

Wendy M. Peace · Malcom W. Wallace
Mark P. Holdstock · John H. Ashton

Ore textures within the U lens of the Navan Zn-Pb deposit, Ireland

Received: 15 May 2002 / Accepted: 17 October 2002 / Published online: 12 February 2003
© Springer-Verlag 2003

Abstract Recent exploration in the vicinity of the giant (>90 Mt) Navan orebody has resulted in the discovery of ore-grade mineralisation to the southwest of the deposit, much of which occurs within the Upper Pale Beds, a horizon that is only weakly mineralised above the main orebody. Within this new U lens, mineralisation preferentially occurs within bioclastic carbonate grainstones and calcareous quartz sandstones, and is dominated by sulphide replacement of the carbonate component of the host sequence. Much of the replacive mineralisation is spatially associated with hydrothermal cavities, which are filled by a variable mixture of brecciated replacement sulphide minerals, space-filling sulphide and gangue cements, and internal sediments. Mineralisation also occurs within veins and dissolution seams, and as disseminated sulphide minerals. Massive mineralisation is typically a complex, chaotic, combination of replacement, cavity, and fracture-filling sulphides. Fluid inclusion analyses of ore-stage saddle dolomite indicate temperatures at the time of sulphide precipitation of ~90–150 °C, with a maximum of 175 °C. These temperatures are lower than those typically proposed for Irish-type deposits.

Keywords Irish-type · Navan · Epigenetic · Zinc

Editorial handling: A.C. Brown

W.M. Peace (✉) · M.W. Wallace
School of Earth Sciences, University of Melbourne,
Victoria, 3010, Australia
E-mail: wendy.peace@woodside.com.au
Tel.: +61-8-93485053
Fax: +61-8 93486644

M.P. Holdstock · J.H. Ashton
Outokumpu—Tara Mines Ltd.,
Knockumber, Navan, Co Meath, Ireland

Present address: W.M. Peace
Woodside Energy, 1 Adelaide Terrace,
Perth, WA, 6000, Australia

Introduction

The Navan Zn-Pb orebody, discovered in 1970, is by far the largest of the known deposits in the Irish orefield, with ~90 Mt of resources (Ashton et al. 2000). The discovery of the South West Extension (SWEX) during the mid-1990s has increased ore resources by more than 10 Mt, and necessitates a reexamination of ideas concerning the processes and controls involved in the localisation of ore at Navan. The discovery of significant mineralisation in part of the stratigraphy previously thought to contain no ore grade mineralisation – the Upper Pale Beds – led to a detailed study of both this new ore horizon (the U lens) and host lithologies (the Upper Pale Beds). In this paper, we describe for the first time the ore textures and their occurrences within the U lens, the most recently discovered ore lens at Navan.

Regional geology

The Navan Zn-Pb orebody is located 1 km northwest of Navan, Co. Meath (approximately 50 km northwest of Dublin), towards the northern margin of the Dublin Basin (Fig. 1). Mineralisation is hosted within a shallow-water Early Carboniferous (Courceyan) carbonate sequence informally known as the Pale Beds (Philcox 1984). The deposit lies on the northern flank of a strongly faulted, northeast-trending anticlinal structure at the southwestern margin of the lower Palaeozoic Longford-Down inlier (Ashton et al. 1986). Most of the mineralisation occurs as a series of stratabound lenses, which dip as much as 20° to the southwest. The stratigraphically highest (and most recently discovered) of these is the U lens (Fig. 2).

In the vicinity of the orebody, deformed lower Palaeozoic rocks are unconformably overlain by a carbonate-dominated Courceyan to Arundian (363–343 Ma) succession deposited in a progressively deepening environment. The geology of the Navan area has been described in some detail by numerous workers (e.g. Philcox 1984, 1989; Ashton et al. 1986, 1992; Rizzi and Braithwaite 1997; Anderson et al. 1998) and is summarised in Fig. 2 and Table 1. Although a formal nomenclature for the units present in the Navan area has been put forward (Strogen et al. 1990), the previously proposed informal nomenclature (Philcox 1984, 1989) is the more commonly used and widely accepted, and thus is favoured here.

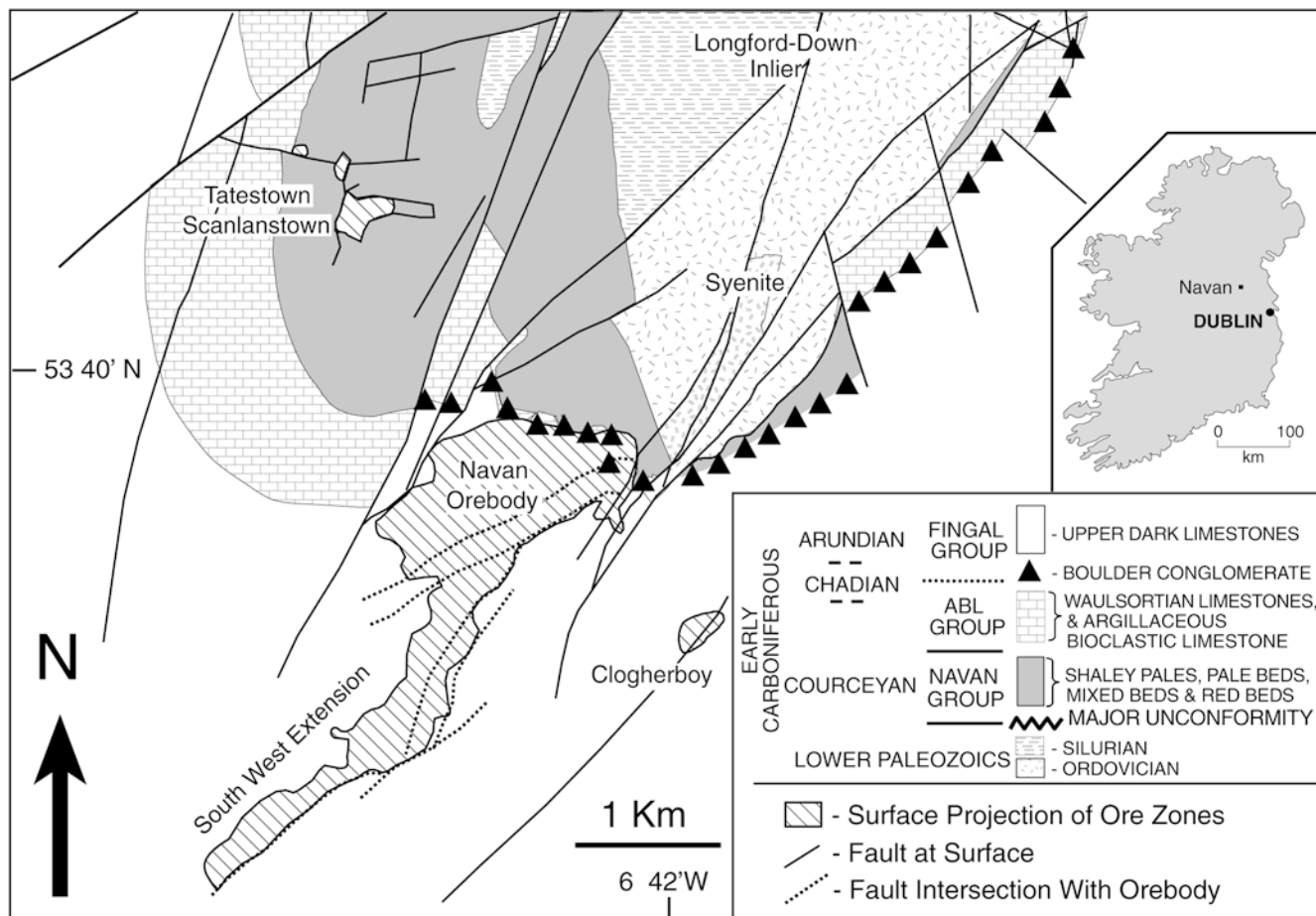


Fig. 1 Geology of the Navan area showing the location of the Navan and satellite orebodies (modified from Ashton et al. 2003)

A majority of the ore at Navan (>97%) is hosted within the Pale Beds (Fig. 2, Table 1), a shallow-water carbonate-dominated sequence of Courceyan age. The Pale Beds grade from micritic mudstones and wackestones at the base to predominantly grainstones and packstones at the top. The detrital quartz content increases upwards through the unit. The Pale Beds have been interpreted as having been deposited in gradually deepening water in a high-energy carbonate ramp or broad tidal belt environment (Andrew and Ashton 1985). Within the Pale Beds, seven stratigraphic marker horizons (numbered upwards from 5 to 0, and U) are defined using stratigraphic marker horizons (Fig. 2). However, in the main area of the mine, mineralisation may be vertically continuous over as much as 80 m.

The remainder of the ore (<3%) at Navan is hosted by the Boulder Conglomerate, a series of chaotic, polymict debris-flow deposits that overlie a southward-sloping submarine erosion surface of pre-Arundian age. This surface has eroded down to the Pale Beds in the deposit area, resulting in areas where the Pale Beds are directly overlain by the Boulder Conglomerate (Fig. 3).

The stratigraphy at Navan is offset by a series of major east-northeast- and northeast-trending faults. These structures have been described in some detail by several authors (e.g. Andrew and Ashton 1985; Ashton et al. 1986, 1992). The faults are the products of reactivation of major basement fault zones during the late Courceyan-Chadian and Hercynian, respectively. These faults are used to divide the orebody into three zones (Fig. 3): north of the B fault (zone 1), between the B and the T fault (zone 2), and south of the T fault (zone 3). The majority of the U lens occurs within zone 3.

Geometry of the U lens and sulphide mineral distribution

The U lens is defined as any mineralisation occurring between the base of the Upper Sandstone Marker and the base of the Lower Shaley Pales (Fig. 2). Along with the underlying (generally poorly mineralised) 0 lens (between the base of the Upper Dark Marker and the base of the Upper Sandstone Marker), the U lens makes up the interval known as the Upper Pale Beds (Philcox 1984). In the area of the main orebody, this interval is commonly absent due to removal by the erosion surface. However, where the Upper Pale Beds are present in this area, the U lens is at best only poorly mineralised. The U lens mineralisation occurs in an area to the southwest of the main orebody (Fig. 3), and together with lesser sulphide mineralisation within the Boulder Conglomerate (e.g. in the 0 and 1 lenses), this area has been termed the southwest extension (SWEX) (Ashton et al. 2000). Recent exploration in the SWEX has resulted in the delineation of a resource of 13.5 Mt at 8.9% Zn and 1.8% Pb (Ashton et al. 2000). Although rare and not ore grade, minor mineralisation also occurs in the Lower Shaley Pales in this area.

