disposition of Carboniferous limestone units with respect to the overlying Pennsylvanian and Permo-Triassic successions. Limestone-hosted ore bodies appear to be largely absent where the Carboniferous limestone succession is separated from the Early Triassic St Bees Sandstone Formation by fine-grained, low permeability Pennsylvanian fluvial-deltaic sediments and units of the Late Permian St Bees Shale Formation (Dunham 1984). Although this relationship has been disputed (Schnellmann 1984), the observation has potentially important consequences for ore deposit formation. The significance of this relationship will be explored in the section on "Hematite mineralization and ore deposit formation" (below).

Ore deposit mineralogy: Ore bodies are dominated by a remarkably simple mineral assemblage consisting of hematite-quartz-dolomite-calcite. Hematite is the only economically significant phase and typically accounts for between 75 and 95 weight % of total production. Quartz constitutes the main gangue mineral (5 to 15 weight %) with subsidiary dolomite and calcite (<10 weight %). Minor amounts (<5 weight %) of barite, fluorite, manganese oxides and sulphides occur locally.

Deposit size: A few individual mines, such as Hodbarrow Point (south Cumbria), are estimated to have produced in excess of 20 million tonnes of hematite ore from a single continuous ore body or a series of interconnected ore bodies. However, the broader ore field is dominated by many smaller deposits with production in the region of 10^5 to 10^7 tonnes.

Total ore production: Although no precise figures are available, estimates for total production from the Cumbrian ore field vary between 200 and 250 million tonnes of high grade ore containing ~50 weight % iron. Unrecovered reserves are thought to be limited to approximately 5 million tonnes (Shepherd and Goldring 1993), but the possibility exists that

undiscovered deposits remain in place in both west and south Cumbria (Rose and Dunham 1977; Dunham 1984).

Style of mineralization: Ore bodies conform to one of three geometric forms. (1) 'Veins'. Narrow, vertical or sub-vertical, ore bodies formed by passive infill of NW-SE (Northern District), N-S (Central District) and NE-SW (Southern District) trending extensional faults or fractures. Mineralization is usually accompanied by brecciation and minor replacement of wall-rocks. Vein-fills form a modest component of limestone-hosted mineralization, but represent the only style of mineralization hosted by Skiddaw Group metasediments and the Eskdale granite. (2) 'Flats'. Tabular, strata-bound and stratiform, metasomatic mineralization formed by replacement of limestones. Typical ore bodies consist of massive, replacive hematite together with passive, cavity filling, botryoidal (kidney ore) and specular hematite. Deposits are invariably linked to fault and fracture feeder systems and are often strongly fabric selective with mineralization essentially confined to specific packages of limestone beds and extending laterally over distances of 100's of metres (e.g. Hodbarrow Point). The architecture and distribution of individual ore bodies may be controlled by contrasts in host rock intergranular, vuggy (?karstic) and fracture porosity, and/or jointing density. (3) 'Sops'. Large (up to 350 m diameter), mineralized, inverted cone-shaped, karst-like collapse features. Deposits are confined to south Cumbria.

Timing of mineralization: Repeated attempts have been made to estimate the timing of ore deposit formation using a range of field relationships, palaeomagnetic analysis and radiometric dating techniques. Dunham (1984), based on relative geological relationships recorded from individual mines and wider field observations, argued for a late or post-Triassic mineralization event with a Jurassic to early Tertiary age being the preferred option. Although Permo-Carboniferous and Permian ages have been proposed on the basis of palaeomagnetic measurements (Evans and El-Nikhely 1982; Evans 1984; Rowe *et al* 1998), the occurrence of mineralization hosted by Early Triassic sandstones of the St Bees Sandstone Formation appears to provide conclusive evidence for a post-Palaeozoic age for hematite emplacement. A more recent palaeomagnetic study (Crowley *et al* 2014), using samples collected from across the entire ore field and an up-graded Apparent Polar Wander Path for the Eurasian continent during the Triassic, inferred a mid-Triassic age. This estimate is, at least, consistent with available field evidence.

Direct and indirect radiometric age estimates in the form of a single (U-Th)/He date (154 ±18 Ma) obtained from a sample of botryoidal hematite collected from Egremont, west Cumbria (Bähr *et al* 1994), and a suite of K-

Ar and Ar/Ar dates (170 to 115 Ma) reported from neoformed clay minerals (illite) associated with fault-related hematite mineralization in the vicinity of Sellafield (Milodowski *et al* 1998), yield Middle Jurassic to Early Cretaceous ages. Unfortunately neither of these age estimates may be viewed uncritically as accurate representations of the timing of mineralization because both dating methods are sensitive to anomalous thermal (resetting) events and potentially subject to significant uncertainty caused by poor retention (leakage and recoil) of unbound daughter noble gas elements (He, Ar) by fine-grained (micron-scale) hematite and clay crystallites.

At present we can only really be confident, despite apparent palaeomagnetic evidence to the contrary, in a maximum, post-Early Triassic (<247 Ma) age estimate imposed by observable field relationships. No useful reliable minimum age estimate is available, but it is probably reasonable to assume that the ore deposits are pre-Neogene (>23 Ma) based simply on the absence of any significant geotectonic activity in the central British Isles that might drive mineralization processes. It is currently not even possible to discriminate between a single mineralization event of limited duration (0 to 10 Ma) and multiple mineralizing events staggered over a prolonged time interval (10 to 100 Ma). Palaeomagnetic data (Crowley *et al* 2014) imply that a mineralization event of limited duration confined to the Middle Triassic is most likely. However, the number (10) and quality of mineralized sites sampled during this study was severely constrained by minimal exposure and the relatively low intensity of mineralization at outcrop. Unfortunately, the interpretation of available palaeomagnetic results is far from conclusive.

Although a single mineralizing event is the preferred option, it is entirely possible that hematite precipitation may have occurred at different times and perhaps repeatedly from the Middle Triassic to Early Cretaceous at individual sites across the ore field. The lack of unambiguous dating evidence remains the biggest single obstacle to understanding ore deposit formation and existing age estimates urgently require independent corroboration using accurate radiometric dating studies.

Fluid inclusions: Possibly the most important data reported from the Cumbrian ore field are fluid inclusion measurements first documented by Shepherd (1973), and supplemented by additional data collected over the following two decades (Rose and Dunham 1977; Shepherd and Goldring 1993). Fluid inclusions (typically 5 to 50 μm in diameter) are small samples of potentially ore-forming fluid trapped within minerals during crystal growth. By observing the behaviour of the inclusion fluid (and any associated solid or vapour phase) under a microscope during freezing, melting and heating experiments it is possible to derive estimates of fluid chemistry and

trapping temperature. The accuracy of resultant compositional and temperature estimates comes with a range of caveats, but essentially the data provide valuable limiting conditions at the time of mineral precipitation.

Although details are sketchy, measurements obtained from ore-stage quartz, fluorite and calcite-hosted inclusions indicate that ore-forming fluids were hypersaline (10 to 25 equivalent weight % NaCl), Na-Ca-Cl (Na/Ca $\approx$ 1) brines. Homogenization temperatures obtained from 2-phase (vapour-fluid) inclusions yield values between 84°C and 121°C. These temperatures are uncorrected for the effects of fluid pressure at the time of precipitation and, as such, represent minimum estimates. At a trapping pressure of ~100 to 150 bars (equivalent to mineralization at a depth of 1 to 1.5 km assuming a hydrostatic pressure gradient) the measured inclusion temperatures would need to be adjusted (increased) by +20 to +30°C.

A much more substantial data set is reported by Milodowski *et al* (1998) from borehole samples collected as part of the Sellafield (NIREX) nuclear waste repository investigation. Milodowski *et al* (1998) identified a (Stage ME6) fracture-hosted, hematite-quartz-calcite-dolomite-barite-fluorite mineralization event which they correlated directly with the major hematite ore deposits of west Cumbria. Although the range of Sellafield data recorded from Stage ME6 minerals is significantly greater than values obtained from ore field samples, the bulk (~90%) of fluid salinity estimates (20 to 30 equivalent weight % NaCl) and homogenization temperatures (80 to 140°C) are effectively identical to hematite ore deposit data.

In summary, ore-forming fluids associated with hematite mineralization appear to have been highly saline (~3 to 8 times the salinity of sea water) and warm (~120 ± 40°C).

Depth of ore deposit formation: Estimates of the burial depth of host rocks at the time of mineralization can be derived on the basis of apatite fission track measurements (supported by sandstone compaction studies) reported by Chadwick *et al* (1994, also see Akhurst *et al* 1997; Green 2002) from samples collected across Cumbria. From an interpretation of these data Chadwick *et al* (1994) were able to derive values for peak burial depth, immediately prior to regional post-early Palaeocene exhumation and cooling, of 1.5 km, 2.0 km and >2.5 km for the central Lake District massif, west Cumbria and south Cumbria respectively. Using estimates for the thickness of the missing Jurassic (~600m) and Cretaceous (~300m) sedimentary successions (Holliday 1993; Akhurst *et al* 1997) suggests possible ore formation depths in the region of 1.5 km during the Middle/Late Triassic, and somewhat deeper (>1.5 km) for Jurassic or Cretaceous mineralization.

Estimates of ambient pre-mineralization host rock temperatures are strongly dependent on contemporary surface temperatures and geothermal gradients. Both variables are likely to have evolved significantly during the Mesozoic (Holliday 1993), but an inferred temperature range of 60 to 80°C, close to minimum temperature values obtained from fluid inclusion analysis, provides a practical reference point.

Iron isotopes: The stable (not modified measurably by radioactive decay) iron isotope composition of hematite samples from the Florence mine (west Cumbria), reported as the ratio ($^{56}Fe/^{54}Fe$) of the amount of the (common) 'heavy' (^{56}Fe) to the (rarer) 'light' isotope of iron (^{54}Fe), has been used as a tracer for the source of iron contained within Cumbrian ore deposits and an indicator of the geochemical behaviour of iron during mineralization (Chapman *et al* 2006 a, b). Of particular importance is the inference that the iron required for hematite mineralization was supplied by the dissolution of iron-bearing minerals present in local granitic rocks (e.g. Eskdale granite) and transported as dissolved iron present in its ferric (Fe^{3+}) oxidation state (see section on "Aqueous iron geochemistry", below, for a summary of the geochemical behaviour of iron in solution). If this interpretation is correct then the implied geochemical processes associated with the generation of ore-forming fluids, iron transport and hematite precipitation has profound consequences for any ore deposit model. Unfortunately, the partitioning of iron isotopes during mineral dissolution, solute transport and mineralization is far from simple (Markl *et al* 2006; Saunier *et al* 2011) and the isotopic data may be subject to alternative interpretation. The implications of iron isotope data for potential ore deposit models will be discussed in more detail towards the end of this contribution.

What should an ore deposit model look like?

Hematite mineralization located on the margins of the East Irish Sea basin falls into the broader, low temperature, epigenetic, base metal (LTEBM) class of ore deposits and shares many similarities with the globally important Mississippi Valley-type lead-zinc sulphide mineralization of the mid-continental USA. The key characteristics of this style of mineralization (see summary of Anderson and Macqueen 1982) are: (1) major deposits occur most frequently in successions of marine limestone or dolomite host rocks, (2) mineralization is younger (often 10's to 100's Ma) than the host rocks, (3) ore-forming fluids are typically warm (80 to 180°C), hypersaline (5 to 10 times the salinity of seawater), Na-Ca-Cl brines, and (4) deposits are often spatially and temporally linked to active orogenic forelands or located on the margins of extensional sedimentary basins.

Although the geological setting and geochemical conditions associated with the formation of individual examples of LTEBM mineralization will vary in detail, all genetic models share three obvious components.

Solute generation: Metals, often dispersed at low to very low concentrations typical of upper crustal averages, are scavenged from large volumes (km^3) of source rock by geochemical reaction between the mineralogical component of aquifer rocks and circulating fluids, to produce potential ore-forming solutions (Bodnar 2009). Base metal (Fe, Zn, Pb) solubilities are enhanced at moderate temperatures (<200°C) under conditions of high fluid chlorinity, low (acidic) pH and low amounts of dissolved sulphur (e.g. Heinrich and Seward 1990; Hanor 1996). Temperature is a modest contributor to base metal solubility over the low and narrow range of values associated with LTEBM mineralization. Formation of inorganic metal complexes, particularly metal-chloride complexes, at high chlorinities (>100g l^{-1} Cl^-, equivalent to a salinity of >170g l^{-1}) may be particularly important in generating (Fe) metal-rich, ore-forming fluids (Heinrich and Seward 1990; Hanor 1996).

Transport: All LTEBM mineralization models are essentially hydrogeological models (Garven and Raffensperger 1997); that is, any genetic model is dependent on the extant, crustal scale, fluid flow (circulation) regime at the time of ore deposit formation. Ore-forming fluids generated by water-rock interaction (see above) must be transported from source to sites of mineralization where metals are concentrated and deposited as economically viable ore bodies. Inherent in this process is the presumption that some form of coupled advective mass and heat transport must occur on scales consistent with the fluid volumes (dependent on the dissolved metal content of the fluid and efficiency of metal precipitation at the site of mineralization) required to account for the amount of metals contained within observed ore deposits. In addition, mineralization is frequently accompanied by localized transient thermal anomalies at the site of ore deposit formation such that the temperature of the ore-forming fluid, at the point of mineralization, is elevated temporarily above the regional (background) temperature of the host rocks. Consequently fluids must migrate from hotter (?deeper) source regions, presumably along high permeability pathways at velocities sufficient to raise local temperatures above ambient values, often by several 10's of degrees Celsius.

Although the physics of fluid, mass and heat transport in porous media is well understood, the interaction of a complex set of variables, including the possible co-existence of differing flow drivers operating at different levels within the upper crust (e.g. Person and Garven 1994; Person *et al* 1996),

makes any attempt to isolate a specific transport model extremely challenging. The most successful approaches to solving the transport problem are based on numerical modelling of fluid flow regimes, often on the largest scales (Person *et al* 1996). We can, however, begin to gauge the importance of transport models at a basic level by examining briefly the effects of a few simple variables – metal content/precipitation efficiency of ore-forming fluids, fluid volumes, flow rates and duration of mineralization events – on ore deposit formation.

Identification of the correct LTEBM mass transport mechanism is particularly constrained by the presumed dissolved metal content of ore-forming fluids. Where mineralizing fluids are assumed to contain relatively low amounts of dissolved base metals (1 to 10 mg l^{-1}) typically reported from analysis of many samples of extant upper crustal fluids then the projected fluid volumes required by some LTEBM models may be considerable – 1 x 10^3 km^3 (e.g. Bethke and Marshak 1990; Garven *et al* 1993). When combined with suitable flow rates required to allow ore deposits to form on geologically reasonable time scales, the involvement of fluids with such low ore-forming potential places severe limits on the choice of possible hydrogeological regimes. In contrast, if the metal content of ore-forming fluids is assumed to be significantly higher (10^3 to 10^4 mg l^{-1}), and the bulk of dissolved metals are precipitated at the site of mineralization, then the volume of fluid and duration of the ore-forming event can be reduced by several orders of magnitude and allow a wider range of possible mass transport processes to be considered. A simplified assessment of the effects of varying dissolved metal concentration, precipitation efficiency, fluid volume and flow rate on the formation of ore deposits of differing size (metal content) can be made by examination of Figure 2 (page 290).

Clearly the link between the metal content of ore-forming fluids, the volume of fluid required to deposit metals in observed ore deposits, the specific hydrogeological regime(s) and the geotectonic setting (regional crustal stress field and thermal structure) is critical in identifying the precise processes responsible for mass (and heat) transport during mineralization. Underlying all of these variables is the precise timing of ore deposit formation which would allow us to define many of these underlying variables and constrain widely differing transport models.

Mineralization: In order to form an economic ore deposit it is necessary to concentrate and precipitate dissolved metals into a relatively small host rock volume. At the same time any genetic model must also be capable of generating space to 'accommodate' newly precipitating ore minerals and the growing ore body..

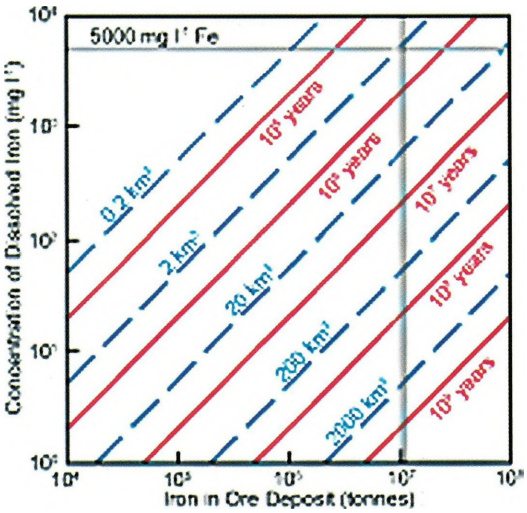


Figure 2. Modified "Roeder diagram" (Roeder 1960; Bodnar 2009) illustrating the relationship between the amount of iron transported per litre of mineralizing fluid (y-axis) and the size (metal tonnage) of the corresponding ore deposit (x-axis). The volume of fluid (dashed blue diagonal lines) and time (red diagonal lines) required to form an ore deposit are contoured onto these co-ordinates assuming a time averaged Darcy flow velocity of 0.5 m yr^{-1} for a cross-sectional flow area of 0.01 km^2 (e.g. 1 km x 10 m). Time and flow volume estimates are shown for a ~$1x10^7$ tonne iron deposit, assuming an ore-forming fluid containing (and precipitating) 5000 mg l^{-1} iron. Under these conditions the plot shows that a large (~ 2 x 10^7 tonne) hematite ore body could be deposited from 2 km^3 (2 x 10^{12} litres) of fluid over a time interval of 4 x 10^5 years.

Most common ore and gangue minerals (with a few notable exceptions such as calcite and dolomite) are characterized by prograde solubilites and this suggests that ore deposits could form by simple fluid cooling. However, cooling of metal-rich fluids alone is probably insufficient to allow significant mineral precipitation to occur over the relatively narrow zones of mineralization associated with small volume, high grade LTEBM ore bodies because the rate of change in the solubility of most ore minerals, with respect to temperature, is generally quite low (1 to 3 mg l^{-1} $°C^{-1}$) below 200°C.

For most ore-forming base metal minerals changes in other geochemical variables are usually more important drivers of precipitation than changes in temperature, unless the temperature gradient is particularly steep (>>10 $°C\ m^{-1}$). Consequently we must consider alternative geochemical processes

that are likely to induce sufficient solubility change so that a large proportion of the dissolved metal content is forced to precipitate from ore-forming fluids within a restricted volume. Chemical reaction between ore fluids and wall rocks, often inducing significant pH change, provides one potential process capable of driving precipitation of ore minerals. A second precipitation mechanism commonly invoked to explain observed economic mineralization is described by the interaction of two separate fluids that are each undersaturated with respect to a specific ore mineral, but when mixed become oversaturated with respect to the mineral of interest.

In addition to establishing the mechanism of ore mineral precipitation, we also need to address the problem of creating space to accommodate the ore body. Unless mineralization is confined to passive filling of pre-existing porosity or evolving dilational brittle faults, fractures and joints then space must be created within the host rock by replacement. Many LTEBM ore deposits are hosted by low porosity ('tight') limestones (or dolostones), so ore-forming fluids must be capable of dissolving carbonate while at the same time precipitating ore minerals and, eventually, removing (or reprecipitating elsewhere) dissolution products (at least to the extent that there is a local nett decrease in total mineral volume). This style of metasomatic replacement mineralization can be an extremely efficient mechanism for inducing ore mineral precipitation because the process of carbonate dissolution can act as a powerful (effectively infinite) pH buffer, allowing minerals that are soluble in acidic fluids to precipitate simply in response to a pH increase driven by the chemical reaction between the ore-forming fluid and reactive carbonate host rock.

The working (R&D) hematite ore deposit model (Rose and Dunham, 1977)

Early (pre-1970) conceptual models for ore deposit formation (see summaries of Rose and Dunham 1977; Dunham 1984; Shepherd and Goldring 1993) were developed on the basis of field observation and strongly influenced by contemporary thinking on ore-forming processes (e.g. see Trotter 1945). Fluid inclusion measurements (see above) allowed some conflicting anomalies apparent in earlier models to be resolved, and field observations and fluid inclusion data were subsumed into a new, internally consistent, paradigm. The resultant genetic model proposed by Rose and Dunham (1977) for the south Cumbria ore district was conditional on the assumption that the model would also explain similar mineralization in west Cumbria and, by implication, could potentially account for other hematite ore deposits located on the margins of East Irish Sea basin.

The essential components of the R&D model are illustrated in Figure 3.

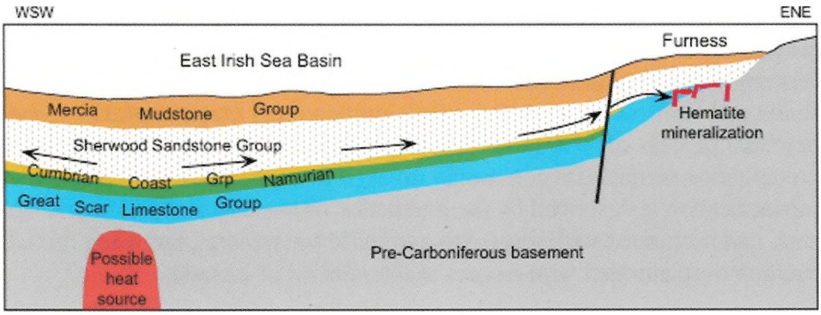


Figure 3. Simplified cross-section illustrating the essential components of the genetic ore deposit model proposed by Rose and Dunham (1977). Warm, iron-rich, hypersaline fluids are generated in the East Irish Sea basin and transferred, possibly involving convection cells driven by igneous intrusion, to basin margins where dissolved iron is precipitated as hematite (after Rose and Dunham 1977).

Rose and Dunham (1977) proposed that 'warm', hypersaline fluids 'leached' iron, originally present as hematite-bearing detrital grains and grain-coating hematite (see Figure 4), from red, fluvial and aeolian Permo-Triassic sandstones located in the deep subsurface of the East Irish Sea basin. The resultant iron-rich fluids were then transported along prevailing hydraulic gradients towards basin margins where dissolved iron was precipitated as hematite within limestone (and pelitic metasediment and granitic) host rocks of western Cumbria. Rose and Dunham (1977) also suggested that ore fluid transport might be influenced by the presence of Tertiary intrusions located within the centre of the basin and argued that a transient heat flow anomaly associated with magmatic activity could drive convective circulation of mineralizing fluids.

In reality although the R&D model postulates a source of metals, a mechanism to transport dissolved metals to sites of mineralization and offers an explanation for the basin margin location of ore bodies, the model is based essentially on very broad generalizations with little explanatory power or insight into the ore-forming process. It notably fails to provide any geochemical basis for the generation of ore-forming fluids or ore deposit formation and presents only a cursory description of possible mass transport. The R&D model does, however, contain numerous testable consequences that would enable the model to be falsified. For example, if Tertiary magmatic heating was a prerequisite for ore deposit formation then the timing of mineralization would need to be consistent with this requirement. In this respect the R&D model does, at least, provide a useful base line for the critical evaluation of potential ore-forming processes responsible for hematite mineralization.

Low temperature (<200°C) aqueous geochemistry of iron

Before proceeding to examine key components of the R&D model in detail, it is worth pausing briefly to remind ourselves of the main factors that control the behaviour of iron in solution under geochemical conditions likely to exist during the generation of ore-forming fluids, transport and mineralization. Iron is the fourth most abundant element in the continental crust (~50g kg^{-1}) and the third most abundant metal. Although iron can exist in a range of oxidation states, chemical behaviour in natural geochemical environments is dependent on the relative propensity of iron to change valence states, from ferrous (Fe^{2+}) to ferric (Fe^{3+}) iron (and *vice versa*), by the transfer of an electron (EQN. 1). This reduction-oxidation (redox couple) process must involve either a suitable reducing (electron donor) or oxidizing (electron acceptor) agent present in the form of a common reactive dissolved gas (H_2S, O_2), species (HS^-, SO_4^{2-}) or compound (organic molecules, metal oxides), usually in the presence of water.

EQN. 1 $\quad Fe^{2+} \leftrightarrow Fe^{3+} + e^-$

The oxidation state of iron is particularly important because, together with pH and temperature, it provides a fundamental control on the stability of iron-bearing minerals and, ultimately, iron solubility (Heinrich and Seward 1990). In unpolluted, oxidizing geochemical environments, such as those encountered at earth-surface or near surface conditions, ferric (Fe^{3+}) iron is the stable state of iron in minerals and solution, but is only sparingly soluble (<0.05 mg l^{-1}) except at low pH (<4). In contrast, ferrous (Fe^{2+}) iron may be highly soluble (>>0.5 mg l^{-1}) at 'normal' pH ($5 \leq pH \leq 8$) providing that geochemical conditions are suitably reducing (Hem 1985).

We can readily confirm that iron can only be dissolved in significant amounts as ferrous (Fe^{2+}) iron, under geochemical conditions (warm, hypersaline, weakly acidic) typically associated with LTEBM mineralization, using a simple thermodynamic model. Figure 5 is an oxygen fugacity ($\log fO_{2(g)}$)-pH ($-\log_{10} aH^+$) diagram that illustrates the stability fields for iron species and common mineral phases with respect to the effective (thermodynamic) concentration (fugacity) of dissolved oxygen gas and (activity) H^+ ions in saline (chloride-rich) fluids at a temperature of 100°C and a total pressure of 250 bars. The diagram shows that ferric iron (as $FeCl_2^+$) is only soluble to any appreciable extent in acidic (pH<3) fluids. Even if such low pH fluids could be generated under geochemical conditions operating in most low temperature upper crustal settings, then these fluids would tend to be buffered rapidly towards higher pH values (pH 4 to 7) by reaction with multi-phase aluminosilicate-carbonate mineral assemblages contained within the rock matrix of the host aquifer.

Under conditions characteristic of many deep sedimentary basins (Hanor, 1994, shaded grey box in Figure 5) it is expected that any dissolved iron will be present in its ferrous (Fe^{2+}) state either as Fe^{2+} and/or iron chloride complexes (e.g. $FeCl^{+}$, $FeCl_2$) dependent on the chloride content of the fluid (Heinrich and Seward 1990). Figure 5 also shows that ferrous (Fe^{2+}) iron solubility is essentially controlled by the stability of a small number of common iron-bearing phases (hematite, pyrite, siderite, chlorite). At low concentrations of dissolved total sulphur and aqueous CO_2 it can be shown (Figure 5, inset) that pyrite and siderite are not significant factors limiting the solubility of iron.