The stratigraphic interval that hosts the U lens was originally 40 to 50 m thick, but is now commonly less due to truncation of the host sequence by the erosion

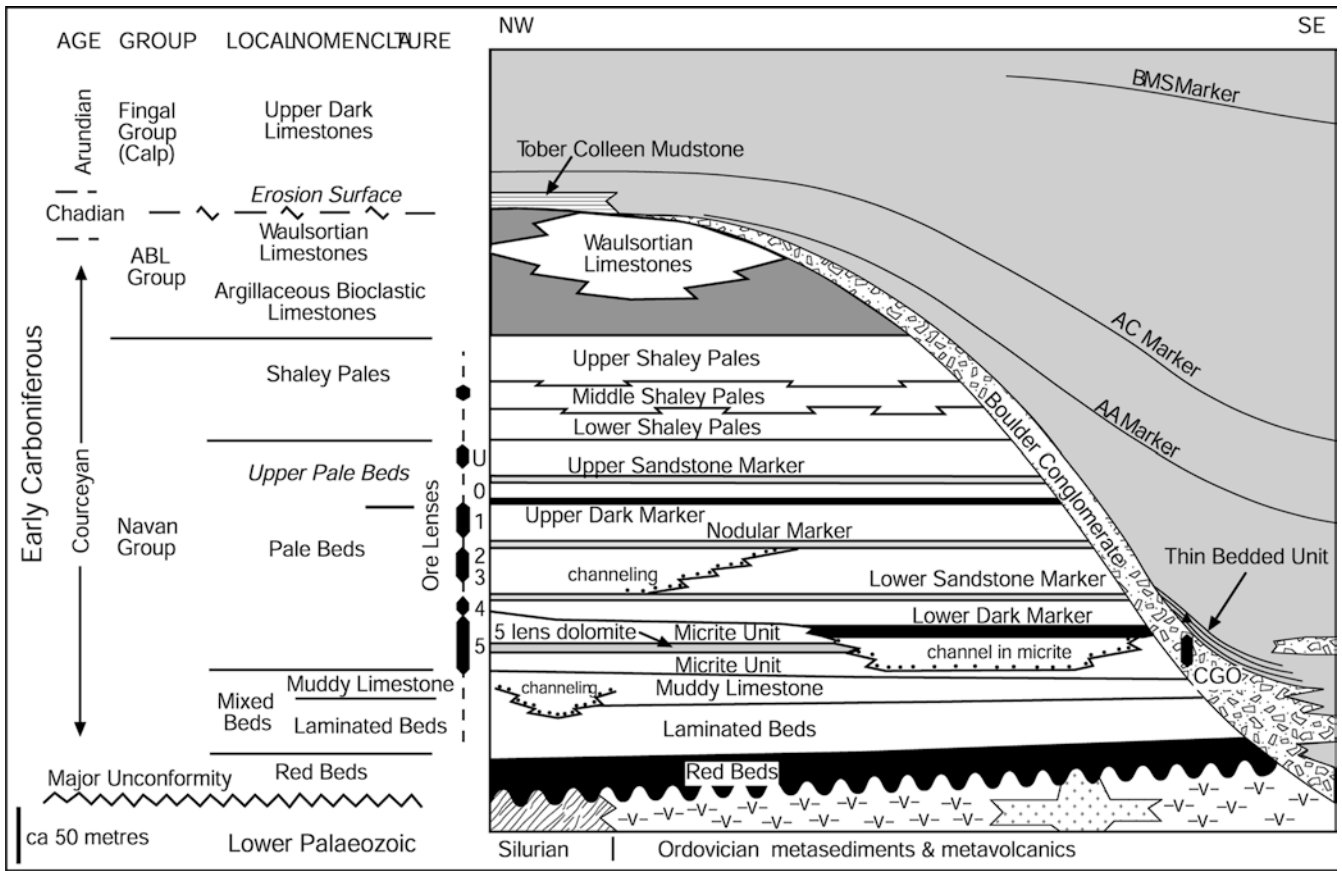


Fig. 2 Stratigraphic sequence in the Navan mine area, including marker horizons used to delineate the ore lenses. Positions of the individual lenses are also shown. CGO Conglomerate Group Ore (modified from Ashton 1995)

surface or post-depositional faulting. The sequence thickens to the southwest and consists of interdigitating lenses of calcareous sandstone, oolite and bioclastic grainstone with marked lateral facies changes. Mineralisation was preceded by dolomitisation of the host rocks, although it varies in intensity, with the entire

Table 1 Formal and informal nomenclature for the lithostratigraphy of the NW part of the Dublin Basin, west Co Meath (after Philcox 1984, 1989; Ashton et al. 1986; Strogon et al. 1990)

		Formal nomenclature	Informal mine nomenclature	Rock types		
Early Carboniferous	Arundian	Fingal Group	Lucan Formation	Upper Dark Limestone	Middle Limest. Group	Limestone and shale turbidites
		Fingal Group	Tober Colleen Formation	Boulder Conglomerate		Debris flows
	Chadian	Cruicetown Group	Feltrim Formation	Waulsortian Limestone	ABL Group	Interbedded 'mudmounds' & crinoidal limestones
			Slane Castle Formation	Argillaceous Bioclastic Limestone		Interbedded crinoidal limestones & shales
	Courseyan	Navan Group	Moathill Formation	Shaley Pales	Navan Group	Limestones, shales, & thin sandstones
			Meath Formation	Pale Beds		Shallow water limestones
			Stackallan Mbr	Micrite Unit		Micrites
			Bishopscourt Mbr	Muddy Limestone		Shaley limestones
			Liscartan Formation	Laminated Beds		Interbedded sand, shale & limestone
	Old Red Sandstone facies	Red Beds	Sandst, mudst, & conglom.			
			Lower Palaeozoic Rocks			Sediments & volcanics

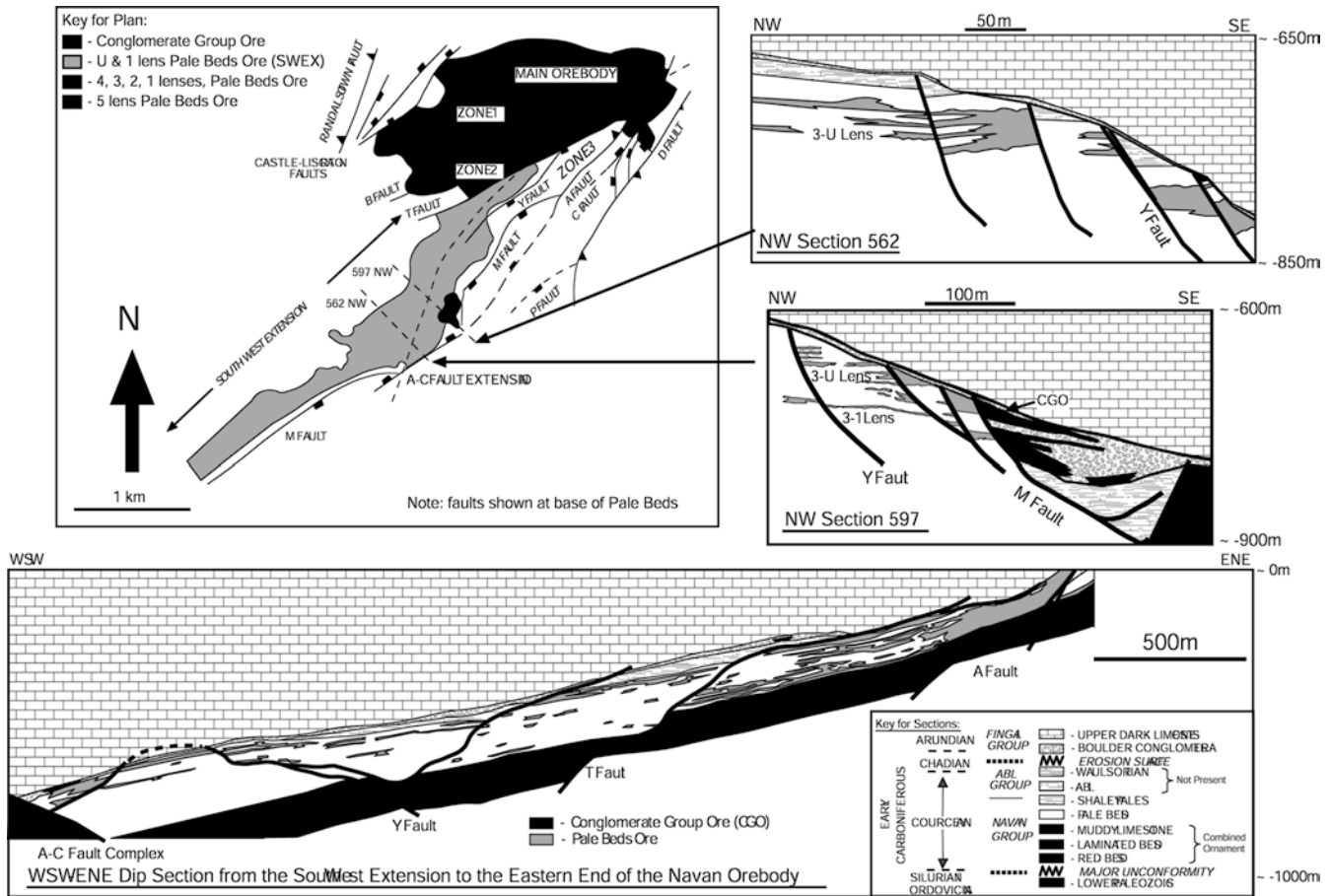


Fig. 3 Simplified plan of the Navan orebody, showing the generalised ore distribution and structure. Sections show the distribution of ore within the southwest extension (SWEX) (modified from Ashton et al 2000. Section 562 NW modified from Peace 1999)

Upper Pale Beds replaced by fine- to medium-grained euhedral dolomite to the northwest of the SWEX, with more minor dolomitisation to the southeast.

The distribution of sulphide minerals within the Upper Pale Beds is illustrated in Fig. 4. Although the overall stratabound nature of the ore is readily apparent in cross section (Fig. 3), it is clear that the ore (especially the low-grade, disseminated sulphides) may also cross-cut the stratigraphy. Thus the ore is stratabound, but not stratiform. Mineralisation can be developed throughout the lens, but in the up-dip section of the U lens (the area studied) it occurs preferentially at three general stratigraphic intervals (Fig. 4). These horizons are just above the Upper Sandstone Marker, immediately beneath the contact with the Shaley Pales, and beneath a sandstone marker (SST) in the upper section of the lens. Sulphide mineralisation appears to display an inverse relationship to the planar replacement dolomite which preceded it, with the highest concentrations of ore where dolomite is minor or absent. However, the correlation is not a precise one, as sulphide minerals do occur within dolomitised beds.

Lateral continuity of the mineralisation has been affected by the offset of the ore-bearing beds by a series of ENE-trending extensional faults, which include the B, T and Y faults. As has also been noted by Anderson (1990), these faults do not display evidence of mineralisation, although the volume of sulphide minerals tends to increase close to the faults. The abundance of sulphide minerals is also generally greater in the footwall of these faults. This observation is most apparent in the footwall of the M fault.