As we will see the geochemistry of iron in solution has potentially profound consequences for the generation of ore-forming fluids and mineralization. An understanding of the behaviour of iron is clearly fundamental to the development of accurate hematite ore deposit models.

Hypersaline basinal fluids as mineralizing solutions

We have already seen from fluid inclusion data that the fluids associated with Cumbrian hematite mineralization (and, in fact, most LTEBM ore deposits world-wide) are characterized by exceptionally high salinities (~ 3 to 8 times the salinity of sea water). The formation of hypersaline brines is evidently significant for tracing the origin of ore-forming fluids.

Hypersaline pore fluids, brines typically containing >100,000 mg l^{-1} total dissolved solids, are a common feature of the deeper (>2km) parts of many sedimentary basins (Hanor 1994), particularly where the basin-fill contains thick evaporite units (e.g. gypsum/anhydrite, halite), such as the Mercia Mudstone Group of the East Irish Sea basin (Jackson *et al* 1995), as part of the sedimentary succession. Sedimentary brines have often been considered as candidate fluids for LTEBM mineralization (e.g. Carpenter *et al* 1974; Sverjensky 1984; Hanor 1994, 1996; Yardley 2005) because they have salinities and major element (Na, Ca, Cl) compositions similar to those recorded in ore and gangue phase fluid inclusions. All brines are chloride-dominated, halite undersaturated, fluids that are thought to form primarily either by evaporation of sea water (up to and beyond halite saturation) or dissolution of evaporitic halite.

Although chloride behaves conservatively during progressive sediment burial (Hanor 1994), the cation component, including economically important base metals, can be modified extensively by reaction between the fluid and common aluminosilicate and carbonate minerals that form host aquifers. It is also evident that as the chloride and, therefore, metal cation content of the fluid increases pH must decrease in order to maintain solute

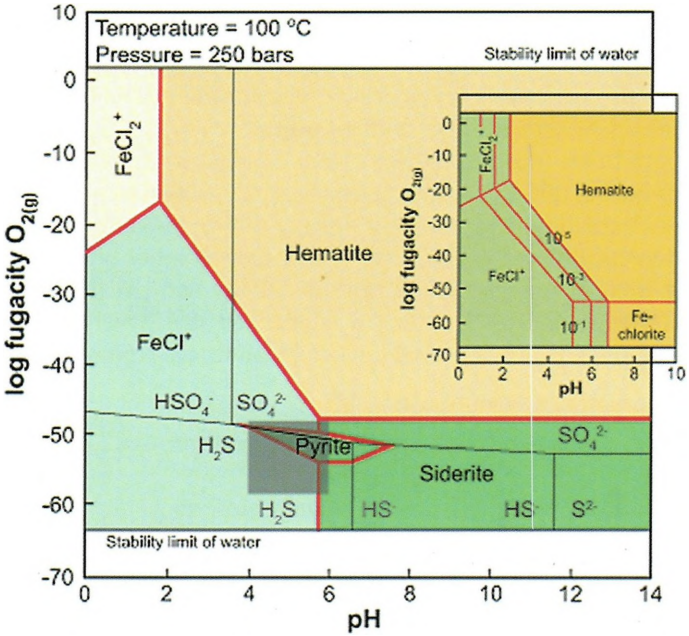


Figure 5. Redox (oxygen fugacity) – pH diagram for the Fe-S-CO_2-Cl-H_2O system at 100°C and 250 bars total pressure. Diagram is constructed using the ACT2 code in Geochemist's Workbench® (Bethke 2013) assuming log activities for total dissolved sulphur (as SO_4^{2-}), iron and chloride of -5, -3 and 0.3 respectively, and log $CO_{2(g)}$ fugacity of 1. The plot is colour coded to highlight the distribution of ferric (Fe^{3+}) iron phases (dark) and species (pale) in yellow shades and ferrous (Fe^{2+}) iron phases (dark) and species (pale) in green shades.

The grey box shows the typical range of estimated $^{log}fO_{2(g)}$ and pH values for oil field brines. The stability boundaries for sulphur species are shown in grey. Inset diagram illustrates the effect of changing the concentration of dissolved iron (as activities) on the stability boundaries for the Fe-Al-Si-S-CO_2-Cl-H_2O system where the activities of aluminium, silica, total sulphur and log $CO_{2(g)}$ fugacity are fixed by the stabilities of kaolinite, chalcedony, anhydrite and calcite respectively. The distribution of soluble iron species is shown in green and solid phases is shown in yellow. Under these conditions (low sulphur and inorganic carbon), Fe-chlorite is stable relative to siderite and pyrite. Note that iron is only soluble in its ferrous state (Fe^{2+}, $FeCl^+$) under 'basinal' geochemical conditions.

cation activity ratios (e.g. aNa^+/aH^+, $aCa^{2+}/(aH^+)^2$) in equilibrium with the extant aquifer mineral assemblage (Hanor 1994). That is, as the concentration (activity) of dissolved cations (Na^+, Ca^{2+} etc) increases the concentration of H^+ must also increase. This process results in progressively more acidic fluids with increasing salinity, with measured pH values in samples of basinal pore fluids (uncorrected for *in situ* conditions) decreasing sharply from near-neutral (pH 6.5) values at salinities of ~100,000 mg l^{-1} to ~ 3.5 at salinities exceeding 150,000 mg l^{-1} (Hanor 1994). Importantly, it can be shown theoretically and experimentally that the solubility of base metals, including iron, is strongly linked to fluid salinity and pH (Heinrich and Seward 1990; Hanor 1996) with the implication that hypersaline brines have the potential to act as potent, low temperature, ore-forming solutions.

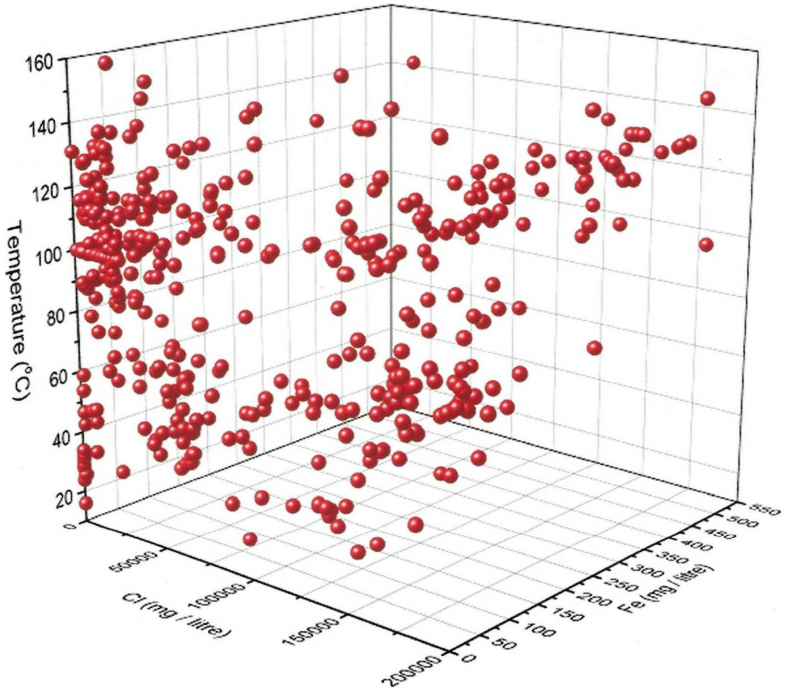


Figure 6. Three-axis plot showing variations in the concentration of dissolved iron with respect to chloride content and temperature for formation fluids collected from hydrocarbon exploration and production wells (462 data points). Although there is considerable scatter, a general trend towards increasing Fe with increasing temperature and salinity may be discerned with high iron concentrations (>100 mg l^{-1}) associated with chlorinities >100,000 mg l^{-1}.

The link between salinity and dissolved base metal content is further supported by measurement of iron concentrations in basinal pore fluids (formation waters) recovered (mostly) from hydrocarbon exploration and production wells. Figure 6 illustrates variations in the iron content of pore fluids (n = 462) collected from a range of sedimentary basins with respect to chloride concentration and temperature compiled from published sources. Despite considerable scatter a crude positive trend may be observed showing that chloride content and temperature act as contributory variables (in addition to oxidation state and pH) influencing iron solubility. The majority of fluid samples containing the highest dissolved iron contents (>400 mg l^{-1}) are associated with elevated chloride concentrations (>100,000 mg l^{-1}) and fluid temperatures (>100 °C).

This general solubility trend is confirmed by thermodynamic modelling (Heinrich and Seward 1990) which shows that the concentration of dissolved iron in reducing, low sulphur and dissolved inorganic carbon systems is essentially dependent on salinity, pH and (to a lesser extent) temperature. Under ideal circumstances the development of warm, iron-rich, hypersaline ore-forming fluids containing dissolved iron concentrations >10^3mg l^{-1} may be possible.

Source of Iron

Perhaps the least intuitive element of the R&D model is the suggestion that the iron required for mineralization was derived from the dissolution of hematite dispersed in Permo-Triassic red sandstones (Figure 4) located offshore within the East Irish Sea basin (Figures 1, 3). Iron cannot be sourced locally because unmineralized Carboniferous marine limestones contain extremely low amounts of iron (typically <0.1 weight %) and other potential sources (e.g. Skiddaw Group metasediments, Eskdale granite, Sherwood Sandstone Group) exhibit little evidence (at least at outcrop) of widespread (selective) dissolution of iron-bearing minerals. Consequently, viable, alternative, non-local, sources must be identified. Rose and Dunham (1977) proposed that Permo-Triassic sediments, specifically offshore equivalents of the Early Triassic St Bees Sandstone Formation, "could readily yield the required quantities of iron", predicted that evidence of dissolution of hematite from red sandstones "might appear at deeper levels below the Irish Sea", and speculated that "the validity of the suggested process would be provided if we were able to indicate that leaching of St Bees Sandstone (Formation) would yield the very simple solutions, more or less free of non-ferrous metals, that are required".

Unfortunately, Rose and Dunham (1977) were unable to provide any observational evidence that hematite dissolution has occurred in East Irish

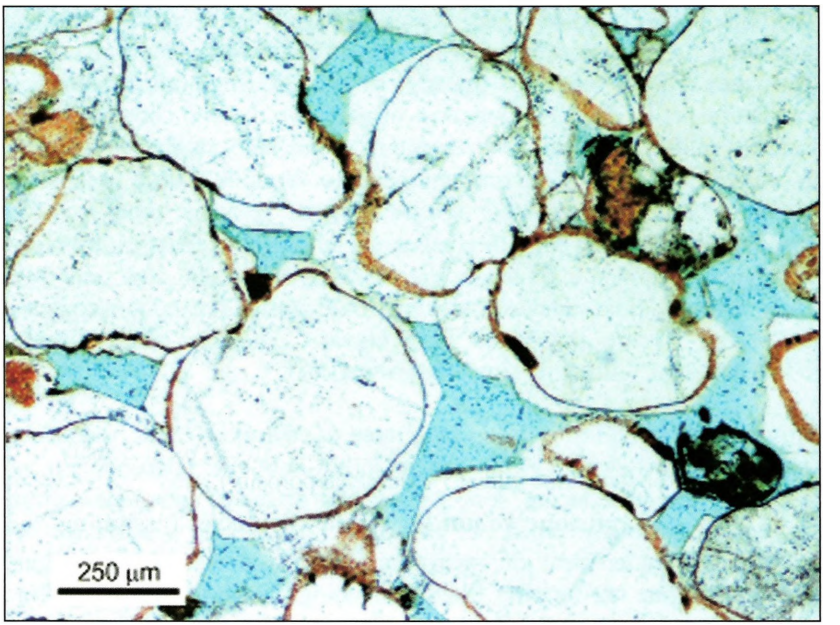


Figure 4. Plane polarized, transmitted light photomicrograph of hematite-bearing sandstone. The distinct red, iron oxide (hematite), coating on well-rounded detrital quartz grains results from the breakdown of detrital ferromagnesian minerals and precipitation of dissolved ferric (Fe^{3+}) iron as microcrystalline hematite under strongly oxidizing geochemical conditions during shallow burial diagenesis. In this example, hematite-rimmed quartz grains are enclosed by pore-filling quartz cement. Blue-dyed resin impregnation shows the distribution of intergranular porosity.

Sea basin red sandstoness, or document any experimental evidence to support the generation of iron-rich, but non-ferrous (Cu, Pb, Zn) metal-poor, fluids by simple geochemical processes. Furthermore, Rose and Dunham (1977) postulated that iron was transported in solution in its reduced, ferrous (Fe^{2+}) state, but failed to offer a plausible mechanism that would produce ferrous (Fe^{2+}) iron-rich fluids by dissolution of hematite where iron occurs in its higher, ferric (Fe^{3+}), oxidation state.

Quantitative estimates of the volume of Permo-Triassic sandstone required to release sufficient iron by dissolution of dispersed hematite can be calculated on the basis of a few simple assumptions (see Text Box 1). A typical large hematite ore body such as Hodbarrow Point would necessitate the dissolution of hematite from a relatively small (~0.28 km^3) volume of sandstone. Extending the calculation to account for hematite mineralization contained within the entire ore field gives a total sandstone volume of ~2.8 km^3.

Similar estimates were reported by Rose and Dunham (1977) using a somewhat higher value for the density of Permo-Triassic red sandstones.

Text Box 1. Iron Source Budget

What volume of sandstone would need to be 'leached' to release sufficient Fe to account for a large (25 million tonnes) hematite ore body?

Early Triassic sandstones of the St Bees Sandstone Fm contain between 1.6 and 2.4 weight % Fe (Rose and Dunham, 1977). A typical Fe content of 20 g kg^{-1} is assumed.

A 25 million tonnes hematite ore deposit with an ore grade of 50 weight % iron contains 1.25×10^{13} g of iron.

Assume a bulk sandstone density of 2.25 g cm^{-3} (typical density estimate for sandstone with 15 volume % porosity).

Minimum sandstone volume in cubic kilometres (assuming that all hematite is quantitatively dissolved and transported to the site of mineralization):

$$\frac{1.25 \times 10^{13}}{20 \times 2.25 \times 10^{12}} = 0.28 \text{ km}^3$$

To supply the Fe budget for the entire Cumbria ore field (250 million tonnes total ore production) would require the dissolution of hematite from **2.8 km^3** of Permo-Triassic red sandstone.

Clearly these volume estimates represent absolute minimum values because the dissolution reaction is unlikely to be completely efficient and a proportion of the total iron released by 'leaching' may reprecipitate as chlorite, pyrite or siderite. Uncertainties associated with other assumptions (iron content and porosity of Permo-Triassic sandstones) are unlikely to lead to significant errors, but it would be surprising if the actual volume of sandstone involved in the 'leaching' process could be less than some low multiple (5 to 10 times) of the minimum estimate, say 15 to 30 km^3.

Precise estimates of the total volume of Permo-Triassic sandstone available within the East Irish Sea basin are problematic. The present-day thickness of the Early Triassic Sherwood Sandstone Group in basins offshore of the Cumbrian coast (Figure 1) typically varies between 500 and 2000m (Jackson *et al* 1995) with an average of ~ 1000 to 1500m. Taking a basin width of 50km and assuming that only sandstones directly offshore of the Cumbrian ore field (stretching approximately 60km from north to south)

yields a volume estimate of 3000 to 4500 km^3. Calculations confirm the assertion of Rose and Dunham (1977) that Permo-Triassic sediments contained within the East Irish Sea basin could easily supply sufficient iron to account for observed hematite mineralization, but calculations also show that the required extent of sandstone 'leaching' is non-trivial and has important implications for other components of any ore deposit model.



Figure 7. Core photograph showing an example of selective hematite 'leaching' in Sherwood Sandstone Group sandstones, East Irish Sea basin (Well 110/2-1, 4079.00-4085.40ft). Variations in the intensity of hematite dissolution are indicated by changes in colour from dull red (hematite-rich) to white (hematite-poor).

Following on from the quantification of apparent source rock volume is the obvious prediction that we should be able to observe evidence of extensive hematite dissolution in offshore Permo-Triassic sedimentary successions. Fortunately, hydrocarbon exploration within the East Irish Sea basin over the last fifty years provides us with a visual record of basin-fill sediments in the form of borehole core photographs. Examination of the British Geological Survey core photographic archive (see BGS GeoIndex Offshore web site at http://mapapps2.bgs.ac.uk/geoindex_offshore/home.html) reveals numerous examples of apparent 'leaching' where presumably originally

red, hematite-bearing, sandstones have been altered to light grey sandstones by hematite dissolution (Figure 7). In fact, examples of hematite 'leaching' have been reported from red bed successions world-wide (Beitler *et al* 2003; Rainoldi *et al* 2014). In specific instances the apparent extent of 'leaching' has proven to be truly spectacular, as exemplified by the Jurassic Navajo Sandstone of southern Utah (USA), where an estimated 14.7 x 10^3 km^3 of sandstone have been affected by 'leaching' processes (Beitler *et al* 2003). These global examples show that under the right conditions hematite 'leaching' is a common geochemical process operating in sedimentary basins that can occur on scales compatible with estimated rock volumes projected for the formation of the entire Cumbrian hematite ore field.

It is clearly evident that the 'leaching' of Permo-Triassic sediments could yield sufficient iron to account for observed hematite mineralization, and core records show that hematite dissolution is a common phenomenon within East Irish Sea basin-fill successions. However, these calculations and observations alone do not provide a plausible quantitative geochemical description of the dissolution process that can account for the dissolution and transformation of ferric (Fe^{3+}) iron-bearing hematite to soluble ferrous (Fe^{2+}) iron in potential ore-forming fluids.

Because of the tendency of detrital minerals containing redox sensitive elements (e.g. amphiboles, pyroxenes) and organic matter to be highly unstable under strongly oxidizing, near surface geochemical conditions, most sediments deposited in arid, terrestrial environments are characteristically devoid of organic carbon, stained red by the formation of authigenic hematite (Figures 4, 7) and contain any redox sensitive inorganic components in their highest possible oxidation states (Walker 1976). As a consequence red bed sediments should have little or no internal electron donor capacity to drive reduction of ferric (Fe^{3+}) to ferrous (Fe^{2+}) iron. Any reduction of ferric (Fe^{3+}) iron present as hematite must, therefore, rely on the introduction of potential reducing agents from external sources.

Possible mobile reducing agents occurring in sedimentary basins, such a $H_2S_{(g)}$ and $H_{2(g)}$, can readily be discounted as likely reactants because under excess $H_2S_{(g)}$ any ferrous (Fe^{2+}) iron would precipitate *(in situ)* as pyrite, while it is unlikely that sufficient $H_{2(g)}$ would be available to release required amounts of iron into solution. The only plausible mobile reducing agents commonly available in significant amounts in sedimentary basins are restricted to petroleum, methane and organic acid anions (Kharaka *et al* 1987; Surdam *et al* 1993; Hanor 1994). Abiogenic reduction reactions involving methane are probably unimportant under low temperature (<160°C) conditions typical of many sedimentary basins (although

bacterially mediated reduction associated with methanotrophic communities is possible below ~ 50°C), and organic acid anions are only likely to be available in amounts sufficient to release a small fraction of the total iron required. This leaves petroleum as the only viable reducing agent that is both mobile and available in suitable volumes.

A variety of reaction schemes have been proposed for the reduction of iron by mobile organic reducing agents (e.g. Kharaka *et al* 1987; Rainoldi *et al* 2014). EQN 2 (Rainoldi *et al* 2014) illustrates one possible description of reaction stoichiometry.

EQN 2. C_9H_{20} (oil) + $4Fe_2O_3$ (hematite) + $3CO_{2(aq)}$ + $16H^+ \leftrightarrow$ $6CH_3COOH$ (acetic acid) + $8Fe^{2+}$ + $6H_2O$

The validity of petroleum-based hematite reduction is further supported by experimental hydrous pyrolysis of hematite-bearing sandstone in artificial mixtures of petroleum + water ± sulphate (Shebl and Surdam 1996), and the occurrence of natural examples of hematite reduction along palaeo-hydrocarbon migration pathways (e.g. Garden *et al* 2001; Beitler *et al* 2003). The reaction should also generate solutions selectively enriched in ferrous (Fe^{2+}) iron, but with no obvious potential to produce anomalous concentrations of other dissolved non-ferrous base metals. This may partly explain the unusual, restricted base metal mineralogy of Cumbrian (and other basin margin) hematite ore deposits. In addition, providing that the total sulphur and inorganic carbon content of the fluid remained sufficiently low, the concentration of dissolved iron, in the presence of elevated dissolved chloride, would continue to increase and be limited only by the availability of reactants and the stability of chlorite (Figure 5).

If we assume that the reduction of hematite by reaction with petroleum represents a viable geochemical mechanism to produce ferrous (Fe^{2+}) iron-rich fluids then we can proceed to estimate the volume of oil involved in the dissolution process. Text Box 2 outlines the necessary assumptions and shows that at least 5×10^9 litres (3×10^7 barrels) of oil is required to release sufficient iron to produce a large (25 million tonnes) hematite ore body such as the Hodbarrow Point deposit, and 5×10^{10} litres (3×10^8 barrels) to account for mineralization across the entire ore field. For comparison typical North Sea Brent oil fields (Heather, Thistle) contain between 4×10^8 and 8×10^8 barrels of oil (in place), while the Douglas and Lennox fields, within the East Irish Sea basin (Liverpool Bay), contain ~ 2×10^8 barrels of oil (in place). Although these estimates are based on the quantitative transmission of iron and represent absolute minimum volumes, it is evident that petroliferous basins are more than capable of generating sufficient petroleum.

Text Box 2. Petroleum Budget

What volume of oil would be required to release sufficient Fe to account for a large (25 million tonnes) hematite ore body?
Assume a molar volume for nonane (C_9H_{20}) of 179cm^3 mol^{-1} (at 25 °C).
Reaction stoichiometry (EQN. 2) shows that 1 mole of petroleum reacts with 4 moles of hematite to produce 8 moles of Fe^{2+} (or 447g of Fe).
A 25 million tonnes hematite ore deposit with an ore grade of 50 weight % Fe contains 1.25×10^{13}g of Fe.
Minimum oil volume (assuming that all Fe^{2+} is quantitatively dissolved and transported to the site of mineralization):

$$\frac{1.25 \times 10^{13} \times 179}{447 \times 1000 \ (\times 159)} = 5.0 \times 10^{9} \text{ litres (or } 3.1 \times 10^{7} \text{ barrels)}$$

To supply the Fe budget for the entire Cumbria ore field (250 million tonnes total ore production) would require 5.0×10^{10} litres or 3.1×10^{8} barrels of oil.

Current hydrocarbon production across the East Irish Sea basin is dominated by methane but many producing gas fields in the northern sector of the basin show evidence of a pre-Early Jurassic (>180 Ma) oil charging event (based on K-Ar radiometric dating of neoformed authigenic illite), with subsequent breaching of reservoirs and formation of bitumen prior to more recent hydrocarbon trapping events (Gluyas and Hichens 2003). In addition, the occurrence of bituminous hydrocarbons in NE Wales, E Isle of Man, SE Scotland and NW England (Jackson *et al* 1995), supported by radiometric dating, also reveals that hydrocarbon migration towards the margins of the East Irish Sea basin had occurred by Early Jurassic times. The long and complex hydrocarbon history of the East Irish Sea basin does not currently allow us to identify the precise timing of oil generation or the volume of oil released, but ample evidence exists to show that liquid hydrocarbons were generated in significant amounts, probably during rapid Middle/Late Triassic subsidence, when potential mid-Carboniferous (Namurian) oil-prone source rocks in the Keys and Tynwald sub-basins (see Figure 1) reached burial depths >3.5 km.

In summary, the hypothesis that dissolution of hematite from red, Permo-Triassic sediments deposited in the East Irish Sea basin could provide a potential source of iron for hematite mineralization at basin margins is

supported by the available volume of basin-fill sediments and the occurrence of numerous examples of 'bleached' sandstone observed in core photographic records. In addition, a mechanism has been proposed, based on the reaction between hematite and mobile hydrocarbons (in the presence of hypersaline pore fluids), that is capable of generating ferrous (Fe^{2+}) iron-rich ore-forming fluids.

Mass Transport

The transport of metals in solution from source to site of mineralization represents a significant and longstanding challenge to the development of accurate LTEBM ore deposit models because of the apparent large volumes of fluid and the comparatively high fluid flow rates required to form observed ore deposits, from fluids commonly thought to contain/precipitate relatively low amounts of dissolved metals (1 to 10 mg l^{-1}) per unit volume of fluid, on realistic geological time scales (Roeder 1960; Bodnar 2009). Even at elevated dissolved metal concentrations (>10^3 mg l^{-1}) large LTEBM ore deposits must still involve the movement of significant fluid volumes (Figure 2). If the source of metals is not local, then these fluids are required to migrate over substantial horizontal and vertical distances. The transport problem is further exacerbated by the fact that ore-forming events are often accompanied by transient and localised thermal anomalies that appear to raise temperatures at the site of mineralization above ambient (regional) crustal geotherms. Consequently ore-forming fluids must also travel at velocities sufficient to prevent significant cooling by conduction in contact with wall rocks/aquifers during transport. These constraints clearly place important limiting conditions on any viable mass transport model.

If we are correct in identifying red, Permo-Triassic sandstones located within the East Irish Sea basin as the principal source of iron then it is evident that ore-forming fluids have migrated over horizontal distances of several tens of kilometres and potentially vertical distances of several kilometres (Figure 1). In addition, any mass transport model should be capable of explaining the apparent transient thermal anomalies required to account for the difference between ambient crustal temperature estimates for the Cumbrian margin (at the time of mineralization) derived from apatite fission track data and the temperature of ore-forming fluids inferred from fluid inclusion measurements. While fluid inclusion measurements suggest maximum fluid temperatures of ~150°C (Shepherd and Goldring 1993; Milodowski *et al* 1998), fission track data implies that a post-Carboniferous peak regional temperature of ~100 to 110°C was only reached prior to Palaeocene cooling and exhumation of the Cumbrian margin (Chadwick *et al* 1994; Akhurst *et al* 1997; Green 2002). Assuming a Mesozoic age for hematite mineralization, with a preference for a Middle/Late Triassic

event(s), and adjusting regional temperatures for the effects of Jurassic and Cretaceous burial, implies an apparent discrepancy between fluid inclusion and ambient crustal temperatures >40°C.