Methods

The Upper Pale Beds (approximately the upper 60 m of the Pale Beds) were logged and sampled from more than 70 mine and exploration drill cores from the mine lease. From these samples, polished thin sections of the Upper Pale Beds and the overlying Boulder Conglomerate were produced, from which descriptions of the sulphide mineral paragenesis were produced. Microthermometric analyses of primary and pseudo-secondary inclusions were performed using a USGS gas-flow heating/freezing system from Fluid Inc. attached to a Zeiss Universal polarising microscope with a final magnification of 500 \times . Temperatures were measured using the methods detailed in Goldstein and Reynolds (1994). Temperatures were measured to a range of 2 $^{\circ}$ C or less, with temperatures

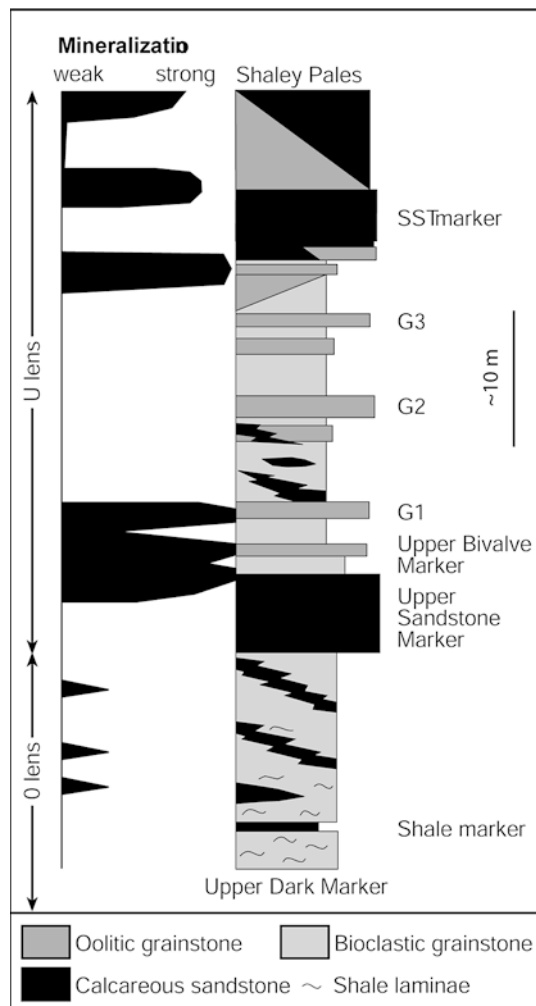


Fig. 4 Generalised stratigraphy of the Upper Pale Beds showing the location of base metal mineralisation (modified from Peace 1999)

reported as the mid-point of the two values achieved within this range. The accuracy of the heating/freezing system is considered to be ± 0.1 °C for homogenisation temperatures (T_H), and thus the total error margin of the values given is ± 1.1 °C.

Sulphide mineral textures

As in the Navan deposit as a whole, sulphide mineralisation within the U lens is dominated by sphalerite. Minor amounts of galena, pyrite, and marcasite are also generally present, with pyrite/marcasite locally dominant on rare occasions. Sulphide replacement of the carbonate host rock is the most important style of mineralisation, accounting for an estimated 70–80% of the ore, with the remaining sulphides occurring as open-space cements (both cavity and fracture-fill) and (to a lesser extent) as internal sediments. Due to the nature of the mineralisation, gangue minerals are relatively rare; where present, the non-sulphide gangue is typically a

combination of barite, saddle dolomite and calcite, with barite the most common of the three.

Mineralisation within the U lens displays a variety of textures and fabrics. These mineralisation textures include: (1) massive replacement, (2) irregular cavity and breccia systems, (3) dissolution seams, (4) dark mosaic breccias, (5) fracture-filling sulphides, and (6) disseminated sulphides.

Massive replacement mineralisation

The massive mineralisation in the U lens is characterised by the virtually complete replacement of the host carbonate by sphalerite (and lesser amounts of pyrite and galena), the exception being fine grains of detrital quartz and feldspar. The detrital quartz grains commonly display euhedral quartz overgrowths, a feature that has not been recognised in unmineralised samples. Replacement varies from mimetic to destructive in nature, with near perfect preservation of bioclast textures in some samples, and complete obliteration of primary sedimentary textures in others (Fig. 5A, B). Massive mineralisation is generally far more complex than simple replacement of the host rock, with space-filling cements and sediments commonly present in at least minor quantities (Fig. 5C), and in places making up more than 50% of the ore.

Where fracturing and brecciation of the sulphide minerals occur (a fairly common feature), the resulting porosity is filled by sphalerite, pyrite and galena cements, with minor amounts of barite, saddle dolomite and/or calcite. The sulphide minerals are typically colloform in nature, although dendritic galena and stactactitic to botryoidal pyrite also occur. Both replacement and space-filling sulphide minerals can be cut by sulphide-filled fractures, which typically feature a final phase of blocky honey-coloured sphalerite overlain by coarse barite and/or saddle dolomite. Several episodes of brecciation and sulphide cementation may be evident, illustrated by clasts of fractured sulphide cements that have been encrusted by subsequent generations of sulphide minerals (Fig. 6A).

Replacement mineralisation can be crosscutting in nature, or may be confined to specific beds. Such bedding-parallel replacement may extend for only a few metres, such as that observed underground in heading 5009F on level 1140, or for tens of metres, such as replaced bedding seen at heading 5001 L1 on the 1140 level. In the 5009F heading, pale pinkish-brown sphalerite has replaced an interval of carbonate approximately 0.5 m thick. Bedding textures are visible within the sphalerite, which die out after a few metres. A large (filled) cavity occurs beneath the bedding-parallel replacement sphalerite (Fig. 6B). In 5001 L1, mineralisation can be traced within a bed that is < 0.5 m thick for more than 30 m. However, the entire thickness is not replaced over most of this distance. Instead, individual layers are replaced within the bed by reddish sphalerite, frequently joined by cross-cutting mineralisation. Several sub-vertical veins

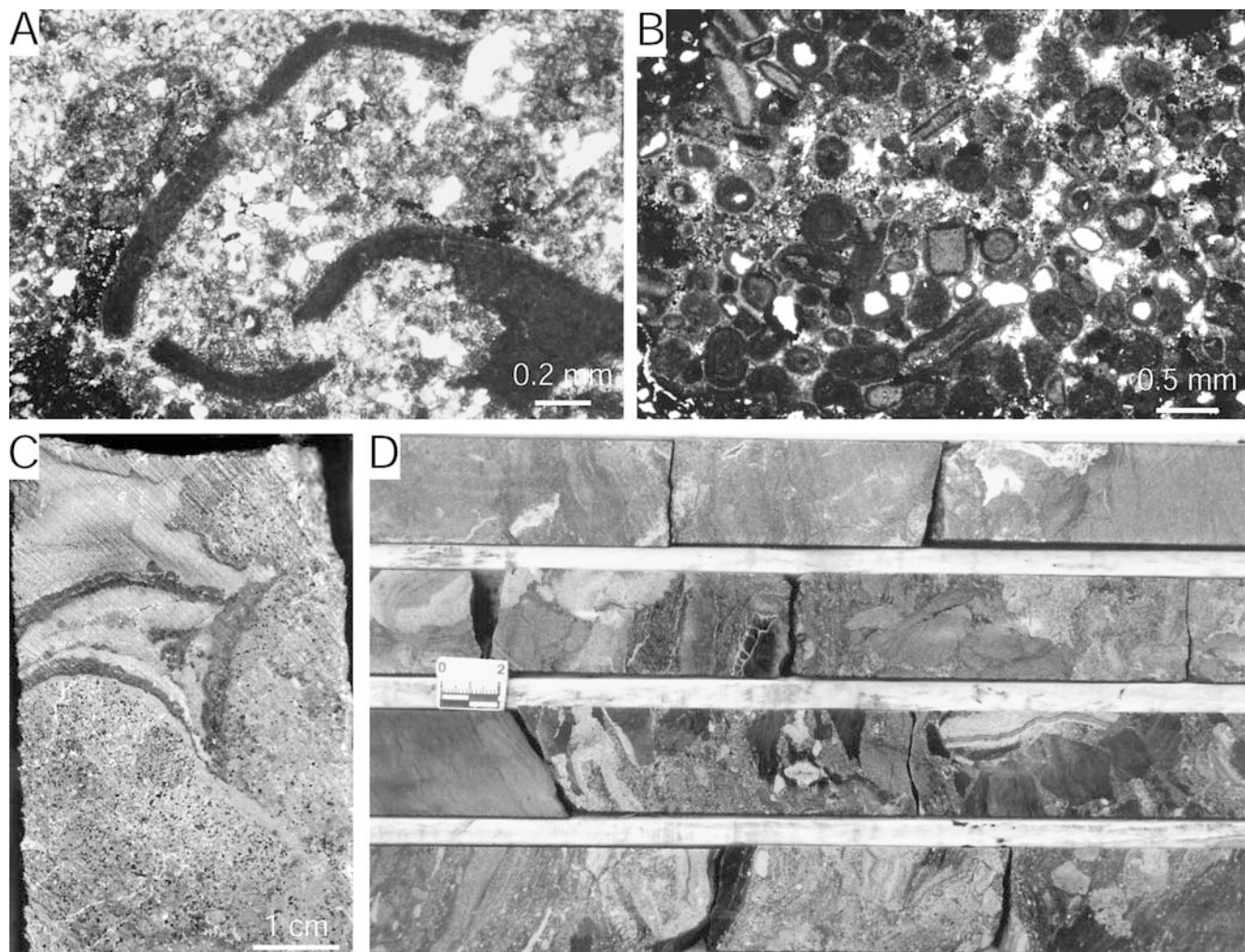


Fig. 5 **A** Replacive spherulite with preservation of bioclast textures. *Bright grains* are quartz. Heading 5009F, 1140 level. **B** Close-up view of a clast of replacive spherulite with mimetic replacement of ooids. N1159, 543.8 m. **C** Massive replacement of sandy carbonate by spherulite. *Fine dark spots* are unreplaced detrital quartz. Small cavity within the replacive spherulite is lined with sulphide cements and internal sediments. N1364, ~462.5 m. **D** Complex, mineralised breccia within a hydrothermal cavity. Clasts are dominantly unmineralised host carbonate and replacive sulphides. The matrix is composed of sandy carbonate, insoluble residue and disseminated spherulite. N1083. Up is to the *left*