Although a variety of hydrogeological mass transport proceses operating in sedimentary basins are available for consideration (see summaries of Person and Garven 1994; Garven and Raffensperger 1997; Bjørlykke 2010), only three fluid flow regimes are thought to be significant in undeformed extensional geotectonic settings: (1) topography driven (forced convection), open-system, meteoric ground water circulation, (2) buoyancy (density) driven (free convection), closed-system, flow cells, and (3) episodic venting of deep basinal fluid overpressures by active, basin-bounding fault systems (seismic pumping). Identifying the correct mass transport regime is dependent on evaluating evolving, time-dependent, geotectonic factors that control palaeohydrogeology on crustal scales. In the absence of direct measurement (not possible in ancient hydrothermal systems) or computer generated numerical hydrogeological flow simulations (not currently available for Cumbria and the East Irish Sea region), hypothesised ore deposit transport models are necessarily qualitative in nature and the selection of a specific model is based essentially on a process of elimination, constrained where possible by observational data. The following discussion briefly considers (and rejects) topography driven and buoyancy driven flow models before arguing that the fault venting model appears to offer the most likely mechanism for transporting metals to sites of hematite mineralization on the Cumbrian margin.

Perhaps the most frequently invoked mass transport model for LTEBM ore deposits is based on the concept of topography driven ground water circulation where fluid flow is induced by differences in the hydraulic head between zones of meteoric recharge and downstream discharge. This model was initially proposed to explain Mississippi Valley-type base metal mineralization in the mid-continent sedimentary basins of North America (e.g. Garven *et al* 1993; Garven and Raffensperger 1997). Numerical hydrogeological modelling (Garven *et al* 1993) suggests that the formation of these deposits was associated with active orogenic forelands that created the topography necessary to drive intense, forced convective, meteoric fluid flow into the deep (>3 km) subsurface. As meteoric fluids migrated along prevailing hydraulic gradients, often over distances of hundreds of kilometres, they interacted with foreland basin aquifer sediments to generate metal-bearing brines before depositing metals in distal foreland-margin locations as fluids cooled and returned towards surface discharge points. Such models are particularly popular because they are relatively unconstrained by apparent fluid volume restrictions evident in some

alternative transport models, and have typical deep aquifer flow velocities (1 to 10m yr^{-1}) that are more than sufficient to allow ore bodies to form on realistic time scales, even at low dissolved metal concentrations.

Intuitively it would seem unreasonable to apply, unmodified, the topography driven ground water model to Cumbrian hematite mineralization. In extensional sedimentary basin settings, such as the East Irish Sea basin, any regional meteoric ground water systems would be expected to flow from topographic highs located at basin margins towards discharge (low elevation) points at basin centres (Person and Garven 1994). A potential modern analogue is provided by the present-day Upper Rhine Rift Valley extensional basin (Stober and Bucher 2015) where ground water systems recharge at basin margins in the Black Forest (east) and Vosges (west) mountains (reaching a maximum depth of 4 km) before discharging as warm, hypersaline mineral springs towards the basin centre. Following the principles of the topography driven model any mineralization and associated thermal anomalies would be expected to occur along the discharge leg of the circulating fluid flow system, with ore deposits, in principle, located offshore of the Cumbrian margin. As we will see in the following section (Hematite mineralization and ore deposit formation) topography driven meteoric ground water is thought to play a role in the mineralization process, but forced convection, as proposed by Garven *et al* (1993), is discounted as a mechanism for the transport of iron-rich, ore-forming fluids from the East Irish Sea basin to sites of mineralization at basin margins.

A second important mass transport process used to explain the formation of some LTEBM deposits is based on the concept of buoyancy driven free convection. In general terms, once confined aquifers are isolated (usually by deep burial) from the dominant effects of topography driven ground water systems, free convective flow can occur in response to dynamic density instabilities generated by any temperature gradient of sufficient magnitude (dependent on aquifer thickness and permeability) to initiate fluid circulation. Free convection results from the fact that progressive thermal expansion of water towards the (hotter) base of an aquifer unit will cause fluid density to decrease, relative to overlying fluids, creating a potentially unstable inverse density gradient. If the density contrast is large enough then the colder fluid will sink into the hotter, less dense, fluid leading to an overturn of the aquifer. Providing that a critical value (Rayleigh number), describing the relative contributions to heat transport by conduction and advection, is exceeded then the fluid system will continue to convect indefinitely unless conditions change. In reality, the scale, geometry and flow characteristics of the resultant convection cells involve a complex

interplay of variables that include the magnitude of the prevailing temperature gradient, and the dimensions, orientation, inclination and permeability anisotropy of the host aquifer.

Free convective mass transport models are attractive because they are effectively unconstrained by potential fluid volume limitations imposed on single pass flow models; that is convection allows the same volume of fluid to be circulated repeatedly, with the possibility that ore-forming components may be dissolved by water-rock interaction along the higher temperature leg of the convection cell and ore minerals can be precipitated as fluids cool along the lower temperature leg of the cell. While some hydrogeologists (Person and Garven 1994) hypothesise that free convective fluid circulation is inevitable once deeply buried porous sediments are isolated from the effects of topography driven flow, others (Bjørlykke *et al* 1988) argue that the absence of perturbations in temperature (near-horizontal geotherms) and salinity gradients (normal density stratification) within many extant sedimentary basins implies that large (km)-scale convection is uncommon unless disturbed by a significant localised high heat flow anomaly (such as the emplacement of a magmatic intrusion). Even if free convective cells are developed, mathematical models indicate that flow velocities may be extremely low (10^{-4} m yr^{-1}) under conditions typical of most sedimentary basins, and convective circulation can readily be disrupted by the occurrence of thin (m-scale), low permeability (e.g. mudstones) horizons.

Although Rose and Dunham (1977) originally suggested that the transport of ore-forming fluids might occur in response to the development of convective cells driven by Palaeocene magmatic intrusions centred on the East Irish Sea basin (Figure 3), there is currently no evidence to support the occurrence of any plutonic igneous body of sufficient magnitude to initiate convection. Alternatively, Palaeocene magmatic underplating of the lithosphere beneath the Irish Sea associated with the evolution of the Iceland mantle plume (Al-Kindi *et al* 2003) might have generated a regional thermal anomaly capable of inducing fluid convection in the overlying continental crust, but the thermal effects is of this event are currently uncertain. It is also possible that convection might have occurred in any Mesozoic sediment overlying the Lake District massif in response to elevated heat flow (higher radiogenic heat production and thermal conductivity of granites relative to surrounding sediments) focussed through granitic basement. However, the effects of convective circulation should be evident within the granites. Any associated mineralization would tend to be concentrated in host rocks directly overlying the granite and be symmetrically distributed around the thermal anomaly. These characteristics are not consistent with the distribution of Cumbrian

hematite ore bodies. We cannot discount entirely the possibility that convection might have occurred within East Irish Sea basin-fill sediments, particularly the thick, permeable sandstones of the Sherwood Sandstone Group, but any convective cells are likely to have been confined to the deep basin and potential cell geometries would be unlikely to extend to shallow basin margins.

While forced and free convective fluid flow regimes are the dominant mechanisms controlling mass transport in sedimentary basins (on the largest scales), it is currently uncertain if either model is capable of: (1) accounting for the transport of iron derived from basin-fill sediments, (2) explaining the location of basin margin ore deposits, or (3) generating localised transient thermal anomalies required to rationalize apparent differences between high fluid inclusion temperatures and lower ambient crustal temperatures on the Cumbrian margin. The topography driven flow model suggests that fluids should be directed from zones of recharge located at basin margins towards the centre of the East Irish Sea basin, with any thermal anomalies generated as fluids migrate from depth towards surface discharge points. Conversely any, relatively slow moving, stable, free convective cells would tend to be confined to thick, high permeability, aquifers within the deep basin-fill sedimentary succession, with minimal potential for transport to sites of mineralization located at basin margins.

However, if ore-forming fluids were instead channelled into relatively narrow, interconnected, high permeability zones such as active dilational brittle faults and fractures that link deep, higher temperature, basinal systems to shallow-buried basin margins then, providing that fluid volumes and flow velocities were of sufficient magnitude (Figure 2), the required mass and heat could be transported from source to sites of mineralization on suitable length and time scales. Such a scenario is provided by episodic venting of deep basinal overpressures along active basin margin fault systems.

Development of overpressures – where pore fluids support some of the stress imparted by an accumulating sediment load – in sedimentary basins is particularly common during periods of rapid subsidence and accumulation of thick packages of fine-grained, low permeability sediments (possibly accentuated by hydrocarbon generation and migration). Within the East Irish Sea basin such conditions are likely to have occurred during the Middle/Late Triassic when between 2 and 3 km of fine-grained, low permeability Mercia Mudstone Group sediments were deposited within the Keys and Tynwald basins (Figure 1) over a time interval of approximately 50 Ma. When the build-up of fluid overpressure exceeds the mechanical strength of the host aquifer excess pressure will be released leading to the

rapid dewatering of the overpressured unit and the migration of fluids, generally, towards lower pressure environments. Overpressured compartments may be particularly prone to failure where tectonically active and mechanically weak faults intersect the overpressured aquifer. Basin-bounding faults that connect deep overpressured units to near-surface hydrostatically pressured aquifers can then act as efficient pressure relief valves leading to rapid upward fluid discharge along transient permeability developed within the fault zone (Sibson 1994; Garven *et al* 2001). Cycles of overpressure, fault zone failure and resealing should lead to the repeated vertical and highly focused discharge of potential ore-forming fluids from sources located in deep basins to shallow basin margin sites of mineralization *via* highly permeable fracture systems.

This concept may be taken a stage further by reference to an example of hydrogeological and thermal modelling of transient fluid flow in response to venting of overpressure triggered by displacement on basin-bounding faults (Boles *et al* 2004). Simulations indicate that high fluid flow velocities (>2000 m yr^{-1}) can be attained within idealised model fracture systems on a time scale of tens of years before decaying towards background rates over a period of a few hundred years. Modelled temperature anomalies associated with venting events follow similar trajectories with peak temperatures exceeding initial (background) values by 80°C before returning to ambient values over time intervals on the order of 10^3 years.

Clearly model calculations are unlikely to represent real geological complexity, but the results suggest that a relatively small number of fluid flow events driven by venting of overpressure along different segments of active fault systems, at differing times during active basin extension, could account for numerous individual ore bodies scattered along the length of the basin margin. In the context of the East Iris Sea basin the Sherwood Sandstone Group provides an ideal permeable aquifer capable of transferring fluids from zones of overpressure towards basin margins. If we assume that ore-forming fluids contained high (>10^3 mg l^{-1}) concentrations of dissolved iron then we are no longer compelled to invoke transport models that are dependent on the transmission of large volumes of fluid. Figure 2 shows that only(!) 2 km^3 (2 x 10^{12} litres) of ore-forming fluid (containing 5000 mg l^{-1} iron) are required to produce a large hematite ore body presuming that all dissolved iron is precipitated within a confined zone of mineralization. While this amount of fluid still non-trivial, such volumes of fluid can plausibly be released from a large sedimentary basin over short (10^7 years) time intervals provided that fluids are released episodically and are highly focussed. If fluid flow was suitably focussed along major basin-bounding faults, such as the Lake District Boundary fault (Akhurst *et al* 1998) and

related, offshore, fault systems (Jackson *et al* 1995), then sufficient volumes of iron-rich fluid could be transported from the East Irish Sea basin to produce basin margin ore bodies and localized temperature anomalies.

Despite a more sophisticated approach to the transport problem, the proposed model for the transfer of ore-forming fluids is still descriptive, qualitative and speculative. The episodic venting of overpressure is, however, compatible with some basic features of mass transport expressed in the R&D model, although it is no longer necessary to invoke a possible Palaeogene heating event to drive fluid flow. Unfortunately, many questions remain; not least of these is the precise geotectonic environment of the East Irish Sea basin at the time of mineralization, since the geotectonic setting will dictate the precise fluid flow regime throughout the entire basin system. A successful quantitative resolution of these issues can really only be achieved by integrating numerical fluid flow models and geological data within the correct temporal framework.

Hematite mineralization and ore deposit formation

Having transported a potential ore-forming fluid to the site of mineralization we are now required to concentrate and precipitate dissolved iron as hematite ore bodies. The stability of hematite is essentially constrained by the oxidation state and pH of the geochemical environment (Figure 5) and solubility of iron can be further reduced by decreasing fluid salinity and temperature (e.g. Heinrich and Seward 1990); effectively the inverse of conditions required to generate ore-forming fluids. If iron is transported in its ferrous (Fe^{2+}) state then dissolved iron must be oxidized to ferric (Fe^{3+}) iron to form hematite by introducing a suitable oxidizing agent (electron acceptor) in amounts sufficient to produce a 250 million tonnes ore deposit. Importantly the transformation of ferrous (Fe^{2+}) to ferric (Fe^{3+}) iron is accompanied by a substantial decrease in iron solubility at near-neutral pH (Figure 5) and oxidation provides a simple geochemical mechanism for inducing hematite precipitation.

Although a variety of inorganic oxidants (e.g. SO_4^{2-}) could be proposed to facilitate iron oxidation, the only plausible oxidizing agent freely available under geochemical conditions required to account for the location and distribution of hematite ore bodies is dissolved oxygen (EQN 3).

EQN 3. $2Fe^{2+} + \frac{1}{2}O_{2(aq)} + 2H_2O \leftrightarrow Fe_2O_3$ (hematite) $+ 4H^+$

Providing that oxygen is continuously supplied to maintain oxidizing geochemical conditions then iron will be precipitated as hematite, with little dissolved iron remaining in the effluent fluid. In effect, iron will be titrated quantitatively from solution in the presence of excess oxygen to form

hematite. Geochemical requirements imply that a separate, oxidizing, fluid must be involved in the mineralization process. Interaction (mixing) of two fluids – one carrying ferrous (Fe^{2+}) iron and a second carrying dissolved oxygen – offers a persuasive mechanism to drive hematite precipitation.Such a dynamic mineralizing environment is possible if oxygenated, surface-derived, meteoric ground waters were transported into the deeper subsurface, perhaps via Early Triassic sandstone aquifers, and able to interact with reduced, iron-bearing, basin-derived fluids. A schematic of the proposed ore-forming model is illustrated in Figure 8.

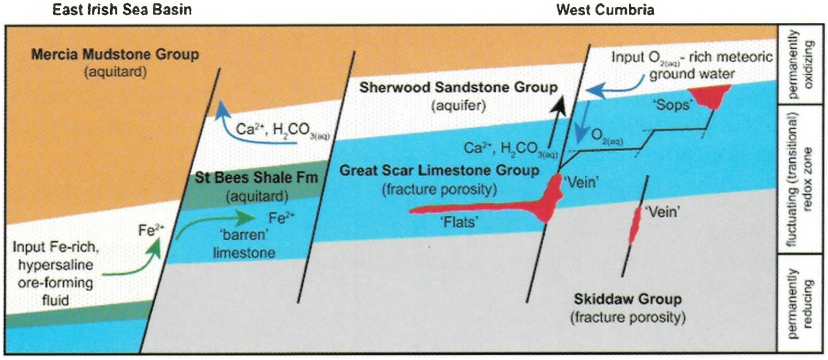


Figure 8. Schematic cartoon illustrating the variables controlling location, distribution and style of hematite ore deposit formation on the Cumbrian margin. Iron-rich, hypersaline fluids (green arrows), transported from the East Irish Sea basin migrate towards the Cumbrian margin via basin-bounding faults and interact with oxygenated, meteoric groundwater (blue arrows) transported towards the basin centre from recharge zones located on sub-aerially exposed basin margin topographic highs. The three ore body (highlighted in red) geometries ('flats', 'veins', 'sops') are shown.

Importantly, this model may explain the apparent relationship between the distribution of hematite ore bodies and the presence (or absence) of low permeability (aquitards) units of fine-grained Pennsylvanian fluvial sediments and Permian St Bees Shale Formation reported by Dunham (1984). Where Early Triassic sandstone aquifer units are in hydrological contact with fractured and possibly karstified Mississippian limestones (or fractured Skiddaw Group, Eskdale granite) then oxidizing conditions can be maintained by the transmission of meteoric ground water. Conversely, where Early Triassic Sandstone aquifers are separated from Mississippian limestones by thick, fine-grained successions then a continuous supply of dissolved oxygen might no longer be possible and hematite precipitation may be severely inhibited or prohibited entirely.

Oxidation of ferrous (Fe^{2+}) iron not only results in efficient precipitation of hematite, but also generates acidity (see EQN. 3) and the potential for wall rock dissolution and replacement. This is particularly important for major limestone-hosted ore deposits of west and south Cumbria because the precipitation mechanism provides a self-sustaining geochemical system that drives carbonate dissolution, propagates space to accommodate the growing ore body and buffers acidity to maintain geochemical conditions within the hematite stability field (Figure 5). EQN 4 describes the simple limestone dissolution reaction.

EQN 4a. $CaCO_3$ (calcite) + $2H^+ \leftrightarrow Ca^{2+} + H_2CO_{3(aq)}$

EQN 4b. $H_2CO_{3(aq)} \leftrightarrow H_2O + CO_{2(g)}$

By combining EQN 3 and EQN 4 and using molar volumes for solid reactants and products it is possible to show (EQN 5) that mineralization generates significant negative volume change (porosity) while maintaining hematite stability.

EQN 5. $2Fe^{2+} + \frac{1}{2}O_{2(aq)} + 2CaCO_3$ (calcite) $\leftrightarrow Fe_2O_3$ (hematite) $+ 2Ca^{2+} + 2CO_{2(g)}$

2 moles of calcite (molar volume = 36.9 $cm^3\ mol^{-1}$) are replaced by 1 mole of hematite (molar volume = 30.3 $cm^3\ mol^{-1}$) giving a nett volume change (Δ_V) of - 58.9%.

Volume change associated with mineralization of carbonate host rocks further accounts for the development of cavernous porosity required to explain the occurrence of the large hematite botryoids ('kidney' ore) that characterize limestone-hosted ore bodies. This process, sometimes termed hydrothermal karstification, may also explain the formation of karst-like 'sop' deposits in south Cumbria where reduced, ore-forming fluids appear to have been transported into close proximity with potential oxygen-carrying aquifer sandstones (Figure 8). Under these conditions the solubility of ferrous (Fe^{2+}) iron should decrease sharply in response to the combined effects of oxidation and cooling of ore-forming fluids, leading to fast precipitation of fine-grained hematite, a rapid decrease in fluid pH, accelerated rates of limestone dissolution, and efficient removal of the reaction products (Ca^{2+}, HCO^-_3) *via* the overlying sandstone aquifer system.

The proposed deep penetration (assuming an ore deposit formation depth of >1.0 km) of oxygen-bearing meteoric ground water also has potentially important implications for the hydrogeology and timing of the mineralization process because it is generally assumed that dissolved oxygen (~12 mg l^{-1}) entering the recharge zone of ground water systems is rapidly depleted by microbial respiration and decomposition of organic matter. Once isolated below the prevailing water table the oxygen content of ground water is further reduced by oxidation of reduced inorganic species

and organic carbon within the saturated aquifer and progressive outgassing of dissolved oxygen (with increasing temperature) until reducing conditions are established (< 0.5 mg l^{-1} $O_{2(g)}$). Consequently it might be anticipated that free oxygen could not be transported into the deep subsurface in amounts sufficient to maintain an oxidizing geochemical environment at the site of mineralization.

Studies of modern aquifers in hot, arid and semi-arid environments (Winograd and Robertson 1982) report measured dissolved oxygen concentrations of between 2 mg l^{-1} and 8 mg l^{-1} to depths of 1000 m (with water table depths of 200 m to 660 m) and water temperatures to 60°C. These observations show that elevated dissolved oxygen concentrations may be sustained where the mineralogical framework of the aquifer consists of redox sensitive species in their highest oxidation states and contains negligible amounts of organic matter. Red, St Bees Sandstone Formation aquifers, in particular, should have little or no capacity to consume dissolved oxygen in meteoric ground water transported into the deep subsurface because the sandstones, originally deposited strongly oxidizing, terrestrial environments, are unlikely to contain any reactive reducing agents.

An ore deposit model that involves the deep circulation of oxygenated meteoric ground water would appear to preclude a Jurassic or Cretaceous age for mineralization because the warm, humid, mid-latitude climatic conditions associated with UK Jurassic and Cretaceous terrestrial environments should result in rapid deoxygenation of meteoric ground water close to the point recharge, with little potential for the transport of dissolved oxygen into the deep subsurface. Further, much of the missing Jurassic and Cretaceous sedimentary record covering Cumbria was probably dominated by marine deposits, thus limiting the potential for topography driven meteoric recharge of deep aquifers. In contrast, hot, arid, desert environmental conditions, coupled with a sub-aerial rift margin topography and low water tables, during the Triassic would provide an ideal setting for the deep circulation of oxygen-rich ground water. This offers indirect geochemical support for a Middle/Late Triassic ore-forming event.

Although oxidation reactions provide a first order control on hematite precipitation, it is evident from the occurrence of coarsely crystalline quartz gangue that fluid cooling must also contribute to the mineralization process. Under assumed conditions associated with the generation of iron-rich, ore-forming fluids the solubility of quartz is predominantly a prograde function of temperature (Rimstidt 1997). Estimates of the likely silica content of ore-forming fluids in equilibrium with quartz yields concentrations of ~ 130 mg l^{-1} (at 140°C). Because the solubility gradient for quartz over the temperature

range defined by fluid inclusion data (140 to 80°C) is relatively low (~1.5 mg $°C^{-1}$) the fluid must undergo an appreciable temperature drop across the zone of mineralization in order for quartz to be precipitated in significant amounts. Assuming a temperature drop of 60°C (from 140 to 80°C) suggests that ~90 mg of quartz could be precipitated from each litre of fluid.

In summary, the location, distribution, style and mineralogy of hematite ore deposits can be explained by the interaction of ferrous (Fe^{2+}) iron-rich, ore-forming fluids with oxygen-bearing meteoric ground water to generate corrosive fluids capable of precipitating hematite and simultaneously causing host rock dissolution (limestones) and wall-rock alteration (fractured metasediments and granite). This simple, coupled geochemical mechanism is particularly important for the formation of the major limestone-hosted ore deposits because it can produce porosity, in otherwise tight limestones, necessary to accommodate the massive, metasomatic ore bodies characteristic of west and south Cumbria.

A new and improved ore deposit model

On the basis of the preceding discussion it is now possible to present a revised version of the R&D model. Figure 9 illustrates key aspects of the model using a highly stylized East Irish Sea basin cross section restored to a Middle/Late Triassic (Mercia Mudstone Group) datum. The model proposes that hydrocarbons generated from organic-rich, mid-Carboniferous (Namurian) marine source rocks during rapid Triassic subsidence migrated into adjacent Permo-Triassic continental red bed sandstones and reacted, in the presence of hypersaline brines, with grain-coating hematite at temperatures of >100°C to produce ferrous (Fe^{2+}) iron-rich, ore-forming fluids containing elevated concentrations (10^3 and 10^4mg l^{-1}) of dissolved iron. These, possibly overpressured, mineralizing fluids were transported up-dip, *via* Permo-Triassic aquifer sandstones, from the central zone of the basin (Keys and Tynwald sub-basins) towards basin margins along prevailing pressure gradients. Tectonically active basin margin faults systems may have acted as conduits for the rapid, periodic, transmission of mineralizing fluids from the deep subsurface to the Cumbrian margin by venting of overpressure (seismic pumping). Hematite precipitation occurred when fluids contacted oxidizing, basin margin, geochemical environments, with major carbonate-hosted ore deposits being formed where oxidation of ferrous (Fe^{2+}) iron resulted in extensive hydrothermal karstification of limestone host rocks and efficient precipitation of iron from solution.

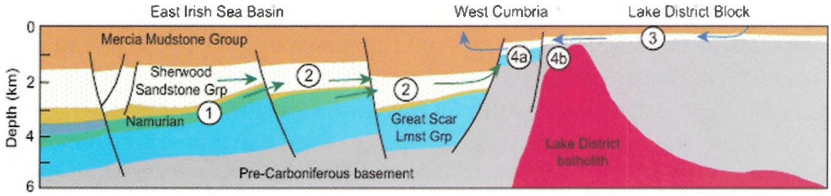


Figure 9. Highly simplified and stylized cross-section of the East Irish Sea basin and Cumbrian margin (after Jackson *et al* 1995; Akhurst *et al* 1997) showing genetic elements of the revised Rose and Dunham (1977) ore deposit model for hematite mineralization on the west Cumbrian platform. 1 – Organic-rich, oil prone, mid-Carboniferous (Namurian) marine source rocks generate oil. 2 – Oil migrates into overlying and laterally contiguous porous, red, Sherwood Sandstone Group aquifers, mix with pre-existing brines and react with grain-coating hematite to produce hypersaline, ferrous (Fe^{2+}) iron-rich, ore-forming fluids that migrate along pressure gradients towards basin margins. 3 – Topography driven meteoric groundwater circulating through St Bees Sandstone Formation aquifers transports dissolved oxygen into the deep subsurface. 4 – Ferrous (Fe^{2+})-rich fluids migrate vertically along seismically active basin margin faults where they interact with oxidizing groundwater to produce: (a) metasomatic and vein-fill hematite ore bodies hosted by Great Scar Limestone Group marine carbonates, and (b) vein-fill hematite ore bodies in fractured Skiddaw Group metasediments and Eskdale granite. The cross-section has been restored to a Middle/Upper Triassic datum (Mercia Mudstone Group) assuming that this represents the peak age of mineral emplacement.

The scale, style and distribution of mineralization reflects the combined effects of the interaction of iron-bearing, basin-derived fluids and oxidizing, meteoric ground water circulating through overlying Permo-Triassic aquifers. Where suitable redox interfaces were maintained mineralization was well developed. In contrast, where potential host limestones were isolated from oxidizing ground water aquifers by fine-grained aquitards (e.g. St Bees Shale Formation) the potential for mineralization was severely impeded. 'Sop' deposits (south Cumbria) are proposed to represent conditions where ore-forming fluids closely approached meteoric aquifer systems resulting in rapid iron oxidation, vigorous hydrothermal karstification of the host limestone, and hematite precipitation.