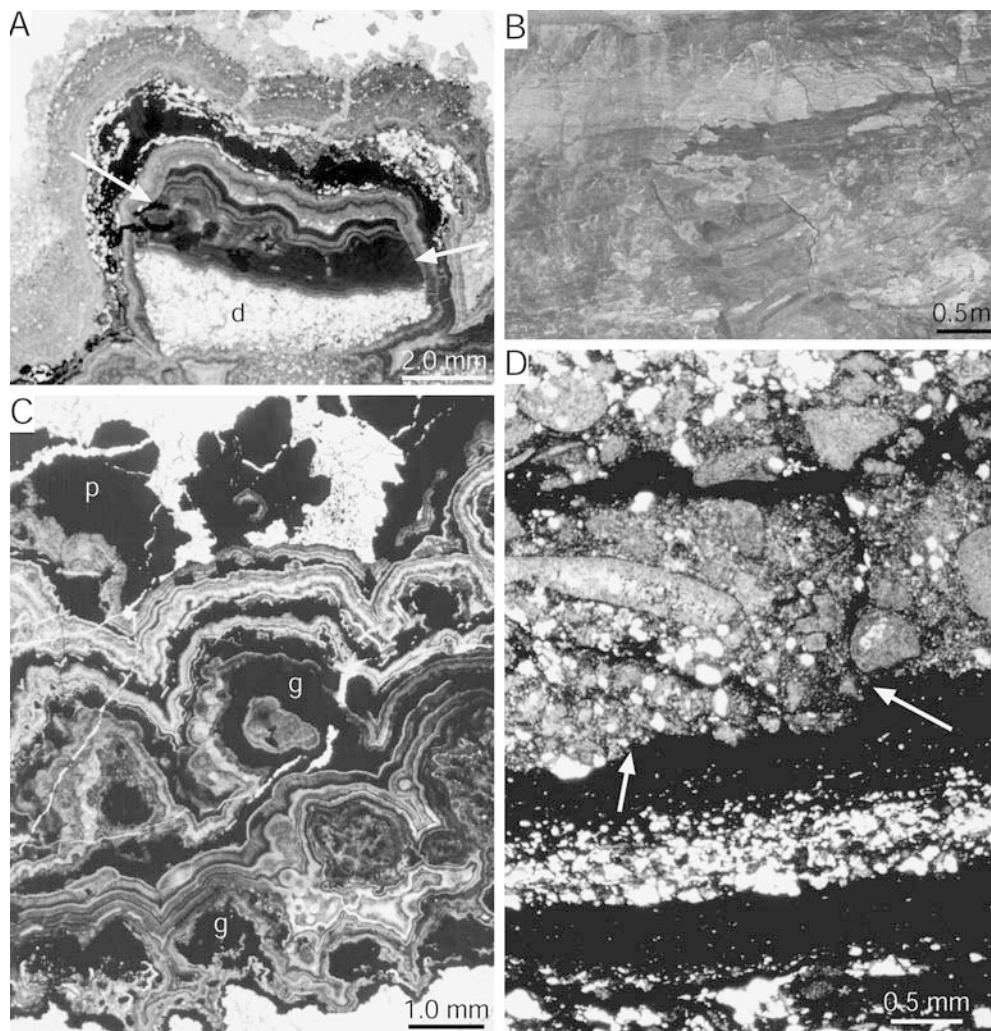
and cavities filled by mineralised breccia connect with this interval from below, which is approximately 3 m above the Upper Bivalve Marker. The morphology of this mineralisation gives the impression that mineralisation has stopped upwards from near the Upper Bivalve Marker, crosscutting bedding until a suitably porous and permeable unit was reached, and has then moved laterally along it, partially replacing the carbonate with spherulite.

Irregular cavity and breccia systems

Ore-filled irregular cavity and breccia systems consist of a combination of mineralised clasts, space-filling

cements and internal sediments. The relative proportions of clasts, cements and sediments vary considerably, with cavities ranging from entirely clast-filled with few internal sediments, to very few clasts and almost entirely filled by internal sediments (e.g., Fig. 6B). Cavities filled with largely unmineralised clasts surrounded by sulphide cements were less commonly observed. As can be seen in Fig. 6B, the irregular cavity and breccia systems are completely filled, with little or no space remaining. The degree to which these cavities are mineralised also varies considerably. Although these mineralised cavities display many variations in the composition of the cavity fill, the cement-rich cavities typically recognised as being associated with epigenetic deposits (i.e. large quantities of coarse calcite/dolomite/fluorite/barite cements and preserved space; e.g., Sangster 1988; Dzulynski and Sass-Gustkiewicz 1989) are only rarely observed here, and only on a relatively small scale within dolomitised units. The large cavities associated with significant sulphide mineralisation tend to be dominated by internal sediments that are typically poorly mineralised, rather than cements, and lack any open space, being completely filled by a combination of clasts, fine- to medium-grained cements and internal sediments.

Fig. 6 **A** Clast of dolomite (*d*) and colloform sulphides which are truncated at the clast edge (*arrows*), coated by later generations of sphalerite and galena. N1372, 723.4 m. **B** Large, bedding-parallel cavity formed beneath bedding-parallel replacement sphalerite. Cavity is filled by internal sediments and minor sulphides. Heading 5009F, 1140 level. **C** Colloform sphalerite cements with minor galena (*g*) and pyrite (*p*). N1377, 786.2 m. **D** Graded internal sediments overlain by a clast of bioclastic limestone replaced by sphalerite. *Arrows* indicate the base of the clast. N1393, 721.5 m



Clasts

As within the massive sulphide horizons, replacement dominates where clasts occur in the cavity and breccia systems. Some clasts of replacement sulphide minerals have few (if any) internal textures preserved, whereas in others, mineralised bioclasts with varying degrees of textural preservation are common, typically replaced by dark to medium-brown sphalerite. Preserved primary textures often indicate that the host rock was originally a bioclastic grainstone. Although less common, some breccia clasts are entirely or almost entirely composed of sulphide cements (Fig. 6A). These clasts are typically found in zones of high-grade mineralisation. The sulphide cements are mainly colloform in nature, although dendritic galena and rosettes of bladed barite were also observed within some breccia clasts. Clasts composed of interlaminated internal sediments and sulphide minerals have also been recognised in underground exposures. The clasts can vary considerably in size, with some mineralised clasts within the larger cavities being 0.5 m in length.

Clasts of all compositions (including unmineralised clasts) are often encrusted on one or more edges by bands of sphalerite (\pm galena and/or pyrite) cements

that are truncated at the clast edge. The truncated edges are normally coated by a later generation of sulphide cement, typically (but not exclusively) the blocky, honey-coloured sphalerite associated with the waning phases of mineralisation.

The textural style of mineralisation making up the breccia clasts appears to be dependent on the mineralisation textures in the surrounding host rock, as this appears to be the source of the clasts. Thus, the mineralisation within dolomite-hosted cavities occurs as medium- to coarse-crystalline cements coating clasts of brecciated dolomite. These clasts feature little or no sulphide replacement. The sulphide minerals within these clasts occur as fine blebs of pale sphalerite. Staining samples of dolomite breccia reveals textures suggesting that this sphalerite either replaced the minor amounts of calcite remaining in the rock or filled the porosity between the dolomite rhombs.

Cements

The styles of space-filling cements within the breccia systems are similar to those associated with massive

replacement-dominated mineralisation: dominantly colloform sulphide minerals (Fig. 6C), bladed barite (either as individual crystals or rosettes), rare dendritic galena, and, in smaller cavities, botryoidal to stalactitic pyrite. Sulphide and gangue cements rarely dominate the cavity-fill, generally comprising less than 40% of the cavity-filling material. The sulphide cements are typically fine- to medium-grained in size, whereas the dolomite and lesser calcite gangue are fine- to coarse-grained. Although small cavities filled almost entirely by pyrite are present, these occurrences are the exception, with cavities on all scales generally dominated by sphalerite.

Sediments

Sulphide minerals also occur within cavities in the form of layered internal sediments. Two types of sulphide sediment occur: carbonate internal sediments that have been replaced by sulphide minerals, and 'genuine' primary sulphide sediments that have settled onto the cavity floor from or through the fluid filling the cavity. True sulphide sediments typically feature alternating bands of fine-grained sphalerite, which vary in colour, and that can normally be correlated with synchronous bands of cements of the same colour on the roof of the cavity (Fig. 5C). Similar internal sediments composed of sulphide minerals and detrital residue have been noted in the Mississippi Valley-type (MVT) deposits of Upper Silesia (Sass-Gustkiewicz et al. 1982). On relatively rare occasions, sphalerite and galena sediments also occur with a 'snow-on-roof' texture, where the sulphide minerals coat only the upward-facing edges of clasts and wall rock.

Some layers of internal sediments within cavities feature more replacement-type textures. Such sulphide 'sediments' are typically a uniform colour and display an intergrown texture, commonly enveloping grains of detrital quartz and feldspar. The quartz grains surrounded by this replacive sphalerite commonly display euhedral quartz overgrowths. Whereas the quartz and other detrital material may display grading or sorting, the accompanying sulphide minerals do not, maintaining a fairly uniform grain size throughout. Finely bladed barite, also a common feature of these replacive sediments, occurs as individual crystals or as crossed, intergrown twins. The barite probably replaced the original carbonate component of the internal sediments.

Dissolution seams

Dissolution features within the U lens are not solely confined to cavities; dissolution seams are also common in the rocks surrounding sulphide mineralisation. Distal to the ore, dissolution seams occur as fine, generally low- to medium-amplitude stylolites, which are overgrown by a few fine blebs of sphalerite \pm galena and/or pyrite, and that differ little from the typical stylolites developed by chemical compaction due to burial. However, closer to the ore, some dissolution seams are obviously not

simply the result of burial processes. Seams become much more abundant proximal to sulphide mineralisation, where they are thicker (as much as several centimetres), have lower amplitudes, and may be bedding parallel or crosscutting in nature. Crosscutting stylolites have not been observed distal to sulphide mineralisation. A similar relationship between dissolution seams and sulphide mineralisation has been noted in the southeast Missouri MVT district by Heyl (1983, p. 53), who stated that "stylolites are everywhere in or near ore, rare or absent elsewhere; they decrease outward."

Dissolution seams may occur as a solitary seam a few millimetres to several centimetres thick or as a band of thin, anastomosing seams. Bedding-parallel seams, ≥ 1 cm in thickness, are common at the contact between different lithologies, such as bioclastic grainstone and oolite. The dissolutional nature of these seams is readily apparent here, with carbonate clearly removed from both the ooids and bioclasts, resulting in sharp contacts between both lithologies and the dissolution seams. Although the dissolution seams quite commonly appear unmineralised in hand specimen, it is not uncommon for fine-grained sulphide masses (mainly sphalerite), ≤ 1 mm in diameter (either as individual, disseminated crystals, or in small clusters), to be present within them. Where seams are quite thick, such sulphides may be visible in hand specimen (Fig. 7A). These mineralisation-related dissolution seams have also been observed crosscutting earlier calcite veins, indicating that mineralisation has occurred after initial burial diagenesis of the host sequence.