Time is of the essence

It is evident from the previous discussion that a succession of geological and geochemical processes must operate within a strict and definable temporal framework in order for model hypotheses to be consistent with observation. This emphasises the key role played by the precise timing of mineralization

in the development of any successful ore deposit model. For example, if the reduction of ferric (Fe^{3+}) iron, present as grain-coating hematite in Permo-Triassic sandstones, in contact with petroleum, is a prerequisite for the generation of iron-rich, ore-forming fluids then the timing of hydrocarbon maturation and migration from oil-prone source rocks must pre-date mineralization. If the timing of ore deposit formation was shown to pre-date hydrocarbon generation then this component of the model would immediately fail.

Many other elements of the revised R&D model could similarly be tested, but perhaps the most important time-dependent constraint is the link between the broader-scale geotectonic setting and the impact of crustal scale tectonics on palaeohydrogeology and mass and heat transport. Although we can be reasonably confident in reconstructions of the geotectonic evolution of the British Isles during much of the Phanerozoic, this vital background information is of minimal value if we cannot use the information to place limiting conditions on the tectonic setting at the time of mineralization.

Ultimately it is difficult to downplay the importance of the timing issue. Despite the current absence of indisputable dating evidence, there are several indirect lines of evidence that might be taken to support a mineralizing event or series of mineralizing events during the Middle Triassic to Late Triassic (~250 to 200 Ma). In the context of the revised R&D model the likely timing of hydrocarbon generation within the East Irish Sea basin, the migration of overpressured ore-forming fluids and the transport of mass and heat to basin margins by rapid expulsion along tectonically active basin-bounding faults, and the development of deep, oxygenated, meteoric ground water systems are all indicative of mineralization occurring sometime during the Middle/Late Triassic.

As we will see in the following section on iron isotope data a potentially very different ore deposit model may be inferred from the interpretation of iron isotope systematics. This currently sketchily developed model would differ fundamentally from that proposed by Rose and Dunham (1977), but it will also inevitably be dependent on temporal constraints. The considerable uncertainty associated with existing age estimates and the development of an accurate ore deposit model that accounts for all components of the ore-forming process highlights the importance of reliable dating. Accurate dating is evidently a vital component of any successful ore deposit model.

Iron isotopes: an alternative perspective?

Using the partitioning (fractionation) behaviour of iron isotopes ($^{56}Fe/^{54}Fe$) as a tracer of geochemical processes during the dissolution, transport and

precipitation of iron, Chapman *et al* (2006a,b) argued that the iron contained within Cumbrian hematite ore deposits was derived by 'leaching' of common iron-bearing minerals (amphibole, biotite, magnetite) present in granitic source rocks. In addition, they proposed that the precipitation of hematite from mineralizing fluids did not involve a change in oxidation state implying that dissolved iron must be transported as ferric (Fe^{3+}) iron. These key conclusions have important implications because the revised R&D model is not consistent with a granitic source for iron and does not predict the transport of iron in its higher oxidation state. If this interpretation is correct then the revised R&D model must be rejected.

The problem with a reliance on iron isotope data is that iron isotope fractionation in hydrothermal environments can be complex. Iron isotope ratios may be influenced by a range of contributory variables related to the source of iron, temperature of the ore-forming fluid, oxidation state and speciation of iron in solution, exchange of iron between fluids and secondary iron phases during transport, changes in oxidation state during mineral precipitation, and mineral precipitation kinetics (Markl *et al* 2006; Saunier *et al* 2011). These variables can cause both positive and negative fractionations that lead to nett isotopic effects that prohibit simplistic interpretation.

A detailed examination of the iron isotope evidence is beyond the scope of this current work, particularly where the data in question have not been fully reported in a peer reviewed publication, but a short appraisal of the isotopic data is presented below. It is also possible to examine briefly some limitations of a potential ore deposit model based on the interpretation of iron isotope data as presented by Chapman *et al* (2006a, b).

Chapman *et al* (2006a, b) report $^{56}Fe/^{54}Fe$ ratios measured from samples collected from the Florence mine (Egremont, west Cumbria) as delta (δ) values relative to the IRMM-014 international reference material ($10^3\delta^{56}Fe_{IRMM\text{-}014}$). Data obtained from samples of ore deposit hematite vary between -1.65 and +0.46. Focusing on samples of 'early', massive, metasomatic hematite, Chapman *et al* (2006a, b) noted that samples recovered from locations close to major fault (feeder) structures are characterized by relatively low isotopic compositions (-0.67 to -0.24) while samples collected from more distal sites yield higher values (+0.24 to +0.37). Chapman *et al* (2006a, b) argued that the change in isotopic composition, from proximal to distal locations, resulted from precipitation-controlled kinetic isotope fractionations (Rayleigh fractionation) – that is, hematite preferentially incorporates ^{54}Fe during precipitation leaving an evolving, residual dissolved iron reservoir that is increasingly enriched in

^{56}Fe as iron is removed progressively from the starting fluid. Under these conditions the earliest formed hematite will have an iron isotope ratio that is lower than the starting fluid while later formed hematite, precipitating from the modified fluid, will be characterized by successively higher iron isotope values. Void-filling, botryoidal and specular hematite exhibit a similar pattern of iron isotope fractionation with proximity to fault structures, but the isotopic compositions of these phases are displaced to lower values by ~ -0.6. This behaviour is attributed to secondary dissolution of 'early', metasomatic hematite and subsequent reprecipitation of dissolved iron as neoformed hematite deposited passively into secondary, vuggy porosity.

Chapman *et al* (2006 a, b) further compared the iron isotopic composition of hematite with two potential iron sources – Permo-Triassic sandstone ($10^3\delta^{56}Fe$ = +0.18 to +0.31) and (?Lake District) 'basement granite' ($10^3\delta^{56}Fe$ = +0.37 to +0.49). They concluded that the iron isotope composition of metasomatic hematite is best explained by iron leached from a granitic source and suggested that there is no evidence for biological or inorganic redox induced isotopic fractionations during hematite precipitation implying that iron was transported as ferric (Fe^{3+}) iron.

Although Shepherd (1973) had previously proposed that Lake District granites may have acted as a source of iron for Cumbrian mineralization based on the trace element composition (notably the occurrence of anomalous concentrations of arsenic) of hematite ores, Shepherd and Goldring (1993) subsequently rejected a granitic source and concluded that the available evidence supported dissolution of grain-coating hematite present in Permo-Triassic sandstones as the most likely source of iron. If Lake District granites did act as a major source of metals then one might expect to see evidence of hydrous alteration of granite associated with the iron 'leaching' process. Not only would fluids need to react, possibly selectively, with iron-bearing minerals, but it would also be necessary for these fluids to circulate freely through largely impermeable granite in order to dissolve sufficient iron. Despite evidence for fluid migration through fractured Eskdale granite in the form of fracture-hosted hematite mineralization (Young 1985), it is questionable whether sufficient permeability could exist to allow extraction of iron from minerals that are dispersed throughout the granite, occupy a small proportion of the total rock volume and have relatively small surface areas.

In the section (above) on the aqueous geochemistry of iron it was argued that iron is only likely to be soluble to any significant extent as ferrous (Fe^{2+}) iron under geochemical conditions associated with LTEBM mineralization.

Ferric (Fe^{3+}) iron can be soluble, particularly as chloride complexes, but this is only possible in strongly acidic (pH<3) environments. (Strongly acidic conditions are unlikely to be maintained for any length of time where fluid pH is buffered by reaction with common aluminosilicate minerals assemblages.) In all other oxidizing, low temperature environments iron is effectively insoluble (Figure 5) and limited by the stability of hematite. There is currently no evidence to support the transport of iron in its ferric (Fe^{3+}) state, apart from the selective interpretation of iron isotope data, and it is presently unclear how the evolution of suitable geochemical conditions could be accommodated within a model that is consistent with other available geological data.

In addition to these simple geochemical objections, it is worth briefly considering if the reported iron isotope data could plausibly be incorporated within the revised R&D model. In an iron isotope study of the formation of iron oxide concretions (Navajo Sandstone, Utah) precipitated from iron derived from the dissolution of hematite present in continental red bed sandstones and transported to the site of mineralization as ferrous (Fe^{2+}) iron, Busigny and Dauphas (2007) observed that the measured isotopic composition of the concretions was much lower than expected based on isotope fractionations associated with the initial dissolution and reduction of hematite originally dispersed in sandstones and the subsequent oxidation of ferrous (Fe^{2+}) iron during concretion growth. They argued that this discrepancy could be explained by the selective adsorption or extraction (by precipitation of new phases) of the heavy isotope of iron from solution during transport, leading to the formation of iron oxide concretions with low iron isotope values. Long distance transport (solute transport distances on the scale of tens of kilometres are envisaged in the revised R&D model), coupled with dissolution and precipitation of a variety of iron-bearing phases (particularly chlorite and Fe-rich carbonates), might realistically account for observed iron isotope values while still allowing Permo-Triassic red beds to act as a source of iron.

Although it is evident that iron isotope studies have the potential to act as powerful tools for tracing a range of geochemical processes, the unambiguous interpretation of iron isotope data in natural hydrothermal systems requires a thorough understanding of all contributions to the partitioning of iron during the evolution of ore-forming fluids. Clearly a more detailed evaluation of iron isotope systematics across the entire Cumbrian hematite ore field is required in order to assess the true implications of the isotopic data. Importantly, iron isotope data cannot currently be shown conclusively to falsify the revised R&D model.

The future?

Flooding of the Florence mine (Egremont, west Cumbria) in 2007 severed our last direct connection to the hematite ore deposits of Cumbria, perhaps forever. To quote Sir Kinsley Dunham (1984): "before long no living geologist with first-hand personal knowledge of the underground detail of these remarkable deposits will remain." That time is close at hand. As a consequence, any future attempts to advance our understanding of ore deposit formation will be dependent entirely on archive (museum) collections or the recovery of mine dump material. This places severe limitations on any new investigations because the context of archive and mine dump samples is seldom recorded. In addition, material not normally considered important by mineral collectors such as altered wall rocks cannot now be sampled.

Despite these limitations it is evident that new insights into ore-forming processes could be obtained using modern geochemical and isotopic analytical techniques coupled with computer-generated reactive transport hydrogeochemical simulations. The ultimate intention would be to produce a fully integrated, quantitative description of the mineralization that accounts for the source, transport and precipitation of iron within the context of the geotectonic evolution of the East Irish Sea basin. Key objectives include: (1) accurate timing of ore deposit formation using radiometric dating techniques such as Sm-Nd (fluorite), (U-Th)/He (hematite), (U-Th)/Ne (hematite), U-Pb (calcite), (2) reconstruction of the post-Devonian subsidence and thermal history of the East Irish Sea basin region and adjacent areas, (3) characterization of ore forming processes based on trace element, stable isotope (carbon, oxygen, iron) and radiogenic isotope (strontium, neodymium) data obtained from ore minerals, gangue and associated wall rocks, (4) determination of ore fluid geochemistry from the chemical analysis of fluid inclusions, and (5) modelling of the interaction between geochemical and hydrological processes at the broadest scale using hydrological and reactive transport computer codes.

An investigation of this scope would inevitably represent a significant undertaking in terms of time, intellectual and financial resources, but an examination of all components are necessary if the tenets of the revised Rose and Dunham (1977) model are to be tested. Unfortunately we are unlikely to see any significant progress along these lines in the near or possibly even distant future.

Conclusions

Although numerous challenges remain, the baseline model proposed by Rose and Dunham (1977) to account for the origin of hematite mineral-

ization in Cumbria, and potentially other locations at the margins of the East Irish Sea basin, contains genetic elements and makes some simple predictions that are supported by observation. By applying simple geological, geochemical and hydrogeological principles it has been possible to scrutinize the model, derive new insights and suggest modifications that allow a better understanding of ore-forming processes.

Clearly new studies, using modern analytical and numerical modelling techniques, are needed in order to test specific components of the revised R&D model and refine the details of ore deposit formation. The most critical requirement remains the acquisition of accurate and precise dating evidence for the timing of ore deposit formation. Once the age of mineralization is defined then many other aspects of the ore deposit model can be evaluated within the context of the geotectonic evolution of the East Irish Sea basin and associated basin margins that host the mineralization.

Many of the challenges encountered in the modelling of Cumbrian hematite mineralization find parallels with the general problem of mass and heat transport during the evolution of extensional sedimentary basins. In this respect hematite mineralization at the margins of the East Irish Sea basin deserves further investigation because a quantitative solution to this problem should contribute to a better understanding of mass and heat transport in sedimentary basins worldwide.

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HOW TOUGH ARE THE LAKE DISTRICT'S ACCRETIONARY LAPILLI?

Clive Boulter

In a general conversation about pyroclastic processes, if you mention that volcanic hailstones/accretionary lapilli can be very tough objects, reactions can be vehement as if you have challenged long held articles of faith about their soft nature. Similar responses can be provoked if you mention that silica-rich magmas can be very fluid or that convection is not the direct driving force of plate tectonics. There is a significant body of evidence to show that many volcanic hailstones are remarkably resilient yet this characteristic is not widely acknowledged. The sentiments expressed by Németh and Martin (2007) typify the views of many researchers when they concluded that volcanic hailstones ".... hardly survive long distance travel, unless the motion is gentle enough, and subsequent deposition is rapid enough, to preserve them against mechanical shear." This viewpoint of a soft/weak nature was engrained in the minds of many workers by a series of early publications such as Moore and Peck (1962) who stated that "Long immersion in water, such as if lapilli fell into the ocean, would surely result in disintegration of the lapilli, held together as they are by water". Early ideas on the topic were strongly influenced by perceptions of the environments and processes that lead to ash particles sticking together. Apart from now discounted processes from 19th Century publications involving raindrops landing on ash, much of the 20th Century published work emphasised formation in Plinian eruption columns followed by fallout from umbrella regions. Towards the latter part of the 20th Century a roll for pyroclastic density current processes entered the discussions (e.g. Schumaker and Schmincke, 1995) and this has helped to change perceptions about the physical attributes of some volcanic hailstones. More recently, careful analysis of repeated distribution patterns of these objects in ignimbrites (Brown *et al.*, 2009) has shown that an understanding of the conditions of their formation can reinforce ideas about their robust nature.

Volcanic hailstones are a widespread, but minor, component of many volcanic successions from the Archaean to the present. The many researchers who believe they are fragile take the next logical step to say that their presence indicates a totally subaerial setting [origin+transport-+deposition]. If volcanic hailstones can be shown to have survived aqueous transport their putative singularity as an environmental indicator has to be discarded. The production of volcanic hailstones typically indicates either a subaerial or very shallow water eruption. Their subsequent fate depends upon many factors and if these lapilli are robust deposits that contain them

are not automatically to be taken as subaerial. What follows is a sample of some of the publications documenting subaqueous deposition of volcanic hailstones. In the exceptionally well preserved Palaeo-archaean greenstone belts of southern Africa several examples of marine sedimentation of accretionary lapilli have been documented including possible examples of deposition in deep-marine environments (e.g. Stanistreet *et al.* 1981). Other South African examples of aqueous deposition are not so well constrained in terms of water depth but still demonstrate reworking in subaqueous environments (e.g. Lowe and Knauth, 1978). In the Neo-archaean of Western Australia accretionary lapilli have been documented sharing fluvial channels with cobbles of granite and gneiss (Boulter, 1987) and in trough cross-bedded units of a fluvial setting. Bateson (1965) has described accretionary lapilli in a greywacke turbidite succession from the Precambrian of South America. Self and Sparks (1979) commented on two occurrences in lake deposits of waterlain accretionary lapilli in areas of recent volcanic activity.

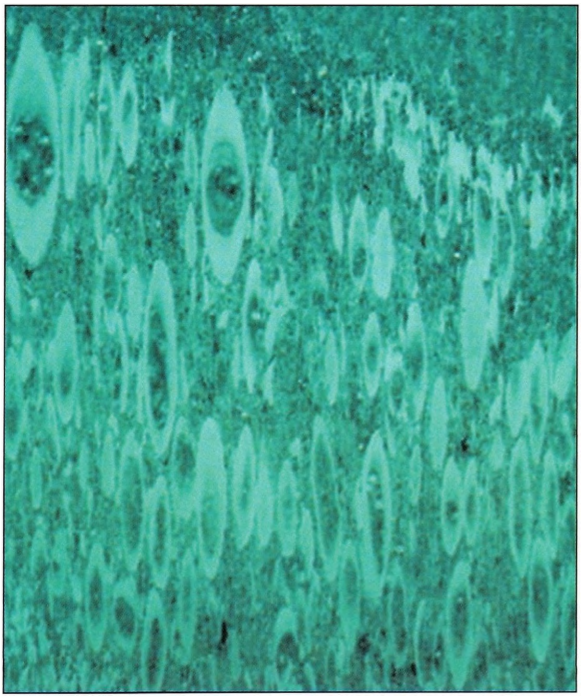


Figure 1. Kentmere: the alignment of the accretionary lapilli defines the slaty cleavage which was superimposed on the layering/stratification. The view is of the highest strain section (XZ). The long axes of the lapilli are mostly between 1 and 1.5cm.

Thus far in this article volcanic hailstone has been used as a term that is readily recognised for an ash aggregate even though it is informal and will not be found in any officially recognised classification scheme. At an even more informal level terms like bird's eye tuff, and even more locally pheasant's eye tuff, are used in the Lake District where these ash aggregates have been distorted (Bell 1981) during the Acadian Orogeny leading to the production of slate (Fig. 1). In journal publications several classification schemes have been proposed essentially involving variations on the theme of coated grains and the latest in this series by Brown *et al.* (2012) may represent a good basis for further discussion but as yet has not been universally adopted. Much of what is termed volcanic hailstone in the Lake District would either be coated pellets or accretionary pellets (formerly accretionary lapilli) in the latest schemes; both sets of objects being cored by ash pellets with no internal structure and coated by very fine-grained ash which in the former has no organisation and in the latter has distinct well-developed concentrically laminations. The removal of the size term lapillus

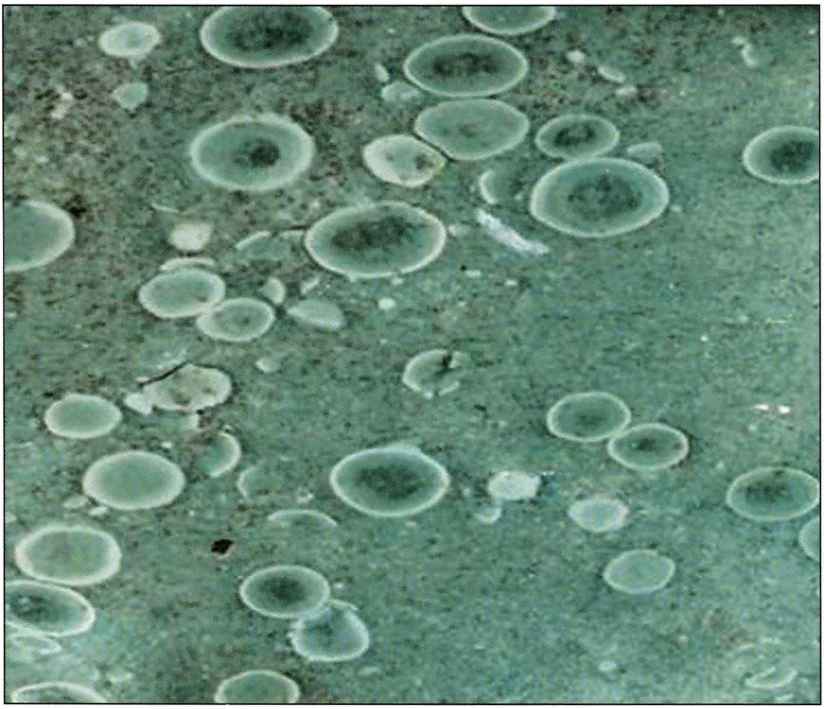


Figure 2. Kentmere: accretionary lapilli as seen on the cleavage plane of a slate sample. The length of the largest lapilli is just over 1cm. Are the concentric laminations in the very fine-grained ash coating sufficiently distinct to place the coated pellets in the Brown *et al.* (2009) accretionary lapilli class?

and use of accretionary pellet for concentric-laminate objects is because not all are in the 2 to 64 mm range reserved for lapilli. Agreement on terms is essential for clear communication but some ambiguity remains because one person's well-developed and distinct laminations in the very fine-grained coatings may be another person's single coating with feint lamination (Fig. 2). Because detailed descriptions are not always provided, in many publications there is uncertainty about which of the Brown *et al.* (2012) categories are involved. What is certain is that the Lake District has large volumes of both coated pellets and accretionary pellets and these have all been referred to as accretionary lapilli in mainstream publications.



Figure 3. A bed of coated pellets in Whorneyside bedded tuff on Stonesty Pike. This eruption was phreatoplinian in nature and the pellets are fallout deposits from the very extensive umbrella cloud. The scale subdivisions are 2cm wide.

The largest occurrence of coated pellets in the Lake District is in the Whorneyside Tuff (Branney 1991) which provides good evidence for the dominant process involved in ash aggregation (Fig. 3). In what may have been one of the largest explosive eruptions of its kind in Earth history, very large volumes (tens km^3) of water gained access to a large andesite magma chamber. In this eruption the moisture in the phreatoplinian eruption column was the main cause of ash aggregation in some combination of moist

adhesion and ice riming. For the very fine-grained coating, electrostatic attraction would have been the main accretion mechanism. Key aspects of the Whorneyside Tuff, in particular its uniform thickness in small erosional rills, show that it was deposited as an airfall tuff despite its poor sorting. However the scale of the subaerial reworking was minimal and hence there is no evidence to demonstrate the resilience or otherwise of these coated pellets formerly referred to as accretionary lapilli.

In a series of recent publications Brown and colleagues (e.g. Brown, *et al.* 2007, 2009) have provided a mechanism for creating very tough volcanic hailstones. The process involves forming ash pellets in a phoenix cloud above a pyroclastic density current (PDC) but a key requirement is that these pellets fall into the pyroclastic current before it wanes significantly or into a closely following PDC. Turbulence in the PDC adds the concentrically laminated very fine-grained ash coating and the heat of the pyroclastic current bakes the whole pellet. In particular the thermal event evaporates any moisture between the ash particles precipitating dissolved chemicals which, for the very fine-grained coating, create a highly lithified accretionary pellet. There is abundant evidence of brittle deformation and very little sign of disaggregation demonstrating the well-lithified nature of the accretionary pellets as they were being deposited in multiple ignimbrite sheets. Brown *et al.* (2009) asked the question "Do accretionary lapilli occur in fallout deposits?" and call into question "the popular assumption that most occurrences of accretionary lapilli occur in fallout deposits". Their discussion concerned accretionary lapilli (sensu stricto) and leaves the possibility of forming coated pellets in Plinian columns and umbrella clouds as observed in the Whorneyside Tuff. It would appear that coated pellets can readily form in phreatoplinian eruption columns/umbrella clouds, but true accretionary lapilli/pellets may not.

In the Borrowdale volcanic sequence true accretionary lapilli are common in stratified pyroclastic rocks that were deposited from dilute pyroclastic currents generated by explosive interaction between silicic magma and external water. These phreatomagmatic eruptions are at the moment best documented for the Scafell caldera succession and only this volcanic assemblage will be considered here. As the caldera evolved, the purely-magmatic eruptions of high density PDCs were punctuated by short-lived episodes of water gaining access to the magma chamber resulting in extreme fragmentation of the magma. Similarly violent eruptions also mark the final stages of the sedimentary infilling of the caldera when silicic magma interacted with groundwater to form the Glaramara tuff (Brown *et al.* 2007). Dilute pyroclastic currents have for a long time been referred to as pyroclastic surges and were regarded as being very different to pyroclastic

flows but the work of Branney and Kokelaar (2002) has provided a conceptual framework that places these processes at points within a pyroclastic current density spectrum. The resulting depositional structures vary along this spectrum with high density PDCs typically creating massive/diffusely stratified units of ignimbrite whereas low density PDCs produce well stratified deposits with common traction current structures such as cross-bedding. The currents at the dense end of the spectrum are from magmatic eruptions and the dilute end tends to be from explosivity involving significant amounts of external water either accessing the magma chamber or magma rising to meet groundwater. In this latter situation the time interval between eruptions is largely dependent upon the rate of groundwater recharge and can be hundreds of years apart. If external water is invading a magma chamber the phreatomagmatic eruptions could be almost continuous. For the Glaramara tuff Brown *et al.* (2007) have inferred that the common accretionary lapilli formed by the same process documented in greater detail on Tenerife (Brown *et al.* 2009), that is, ash pellets being created in ash clouds lofted above ground-hugging pyroclastic currents and very fine-grained ash coatings added when the pellets dropped into the PDCs. This requires that the pulses of eruptive activity were more or less continuous. It is possible that ash pellets formed above one dilute PDC fell into the next pyroclastic current to be coated but, given fall velocities of lapilli sized pellets, the single pulses would have to be separated by intervals of minutes to tens of minutes.

Still restricting the account to the well-documented Scafell caldera, during the explosive eruptions that formed the caldera it was infilled by up to a kilometre of ignimbrite (Airy's Bridge Formation). In this pyroclastic infill there are three very thin, but laterally persistent, accretionary lapilli-bearing horizons. At the base of the Airy's Bridge is the Stonesty Tuff which is typically only between one and two metres thick but which can be traced laterally for 20km. Within the Airy's Bridge, the Cam Spout and Hanging Stone tuffs (Fig. 4) have similar thickness and lateral persistence characteristics. These horizons are believed to mark incursions of water into the magma chamber(s). Concentric laminations are prominent in many of the accretionary lapilli and fragments of the very fine-grained rims are common. Features such as the low-angle cross-stratification support the proposal that these thin units were deposited by dilute pyroclastic currents. The angular rim fragments provide graphic evidence of the lithified state of the accretionary lapilli almost immediately after formation because the fragmentation was a result of high-energy impacts within the turbulent pyroclastic current that provided the very fine-grained ash for the coating. Rim fragments are particularly abundant in an accretionary lapilli-rich horizon (locality 8, Branney 1990) in the Side Pike Complex, a unit which



Figure 4. Stonesty Tuff, Hanging Stone, Seathwaite. Accretionary lapilli with distinct concentric laminations showing the effects of compaction and the angular fragments of rims show they were brittle whilst still in the pyroclastic current. The largest lapillus is 1.3cm long.