Dark mosaic breccias

The dissolution seams described above commonly grade into particulate crackle breccias (non-genetic breccia

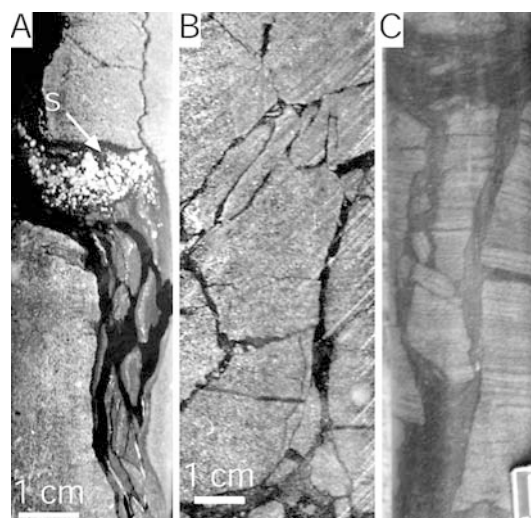


Fig. 7 A Large subvertical dissolution seam containing fine blebs of sphalerite (s). U11490, 78.7 m. B Hydrothermal particulate crackle breccia in bioclastic limestone. N1159, ~521.7 m. C Hydrothermal particulate mosaic breccia in the sst marker. N1348, ~542.3 m. Scale bar is 1.0 cm

classification of Morrow 1982), where the rock is crosscut by dissolution seams. Individual clasts, however, are still tightly packed together and show little or no rotation (Fig. 7B). In turn, these particulate crackle breccias typically grade into clast-supported to matrix-supported particulate mosaic breccias of subrounded to subangular clasts of host rock within a fine-grained black matrix (Fig. 7C). These 'dark mosaic breccias' bear a strong resemblance to the Black Matrix Breccias observed at Lisheen, Silvermines and Galmoy (M. Hitzman, personal communication 1998 and observations of principal author).

The particulate crackle breccias and dark mosaic breccias, as well as the dissolution seams themselves, occur in both dolomitised and undolomitised horizons. The dark material that forms the matrices of these breccias and fills the dissolution seams is the same material that fills the irregular cavities discussed previously: detrital quartz and feldspar grains, dark, opaque material, and, where the host is dolomitised, fine-grained dolomite rhombs. As illustrated in Fig. 8, the dissolution seams and brecciation appear to have a zonal relationship to mineralisation, with fine (<1 mm), stylolitic seams the furthest from, and dark mosaic breccias the closest to, sulphide mineralisation.

Fracture-filling sulphide minerals

Mineralised fractures occur both as interconnecting networks, which are typically below or between massive or cavity-filling mineralisation, and as individual veins.

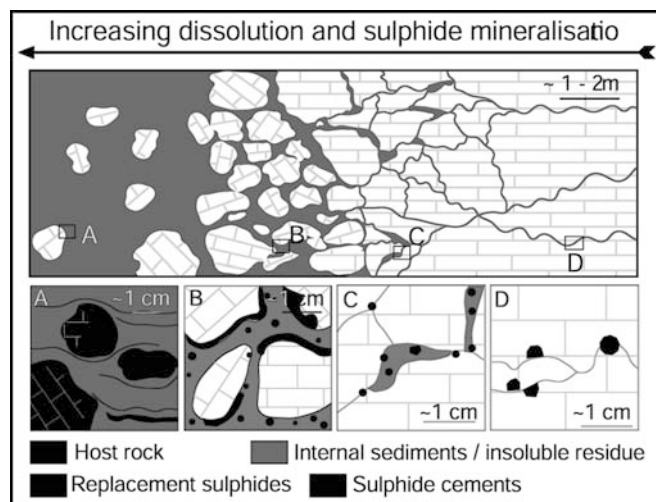


Fig. 8 Simplified representation of the relationship between dissolution and sulphide mineralisation: within the cavity, clasts of both replace and space-filling cement sulphides occur (A); as distance increases from the centre of the cavity, less mineralised clasts occur (B). Further removed, small cavities are connected by fine dissolution seams (C), with associated fine, disseminated sulphides, until furthest removed (D), dissolution is present as fine dissolution seams with occasional fine, mm-sized sulphide crystals (modified from Peace 1999)

Networks of fine (<1–2 mm) sphalerite and galena veins forming crackle breccias were observed in both drill core and underground exposures. Similar crackle breccias occur associated with the five-lens mineralisation. These veins vary in thickness, with individual veins of as much as several centimetres in diameter, whereas those forming fracture networks tend to be thinner, typically on a millimetre scale. Fractures tend to be filled initially by sulphide minerals, with gangue phases (dolomite, barite, \pm calcite) filling the centre of the vein (Fig. 9A). Fine-grained, disseminated sulphides (dominantly sphalerite) may also be concentrated in the host rock surrounding the veins, becoming less abundant and more disseminated outwards.

Individual fractures filled with sediments rather than cements are common (Fig. 9B). This variety of mineralised vein features a thin coating of pale yellow or yellow-brown sphalerite, followed by finely laminated sediments that invariably reflect the composition of the surrounding host rock. Lath-shaped barite is commonly present within the sediment. Any porosity at the top of the fracture is filled by medium- to coarse-grained saddle dolomite and barite laths. These sediment-filled fractures tend to narrow upwards, pinching out to thin, millimetre-wide sphalerite veinlets, whereas downwards they

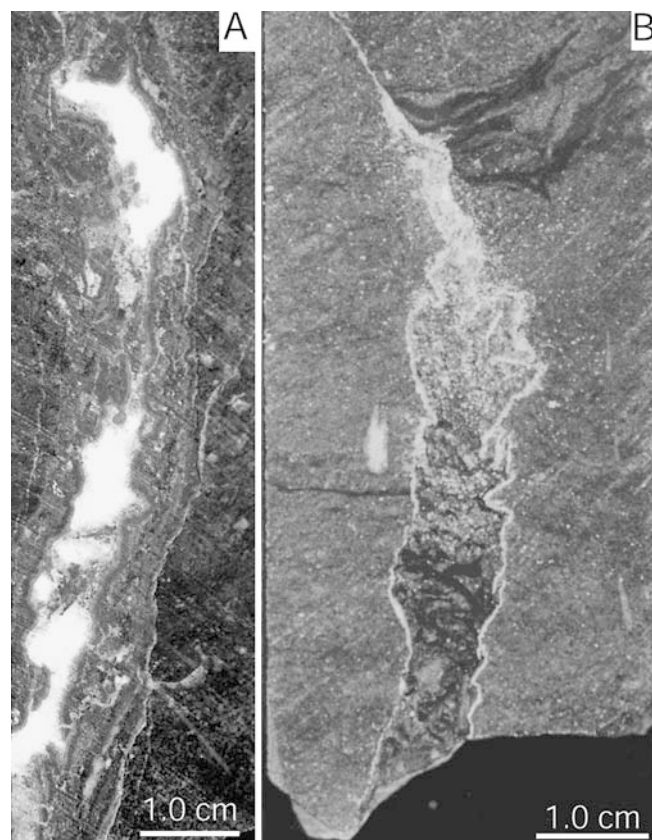


Fig. 9 A Mineralised vein cutting through bioclastic limestone. Centre of vein is filled by coarse white saddle dolomite. N1393, 708.3 m. B Sphalerite-lined vein filled with internal sediments. The vein opens into a mineralised cavity down hole. N1146, 577.2 m

tend to merge into larger, sediment-filled cavities. More unusual are the rare fractures that appear to have taken advantage of dissolution seams as a zone of weakness. These fractures give the impression of being surrounded by a 'dissolution halo', although on closer examination, the vein can be seen to be crosscutting the dissolution seam.

Disseminated sulphide minerals

Fine-grained, disseminated sulphide minerals (predominantly sphalerite) are quite common throughout the U lens. Even drill holes that are considered economically barren frequently contain disseminated sulphide minerals over significant intervals. Disseminated sulphides are common in bioclastic grainstones, quartzose sandstones and dolomitised units, but are much less common in the earlier cemented oolites. The disseminated sulphides are typically pale pink or yellow-brown sphalerite, more rarely accompanied by lesser, fine-grained pyrite and/or galena. As mentioned previously, the irregular cavities commonly feature a halo of disseminated sulphide minerals. Similarly, halos of disseminated sulphides are common surrounding fractures and dissolution seams, with the sulphide minerals most abundant adjacent to the vein or seam and becoming more disseminated outwards, becoming less voluminous into unmineralised host rock. The distance over which the sulphide haloes occur correlates with the degree of mineralisation; disseminated sulphides may extend outward for only a few centimetres or less surrounding a thin vein, but may extend for several metres adjacent to a large sulphide-filled cavity.

Although disseminated sulphides are commonly associated with high-grade mineralisation, this is not always the case. Frequently, there is no apparent relationship with areas of more intense mineralisation. The Upper Sandstone Marker (USM) in particular frequently contains fine-grained, disseminated sphalerite, especially where it has not undergone dolomitisation, despite the fact that the surrounding carbonates are only poorly mineralised.

Disseminated sphalerite occurs within dolomitised horizons filling fine-grained, millimetre-scale moldic porosity, and where present, larger vuggy porosity. Although some vugs are completely filled (mostly by later, post-mineralisation calcite), many are not and are lined instead only by fine-grained, individual crystals of sphalerite, barite and/or saddle dolomite. The type of sphalerite present is typically small (mm size) crystals of honey-coloured sphalerite, although rare crystals of deep-red sphalerite also occur.

Relationship between sulphide minerals and host rock

Through logging of exploration and mine drill core, it is apparent that a majority of the U lens mineralisation favours particular horizons or, more specifically,

particular lithologies. The massive replacement and cavity-hosted mineralisation, which makes up a majority of the ore, is predominantly hosted by bioclastic grainstones. These mineralisation styles (and the associated dissolution features) commonly occur at the contact between bioclastic grainstone and overlying oolitic grainstone or dolomitic siltstone. Disseminated sulphide minerals, although most widespread within bioclastic grainstones and sandstones, occur to some degree within all lithologies. Fracture-filling sulphide minerals occur within most lithologies, as do the dark mosaic breccias, although these particular mineralisation styles are less common within oolitic grainstones and more argillaceous units. As with replacement and cavity-hosted sulphide minerals, both the dark breccias and dissolution seams are frequently located at the contact between bioclastic and overlying oolitic grainstones and dolomitic siltstones.

Sulphide mineral and gangue paragenesis

As with the replacement sulphide minerals, sphalerite is by far the most common sulphide cement in the U lens. Lining both cavities and fractures, as well as encrusting breccia clasts, the sulphide cements are typically colloform in nature, but may also be dendritic (galena), stalactitic (pyrite/marcasite, with or without sphalerite), or botryoidal (pyrite/marcasite). The paragenetic order of the sulphide cements is shown in Table 2, and represented diagrammatically in Fig. 10. As can be seen in Table 2, the sulphide mineral paragenesis consists of several cycles, each with a similar order of sulphide precipitation. The basic cycle is initially iron sulphide, followed by sphalerite that grades from dark to light in colour over time, and finally galena. Barite did not begin to precipitate until the end of the mineralising sequence, with none present in the initial cycles. Except for the final blocky phase of sphalerite, the sulphide minerals are typically fine- to medium-grained in size and appear overall to feature less coarsely crystalline material than the lower lenses (cf. Anderson 1990; Anderson et al. 1998).