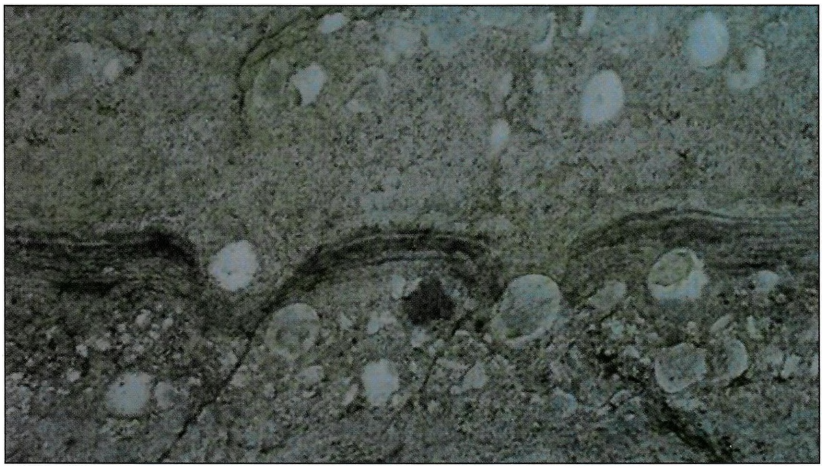


Figure 5. Impact sags created by accretionary lapilli, Wrengill Quarry, Longsleddale. The largest lapillus is 1cm long. This is a view onto the cleavage plane which a low strain orientation and hence the asymmetry of the sags will not have been changed much. Similar structures are seen on Pinch Crags, Scandale Valley near Ambleside and are documented throughout the Glaramara Tuff Member by Brown *et al.* (2007). These are equivalent to bomb sags where the impactor is much larger.

probably represents a lateral collapse deposit during the evolution of the Scafell caldera. This disruption of the stratigraphy during the collapse process means that correlation with the main caldera sequence is uncertain. Further evidence for an advanced state of lithification of Lake District accretionary lapilli is seen in the common impact structures (Fig. 5) they have created in the Glaramara tuff as noted by Brown *et al.* (2007) and seen in correlates on Pinch Crags in the Scandale Valley and in the slate belt along the southern outcrop of the Borrowdale volcanics. In these examples the accretionary lapilli did not disintegrate upon impacting, and penetrating, an admittedly soft substrate. Unequivocal sedimentary reworking of accretionary lapilli in the Borrowdale volcanics is rare; these examples are of ornamental slate in turbidites from the Seathwaite Fell Formation which is a lacustrine succession that infilled the Scafell caldera after the peak of explosive volcanism.

Conclusions

The case for weak volcanic hailstones appears to have been unduly influenced by reports of ash aggregates disintegrating on impact both in modern observed eruptions and rock record examples. The multiple examples of robust properties during transport as they formed, deposition, and reworking, have either been ignored or underplayed. Accretionary pellets with concentric-laminated coatings are very robust, being capable of surviving (1) severe turbulence within pyroclastic currents, (2) impacts as they ballistically indented fine ash, and (3) subaqueous reworking in conditions ranging from fluviatile to deep-marine. As yet the resilience of coated ash pellets without the multiple concentric laminations is not defined. In the Whorneyside Tuff they are found in beds with no signs of soft-state distortion or disaggregation but no reworking has been observed.

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A NOTE ON THE OCCURRENCE AND CHARACTERISTICS OF SMALL-SCALE SORTED PATTERNED GROUND IN TARN-BED GRAVELS AT BECK HEAD, WESTERN LAKE DISTRICT

Peter Wilson

Introduction

The term small-scale sorted patterned ground refers to the more or less regular arrangements of moderately-sized surface clasts in the form of circles, polygons, nets and stripes. Separating these concentrations of clasts are cells or stripes containing smaller and/or fewer clasts and fine-grained material. These features have been widely reported in alpine and subpolar environments and are generally attributed to contemporary ground frost activity, in particular the repeated freezing and thawing of frost-susceptible regolith that leads to frost heave and the sorting of small clasts according to size (Ballantyne, 2013). Examples of similar forms have been recorded at many locations in the uplands of Great Britain (Ballantyne & Harris, 1994). Attention was apparently first drawn to them in the nineteenth century by Clifton Ward who noted, in the northern Lake District, the "…..strange appearance upon many of the fell-tops, or sides, as if stones had been sown along regular lines close together." (Ward, 1876). He regarded them as products of heavy rainfall. Since then instances of small-scale sorted patterns have been found and described from many Lake District locations (Hollingworth, 1934; Hay, 1936, 1937, 1942, 1943; Caine, 1963a, 1963b, 1972; Tufnell, 1969; Warburton, 1985, 1987; Warburton & Caine, 1999; Wilson, 2010) and are considered to result from ground frost activity rather than heavy rainfall.

Warburton (1985) provided a summary of Lake District sorted stripe dimensions based on details provided in earlier reports. Fine stripes are generally within the width range of 8-45 cm, with coarse stripe widths of 5-23 cm. Clasts in fine stripes are usually less than 4 cm in length, whereas clasts in coarse stripes range from 5 cm to 20 cm in length. Larger clasts (up to 98 cm in length) have been reported in coarse stripes but these are unlikely to have undergone frost sorting; rather they have probably remained static as the heave and sorting of smaller clasts occurred around them. Stripe spacing is also variable with fine-to-fine spacing in the range 18-52 cm and coarse-to-coarse spacing of 13-100 cm. Sorted circles are less well documented features but fine-centre diameters of between 12 cm and 60 cm and coarse border widths up to 25 cm have been recorded (Hollingworth, 1934; Caine, 1972; Wilson, 2010).

Although most published details of small-scale sorted patterned ground relate to sites on Skiddaw, Grasmoor and Helvellyn, Caine (1972) reported its occurrence on many other mountains. His map indicates an almost equal distribution of sites between rocks of the Skiddaw Group and those of the Borrowdale Volcanic Group; one site was recorded as being on the contact between the Skiddaw Group and the Ennerdale Granite, in the vicinity of Starling Dodd. In terms of elevation this latter site was the lowest observed (~ 600m OD); the lower limit on the Skiddaw Group was 675 m OD and on the Borrowdale Volcanic Group it was 630m OD. Statistical analysis of site aspect data established that patterns had no significant preferred orientation (Caine, 1972).

Generally, patterned ground locations are devoid of vegetation or, at most, there is a sparse covering. Polygons, circles and nets are normally restricted to gradients of less than 5°, whereas stripes occur on slopes of up to 19° (Warburton and Caine, 1999). A transition from polygons, circles and nets into stripes with increasing slope gradient has frequently been claimed in the wider literature, but Caine (1972) was only able to identify one such occurrence in the Lake District.

To-date small-scale sorted patterned ground has not been reported from gravels on the beds of small upland lakes in Great Britain. In present-day alpine and subpolar regions large-scale patterned ground (polygons in particular) is present on lake beds (e.g. Jennings, 1960; Seppälä, 1972; Ray *et al.,* 1982; Krantz *et al.,* 1988) and is believed to develop where shallow water freezes onto the lake bed. Therefore, if present, submerged patterns are usually restricted to the shallow margins of lakes.

This note reports the occurrence and characteristics of small-scale patterned ground (stripes and circles) at a tarn bed site in the western Lake District. It is believed that this represents the first such setting in Great Britain at which patterns have been identified.

Beckhead Tarn

Beck Head (Grid Reference NY 206107; 625 m OD) is the col separating Great Gable from Kirk Fell in the western fells of the Lake District. There are two small tarns in the col the larger of which is named as Beckhead Tarn. On the southern and eastern margin of the tarn, where water depths are normally less than 10-20 cm when the tarn is full, small-scale patterned ground is present in the gravels of the bed and extends over areas of several square metres (Figure 1). The site occurs on the Birker Fell Formation of the Borrowdale Volcanic Group (British Geological Survey, 1999) and although the gravels are clearly derived from volcanic lithologies they may have been glacially or periglacially emplaced.

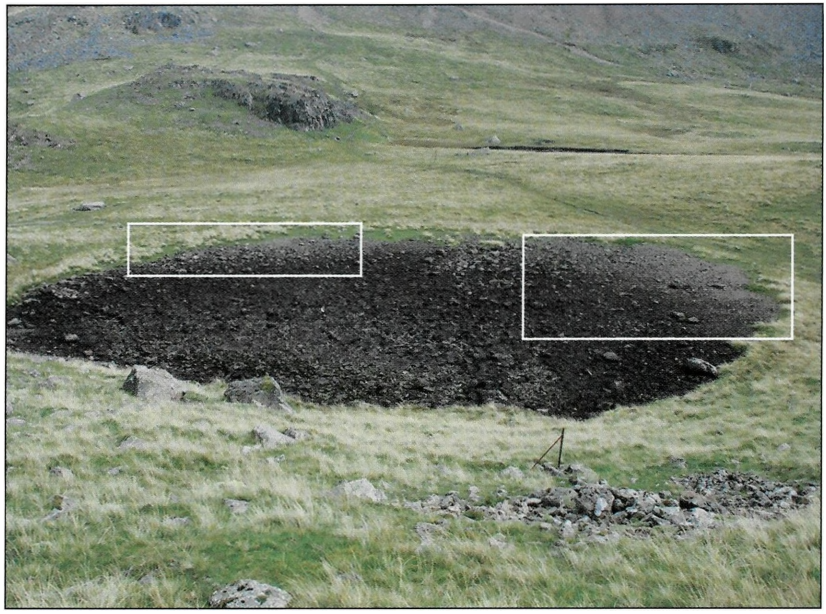


Figure 1. The dry bed of Beckhead Tarn, boxed areas are those with patterned ground.

Pattern characteristics

The areas of tarn bed that display pattern development differ from adjacent non-patterned areas in having clasts of smaller calibre. Very few of the clasts incorporated in the patterns exceed 20cm in length; in contrast, non-patterned areas are dominated by clasts of greater lengths.

Sorted stripes are the principal form of patterned ground in the Beckhead Tarn gravels (Figures 2 and 3). On the south side of the tarn they are present along a 15m length of shoreline, orthogonal to the shore, and can be traced for up to 5m in towards the centre of the tarn. On the east side stripes occupy a 2m length of shore and extend 2m in towards the centre. At both locations surface gradient does not exceed 2°. Coarse stripes range from 25cm to 50cm in width (mean = 36cm, n = 50) with fine stripes between 20cm and 54cm wide (mean = 30cm, n = 50). Coarse stripe clasts range from 6cm to 24cm in length with a mean of 9.1cm (n = 100) and fine stripe clasts range from 1cm to 5cm in length with a mean of 2cm (n = 100). Shallow excavations in coarse and fine stripes showed both to be 12cm deep and underlain by grit with 15-17% organic matter as determined by loss-on-ignition at 430°C.



Figure 2. Sorted stripes at the margin of Beckhead Tarn.
The survey pole is 1m in length.



Figure 3. Sorted stripes at the margin of Beckhead Tarn.
The scale bar divisions are 6cm.



Figure 4. A sorted circle on the bed of Beckhead Tarn.
Scale bar divisions are 6cm.

Sorted circles are also present within the gravels although only three examples were seen (Figure 4). All three were adjacent to places where the sorted stripes terminated towards the centre of the tarn. Circle fine centres range from 17cm to 35cm in diameter with clasts between 2cm and 6cm in length (mean = 3cm, n = 45). Clasts in circle coarse borders range from 6 cm to 30cm in length (mean = 16cm, n = 30).

Discussion

In most respects the dimensions and clast sizes of the Beck Head patterns conform to those given in previous reports of Lake District patterns. One apparent anomaly, however, is that the width of the coarse stripes (25-50cm) at Beck Head exceeds the range (5-23cm) noted for coarse stripes by earlier workers (Warburton, 1985). There is no immediately obvious reason why this should be, but because the coarse stripe dimensions at Beck Head are within the range of those reported from sites elsewhere in Great Britain (Ballantyne & Harris, 1994) it is inferred that they are not sufficiently anomalous to require exceptional formative circumstances.

The development of small-scale sorted patterned ground has been explored both mathematically and through field observation and measurement, and is generally attributed to shallow ground freezing and the associated differential growth of needle ice (Ballantyne, 1996; Matsuoka *et al.,* 2003).

In contrast the origin of large-scale patterned ground has been regarded as a product of either buoyance-driven circulation of water and/or soil (see Ballantyne, 2013 for review of these mechanisms), in addition to differential frost heave.

Thus, given their scale, the Beck Head patterns are probably best considered as a function of frost heave and needle ice growth, and such processes are prevalent throughout the range of Lake District altitude during the winter months (Wilson, 2010). However, what is not completely clear is whether the patterns were created at a time when the tarn margins were covered by water or when they were dry.

Reports of patterned ground on the beds of shallow lakes have led to the suggestion that it is likely to form in submerged settings as a consequence of the free convection of water and/or soil consequent upon the lake waters freezing to the bed in winter and thawing in summer (Ray *et al.,* 1982; Krantz *et al.,* 1988), and McKay & Straughan (1993) have developed a convection-based mathematical model for pattern formation under water. Other examples of lake-bed patterns may have developed when lakes were dry (Jennings, 1960) or when lakes were dry but the water table was coincident with the ground surface such that the substrate remained saturated (Seppälä, 1972). In all cases of lake bed patterns circular or polygonal forms have been observed rather than stripes.

At Beck Head, the small-scale of the patterns and the dominance of stripes over circles suggests that they did not develop in a submerged context. Rather, subaerial frost heave and the differential growth of needle ice probably account for their development. If correct, this implies that the patterns formed when the tarn bed margin was dry. Given that the patterns developed in association with winter frosts, then the tarn bed margin must have been exposed during that season in order to facilitate pattern formation. This in turn indicates that pattern formation is a consequence of the combined effects of low water levels, suitably sized clasts, and frost processes of sufficient magnitude and frequency to sort the clasts into stripes and circles. The presence of organic material below the clasts likely aided pattern development through its ability to retain moisture, after the tarn level had fallen, and act as a moisture source for needle ice growth.

Conclusions

The tarn-bed patterned ground at Beck Head is dimensionally consistent with examples recorded at other Lake District and wider British locations; in this respect it is of limited importance. However, the significance of the patterns lies in the surface context of the site, with these being the first

small-scale patterns to be reported from the gravels of a tarn bed. Their development is related to a combination of local circumstances including the need for the tarn margin to be dry at times when the expectation for water levels would be otherwise. Given the number of shallow tarns in the British mountains with gravel beds the Beck Head patterns are unlikely to be the only examples.

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LANDFORMS, TECTONICS AND MINES OF THE HARTLEY FELL SIDE

Noel Pearson

This paper brings together material prepared for the Society fieldtrip to the Hartley area led by the author on May 17th 2015. It incorporates helpful observations from members taking part in the excursion and additional research material by the author from the subsequent exploration of Hartley Quarry Cave. It is presented in an excursion format enabling anyone wishing to repeat the walk.

Geological and Topographic Setting

Hartley Fell-Side lies due east of Kirkby Stephen in the Upper Eden valley. It rises from the river Eden, at 170m, for 6k, until it reaches Nine Standards at 662m. Those trudging along the 'Coast to Coast Walk' will have their efforts rewarded, as Wainwright remarks, because "the views from the road are good and progress along it is fast". At around 300m those hasty travellers will enjoy a brief respite in the slope, below Fell House, courtesy of the Dent Line fault - a major fault in N England linking the Dent and Pennine faults. Doubtless most will also be unaware of the spectacular scenery left by mining behind Hartley Birkett Hill and the limestone karst under their feet. To reach this point they will have missed the alternative route from Kirkby Stephen via the hidden and fabulous Ewebank Scar. Less assiduous than usual, Wainwright had, in his 1st edition, directed walkers over private land instead (part of the return route) and later issued an apology.

Figure 1 illustrates the essential geo/tectonic context of the walk. The syn-form of the Permo-Triassic outlier between Kirkby Stephen and Nateby lies unconformably over the Great Scar limestone which takes the form of a NE-SW anticline trending in parallel to the Dent Line. The 'Hartley half-dome' plunges to the NE but is also represented in High Out Wood. Mineralisation occurs along the Dent Line fault and particularly when crossed by NW-SE faults at Hartley Birkett Hill. Yoredale-type cyclothems, dipping ever more gently east, underlie the rising ground of the high fell. Completing the description, it is also worth mentioning the renowned Stenkrith Brockram north of Nateby and the Holkerian-Asbian boundary running to the south.

The main contributors to mapping this area include Dakyns *et al* (1891), Turner (1935), Burgess (1965), Dunham and Wilson (1985) and Underhill *et al* (1988). Finally, some amendments were made by BGS in 1996 prior to publication of the 1:50,000 Kirkby Stephen, Solid and Drift Edition Sheet

40. This map is still 'Provisional' and according to the BGS Library likely to remain so. The relevant section is reproduced here as Fig 20.

While there is room then to add some more geological detail, the geomorphology is relatively neglected. Hollingworth in the 1920's and most recently Mitchell, in the 1990's have looked at drainage patterns and the effect of glacial ice on the current landforms, but mostly to the south-west of this area around the Howgills and in the Upper Lune valley to the west. This walk highlights, on a more local level, a comparison between two watercourses as they flow over (and under) karst 'country'. To this end Fig 1 shows the catchment areas of Ladthwaite and Hartley/Birkett Becks and the marked difference between them.

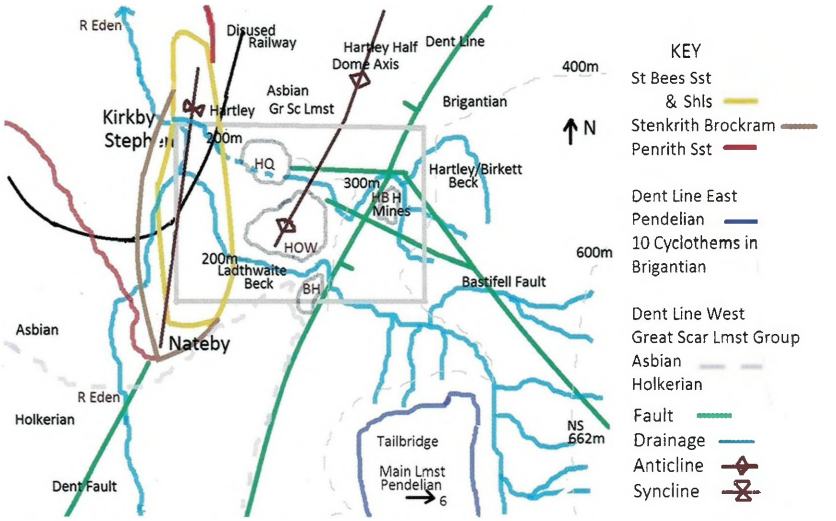
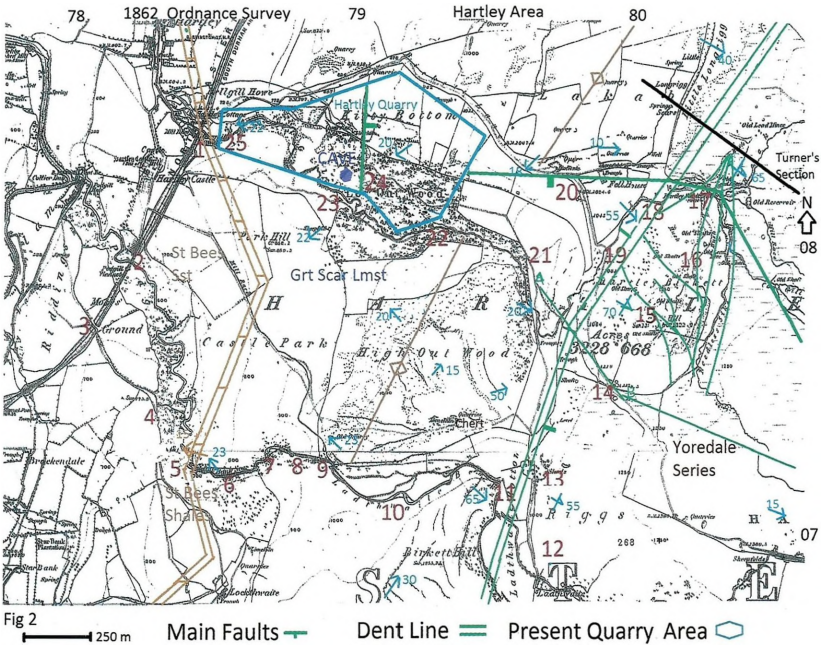


Fig 1 Hartley Area Geological/Tectonic Context ⟵⟶ 500m
HQ Hartley Quarry. HOW High Out Wood. BH Birkett Hill.
HBH Hartley Birkett Hill. NS Nine Standards

Hartley Round Walk – Viaducts

The starting point for the walk is the parking point at NY 784084 outside Hartley Quarry (OS Explorer Map OL19). Figure 2 shows points of interest (numbered in red) and directions. Moving on to Merrygill viaduct (1), completed around 1860, and looking east, the Hartley Beck resurgence from the limestone can be glimpsed through the trees. Looking west, the steep-sided valley carved in the St Bees Sandstone narrows, where the beck bends



to the north following the strike. The stream has also cut a 3m deep channel in the valley floor, which is flat, apart from some 'lumps and bumps' indicating the site of the former Corn Mill (Hartley High Mill). An overgrown channel under the viaduct has previously fed water to the Mill from the resurgence, but it has been supplemented by a supply via a constructed Mill Race from Ladthwaite Beck (it can be seen crossing Castle Park).

Whereas Merrygill viaduct is 366ft long, Podgill viaduct exceeds it by 100ft. Fig 3 shows a wide valley with a shallow, fairly meandering stream bed. The contrast with Hartley Beck reflects the extra distance the stream has travelled through the easily eroded sandstones from its limestone contact and also the extra volume of water gathered from a greater catchment area (4 -5 x). Looking west, from the viaduct, the valley narrows, but this time the more powerful beck has cut through the St Bees Sandstone cliff on its way to the R. Eden, (Fig 4.)

The route moves on to and up over the next bridge (3). Lime deposited as flowstone coats the sides of the arch, forming over the last 150 years since it was built. On the way up the hill the Mill Race can be seen on the crest of the slope opposite, marked by a line of gorse. At (4), left and left again over two stiles, the descent is made into a deep wooded valley.



Figure 3. Ladthwaite Beck looking East from Podgill Viaduct



Figure 4. Ladthwaite Beck looking West from Podgill Viaduct

Ewebank Scar

The path continues, noticeably 'claggy', down to the stream's edge (5) where, if fordable, Fig 5 may be seen. The fluvial St Bees Sandstones and Siltstones form the west bank above a hard sandstone band in the bed of the stream. The dip is N of West around 10 to 15°. Further along the east bank the St Bees group lies unconformably on the Great Scar Limestone dipping 23° to the NW (Penrith Sandstones, Eden Shales and the rest of the Carboniferous Series are missing). The discordance of some 8-10° is similar to the exposure below Merrygill Viaduct (25). Some scrambling amongst fallen branches will reveal red shales interbedded in the limestone under a basal conglomerate. Burgess (1965) discusses this in detail at the Merrygill exposure and how in the 1890's these beds were thought to be Permian rather than within the Carboniferous.



Figure 5. St Bees Ssts & Slts in Ladthwaite Beck

For the next 600m Ladthwaite Beck has created an unresearched geomorphological 'marvel'; Ewebank Scar and Gorge. Over this distance it drops 85 m in two sections of 40 m by 250 m. The lower section Fig 6, is the more spectacular as the stream is usually running along a ledge 2-3 m above the base of the gorge and some 20 m below the cliff top. Doubtless the Victorians saw this as a beauty spot, but the only celebration in literature



Figure 6. Ewebank Scar

(local or otherwise) is in Charlie Emett's Walker's Guide to the Eden Valley. Pondering on the cause of this phenomenon, he refers to "Orgogenisis (sic), the intense vertical movement of the earth's crust, caused this local fracture in the Carboniferous limestone of the Askrigg Block, one side of which was pushed upwards, forming this dramatic scar". Clearly a Catastrophist at heart, Emett was an unwitting follower of Adam Sedgwick.

Although some jointing is evident, there is no evidence of dislocation. The simpler process of Uniclinal Shifting, (thanks to Alan Smith for this terminology), whereby a stream eroding down dip will account, given sufficient time, for a scarp slope. Here the stream is eroding the shale partings that underpin the limestone beds dipping into the cliff. In times of spate – particularly post-glacial – dissolution, collapse and mechanical removal of debris, in addition to quarrying by ice, are likely to have created the cliff, leading to overdeepening below the current stream level.

On the way to the upper section, the gorge widens briefly (7). Probably this is due to a resurgence half-way down the slope. This was contributing a considerable to the volume to Ladthwaite Beck at this time. Collapse in the overlying soil surface indicates how active the dissolution processes are below. Just above (9) and noticeable when the stream level is low, most of the beck disappears down a rift, leaving the main, lower stream bed virtually dry. It is likely that Ladthwaite Beck is then replenishing itself downstream at the resurgence. The upper gorge is a youthful version of its counterpart downstream.

Before emerging from the gorge, passing mention should be given at (8) of the first chance to examine a fossil assemblage, corals and crinoids. At (9), on the other bank, the feed to the Mill Race can be seen with pipework buried within a 10m long by 2m deep rock cutting. A culvert can be traced upstream, parallel to the beck, to a weir. This point is 100m higher than Hartley High Mill. A tribute to the surveyor who negotiated so few obstacles.

High Out Wood and Birkett Hill

Keeping Ladthwaite Beck to the left and below at (10), the west side of Birkett Hill can be viewed (Fig 7). The hill is regarded as Robinson Limestone (youngest of the Great Scar Limestone Group) the southerly end dipping 30° to the NE. The (obvious ?) fault, against the apparently horizontal beds, is not recorded on the BGS map. Birkett Hill is a SSSI, as is High Out Wood, both for botanical reasons. Late spring brings forth an abundance of orchids.

Turning to the north, looking across the floor of the alluvial valley, the structure in High Out Wood comes into view. From the top of Birkett Hill,







Top:
Figure 7.
Birkett Hill
West Side

Centre:
Figure 8.
View NW to NE
from Birkett Hill

Left:
Figure 9.
Branded Chert
below quarries
at High Out
Wood

(Fig 8), a succession of limestone scarps, beds dipping increasingly steeply ENE, indicate the east flank of the anticline. (The hill beyond to the right, against the horizon, is Long Rigg, the east flank of the Hartley Half Dome). The anticlinal axis of High Out Wood meets Ladthwaite Beck, left of photo, at the top of the gorge. It is interesting to speculate whether the change in dip to horizontal means the stream is less able to 'shift' or erode as at Ewebank. Given time the stream will presumably cut back, possibly with a greater emphasis on collapse by dissolution.

High Out Wood is on private land. If permission is obtained it is worth visiting the area of limestone pavement on top (can be seen on Google Earth). Also below the quarries is a significant band of chert, (Fig 9). Close examination shows limestone replacement.

On the right of Fig 8 can be seen Hartley Birkett Hill and tracing the valley to the bottom right leads past the scree of the Ladthwaite Mine. In the centre far horizon is the Pennine escarpment above the Eden valley half-graben.

The footpath now swings round the north side of Birkett Hill, through a hay meadow to Ladthwaite. The beck used to run under the imposing east face (11), which according to Wainwright, looks 'aggressively attractive'. Sink holes in the former course presumably drain under Birkett Hill to the west. Best seen, as in Fig 10, this side is a complete contrast to the other. The limestone is steeply dipping 65°+ to the SE (again not recorded on Sheet 40, but on the 1889 map). Here the effect of the Dent Line, valley wide, fault zone is apparent.



Figure 10. Birkett Hill looking SW.



Figure 11. Hartley Birkett Main Hush, view to NE.



Figure 12. Hartley Birkett Main Hush, view to NE.



Figure 13. Hartley Birkett Main Hush, view to SW.

An examination of the beds at Ladthwaite (12) shows a dip, this time 55° to the NW. Darker limestones with more shaley partings lie to the east of more massive pale limestones, above the road, further north. A cursory examination of a brachiopod in the former and corals in the latter did not lead to any firm ideas about 'wayupness' at this point. However some conclusions can be extrapolated from evidence further on. In the background of Fig 10 can be seen the Upper Rawthey valley, between the Howgills and Wild Boar Fell, where it follows the line of the Dent Fault. On the left, much glacial till obscures the SW continuation of the Dent Line.

Ladthwaite and Hartley Birkett Mines.

The route from Ladthwaite to Hartley Birkett, along a ridge, passes to the left of an alignment of shafts and above the Mine scree, now enclosed, and is taken to be the direction of the Ladthwaite Level. It meets Fell Lane at (14). It is best to climb to the top of Hartley Birkett hill (15) to truly appreciate the view of the mine workings. Fig 11, looking north, taken from (16), shows just how much has been removed and hushed since the earliest records in the 14th century (for the most complete account and map see Smith and Murphy 2011). The major exploitation took place in the 17 and 1800's mostly of galena. Subsidiary minerals – fluorite, barite, calcite, dolomite, sphalerite and chalcopyrite were also present.

Hartley Birkett hill is structurally complex. Early geologists used the words 'very doubtful' and 'very disturbed' to qualify the word 'ground' on their field notes. By 1985 Dunham records in the BGS Memoir that short of drilling, the general view is the hill "is a mass of tectonised limestone . . . interpreted as Robinson Limestone, thickened by faulting . . . and the clastic partings faulted out". "Every contact has been jostled, and parts of the limestone converted into microbreccia, now mainly cemented with calcite". Fig 13, looking southwest is the reverse of Fig 11 showing the main mining area and the limestone beds dipping steeply to the NW.

The Dent Line

The key to the structural explanation of this area can be seen left of centre in Fig 11 and in close-up in Fig 12. This is just to the left of the remains of the old Smelt Mill (17). The limestone is taken to be Hawes Limestone (due to Girvanella being found by Turner in the 1920s) with an "inverted dip of about 70° to the west, well attested by the fact that the shale member of the cyclothem overlies the sandstone" (Dunham 1985). Thus dips to the northwest on Hartley Birkett and at Ladthwaite, all on the east side of the Dent Line valley, may be interpreted as being overturned. Turner (1935) stated that this is a "structure, essentially that of a monoclinal fold, the middle limb constituting a belt of vertical or inverted beds, often marked by

faulting. The line of steeply dipping strata, whether actually faulted or not, may be termed the Dent Line". Fig 14 shows Turner's Section to the north of 'Dunham's Outcrop', where the beds have resumed a more vertical position.

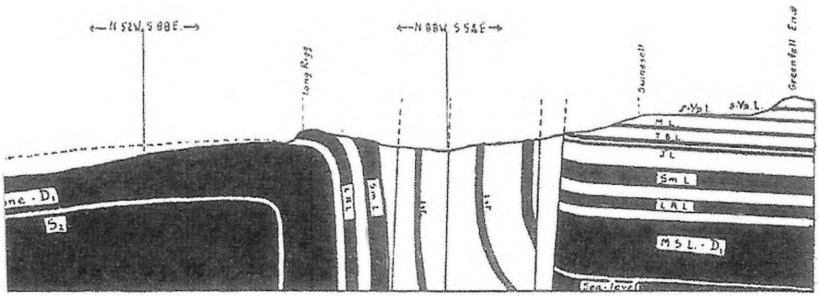


Figure 14. Turner's Section across the Hartley Half-Dome and Dent Line (1935)

Underhill *et al* (1988) in resurveying the whole Dent Fault brought a more detailed understanding of kinematics to the Dent Line area. The fault along A-B SW of Hartley Birkett Hill is taken to be a fulcrum, south of which beds are downthrown to the east of the Dent Line. To the north the downthrow is to the west, resulting in the plunge of the Hartley Half Dome to the north. Relative to this the overturned beds of 'Dunham's Outcrop', the Hawes Limestones, are raised. This would explain his observation that beds, including and between the Robinson Limestone of Long Rigg and the Hawes Limestone (both dipping towards the Dent fault Line from either side) may be repeated.

The age of monocline formation is thought by Underhill *et al* (1988) to be due to late Carboniferous/early Permian Variscan NW-SE compression. In addition, this imparted a sinistral oblique-slip motion which was further enhanced in the Permian by NE-SW transtention as the Vale of Eden half-graben developed. (Possibly the area around NY 797075 on Sheet 40 conveys this complexity - Fig 20). This may have controlled sedimentation events including the initiation of the Stenkrith Brockram alluvial fan.

Hartley Beck

The upper reaches of this stream are referred to as Birkett Beck, on the 1862 OS map, as far as the mine area. This is the source of Hartley Beck the name now used for the rest of watercourse. Fig 13 shows the stream emerging from the mine area and meandering to the SW. Limestone beds dipping SE at 55° (18), guide the stream, over several sink-holes, the distance achieved being directly related to the flow volume. After heavy rainfall, 2-3 times/year the stream will reach the bridge (19). Downstream the map

shows the course of the beck turning through nearly 180° to flow north. Here as the beck enters High Out Wood the dip becomes 20° to the SE. It is possible that preferential 'shifting' down dip is a sufficient explanation once the stream found this easier route. Possibly its route to the south was redirected north by ice or blocked by 'drift' at some stage or it follows a fault. Perhaps Ladthwaite Beck did not always flow west around Birkett Hill. Whatever the cause we have two streams entering a Karst environment (similar height of 310m): one continuous, the other sporadic and reflecting the difference in their catchment areas.

Without the relevant permissions, the walk would follow the road back to Hartley with views of the quarry to the south. However, below Fell House (20) a gate gives access across two fields to a small outcrop by a stile (21). Here and in the stream bank below the wall chert can be found, but in nodular form rather than banded. The walk then follows the dry channel for around 750m past (22), the crest of the anticline and some 'dry' waterfalls, to the point where it enters the quarry (23). It has here been tunnelled and piped through settling tanks to its exit near the resurgence below the viaduct (25).

Hartley Quarry and Cave

Although there has been some quarrying since 1860, the main operation started in 1925 with a limeworks making high quality lime for steel and for agricultural use. This could be transported via rail until the track closure in 1974. The quarry changed hands twice and was bought by Cemex in 2004, but mothballed in 2008 when extraction ceased. The screening, crushing and coated roadstone plants were removed and the lime plant is now used by a Belgian based company L'hoist, who now transport a higher quality limestone from Buxton.

The Society gained permission to have a general look around on a promise not get too close to cliff faces. Across the quarry there is a general dip from NE to SW, typically of 20°, unless disturbed by faults. The main fault crosses from N to S, downthrow to the west, with a central remnant showing slickensides. Fig 15 shows a typical quarry face, a mixture of competent beds, and rubbly pseudobreccia, stained and oxidised red, with some minor faulting. Corals, crinoids and brachiopods were abundant in many of the large limestone blocks lining the roadways.

On the south side of the quarry is a cave entrance about 2-3m in diameter with access up some fallen blocks, (Fig 16). It must have been revealed, after blasting, before 1991 as this date can be found a short way inside. Until now it had not been explored and this venture was enthusiastically taken on



Figure 15. Hartley Quarry West Face.



Figure 16. Hartley Quarry Cave entrance.

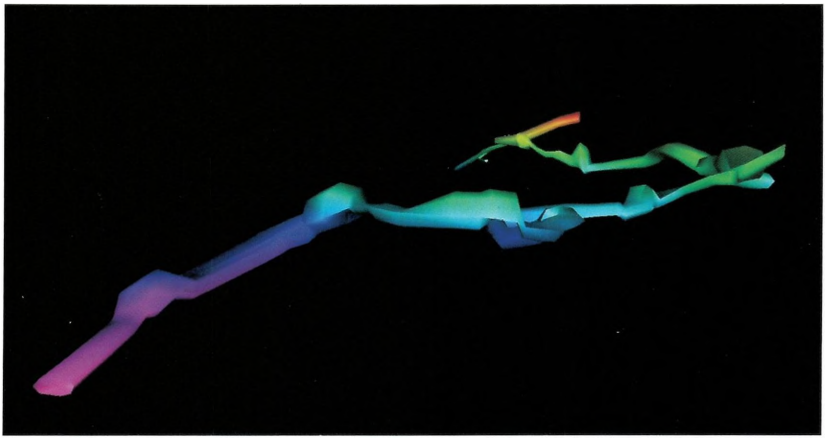


Figure 17. Hartley Quarry Cave. Entrance far right. Red = high, Purple = low.



Figure 18. Hartley Quarry Cave, Aven in left branch.

by local cavers and a diver in 2015, when the water levels were low. Fig 17 is the result of initial surveys. There are two branches – 137m to the left and 124m to the right so far, with sumps in each. Passages range up to 2m across and over 4m high in chambers (Fig 18). Walls are mostly covered in a fine, dried clay with occasional flowstone. The assumption is that this cave entrance is a breach in a left to right (E to W) trending downstream passage, rather than a double passage leading out into the now excavated quarry area. Hopefully there will be a chance to refine this information later this year, but the story so far can be seen on You Tube.

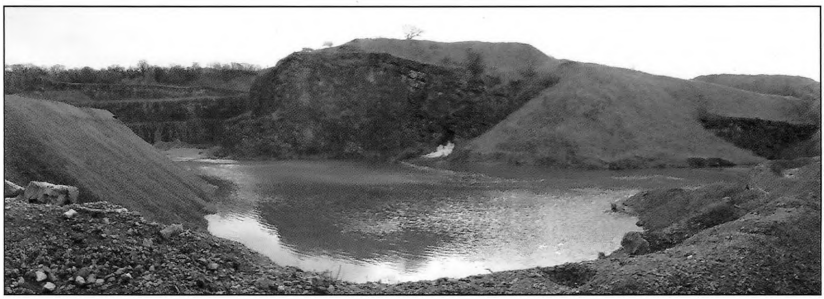


Figure 19. Hartley Quarry/Cave, 6th December 2015 after storm Desmond

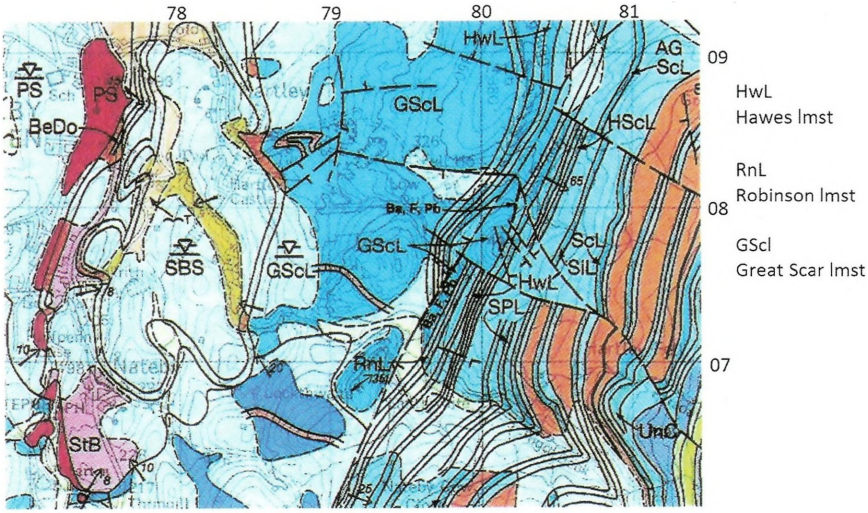


Figure 20. Geology of Hartley Fell showing effect of Dent Line on lmst formations

Lastly what does this cave tell us about the hydrology of Hartley Beck? Fig 19 illustrates that it is still active. When water issues from the cave mouth it is an indicator that the overground channel is full and discharging into the quarry. This is a rare event (although twice in Dec to Feb 2015/16) and can be correlated with the River Eden level at the Kirkby Stephen gauge exceeding 1.7m (1m above an Environment Agency warning!).

However, the normal situation is for water sinking near Fell House to find its way below the explored part of the cave to emerge at the resurgence. Hartley Cave once phreatic, is now a relic above the water table which a nearby borehole gives as 197m OD (the quarry floor has a limit of 220m OD). So how did Hartley Beck develop its watercourse? Clearly the water table was higher just post-glacial. But as with the development of caves in the Yorkshire Dales, for instance, White Scar Cave, where the change from phreatic to vadose occurred after valley development lowered the water table in Chapel-le Dale. Here, it likely that the cover of the Permian red beds was eroded back by the ancient Hartley Beck to the limestone contact. The water table dropped and Hartley Cave, once a major conduit became 'dry'. Dating of this event might be possible if suitable cave deposits could be located. Although the cave is subject to sporadic flooding, there are flowstone deposits that are clearly higher than the maximum flood event, even before the cave was unsealed by quarrying. Undoubtedly, valley deepening of the river Eden at Kirkby Stephen has exerted control on the development of Ladthwaite and Hartley Becks. Possibly, clues here within the cave would have some relevance to understanding the geomorphology of the wider area.

Notes on Maps and Mapping

The BGS website has 2 of the 3 Sheet 40 Kirkby Stephen Maps available on its viewer. 1997 Solid and Drift Provisional 1:50 000 and 1889 Solid 1:63 360. Field slips are available to view at the library, used as the precursors for both these final editions. But, for the recent map, the library could find no field recordings for the critical area around Hartley Birkett. In Underhill *et al* there is a small map of Hartley Birkett but contact with a co-author has not elicited a source. Reference in this text to field maps being available at the British Library proved unfounded as the microfiche had been lost.

1862 Ordnance Survey Map. Permission obtained from Kendal Archives.
Permission obtained from BGS to reproduce extract from Sheet 40 (Permit No. CP16/010).

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And lastly many thanks to Fred Rattray and friends for the following:
You Tube.
Hartley Quarry Cave
https://www.youtube.com/watch?v=d5e EOUIQwHw

Noel Pearson
[redacted]

LOCAL GEOLOGICAL SITES (LGS) AND THE WORK OF THE CUMBRIA GEOCONSERVATION GROUP (CGC).

Sylvia Woodhead

Cumbria GeoConservation (CGC) is a new group, formerly known as the Cumbria RIGS Group. RIGS started as de-notified SSSIs after the Earth Science Conservation Strategy was published in 1990. They are non-statutory, and as the name implies they have a regional focus. Through the local planning system the designation is intended to protect important geological sites from inappropriate development. Cumbria RIGS group was set up in 1992, to conserve regionally important geological and geomorphological sites (RIGGS) in Cumbria. Its remit was to encourage the appreciation, conservation and promotion of geological heritage for education and public benefit. CGC now operates as a special interest group of Cumbria Wildlife Trust (CWT), and meets formally three times a year at Plumgarths, the CWT HQ in Kendal.

The first (RIGS) sites were identified by members of the Cumberland and the Westmorland Geological Societies and were designated in January 1993. The site data records were formerly held by Cumbria County Council, an arrangement which is now ending. Site records are numbered according to the planning district within Cumbria, which now has around 280 LGS. This large number reflects the geological richness of Cumbria's geology. In total they represent considerable geodiversity, hard rocks of all ages, structures such as faults and anticlines, fossil sites, glacial, post glacial and recent sites such as deltas and waterfalls.

280 LGS coded 1-8

Code	Cumbrian District	No of LGS	Example	LGS name
1	Carlisle	12	1/009	Irthing Gorge
2	Allerdale	14	2/021	Maryport Foreshore
3	Eden	57	3/010	High Cup Nick
4	Copeland	42	4/042	Hodbarrow Point
5	South Lakeland DC	41	5/022	Benson Knott
6	Barrow	9	6/009	Walney Island South
7	Lake District NP	166	7/005	Rosthwaite moraines
8	Yorkshire Dales NP	12	8/013	Carlingill, Tebay

Since Defra Local Sites Guidance 2006, RIGs became known as Local Geological Sites, LGS, as the geological equivalent of Local Wildlife Sites. GeoConservation UK Guidelines for LGS selection are:

- Value for education and lifelong learning
- Intrinsic scientific interest
- Landscape value and promotion of public awareness
- Historical vale and context of geological thinking

Since 2014 the CGC group has met informally in the field to visit and review geological sites, and has recently started negotiating more active conservation management with landowners and managers. Unlike SSSIs LGS are not required to have management plans. When designated they were mostly not seen to be at risk. However over time many sites have become overgrown, their access lost or quarries infilled. Recent discussions with Lake District National Park Rangers will hopefully lead to clearance of obscuring vegetation from some LGS quarries, such as Brown How, Blawith, LGS 7/027 (SD 289901) and Scout Scar Quarries near Kendal LGS 7/128 (SD 487925).

The CGC group has plans to include with LGS designations an account of the geology which is more accessible to the general public and to planning officers, and to design more geological trail leaflets.



Positioning of the geology interpretation panel on the prepared sandstone block at Birkhams Quarry. March 2016.

An excellent current example of active co-operation with site owners and managers is with Marshalls Stone at Birkhams Quarry, St Bees Head, LGS 4/030 (NX 956154), where St Bees Sandstone is extracted, outside the bird nesting season, for conservation building stone. Blocks of stone for two interpretation panels, one for geology and one describing the quarrying process, were chosen by CGC and later taken to Locharbriggs quarry, Dumfries where the blocks were cut. The panels will be placed at a viewing area just outside the quarry, at a strategic location on two long distance footpaths; 'Wainwright's Coast to Coast Walk' and the 'England Coastal Path'.



Rock faces to be retained in Phase 1 of the progressive restoration at Birkhams Quarry

Great progress has also been made at Eycott Hill, CWT's latest reserve. Geological interpretation panels, designed by Elizabeth Pickett, are in preparation, for the car park and for a viewing platform on the first lava outcrop encountered on a waymarked trail around the reserve. A geological trail is also planned. A short video film featuring John Rodgers of the CGS explaining the geology for non-specialists has stimulated a wish list for similar videos for other LGS.

Recent survey visits have identified certain problem sites, for example Hodgson Howe Quarry, Keswick LGS 7/079, (NY 242236) where despite being by the roadside, access and visibility is made next to impossible by a locked gate, barbed wire, soft slurry in the quarry floor and abundant tree growth obscuring the steeply dipping Skiddaw Group exposure in the quarry walls. This quarry is the site where Edgar Shackleton found the graptolite *Azyograptus lapworthii,* which was formerly displayed in Keswick Museum. For this reason the site should not be archived, but it cannot be recommended for a visit. At Park Wood, Isel LGS 7/125 (NY 167348), the contacts between the underlying Skiddaw Group, the Cockermouth Lavas and Carboniferous Limestone can no long be seen, while the adjacent site of Clints Crag and Thackray Wood LGS 7/067 (NY 160357) also needs a thorough review. A good LGS site for Cockermouth Lavas is required and Gill Beck, Blindcrake (NY 149343) is currently being considered.

LGS designations are currently stored as PDF files. The CGC group is currently working on how to make these more accessible. It is hoped to work closely with CGS members in the future, particularly in the joint tasks of field surveys and site reviews.

FIELD RESEARCH PROJECTS

Summer 2015

The Society instigated a new venture during the summer of 2015 with two field research projects. The idea behind these was to use the collective resources of a group of members to collect data and information on sites of geological/geomorphological interest. The material collected would be recorded as a lasting record in Proceedings as an aid to further researchers. It is not easy to continue finding field excursion leaders able to conduct members to new sites of interest and to provide new material on the geology of the county. These new research projects therefore are an attempt not just at 'do it yourself' project work, but they will supplement our future summer excursion programmes. They enable sites to be investigated more fully by a group and they can potentially provide large amounts of useful data. Detailed measuring, counting, recording and listing of information can be achieved by group research.

Reports of the two sites visited in 2015 follow. Members are asked to come forward with future ideas on site/locations that could be investigated – please let the Excursion Secretary have your thoughts.

THIRLMERE

August 12th, 2015

This location was brought to our attention by Society member Alan Wise. Not only did he find the site, but he liased with staff from *United Utilities* who are the owners of the area and completed some site clearance on the exposure before members gathered there on a nice summer evening.

The exposure is on the side of a forest track in the Thirlmere reservoir planations approximately 1km south of the Swirls Car Park on the A591 along the eastern side of the Thirlmere valley. The exact location is at NY 32112 15435. It stands at around 260m OD and is in the Seathwaite Fell Formation of the Borrowdale Volcanic Group as mapped by BGS. (Keswick Sheet 29, 1:50,000).

The exposure shows a series of ovoid structures originally postulated to be volcanic bombs by Alan Wise. 10 ovoids were measured along a section about 19m long. They were all confined to one steeply dipping 'bed' in the

rock, approximately 15cm thick (Figures 1 and 2). The individual ovoids are largely oval or almost circular in shape and measure between 50 and 200 cm along their long axes (Figure 3). They are quite regularly spaced along the section. At detached specimen of the ovoids from the base of the exposure was collected and subsequently sliced to reveal the internal structures on a clean surface (Figure 4).



Figure 1. Site of ovoids in dipping bed - left half of picture.



Figure 2. Chain of 10 ovoids measured along section

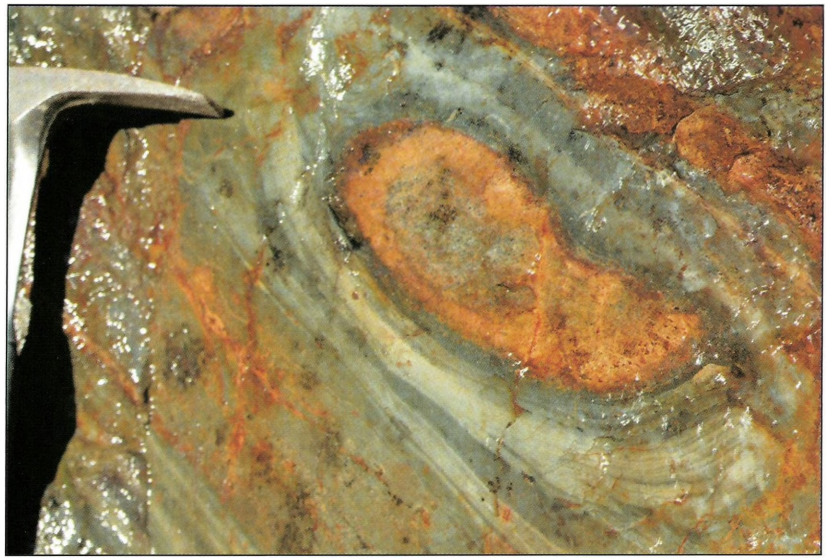


Figure 3. Close up of one of the major ovoids – hammer point for scale. Layers of pale grey mudstones wrapping round ovoid structure.



Figure 4. Sliced ovoid sample. Dark clastic ovoid at extreme left, part of second ovoid (lighter in colour) at right. Central area of sample shows contorted and folded mudstone structures. Darker spots?

Discussion in the field between members cast doubt on their origin as volcanic bombs. No impact features were apparent beneath the ovoids. Finely laminated mudstones surround the ovoids and appear to be heavily contorted between them. The site has subsequently been revisited by Clive Boulter who has contributed the following interpretation of these features.

ENIGMATIC THIRLMERE OVOIDS

Clive Boulter

Multiple working hypotheses:
1. Volcanic bombs – subaerial/very shallow water.
2. Volcanic bombs – deeper water (water settled and no impact sags)
3. Soft-sedimentary deformation – e.g. pseudo-nodules

Basic tests:

- In 1 and 2 the proposed bombs would have igneous textures.
- In 3 the textures would be clastic sedimentary.

The ovoids have a clastic texture similar to the entire bedded sequence – dominated by highly angular quartz grains, which fits the assignment to the Seathwaite Fell Formation.