Galena

Galena occurs as dendritic/skeletal open-space growths, bands of subhedral to euhedral overgrowths of colloform sphalerite, internal sediments, and as a replacement of earlier sulphide minerals. Three generations of galena have been distinguished (Table 2). The dominant morphology varies with each generation. The first generation of galena (G1) most commonly occurs as 2–5 mm diameter dendrites composed of branches of anhedral crystals <0.05 mm thick (Fig. 11A). The G2 galena typically occurs as discontinuous bands of subhedral to euhedral crystals overlying S2 sphalerite, whereas G3 predominantly occurs as a replacement

Table 2 Sulphide paragenesis of the U lens (Peace 1999)

Order of occurrence (first to last)	Generation
Marcasite/pyrite	M1/P1
Banded phalerite. Pale yellow-brown to dark orange-brown	S1
Galena	G1
Marcasite/pyrite	M2/P2
Banded phalerite. Pale to dark to pale orange-brown (partial replacement by opaquegrey sphalerite)	S2
Galena	G2
Pyrite	P3
Barite	B1
Banded sphalerite. Dark orange-brown to pale yellow-brown with fine/brown bands	S3
Galena	G3
Pyrite	P4
Blocky paleyellow-brown sphalerite	S4
Barite	B2

of earlier sulphide minerals. Several samples of galena have elevated concentrations of antimony, with microprobe analyses revealing as much as 18.9 wt% Sb in some samples. These analyses may represent inclusions of Pb-Sb sulphosalts such as semseyite, which have previously been reported from the main orebody (Steed 1981).

Iron sulphide minerals

Two marcasite and four pyrite generations are recognised within the U lens. Both generations of marcasite (M1 and M2) occur together with P1 and P2 pyrite as botryoidal to stalactitic growths 2–3 cm in length on the roofs and floors of cavities (Fig. 11C). These botryoidal and stalactitic forms are most common where pyrite/marcasite are locally the dominant sulphide minerals, with botryoids more prevalent than stalactites. The stalactites and botryoids both feature alternating layers of marcasite and pyrite. These stalactites appear similar to those described by Ashton et al. (1986) and Anderson (1990) for the lower ore lenses, in particular the 1–5 lens.

The P2 pyrite (not associated with marcasite) also occurs as a generally fine-grained cement overlying colloform S1 sphalerite or G1 galena. The later generations of iron sulphide minerals occur only as pyrite, with marcasite apparently confined to the earlier stalactitic to botryoidal growths. The P3 and P4 pyrite both occur as fine-grained cements in cavities and veins, coating earlier sulphide minerals. The P2, P3 and P4 pyrite may all partially replace earlier sulphide phases.

Pyrite also occurs as replacement of bioclasts, and, less commonly, the surrounding cements. The shapes of these bioclasts (which include ooids, bivalves and brachiopods) have been preserved, although typically internal textures have not. This replacive pyrite commonly occurs in poorly mineralised carbonates, usually associated with dissolution seams. As the host is poorly mineralised, the timing of this replacive pyrite relative to the other pyrite generations cannot be accurately determined. It is thought to most likely be correlative with

P3 or P4, but earlier formation cannot be definitively ruled out.

Sphalerite

Using transmitted light microscopy, four phases of sphalerite mineralisation can be distinguished. These phases have been determined based on sphalerite colour, which is caused by the trace element chemistry, and the relationships of the sphalerite to other sulphide minerals. The first three sphalerite phases (S1, S2 and S3) can be further subdivided into more than ten individual bands of varying colour and thickness. However, as variations in the order of these bands tend to exist between samples and not all bands are always present, discussion is largely confined to the four broader phases of sphalerite. Secondly, although not all ten bands of sphalerite are easily distinguished in hand specimen, the four broad phases are generally readily identifiable. Except for the final phase (S4), the sphalerite cements are typically colloform in nature with fine internal

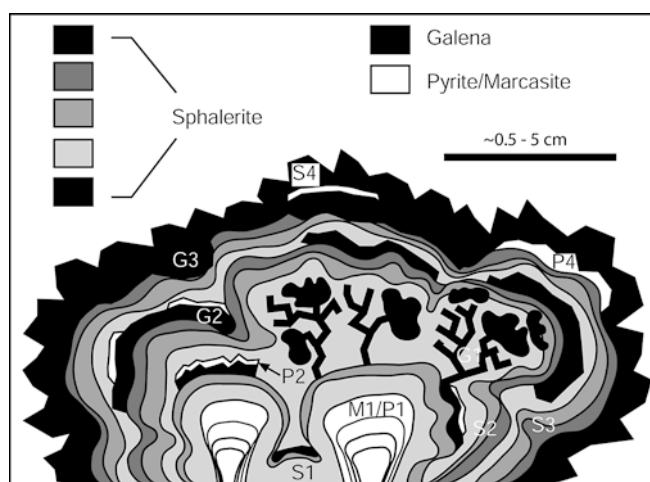


Fig. 10 Diagrammatic representation of the U lens sulphide paragenesis (see Table 2), showing the four broad phases of sulphide mineralisation identified. Note the colloform nature of generations 1–3 of sphalerite, which are followed a final coarser, blocky sphalerite phase (S4) (Peace 1999)

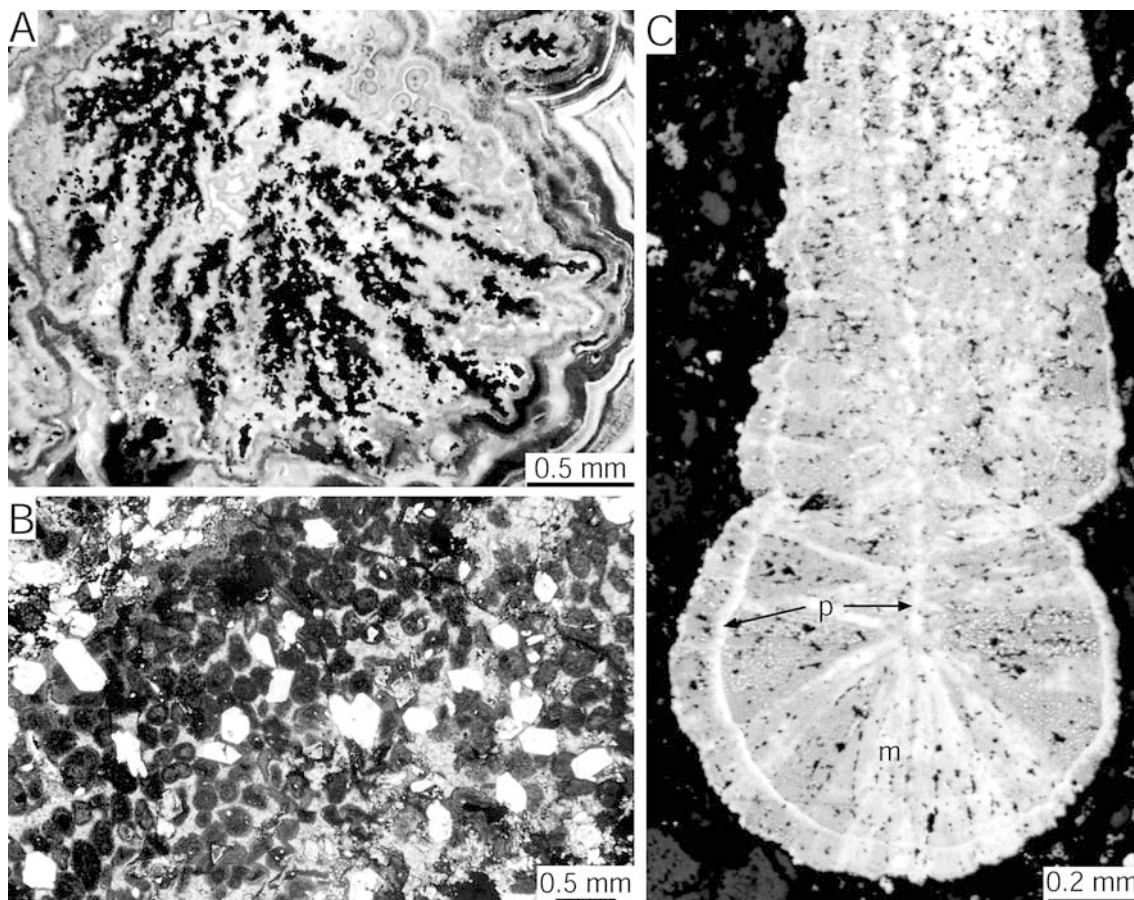


Fig. 11 **A** Cavity-filling dendritic galena (black) overlain by colloform sphalerite cement within a cavity. N1377, 786.2 m. **B** Sandy oolite mimetically replaced by sphalerite. Detrital quartz grains feature euhedral quartz overgrowths. U11390, 39.7 m. **C** Stalactitic pyrite and marcasite. Specimen is largely marcasite (*m*), with a thin pyrite core and band (*p*). Up is to the top of the diagram. N1383, 723.2 m

banding and may cycle through part of a sequence two or three times. All three colloform phases of sphalerite become lighter with time, from dark, almost opaque brown, through shades of orange-brown, to pale yellow-brown 'honeyblende' (Figs. 6C, 11A).

The final phase of sulphide mineral deposition (S4) is formation of a blocky, yellow-brown or honey-coloured ('honeyblende') sphalerite, the most distinctive phase of sphalerite (Fig. 6A, C). Although not the most abundant sphalerite phase, it is probably the most widespread, being readily identified in all the Pale Beds-hosted lenses and the Conglomerate Group ore. In hand specimen, this phase is translucent pale-yellow to yellow-brown. It occurs as blocky, anhedral to euhedral crystals ≤ 5 mm in diameter, and forms a thin isopachous coating over all previous sulphide minerals, including internal sediments, brecciated sulphide clasts, and fracture-filling sulphides. The S4 sphalerite also occurs where no earlier phases of sulphide minerals exist, such as coatings on clasts in dolomite breccias and as linings of small vugs within dolomitised units.

Replacement sphalerite

As emphasised earlier, the most common style of mineralisation within the U lens is replacement of the carbonate component of the host rock with replacement sulphide minerals, almost exclusively sphalerite. Replacement sphalerite is generally fine- to medium-grained (≤ 2 –3 mm), the exception being where sphalerite has replaced fragments of monocrystalline bioclasts such as crinoids, which are typically mimetically replaced by a single sphalerite crystal. Although ooids may be completely replaced by sphalerite, which mimics much of the internal texture of the ooids, bioclasts are most commonly preserved as ghosts with no visible internal textures. Such clasts can be distinguished in thin section or hand specimen by the colour contrast between the sphalerite phase, which has replaced the allochems, and that which has replaced the surrounding cements (e.g. Fig. 5A).

The style of replacement varies from fine-grained, disseminated crystals 'bleeding off' a dissolution seam through to complete, massive replacement over an interval of several metres. Where replacement is relatively minor, it may be obvious that it was initiated along one or more dissolution seams, with replacement decreasing in intensity away from the dissolution seam. However it is as common for there to be no obvious starting point for the replacement process.

The sulphide minerals have generally initially precipitated as a cement and/or replaced the matrix, and then replaced carbonate grains as mineralisation progressed. However, in some instances, the carbonate grains have been replaced in preference to the cements/matrix. In these cases, replacement sphalerite occurs within allochems as very fine-grained (micron-sized), disseminated crystals that are typically concentrated near the edges of the clasts and decrease in intensity inwards. Within brachiopod shells, very fine-grained sphalerite has replaced individual shell-growth zones. Such delicate replacement seems to occur within units that contained little porosity at the time of mineralisation.

It has been shown that replacement ranges from fine-grained, delicate replacement to coarser-grained, more destructive replacement where no primary textures are preserved. Hence, the bioclasts that have been preserved by sulphide replacement, together with detrital quartz and (lesser) feldspar grains, are commonly the sole remnants of the original host rock, and yet these provide strong evidence for replacement as the means of mineralisation.

The majority of replacement sphalerite is pale pinkish-brown, red-brown or pale to dark brown in colour. As most of the replacement sphalerite has not been observed as cements or is obviously cut by or has replaced distinctive sulphide phases, its relative position within the paragenetic sequence is unclear. The pale brown phase of replacive sphalerite is relatively early in the paragenetic sequence, as it is cut by first generation galena (G1) and second generation pyrite and sphalerite (P2 and S2). As replacement sphalerite forms the majority of the ore, replacement has probably occurred throughout the mineralising event.

The only replacement phase that can be reliably placed within the paragenetic sequence is S2OG (S2 OpaqueGrey; see Table 2), a distinctive phase that is opaque grey in transmitted light and pink-cream in hand specimen. Although replacive in nature, it is relatively easily located within the paragenetic sequence as it has replaced only earlier sulphide cements, not carbonate, and is observed only prior to G2 galena in the paragenetic sequence. This phase has most commonly replaced colloform sphalerite overgrowths on dendritic G1 galena, but also occurs as patches replacing distinct bands within colloform sphalerite. The S2OG is itself commonly partially dissolved, especially where it is associated with dendritic galena. This dissolution has obviously been a later event, as the colloform shape of the sphalerite is preserved. Where S2OG surrounds dendritic G1 galena, the molds within the sphalerite are typically filled by bladed barite or more rarely by coarse, poikilotopic post-mineralisation calcite. The presence of the barite and calcite suggests that the dissolution of S2OG occurred towards the end of sulphide mineralisation. Where S2OG occurs within banded, colloform sphalerite, it has typically replaced pale-to medium-brown S2 sphalerite.

Barite

Two phases of barite growth have been identified, with the second phase (B2) the more abundant of the two. Both phases of barite have the same growth habits and are divided into two generations based on their relationship to the sulphide mineral paragenesis described above.

Barite tends to fill most of the porosity that remained after sulphide mineral precipitation, growing as twinned laths ≤ 1.5 cm in length, and as rosettes or partial rosettes composed of much smaller laths. Barite also occurs as fine-grained (< 2 mm) laths (either individually or as crossed twins) within internal sediments. These laths may be a minor constituent or comprise more than 90% of the sediment layer. It is debatable whether these laths settled out from solution or grew within the unlithified internal sediments. However, the interlocking texture of much of the barite suggests a replacive origin. Rarely, post-mineralisation calcite pseudomorphs after barite are observed within sphalerite sediments. At least some barite is replacive, as indicated by clasts of barite within mineralised cavities having remnant textures of ooids.

Saddle dolomite

Saddle dolomite is a common ore-stage mineral, first occurring immediately prior to initial sulphide mineral precipitation and last forming shortly after the end of barite growth. Saddle dolomite occurs dominantly as fine-to coarse-grained euhedral crystals filling primary porosity, secondary moldic and vuggy porosity, and fractures. Many crystals display a distinctive crystal curvature, which in larger crystals is visible in hand specimen. With barite, saddle dolomite typically fills any porosity remaining after the precipitation of the sulphide cements. Although widespread, saddle dolomite is overall only a relatively minor constituent of the ore.

Calcite

Ore-stage calcite, recognised by its bright luminescence in cathodoluminescence, is a very minor component of the U lens mineralisation. It is typically present as a growth band in blocky burial calcite occluding porosity in the enclosing host rock and only rarely occurs as individual crystals intergrown with sulphide minerals. Where present, ore-stage calcite generally forms fine-grained crystals less than 1 mm in diameter. Where porosity remained unfilled within cavities and veins after sulphide mineralisation, it was filled by late, post-mineralisation blocky calcite that is recognised by its mottled, dull/bright luminescence. As with ore-stage calcite, the post-mineralisation calcite is only present in minor quantities.

Quartz

Quartz is quite rare, observed in less than 5% of the sections examined (excluding the slightly more common euhedral overgrowths on detrital quartz). Most, but not all, of the quartz occurs not in the U lens itself, but within the overlying mineralised Shaley Pales. Medium- to coarse-grained, euhedral and subhedral quartz crystals filling porosity in the Shaley Pales contains many fine-grained, high birefringence crystal inclusions and ghosts of crystals, giving the appearance that the quartz had overgrown and replaced a pre-existing mineral, possibly anhydrite.

Chalcedonic quartz has replaced bioclasts, typically crinoid and echinoid fragments. This quartz has a fine-grained, dusty-brown appearance in plane light, and is also more common within the Lower Shaley Pales than the Upper Pale Beds. As some chalcedonic quartz is overgrown by subhedral to euhedral quartz, both types probably precipitated synchronously. Chalcedonic quartz was observed in dissolutional contact with the underlying brightly luminescent calcite, and is overlain by both saddle dolomite and barite. Dusty, chalcedonic quartz also fills the porosity between botryoidal and stalactitic pyrite. Thus, although its exact place in the sulphide mineral paragenesis cannot be determined, this quartz is definitely an ore-stage mineral, occurring later than the brightly luminescent calcite, but earlier than barite. The quartz overgrowths on detrital quartz grains found in several samples of sulphide-rich material (Fig. 11B) are thought to have formed at the same time.

Fluid inclusions

Fluid inclusion analyses were attempted on sphalerite, barite and ore-stage saddle dolomite. Obtaining inclusion data from sphalerite proved difficult; only a solitary homogenisation temperature (147 °C) was obtained from U lens sphalerite. Most sphalerite phases were too dark for reliable identification of inclusions, leaving the pale blocky S4 sphalerite as the principal phase to be studied. Although primary inclusions were observed in the S4 honeyblende, with the single exception stated above, none were suitable for microthermometric measurement. The majority of inclusions were less than 1 µm in diameter and thus too small to accurately measure. Rarer, larger inclusions (~2–10 µm) were dark, opaque and apparently single phase. These inclusions were non-reflective under reflected light, and exhibited a blue luminescence under UV, suggestive of trapped hydrocarbons. Similar problems were encountered with sphalerite from the 5 lens.

Fluid inclusions within barite are common. However, these inclusions produced highly variable, unreliable data: inclusions only a few µm apart produced homogenisation temperatures that differed by as much as 40 °C, whereas other inclusions displayed obvious

evidence of stretching and necking down. Barite was, therefore, ignored in the fluid inclusion study.

Measurement of fluid inclusions within saddle dolomite proved more successful, and as saddle dolomite is an ore-stage mineral, the data obtained from saddle dolomite can be used as a proxy for sulphide precipitation temperatures. The majority of the fluid inclusions observed within saddle dolomite were two-phase with small vapor bubbles. The remainder (much rarer) were opaque, apparently single-phase inclusions. Homogenisation temperatures recorded from primary and pseudo-secondary inclusions range from 90 to 175 °C (Fig. 12), with an average of 132 °C.

Discussion

Sulphide mineralisation is developed preferentially within bioclastic grainstone units, commonly at the contact between the bioclastic grainstone and the overlying oolitic grainstone or dolomitised horizon. Concentration of sulphide minerals beneath dolomite horizons in the lower lenses has been attributed by Anderson et al. (1998) to competency contrasts between the limestone and dolomite resulting in the preferential development of secondary fracture porosity within the limestones. However, cathodoluminescence studies of the Upper Pale Beds (Peace 1999) suggests that porosity differences between the lithologies have been the significant control, with oolitic grainstones and dolomitised horizons becoming essentially impermeable prior to mineralisation. These impermeable horizons have probably acted as stratigraphic traps, confining the mineralising fluids to the more porous bioclastic grainstones. This conclusion is supported by the observation that the majority of the U lens ore is concentrated within bioclastic grainstone units. Argillaceous and dolomitised units within the main orebody may have acted in a similar fashion.

The ore textures within the U lens are indicative of an epigenetic origin for mineralisation, with many textures commonly observed within MVT deposits also noted in the U lens (stalactitic pyrite, skeletal galena, snow-on-roof textures, internal sediments, etc.). Space-filling

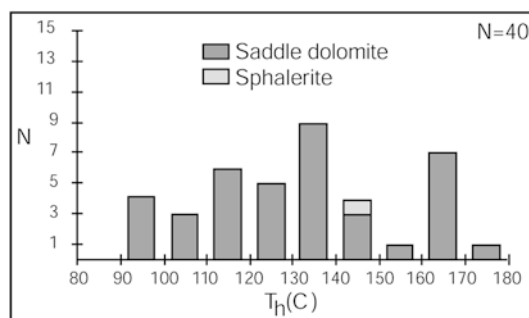


Fig. 12 Histogram of fluid inclusion homogenisation temperatures from dolomite and sphalerite

cements, such as stalactitic pyrite and marcasite and skeletal galena within cavities, indicate that the host units were lithified enough to retain open spaces. Sulphide minerals post-dating burial cements, including truncating veins filled by burial cements, indicate that sulphide mineralisation was late diagenetic in origin (Peace 1999; Peace and Wallace 2000).

As has been noted, the sulphide ore at Navan, in general, and within the U lens, in particular, lacks significant amounts of coarse, space-filling cements (particularly gangue cements) compared to many other epigenetic carbonate-hosted (i.e. MVT) deposits, which may feature a combination of replacement, internal sediments and coarse cements (e.g. Silesia-Cracow, Poland: Leach et al. 1996; Sass-Gustkiewicz 1996; Cadjebut, Western Australia: Bradley 1994). However, *none* of the ore textures observed in the U lens are unknown in MVTs.