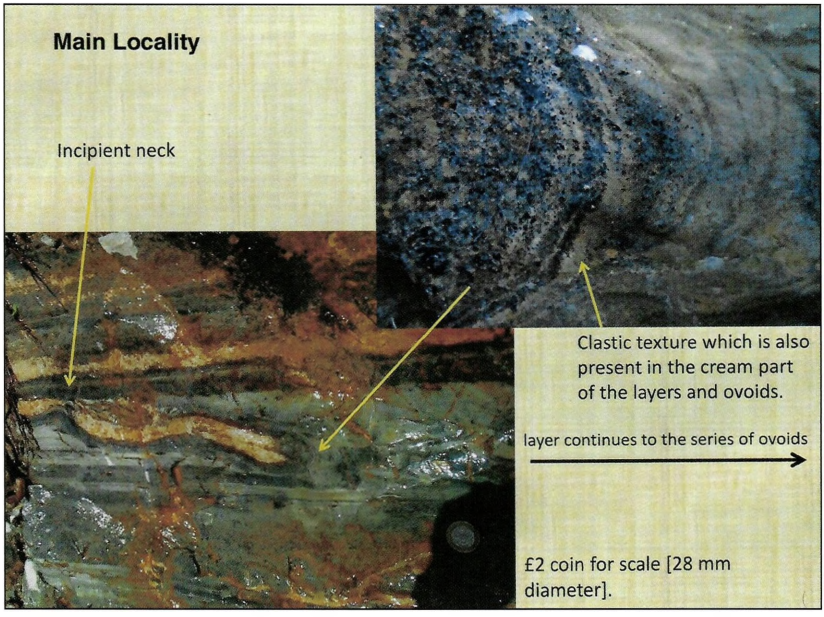


Figure 5. Interpretation of neck structures (Clive Boulter).

The layer the ovoids were derived from is semi-intact at the head of the exposure where an incipient neck shows the early stage in the segmentation process (Figure 5).

The ovoids are therefore interpreted to be the result of soft-sedimentary deformation but the exact sequence of events is not clear. There are other locations in the lake deposits of the Scafell Caldera where vertical sinking of the sand-grade deposits was the main process; that is load and flame structures progressing to pseudo nodules (on Bowfell and Scout Crag, Great Langdale).

The layer of isolated ovoid bodies is challenging and I do not recall seeing anything like it. These features typically involve sinking of the coarse layer into the finer-grained layer. Is this enough to explain the wrapping of the ovoids by the lamination in the former mud? Liquefaction was certainly involved but do we need to go further to explain the complexity of the folds in the mudrock surrounding the ovoids; there is a temptation to invoke some downslope movement at the time of liquefaction and perhaps sheath like folding?

LINEWATH AND ROSPOW HILLS

September 13th, 2015

Alan Smith

To the north of the village of Mosedale, on the north-eastern edge of the Lakeland Fells, there is an area of relatively low ground with an array of quite complex landform features. Frequently referred to as 'Carrock End' it stands beneath the towering slopes of the igneous complex of Carrock Fell to the west. The landforms result from the interplay of glacial, periglacial, fluvial and hillslope processes. The area is drained by the River Caldew which has its headwaters in the heart of the Skiddaw-Blencathra fells. Where the river emerges from the fells at Mosedale, it turns abruptly through 90° to flow northwards, cutting across this complex 'Carrock End' area. Overlooking the area on the east, is the low scarp of the Carboniferous limestone, marking the true eastern edge of the Lake District Lower Palaeozoic inlier. To the south of Mosedale, the interfluve with the southward flowing Glenderamackin, which emerges from the Skiddaw-Blencathra fells at Mungrisedale, is indistinct and is hardly discernible in the low lying relief.

An interpretation of the area was offered by Boardman in 1996. Subsequently, papers by Clark, R and Wilson, P. (1997 and 2002) also relate to this Carrock End area and offer further details of some of the landforms. Many of the features of this area have so far only been described in terms

of their morphology. Sampling of the deposits and analyses of the materials making up many of the features is still in need of much further work. In particular, a detailed chronology of landform formation is far from clear.

Thirteen Society members spent a full day examining two areas close to Linewath at the northern end of this complex. First, some measurement and sampling of the fabric of Long Hill, a suggested esker feature was attempted, followed later by a brief investigation of the Rospow Hills.

LONG HILL, LINEWATH (NY 354345)

The basic topography of Long Hill was measured using GPS, *Google Earth* and simple surveying instruments. Figure 1 shows the detail. The ridge is just over 200m long and around 40m wide along much of its length. It is very gently curved and is aligned roughly on a bearing of between 27° and 35° east of north. In profile it is a curved convex feature, slightly asymmetrical, with slopes of around 19° along its western face, but between 20° and 22° on its eastern side. The wet, level ground to the west of the ridge is very slightly higher than the area on the other side, which falls way gently eastwards. The northern end of the ridge has been quarried for stones and gravel (areas C and D on Figure 1). The very northern end of the feature is a little indistinct as material has also been removed from this area.

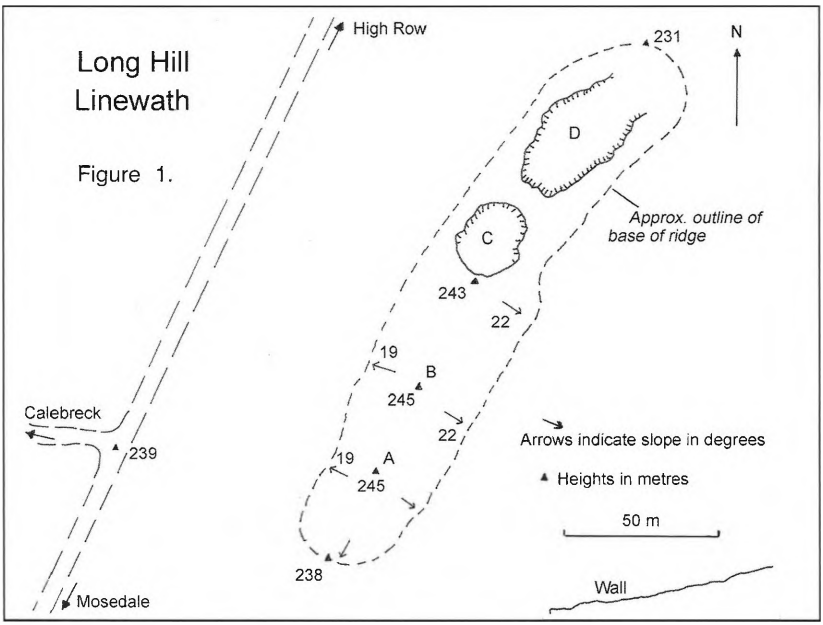
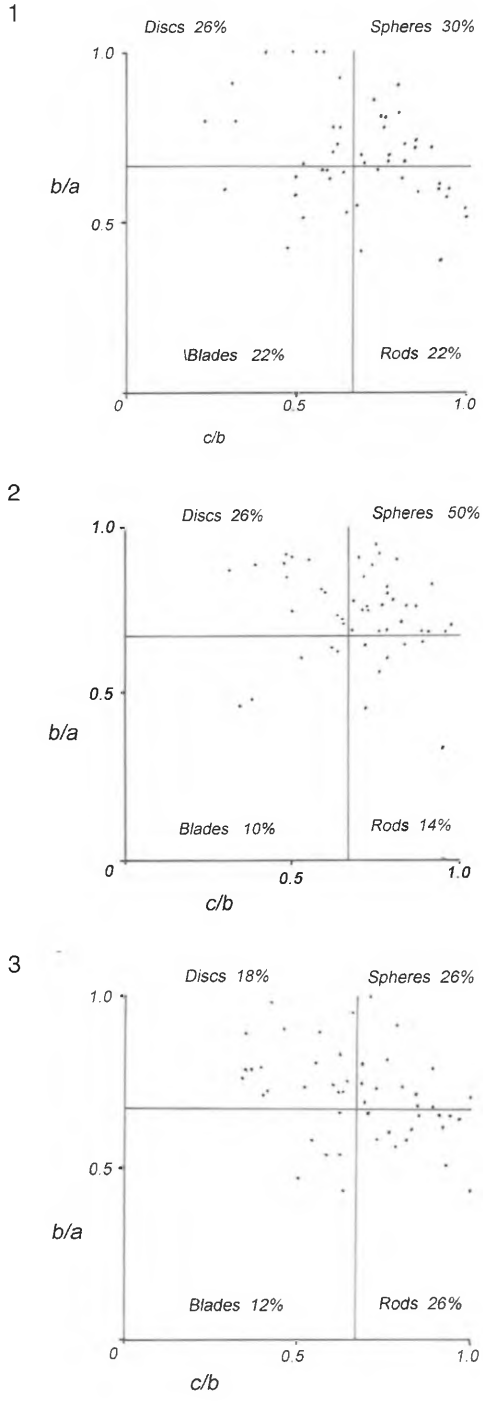


Figure 2.
Zingg and Krumbein data on the three samples.



Overall data from Zingg analysis.

Total 150 samples.

Discs	29%
Spheres	35%
Blades	15%
Rods	21%

Krumbein Analysis.

Sample 1	0.681
Sample 2	0.719
Sample 3	0.742
Overall mean	0.714

Three random samples (each of 50 clasts) from the quarried area D were measured – recording the three axes – *a, b and c*. Zingg and Krumbein calculations were completed on each of the samples. The data is shown in Figure 2 and associated tables. All the samples show a high degree of rounding of the clasts, indicating transport of the material over some distance. The Zingg calculations show overall 35% of the clasts have a spherical form. The scatter diagrams (Figure 2) also point to overall high concentrations in the NE quadrants, confirming high rounding characteristics in all four shape forms. The Krumbein sphericity index figures confirm this overall rounding of the clasts. A mean index of 0.714 is a relatively high figure. Krumbein has a single sphericity value range from a minimum of 0 to a maximum of 1. A true sphere has a value of 1. Such a high value for sphericity, when put together with the Zingg data, which is a measure of particle form, points to a high degree of particle modification by transport attrition.

Figure 3.

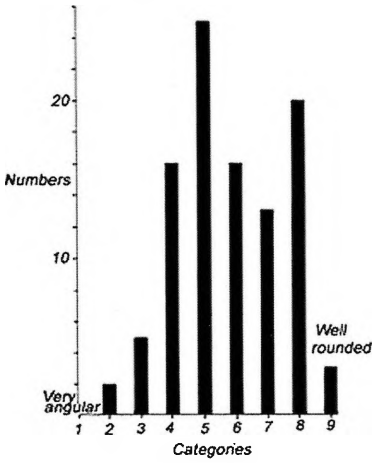
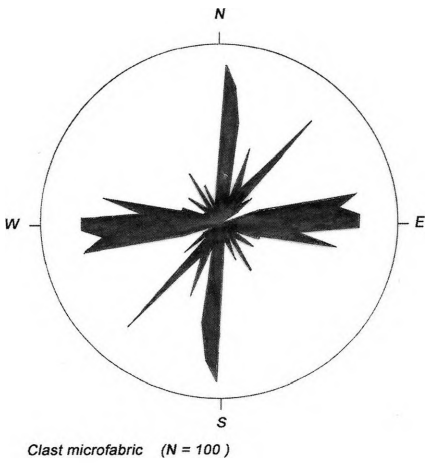


Figure 4.



A further random sample of 100 clasts was measured using a visual comparison chart (similar to the one devised by Powers 1953). On a 1-9 scale each clast was visually assessed for roundness (1 being very angular, 9 being well rounded) (Figure 3). Interestingly whilst this further confirms the highly rounded state of the deposit as a whole, it does suggest a bi-modal grouping. Significantly no clasts were recorded with an index of 1 (very angular).

Particle orientation was not measured by the group, but subsequently Fiona Harwood has kindly made available some data collected from the same site by a school group. This has been plotted out and is shown in Figure 4. Whilst the data is only two dimensional, the rose diagram of orientation on

the long *a* axis of the clasts is of interest. There is substantial orientation along the long axis of the ridge, but more than equally strong is a fabric aligned almost west-east.

Some considerable time was spent by a group of members deciphering the lithology of the clasts. Lithology was not considered along with the shape measurements. Intensive searching however produced a large array of rock types. Material was identified from the following locations:

- Large array of gabbros, microgranites and other igneous material from the Carrock Fell intrusion.
- Many specimens of Skiddaw Group material, including the hornfels and cordierite hornfels from the Skiddaw Granite aureole.
- A variety of lavas and pyroclastics from the Borrowdale Volcanic Group – importantly including the augite-andesite found at Bramcrag, St John's in the Vale.
- Several pieces of the Purple Breccia which is the basal conglomerate of the Borrowdale Volcanic group (Thirlmere Formation) which occurs both in the Naddle Valley and at the BVG/SG junction on the eastern shore of Derwent Water.
- Threlkeld microgranite.
- Microgranite from the Armboth dyke, Thirlmere.
- Mell Fell Conglomerate.
- Lower Carboniferous Sandstones (precise locations not known).
- Eycott Hill porphyritic andesite (Eycott Group).

Interestingly no Penrith Sandstones or Carboniferous limestone material was found.

This array of material generally confirms Boardman's conclusion of northward moving ice over this terrain, with sources in the Thirlmere/ Threlkeld areas and from the eastern Helvellyn fells to the south. Ice was clearly also feeding erratic material down the Caldew Valley from the heart of the Skiddaw-Blencathra massif. This conclusion is important as the pattern of ice movement has recently been question by Jarman (2015). He has pointed out that the 'Threlkeld basin' *(his term)* although suggestive of glacial overdeepening, is at odds with the present topography. It would require easterly and north-easterly moving ice to re-ascend some 100m and take a rather circuitous circular route round the eastern end of the Skiddaw-Blencathra massif when a more direct alternative, downhill all the way, is available north-westwards to Keswick and Bassenthwaite. Our count of the material in the Linewath ridge clearly contains Armboth dyke, BVG basal conglomerate, Bramcrag material and Threlkeld microgranite all supporting eastward moving ice from the Thirlmere/Threlkeld area.

Our results from Linewath add a little more information on these landforms. All the results are commensurate with an interpretation of Long Hill as an esker feature. Its relationships with the other features nearby need further work.

ROSPOW HILLS

For the latter part of the afternoon the group moved a short distance northwestwards up to the southern side of Carrock Beck to Rospow Hills. Here there are two, smooth conical hill features (Figure 5), apparently composed of superficial deposits of some kind. Scattered on the flanks are occasional boulders of Carrock Fell gabbro and microgranite. There are no exposures on the summit areas of the two hills which are covered with short moorland vegetation, dominated by mat grass *(Nardus stricta)*. Small gullies and some bankside erosion scars on the northern edges of the features close to Carrock Beck reveal glacial till, dominated by assorted semi-angular clasts of Carrock Fell material. The BGS mapping of this area (Cockermouth Sheet 23, solid and drift Edition) simply shows the area as undifferentiated boulder clay.







Figure 5.
Rospow Hills:
Above left: General view of Rospow Hills looking east;
Left: Society group at summit;
above: soilpit with horizons marked.

The group were introduced to the technique of soil chronosequencing. This has very recently been first used in the Lake District (Bickerdike *et al* 2015) as a possible way of dating moraines and evaluating the extent of the Younger Dryas Glaciation. The pioneering work on this technique was done on Canadian arctic landforms (Birkeland 1978 and Bockheim 1979), but its application to Lakeland landforms has proved potentially useful within the last year in giving relative dates on several valley moraines. The technique employs soil development as a relative indicator of age. The thickness of soil B horizons has proved a particularly useful index in age differentiation. In the Lakeland work done so far, Younger Dryas moraine B horizons depths lie in the range 9 - 17.5 cm, whereas those of older valley moraines are in the range of 23 - 33 cm.

A single pit was dug on the summit of the most northerly of the Rospow Hills at grid ref. NY 34438 34961. It is at 330m OD. A very clear conclusion was reached. A very deep profile was discovered.

Horizon A = 8cm deep
E = 4cm
B = 50 + cm (base not seen)

The B Horizon material was noticeably gravelly and almost granular with very few stones present. This suggests (albeit on the basis of just one profile and without further data on Munsell Colour, pH, silt/clay concentrations and iron content; all required for a full analysis) that these features are older than Younger Dryas.

Thanks are due to all the members who took part in this day and spent several hours doing the mapping and measurements. Those taking part were – Susan and David Beale, Peter Davis-Merry, John Lackie, Margaret Bennett, Bill Richardson, John Lawson, Fred Lawton, Fiona. Harwood, John Rodgers, Tricia Brown, Margaret Fox and Alan Smith.
All the data collected on both of these projects is with the Editor (Alan Smith). We can also give access to any of the reference material.

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Long Hill, Linewath.

General view of the ridge from northern end with Carrock Fell in background.



Group sampling in area D.

EXCURSION REPORTS 2014

EYCOTT HILL AND HIGHBANKS WOOD

Leader: Mike Dewey
(Westmorland Geological Society)

April 27th 2014

The excursion was in two distinct halves, reflecting the location astride the major unconformity on this NE side of the Lake District. Eycott Hill is the type section of the Eycott Group of volcanic rocks (Ordovician, Caradoc); overlapping them to the east are thick beds of Carboniferous Limestones.

The morning was spent traversing Eycott Hill looking essentially at two rock types. The fine grained basaltic-andesitic flows outcroping over much of the summit area and forming distinct ridges running almost N-S across the hilltop. More distinctive, however, were the classic keratophyre tuff outcrops forming a craggy ridge feature lower on the western side of the hill. Time was spent examining the large phenocrysts of bytownite in the tuff, a relatively rare calcium rich member of the plagioclase group of feldspar minerals.

Very good exposures of Carboniferous Limestone are seen alongside the minor road on the eastern side of the reserve. Small abandoned quarries in the 7th Limestone were visited on the eastern side of the road (NY388306).The leader explained the differences between the limestones in this northern section of Lakeland as compared with the more extensive and thicker beds exposed in the south of the County. Limestones were deposited on a southerly sloping ramp feature – here therefore, seas were shallower and muddier than further south. This 7th Limestone equates with the Park Limestone of S. Cumbria. It is a very fine grained wackstone, the grains supported by calcite muds. Just to the west of the road further outcrops were exploited for lime burning in the now derelict lime kiln which stands overlooking the acid pastures of Eycott Hill.

The afternoon was spent walking into Highbank Wood and traversing the extensive areas of pavement and open limestone grassland to the east. Much discussion ensued regarding the nature of the pavement surfaces. The leader was of the opinion that many of these very dark limestone pavement

exposures (6th Limestone) were palaeokarst features. Much of the limestone has a glacial till cover and several irregular sink hole features pit the surface and are aligned with the strike of the very gently NE dipping beds.



One of several shallow sink hole features on the limestone surface.



The distinctly 'nobbly' surface texture of the 6th Limestone pavement surfaces – possible palaeokarst (NY 391313).

EMBLETON MICRODIORITE QUARRY

Leader: Jim Samson

May 14th 2014

Eleven members and friends walked the short distance from Embleton village into the hidden world of one of the most impressive quarries of Cumbria. The quarry displays in fine style the Embleton diorite intrusion that was worked for road stone and exported via a waggonway to exchange sidings on the Penrith-Workington railway. The quarry faces are clean and visible, if inaccessible due to screes and their sheer height. After examining the screes for evidence of fresh samples of the diorite, the group followed the old mine paths to explore the industrial archaeology including a mine entrance and gantry.

GEOLOGY OF THE SHAP FELL AREA

Leader: John Rodgers (Senior Vice President, CGS)

June 7th 2014

16 members gathered on the A6 at Shap Fell summit on a morning of threatening thunderclouds and grey skies. The first location was the fine roadcut section 500m down the road at Crookdale Crags, where Silurian sediments of the Windermere Supergroup have been folded in the Acadian Orogeny into a series of minor anticlines and synclines. A number of features in the interbedded cleaved mudstones and dark sandstones (greywackes) were examined including vertical cleavage in the mudstones, flute casts and ripple marks indicating the current action of turbidite flows, and changes in thickness around fold axes in the fine-grained beds. A short but sharp shower caught out those who had chosen to walk down to the section, while others escaped the worst by reaching cars that had been brought down to the nearby lay-by.

The second location was an arranged visit to the Shap Granite quarry which at the time of the trip had not been worked for many years. Access had been organised with the Shap Granite Co. and members looked at the variation in texture, colour, content and structure, such as joint frequency. There was also focus on the numerous dark inclusions, usually thought of as xenoliths (pieces of the surrounding country rock that have been taken into the intruding magma). Other theories were mentioned for their origin and

which recognised the enigmatic enclosure of feldspar crystals entirely within the dark rock of the inclusions. There are two developments relevant to mention here. Since summer 2015 the granite has been worked on a large scale, mainly for large blocks of 2-3 tonnes which are being taken to coastal locations such as Fleetwood and Hartlepool and used for protective rock armour. The company, in response to requests from a Cumbria GeoCon group visit, agreed to place a number of representative large blocks at the quarry entrance so they could be studied by visitors without having to enter a working environment.

A short visit to an exposure in a stream bed (NY562101) on the north edge of the Shap intrusion was then made to look at thin veins of pale, fine-grained granitic rock (aplite) which had been injected into the metamorphic aureole of the surrounding volcanic material. On leaving the track back onto the A6, piles of crushed Shap Blue rock were examined for any of the minerals that had been introduced into the surrounding andesites by metasomatic high temperatures fluids emanating from the late-stage granite magma. These included garnet, pyrite, chalcopyrite, and epidote along with quartz and calcite. Much of the Shap Blue that is quarried from this area is crushed and used as aggregate around the region.

The group then moved on to the Shap Wells Hotel where there is an exposure (NY578095) of the unconformity between the underlying Lower Silurian black siltstones and mudstones and the overlying near horizontal pink conglomerate of probable Devonian age. The former occurred in stream bed outcrops above the waterfall, striking diagonally across the stream at dips of c. 60 to the south east. A notable feature of the conglomerate was the content of large grains of pink orthoclase, clearly from the Shap granite and so identifying the age of intrusion on field evidence alone as being post Lower Silurian and pre-conglomerate. But how old is the conglomerate?

To answer this question the group proceeded to a small limestone exposure in a roadside trench (NY581135) about 1 km south of Hardendale, and in beds that were obviously at a higher (younger) stratigraphic level than the conglomerate. Blocks containing fine corals and brachiopods were extracted (and replaced) to demonstrate the Carboniferous age of the fossils and so the conglomerate must be somewhere between this and Lower Silurian. Our final observation was of the fellside spread of granite boulders that are part of a train of such erratics that occur all the way to the east coast at Flamborough Head. By this time we were thoroughly soaked by the last of the showers but the leader was impressed by the tenacious spirit shown by the group who seemed to have enjoyed the day.

DRYGILL MINES, CALDBECK FELLS

Leader: Jim Samsom

June 28th 2014

Advantage was taken of dry weather to walk from Calebreck into this world renowned area, famous for its range of minerals. As is usual on Society field trips experienced members gave freely of their time to help participants identify finds (collecting is very restricted under the mineral policy of the Caldbeck Commons). Progress was gratifyingly slow, as there is plenty to see on the ground. The route took the group via Driggith mine to the summit and return via Drygill. Especially close attention was paid to sedimentary features in Drygill, which was the icing on the cake.

ALSTON – NENTHEAD – MINES AND GEOLOGY

Leader: Elizabeth Pickett

July 6th 2014

Unfortunately this excursion had to be cancelled as the leader was unavailable on that date.

THRELKELD MINING MUSEUM AND THRELKELD QUARRY

Leader: Rosemary Vidler

July 16th 2014

A group of about a dozen, including visitors, met in the car park at Threlkeld Quarry Museum on a dry evening. We walked into the quarry along the railway line to look at the contact between the Threlkeld micro granite and the country rock of Skiddaw Slates. The lack of a chilled margin and only the suggestion of a metamorphic aureole suggests the granite was cool when intruded.

Further along the railway line into the quarry there were some further glimpses of the contact in the form of 'fingers' pushing up into the surrounding rock. This site has deteriorated over the years due to its steepness

and falling scree so these were not easy to see. There were some suggestions that perhaps the site could be cleared but it is very unstable.

Moving further round the quarry we then looked at some mineralisation. It is reputed that magnetite, malachite, calcite and chalcopyrite has been found there however the dangerous condition of the scree there precluded searching for these. Some minor mineralisation could be seen in some of the more easily accessible rocks.

Towards the western end of the quarry we looked for further mineralisation and xenoliths in the rocks and boulders. It has been reported that some of the xenoliths previously examined have exhibited folding and copper mineralisation has been seen. Recent development work in the area to install railway lines for the quarry museum has dispersed some of the specimens that had previously been visible and time was spent looking for other examples with limited success.

We then made our way back to the car park looking at some of the abandoned machinery and other mineralised rocks. The intrusion of the laccolith has arched the beds of the overlying rocks but the weathering of the quarry faces now makes this difficult to observe.

GEOLOGY OF THE AREA AROUND GRIZEBECK AND GAWTHWAITE

Leader: David Kelly

September 6th

This was an excursion devoted to exposures of the Silurian rocks of the Windermere Supergroup. In this part of Lakeland they are around 5km thick and rest unconformably on the Borrowdale Volcanics to the NW. The rocks generally dip steeply towards the SE, although locally there are several NE-SW trending plunging folds and the rocks may be vertical or overturned. The succession is also displaced by major faults.

The rocks are turbidite sandstones and inter-turbidite mudstones, with the mudstones showing a strong cleavage. In some of the units sandstones dominate, whereas in others the mudstones dominate. There are some graptolite localities. There have also been some rare crinoid and bivalve finds (see Proceedings Vol. 8, part 2).

Bank End Quarry, Grizebeck SD 233851
This small disused quarry is in the sandstone dominated Poolscar Formation. Beds up to about 1m thick dip steeply to the SE. The thin mudstone layers show a strong, roughly vertical cleavage whereas a weak cleavage 'refracts' across the sandstone beds. The waste tips of the Kirkby Slate Quarries are seen from here to the SE. These are in the older Wray Castle Formation. Between the two localities is the Kirby Tear Fault, a sinistral tear which displaces the outcrops by several kilometres.

Pen Hill, near Chapels SD 237842
The Poolscar Formation is seen in small exposures in the fields. Cleavage dipping more gently than the bedding shows the inverted limb of a fold. A considerable time was spent in looking at the relationships between bedding and cleavage on the exposed rocks either side of this hill.

Great Burney Fell, near Burney Farm SD 256858
The rocks to the east of the road dip steeply to the NW and are turbidites of the Yewbank Formation. Spectacular flute clasts at this locality are highly distorted. This is probably because the exposure lies close to the axis of a syncline which plunges to the NE. The fold can be demonstrated by walking a short distance along the road to the north. The fold axis is crossed and the dips are then to the SE.



Flute clasts at Burney Fell.

Fellside NE of Heathwaite around Pewet Tarn
SD 251887715 to SD 25468743

The BGS map of the area shows many small NE-SW trending early Devonian microgranite dykes. One of these was traced across the fellside. It is intruded into the Yewbank Formation. Spectacular views form this location on a perfect day illustrated the major rock groups of the panorama round from Black Coombe, across to the Helvellyn fells.

Lakenhow Quarry SD268843

This small disused slate quarry is in the fine beds of the Wray Castle Formation which lies below the Gawthorpe Formation. Cleavage is well developed here, close to the axis of small plunging anticline which is seen on the quarry. This mirrors the situation in the large commercial Burlington Slate Quarries to the SW.

Bush Green Quarry (donkey Rock) near Foxfield SD 210868

The excursion concluded with a visit to this well documented site within the Gawthwaite Formation. This formation is of well-bedded turbidite sandstones inter-bedded with thin mudstones. The near vertical bedding planes strike roughly NE-SW. They show a large number of extremely well developed flute clasts, which indicate a palaeocurrent from the NW. This is consistent with other palaeocurrent indicators in the area.

EXCURSION REPORTS 2015

THE SKIDDAW METAMORPHIC AUREOLE.

Leader: Rosemary Vidler

Saturday 11th April, 2015.

A small group of members completed this 7km walk across the aureole from the Blencathra Centre, The geology is not reported as this is a widely known and well documented location.

LANDFORMS, TECTONICS AND MINES: THE GEOLOGY OF THE AREA EAST OF KIRKBY STEPHEN

Leader: Noel Pearson

Sunday 17th May 2015

The report of this excursion is presented as a separate paper in the original papers section of these Proceedings.

THE BUILDING STONES OF COCKERMOUTH

Wednesday 3rd June 2015

Leader: Jim Samson

Starting by the River Derwent the shallow water revealed the glacially derived cobbles that are found in many of Cockermouth's walls. The tour contrasted the use of local stone, largely Hensingham Grit from quarries in Brigham said to have been worked since Roman times with imported examples, both historical (Portland stone for Mayo's statue) and modern (imported granite for the Market Place restoration). The tour showed that geology in towns can be as informative as on the fellsides. The tour finished by the new hospital whose garden walls have much of interest in them, with the leader having learnt as much from the participants as he imparted to them.

PARTON BEACH

14th June 2015

A group of Society members led an exploration and collecting day on the beach at Parton for Petrostrat Ltd, a sedimentology and micro-palaeontological company based in Conwy, N. Wales. Details of the fossil finds and the exposures examined were not recorded, as this was a private visit which the Society was happy to host and provide local expertise.

PARTON DRIFT, COUNTESS PIT AND HARRINGTON COLLIERIES

Leader: David Powell

Sunday 14th July 2015

GR. NX 982 216

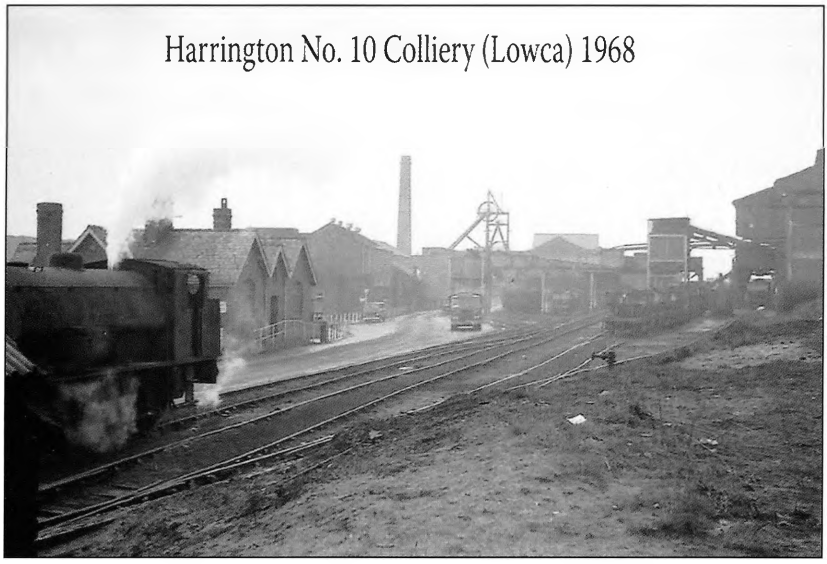
A group of members met at the old Harrington No. 10 Pit Yard in the village of Lowca. The purpose of the visit was to examine the remains of the coal mining industry that one time was extensive and stretched from Maryport in the north all along the coast to Whitehaven in the south.

Unfortunately there is very little bedrock geology to be seen on the surface due to a covering of glacial drift although the whole area is heavily faulted. Coal crops out in Distington Beck and along the cliffs north of Parton. Most of the knowledge of the faulting has come from the numerous mines sunk in this area and the faults are named after the mines.

The meeting point, (once the pit yard of Harrington No. 10) was the site of 3 shafts Nos. 7, 9 and 10, the only remains of these shafts is the gas vent out of No. 10 shaft. This shaft was the main down cast shaft and No. 9 was the up cast shaft. These shafts were the main ventilation of other mines in the locality No. 5 (Micklam Pit) and No. 11 pits.

Number 10 was opened in 1910 and sunk to the six quarter seam at a depth of 648 feet from the surface and most of the workings are under the sea. The other seams worked at No. 10 were Harrington Four Foot, Sixquarters, Lickbank, Eighteen Inch (potash), Little Main, and Main Band. All these seams were worked at sometime in this pit's life. The whole area has been reclaimed since the closure in 1968 and is very difficult to visualise as it was when it was a working mine.

It was explained that the whole area of the Pit Yard where the playing field and club house now stands has been raised some meters above the original surface. This was achieved by using the waste material from the mines old pit heap.



The party followed the old Lowca Light Railway along the cliff top to No. 5 pit (Micklam). This pit was mainly producing Fire Clay associated with the seat earth of the Lower Threequarters coal. Associated with this mine was the Micklam brick works that produced refractory for the steel industry. The only remains to be seen of the mine are the sealed up entrance to the drift and some railway lines sunk in to concrete and the foundations of the kilns.

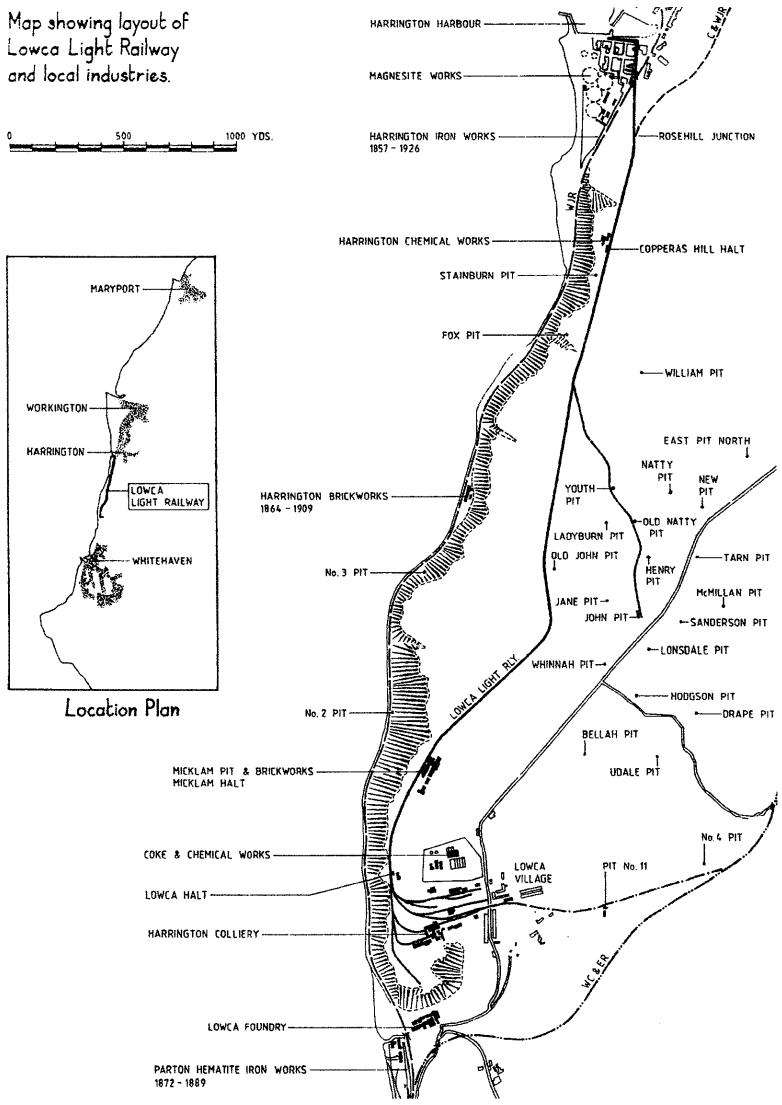
The rubble from the works has been used to consolidate the slope down to the railway and the shore.

North of this area (which we did not visit) are many old mines that have all but vanished but for some disturbed ground.

Generalised vertical section of the Coal Measures in the district
In 1673 Henry Curwen (Lord of the Manor) owned the mines in Harrington and Lowca. In 1783 the Harrington colliery consisted of 7 pits between 1783 and 1792 John Christian Curwen (Lord of the Manor) sank a further 11 pits. By 1812 only three pits were still working, and in 1847 only two were left working producing 25,000 - 30,000 tons a year. The colliery was taken over in 1865 by Bain Blair and Paterson of Harrington Ironworks who

were to eventually sink ten pits and the pits were numbered 1 to 10 but there was actually a number 11 and 12. The colliery was taken over by the Workington Iron and Steel Company in 1909.

On the 9th December 1946 an explosion occurred at No. 10 Pit which claimed the lives of 15 men. This was only three weeks before the pits were nationalised, the owners being United Steels Company.



We returned to the car park and then drove down into the village of Parton and had lunch on the shore in beautiful sunshine. Sometimes after storms and high tides you can find seat earth out cropping on this beach but on this visit we were unlucky.

The party made its way to the south end of the beach to the location of the remains of Parton Pit where there are remains of some buildings that were part of the mine. The drift and mine were all part of the Moresby Colliery which was established by the 17th century, and the owners shipped the coal out through Parton. The drift was driven for drainage purposes from the Moresby mines in 1827 and the Countess Pit shaft was started in 1832 and the mine was completed in 1836. The Countess pit is known locally as Lambhill and had been pointed out to the party when the members were on the high ground at Harringtron No. 10. It's on the east side of the Whitehaven to Workington road. The same location is well known for the Countess Sandstone quarry that in the years past has sent this stone to America. The coal from the Bannock Band, Main Band and the 6/4 seams were raised to the level of Parton Drift and then hauled along the drift to be raised 60 feet to the surface at Parton Pit. The party paused on the old wagon way and admired the Countess Sandstone cliffs before returning to the cars and home.

SMARDALE GILL

Leader: Sylvia Woodhead

23 August 2015

A large group of CGS members met at the western, Newbiggin-on-Lune, end of the CWT Smardale Gill Reserve (NY 703055) in glorious sunshine. The leader explained the location in the Lune Gap, near the source of the river Lune which flows north from Green Bell on the Howgill Fells, then west and south to Morecambe Bay. The area was glaciated; a small snowfield and glacier formed on the flat top of Wild Boar Fell over 600m and moved north, depositing glacial drift, in places shaped into drumlins. Smardale is drained by the Scandal Beck, a tributary of the River Eden. The present Scandal Beck rises on the slopes of Wild Boar Fell, to the south of Ravenstonedale, and flows north to join the river Eden north of Kirkby Stephen, but in the past joined the river further south. Several stages of river capture have been identified in classic research by Cuchlaine A. M. King. (1976). Smardale Gill is now interpreted as a post glacial river gorge.

Part of the purpose of the visit was to try out the draft text for a planned geological trail of the Smardale Reserve, designed for a non-geological audience, and also to visit two Local Geological Sites (LGS) for Carboniferous Limestone and Ashfell Sandstone.

The walk began along the former Stainmore railway line, which ran from Tebay to Darlington, transporting coal to Barrow and ores and limestone for the Teesside steel works, from 1861-1962. In the distance to the south the rounded outlines of the Howgill Fells can be seen, while the fields near the village of Newbiggin have a cover of glacial drift.

Rock exposures are now rare in the first railway cutting encountered, (NY 707056), particularly in August when the embankment sides are bright with wild flowers. However, on the north side, there were signs of small outcrops of thin, near horizontal layers of Great Scar Limestone and sandstone. At the end of the cutting, the construction of the railway bridge over the footpath was examined, and two sets of building stone could be distinguished, each a slightly different colour. One rock was coarser, feeling rough, with some pebbles and the other finer, with many masons' marks. Some blocks showed graded bedding.

An exposure of the Ashfell Sandstone (NY 709058) showed a 2 metre outcrop of sandstone, thin layers, with some cross bedding, pebbly below with the layers becoming finer grained higher up, of the Brownber Formation. Ashfell Sandstone makes strong building stone which can be seen in the adjacent barn. It is a red sandstone, well cemented with silica and has good load bearing properties.

Along the next section of path there are good views of drumlins, here showing ice movement from west to east. Continuing along the railway line, a small outcrop of Ashfell Limestone may be glimpsed on the hillside to the north.

Past Sandy Bank signal box, (NY 715062), the railway line cuts through a drumlin, and as the glacial drift is liable to slumping, the cutting has been strengthened by a stone wall, using both the Ashfell Sandstone and Ashfell Limestone blocks. Good examples of fossil corals and brachiopods may be seen in the limestone, together with chert nodules. There is a striking difference in weathering effects on the north and southern sides of the cutting, with a more recent (August 2015) sandstone wall over an arc-shaped landslip scar on the north facing side.

Beyond the cutting, the hillside to the north shows a series of benches, reflecting layers of limestone with a layer of reddish Ashfell Sandstone between. There is soon the first view, looking south-east, of the Smardale Bridge sandstone quarry, and Smardale Beck, flowing north from Wild Boar Fell, and then through Smardale Gill.

Lunch was taken at the Smardale Gill Limestone Quarry Local Geological Site, NNR (GR) and Lime kilns, a Grade II Listed Building (NY 724064). The rock exposed in the quarry is the Ashfell Limestone Formation, near the base of the Lower Carboniferous in this district on the Alston Block. It is now believed to lie above the 80m thick Ravenstonedale Limestone Group, and to comprise the oldest truly marine strata in the district. The rock is evenly bedded with siltstone partings. It was deposited in a warm clear shallow sea, rich in life, which is now preserved as fossil corals (Lithostrotion occurs both as fragments and in its living position). Productid brachiopods and Fenestella, a bryozoan, also occur. There are also layers of quartz and nodules of chert, which may be stromatolite algal mounds.

The fossils are silicified, making them stand out clearly from the weathered rock surfaces. Further along the railway line, the Ashfell Limestone Formation can be seen to be dipping about 30° to the NE, and to be dolomitised.

At NY 726068 the limestone is disturbed by a number of faults, trending mostly north to south, which have caused some gullying. The valley of the Scandal Beck south from here may also be fault guided. The view ahead shows dipping layers of limestone, known as the 'Witches Stride', on either side of Smardale.

The 80 foot high Smardalegill viaduct, now maintained by the Northern Viaduct Trust, was built from Ashfell Sandstone in 1860. Over 60,000 tons were said to have been used. The arch quoins are of Millstone Grit. The viaduct took 5 years to construct and opened in 1865. The large sandstone blocks forming the present edge of the viaduct were added in the late 1990s as a safety measure, and are not local rocks. The valley sides are steep and are still slipping down under gravity.

From the viaduct, the permissive path on the west side of the valley was taken. The path was once the track taking sandstone blocks to the viaduct: some have been left beside the path. Small exposures were seen of the thin cover of glacial drift, which here is dominantly reddish and sandy, suggesting that the ice eroded the local Ashfell Sandstone. Some if it

contains rounded pebbles, suggesting deposition by meltwater streams, and a fluvioglacial origin. The path is nearly along the junction of the Ashfell Limestone above and the conformable well-cemented Ashfell Sandstone below, the contact being often revealed as a spring line.

The Smardale Bridge Local Geological Site includes several sandstone quarries (NY 724060), with good sections through the Ashfell Sandstone. Members discussed the views expressed in a recent article about the environment of formation of the sandstone (Stanway *et al* 2015). Its red colour suggests proximity to an arid area, probably a shoreline where periodic storms generated rivers which deposited the sediment. A mixture of part-cut sandstone blocks can be seen, some of which may have been brought from elsewhere to be cut here.

From the quarries a view across the valley shows an alluvial fan, probably formed by post glacial meltwater channels. Downslope movements under gravity are exposing reddish material, derived from the Ashfell Sandstone. Examination of the stone walls and gate posts revealed coral colonies, brachiopods, chert nodules and erratics, together with firebricks from the former inn at Smardale Bridge.

Just alongside the footpath to Ravenstonedale is the main Smardale Bridge Local Geological Site.

Several small quarries operated in the 1860-70s, extracting the local Ashfell Sandstone Formation for building stone. The LGS designation describes the rock as 'rusty red fine-grained sandstone, well bedded with strong cross bedding. The sandstone is well cemented with secondary silica, making a strong building stone'. A rock fall showed bright red unweathered sandstone blocks in March 2015, but by August 2015 was already weathered to a duller colour. Lower down, a closer examination shows solution holes where lime has dissolved out. Tufa has been deposited from lime rich water percolating down from the limestone above, and trace fossils may be seen. The sandstone lies conformably below the Ashfell Limestone, forming lenses and layers, lower in the succession. The sandstone formed at the beginning of Carboniferous marine deposition here, when rivers from the north east intermittently emptied land-derived deltaic deposits into the ancient sea.

Smardale Bridge – a Grade II listed structure. This old packhorse bridge was probably constructed of rock from the smaller quarry just to the west of the bridge. Various other types of sandstone have been used later to strengthen its parapet. The ground is wetter here as the Ashfell Sandstone is well

cemented and does not let water in. Walking back up to return along the railway line, members enjoyed the wealth of flowers and Scotch Argus butterflies in the reserve. The leader thanked all for their help and suggestions towards producing a geological leaflet for Smardale.



Dipping layers of Carboniferous Ashfell Sandstone Formation and wall of lime kiln at Smardale Gill LGS



Dressed Sandstone blocks in railway bridge at NY 707056



Carboniferous Ashfell Sandstone Formation at the Smardale Bridge LGS, quarried for use in the Smardale gill viaduct on the former Smardale railway.

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NANNYCATCH GORGE AND UPPER CALDER VALLEY: GLACI-FLUVIAL FEATURES AND LOWER PALAEOZOIC ROCKS

Leader: David Kelly

20th September 2015

The party met at Kinniside Stone Circle where it was noted that the area had been visited by the Society on various occasions and accounts of these visits can be found in the Proceedings. The area was also the subject of a chapter in Edgar Shackleton's Geological Excursions in the Lake District (1975). The area is covered by the BGS sheet 28 (Whitehaven) and the accompanying memoir published in 1931. The re-surveyed Whitehaven sheet was published in 2005 and covered by the *Geology of the West Cumbria District memoir* (1997).

The morning was spent in the upper part of the catchment of Kirk Beck, a tributary of the Ehen. On descending the path from Kinniside Stone Circle to the Nannycatch Gorge, various exposures of the Buttermere Formation (BF) of the Skiddaw Group (SG) were examined. These were generally laminated siltstones and mudstones with sandstone lenses. Dips were variable and there is was cleavage. The BF has been interpreted as an olistostrome – a mélange formed by gravitational sliding under water. In the bed of the footpath itself, some highly contorted beds were examined. One olistolith almost a metre across was seen in deposits to the right of the path.

It was noted that the Nannycatch Gorge is one of two gorge sections along Kirk Beck, the second being the Stockbridge-Haile Gorge. The gorge starts at a col above the Ennerdale valley, has various tributary valleys joining, a steep gradient and meanders. It is interpreted as a sub-glacial meltwater valley.

On the east side of the gorge, terracettes were seen along the north facing sides of spurs and not the south facing sides, showing stronger soil creep.

On the Flat Fell side of the gorge there were views of river cliffs, slope failures and deposits of stratified scree which typically mask the slopes of hills in the SG.

The BGS Whitehaven sheet shows various intrusions at the end of the valley which separates Flat Fell and Dent. Some members climbed the steep, lower slopes of Dent next to Raven Crag to examine a felsite intrusion consisting of a fine-grained, pale grey rock with occasional pink feldspar crystals. Its

relationship with the BF was complex and the steep slopes made viewing difficult.

The afternoon was spent in the upper part of the River Calder catchment. On the ridge on the south side of Swarth Fell there was a small exposure of the Devoke Water Member within the Birker Fell Andesite Formation of the Borrowdale Volcanic Group (BVG). Matrix supported conglomerates suggested a mudflow type deposit.

On the side of the Calder Valley, north east of Ghyll Farm there was a good opportunity to see easily accessible exposures of the Latterbarrow Sandstone, a coarse, siliceous, purple or grey sandstone. The recent BGS mapping shows the Latterbarrow Sandstone Formation as the basal formation of the BVG in this area. It is shown lying unconformably on the Buttermere Formation of the SG, and succeeded unconformably by the Devoke Water Member of the Birker Fell Formation.

The view across the valley showed a considerable thickness of till into which the Calder has become incised. The West Cumbria memoir records a possible weathered diamicton of possible Wolstonian age in a river cliff at NY 069119 (Nirex).

There was also a good view of a river cliff on the left bank. Here, close to a pronounced bend (NY 068117), are dark, mudstones with paler silty laminations of the BF, very different to the overlying Latterbarrow Sandstone. The bedding is contorted and a small, recumbent antiform was visible. Beds of sandstone are more common towards the southerly edge of the exposure. Two fine-grained dykes were identified crossing the stream a few metres south of the sharp bend in the river. It was at this location that, during the re-survey of the Lake District by the BGS, Molyneux and Rushton (1984) discovered trilobites and acritarchs of late Tremadoc age close to, and apparently immediately beneath, the Latterbarrow Sandstone.

In deteriorating weather a few members ascended the slopes between the summit of Blakeley Raise and Kinney How. Here a small dry channel, possibly eroded by an ice-marginal stream was examined. Further down this channel before it reaches Comb Beck a small felsite dyke was seen intruding the BF.

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LECTURE MEETINGS

WINTER 2014/15

24th September 2014
Queen Elizabeth Grammar School, Penrith
'Extreme avalanche mobility: theories, experiments and for real'
Prof. Peter Kokelaar (University of Liverpool)

22nd October 2014
Friends Meeting House, Cockermouth
'The Earth's Enigmatic Inner Core'
Dr Jon Mound (University of Leeds)

12th November 2014
Community Hall, Bolton St. Workington
'Rootless Cones and Scoria Cones: insights into basaltic volcanism'
Peter Reynolds (Durham University)

3rd December 2014
Friends Meeting House, Keswick
Members Evening

21st January, 2015
Tullie House Carlisle (Joint meeting with Carlisle Natural History Soc.)
'Hematite Mineralization at the margins of the east Irish Sea Basin: Cumbria, N. Wales, Isle of Man.
Dr Stephen Crowley (University of Liverpool)

11th February 2015
Queen Elzabeth Grammar School, Penrith.
'The Evolution of an Upper Permian Evaporite Basin in Northwest England'
Dr Noel Worley

11th March 2015
Friends Meeting House, Keswick.
Members Night

LECTURE MEETINGS

WINTER 2015/16

23rd September 2015
Queen Elizabeth Grammar School, Penrith.
'The New Crofton Co-operative Colliery'
Toby White.

14th October, 2015.
Community Hall, Bolton St. Workington.
'Volcanoes of Iceland and Chile'
Dr. Dave McGarvie, Open University.

11th November 201.
Friends Meeting House, Cockermouth.
'Current Exploration for coal off the west coast at Whitehaven'.
Eric Gozlan (West Cumbria Mining).

9th December, 2015
Members Evening.
Unfortunately this event had to be cancelled because of the severe flooding of the venue at the Friends meeting House, Keswick.

20th January 2016.
Tullie House, Carlisle (Joint meeting with Carlisle Natural History Soc.)
The Lake District's tuff hailstones.
Clive Boulter.

10th February 2016.
Friends Meeting House, Cockermouth.
'The Biggest Hole in the World'.
Richard Walters.

2nd. March, 2016.
AGM and Members Night.
Braithwaite Institute, Keswick.

ANNUAL DINNERS

The 2015 Annual Dinner was held on Saturday 21st March at the Skiddaw Hotel, Keswick. The speaker was Dr. Alan Smith who spoke on 'Six Men – the contribution of amateurs to the geology of Cumbria'.

The 2016 Annual Dinner was also held at the Skiddaw Hotel in Keswick on Saturday, March 19th. Professor Chris Wilson, Emeritus Professor of Geology from the Open University was the guest speaker. His talk was entitled. 'Geology at the Open University: How it all began'.

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O B I T U A R Y

Gordon Ritson

Gordon died on 19th October 2015 aged 78 years.

He came to The Barns at Moat, Longtown from Consett, Co. Durham where his father worked at the Consett Iron Works. The Ritsons had once been in the lead industry in Upper Allendale and moved to the Durham coalfield.

In 1947 they moved to the Moat, near Longtown, to a farm, known as The Barns (originally a tithe barn for Arthuret). The furniture lorry was delayed for four weeks because of the snow drifts between Consett and Longtown. In 1949 Gordon went to Carlisle Grammar School, the journey by train on the now defunct Waverley Line. There he got a taste for reading which lasted all his life – interested in travel, natural history and geology. After school he joined the RAF as a radar technician, serving in Germany and Holland and often with American units. He travelled to Denmark and other parts of Europe. Leaving the RAF he served a long time at the RAF Maintenance Unit 14 at Carlisle, repairing electronic equipment of all kinds.

He was interested in the countryside, often with a geological bent – with local sites such as Penton Linns and Carwinley Burns, St Bees sandstones, shales, conglomerates and a fossilised Carboniferous anticline beneath the New Red Sandstones, all on his doorstep. This led to his varied reading on geological matters and a greatly informed interest and knowledge in them. He enjoyed Yorkshire Geological Society excursions to Northumbria, Derbyshire and Yorkshire and particularly the Cumberland Geological Society lectures and visits. He helped on the farms of his brother and nephew all his life, latterly doing the bookwork and stewardship schemes.

His death was sudden and not expected. He was a good friend and companion for over 60 years.

Iver Gray

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