The conditions at Navan obviously favoured replacement of the host carbonate over the precipitation of space-filling sulphide and non-sulphide cements. Possible reasons for this preference include: (1) the amount of metals in solution being greater than the open space available for sulphide cement growth, (2) fluid conditions favouring replacement over cementation, or (3) that the extended periods of uninterrupted cementation necessary for the growth of coarse cements were prevented by the continued brecciation of the cavity walls, resulting in continual re-nucleation of cements on the new cavity edges. A combination of these or other factors may have been responsible for the dominance of replacement sulphide minerals.

As well as the overall dominance of replacive mineralisation, the apparent lack of cements associated with mineralised cavities is partially explained by the presence of internal sediments leaving less space for cement growth (and potentially being less favourable sites for sulphide nucleation), and by the likelihood that the deposit was still undergoing compaction due to burial at the time of mineralisation. This compaction, evidenced by soft-sediment deformation textures within cavity-filling internal sediments, may have severely reduced the open space remaining within the cavities. The relatively high volume of internal sediments within the mineralised cavities is due to the high siliciclastic and organic content of the Upper Pale Beds. A second possible reason for the lack of carbonate gangue is that, during mineralisation, the fluids were largely undersaturated with respect to calcite and dolomite. Dissolution of the host carbonate obviously produced the cavities, implying carbonate undersaturation of the fluid.

Cavities beneath layers of replacement sphalerite, which contain clasts of the overlying sulphide minerals, suggest that replacement of carbonate horizons by sulphides began prior to the initiation of cavity formation. Replacement of the host carbonate possibly caused the mineralising fluids to be trapped beneath them, resulting in dissolution and cavity/breccia formation.

The role of dissolution in the mineralising process is beyond the scope of this paper. However, the presence of

dissolution seams, particulate crackle breccias and dark mosaic breccias is evidence that dissolution has occurred. Unlike dissolution seams related to burial compaction, these dissolution seams are related to mineralisation because: (1) they are distinctly different in appearance from the burial-related stylolites within the Upper Pale Beds; (2) they are commonly mineralised; and (3) they are more abundant near mineralisation. Similarly, the particulate crackle breccias and dark mosaic breccias are not observed distal to the ore and they generally feature fine-grained, disseminated sulphide minerals within their matrix. The relationships of these breccias and dissolution seams to the irregularly mineralised cavity and breccia systems suggest that they formed coevally as a result of the same mineralising processes.

Temperature of ore formation

The raw fluid inclusion data indicate temperatures during sulphide mineralisation of 90–175 °C. However, the strong evidence of stretching (thermal reequilibration) of inclusions within barite suggests that these temperatures should be viewed with a degree of caution. The wide range and irregular distribution of these temperatures, which are suggestive of stretching, may have led to some erroneously high homogenisation temperatures.

Inclusions within the same growth zone of a given sample of dolomite may give temperatures varying by as much as 20–30 °C. Goldstein and Reynolds (1994) state that for inclusions to be considered as part of the same group, 95% should not vary by more than 10–15 °C. Clearly this is not the case at Navan. Some of the fluid inclusions have probably undergone thermal reequilibration (stretching) some time after sulphide mineral deposition (possibly during further burial). The resulting thermal reequilibration results in homogenisation temperatures that are thus higher than true trapping temperatures. Alternatively, however, inclusions within stronger minerals are less easily stretched than those in softer minerals, and so the data for the dolomite inclusions (e.g. Fig. 12) may approximate the original trapping temperatures. Assuming only some of the inclusions in dolomite have undergone stretching, the lower recorded values are taken here to indicate the true trapping temperatures. As such, sulphide mineralisation is likely to have occurred at temperatures of approximately 90–150 °C.

It should be noted that even if stretching of inclusions is not considered for the highest temperature inclusions, the maximum temperature recorded is 175 °C. This temperature is significantly less than the previously published data from Navan of 120–250 °C and 187–220 °C (Rizzi et al. 1990 in Hitzman and Beaty 1996; Everett et al. 1999), respectively. However, the validity of the previously published data must be questioned because the solitary sample used by Everett et al. (1999) is, as admitted by the authors, of dubious origin and potentially not from Navan at all, whereas the data

presented by Rizzi et al. (1990) were corrected by Rizzi (1992) to a maximum of 159 °C. It is therefore suggested that the temperature of sulphide mineralisation has previously been overestimated. The current data suggest mineralisation temperatures of approximately 90–150 °C, and certainly no greater than 175 °C, in agreement with the data of Rizzi (1992).

Summary

The mineralogy of the U lens ore is relatively simple, being composed predominantly of sphalerite with lesser amounts of pyrite, marcasite and galena. Non-sulphide minerals include barite, saddle dolomite, minor calcite and rare quartz, although the amount of gangue is low compared to many other epigenetic carbonate-hosted Zn-Pb deposits. Sulphide mineralisation within the U lens is dominated by sphalerite replacement of the host carbonates. However, a significant proportion of this replacement mineralisation is associated with irregular cavity and breccia systems, either as breccia clasts within the cavities or as replacement of the surrounding or overlying host rock. A network of dark matrix-filled breccias and dissolution seams, which surround both massive replacement mineralisation and filled cavities, indicates (along with the cavities themselves) that dissolution of the host carbonate played an important role in the mineralising process.

Both massive replacement and cavity-related ores predominantly occur within bioclastic grainstones, especially near the contacts with overlying less permeable units. These units, which have been shown to have been well-cemented by the time of mineralisation, acted as barriers to the mineralising solutions. These impermeable beds caused the metal-bearing fluids, which were warm brines of approximately 90–150 °C, to be focused within the more porous lithologies.

The U lens at Navan is a replacement-dominated, epigenetic deposit that formed at low (<175 °C) temperatures. There is no evidence from the U lens of syngenetic or early diagenetic mineralisation, as described from other Irish carbonate-hosted deposits (Hitzman and Beaty 1996).

Acknowledgements The authors gratefully acknowledge the permission of Outokumpu-Tara Mines Ltd. to carry out and publish this work. Thanks are also extended to the Tara mine and exploration geologists, especially J. Geraghty, B. O'Keefe and E. Hyland. We are also grateful to D. Sangster and J. Gregg for their helpful comments in reviewing this manuscript. This research was carried out as part of a Ph.D. project by W.M.P. whilst in receipt of an Australian Postgraduate Award. M.W.W. was funded by an Australian Research Council Large Grant.

References

- Anderson IK (1990) Ore depositional processes in the formation of the Navan zinc/lead deposit, Co. Meath, Ireland. PhD Thesis, Strathclyde University, Glasgow
- Anderson IK, Ashton JH, Boyce AJ, Fallick AE, Russell MJ (1998) Ore depositional processes in the Navan Zn-Pb deposit, Ireland. *Econ Geol* 93:535–563
- Andrew CJ, Ashton JH (1985) The regional setting, geology and metal distribution patterns of the Navan orebody, Ireland. *Trans Inst Min Met (Sect B: Appl Earth Sci)* 94:66–93
- Ashton JH (1995) Guide to the geology of the Navan orebody. In: Anderson IK, Ashton J, Earls G, Hitzman M, Tear S (eds) Irish carbonate-hosted Zn-Pb deposits. Society for Economic Geologists, Fort Collins, pp 151–168
- Ashton JH, Downing DT, Finlay S (1986) The geology of the Navan Zn-Pb orebody. In: Andrew CJ, Crowe RWA, Finlay S, Pennell WM, Pyne JF (eds) Geology and genesis of mineral deposits in Ireland. Irish Association for Economic Geology, Dublin, pp 243–280
- Ashton JH, Black A, Geraghty J, Holdstock M, Hyland E (1992) The geological setting and metal distribution patterns of Zn-Pb-Fe mineralization in the Navan Boulder Conglomerate. In: Bowden AA, Earls G, O'Connor PG, Pyne JF (eds) The Irish minerals industry 1980–1990. Irish Association for Economic Geology, Dublin, pp 171–210
- Ashton JH, Holdstock MP, Geraghty JG, O'Keefe WG, Martinez N, Peace W, Philcox ME (2000) The Navan orebody discovery and geology of the SW extension. Europe's major base metal deposits, IAEG Conference, Galway, Abstracts Volume
- Ashton JH, Holdstock MP, Geraghty JF, O'Keefe WG, Martinez N, Peace W, Philcox ME (2003) The Navan orebody discovery and geology of the SW extension. In: Proceedings of the IAEG Millennium Conference on Europe's base metal deposits, Galway, May 2000, Irish Association for Economic Geology, Dublin
- Bradley AJ (1994) The origin of Mississippi Valley-type sulfide mineralisation at Twelve Mile Bore, Lennard Shelf, Western Australia. Honours Thesis, University of Melbourne, 70 pp
- Dzulynski S, Sass-Gustkiewicz M (1989) Pb-Zn ores. In: Bosak P, Ford DC, Glazek J, Horacek I (eds) Palaeokarst. A systematic and regional review. Elsevier, Amsterdam, pp 377–397
- Everett CE, Wilkinson JJ, Rye DM (1999) Fracture-controlled fluid flow in the Lower Palaeozoic basement rocks of Ireland: implications for the genesis of Irish-type Zn-Pb deposits. In: McCaffrey KJW, Lonergan L, Wilkinson JJ (eds) Fractures, fluid flow and mineralization. Geological Society, London, pp 247–276
- Goldstein RH, Reynolds TJ (1994) Systematics of fluid inclusions in diagenetic minerals. Society for Sedimentary Geology, Tulsa, 199 pp
- Heyl, AV (1983) Geologic characteristics of three major Mississippi Valley districts International conference on Mississippi Valley type lead-zinc deposits. In: Kisvarsany G, Grant SK, Pratt WP, Koenig JW (eds) Proceedings volume, Rolla, Missouri, pp 27–60
- Hitzman MW, Beaty DW (1996) The Irish Zn-Pb-(Ba) orefield. In: Sangster DF (ed) Carbonate-hosted lead-zinc deposits. Society for Economic Geologists, Special Volume no 4, pp 112–143
- Leach DL, Viets JG, Kozłowski A, Kibitlewski S (1996) Geology, geochemistry, and genesis of the Silesia-Cracow zinc-lead district, southern Poland. In: Sangster DF (ed) Carbonate-hosted lead-zinc deposits. Society for Economic Geology, Special Volume no 4, pp 144–170
- Morrow DW (1982) Descriptive field classification of sedimentary and diagenetic breccia fabrics in carbonate rocks. *Bull Can Petrol Geol* 30: 227–229
- Peace WM (1999) Carbonate-hosted Zn-Pb mineralisation within the Upper Pale Beds at Navan, Ireland. PhD Thesis, University of Melbourne, 284 pp
- Peace WM, Wallace MW (2000) Timing of mineralization at the Navan Zn-Pb deposit: a post-Arundian age for Irish mineralization. *Geology* 28:711–714
- Philcox ME (1984) Lower Carboniferous lithostratigraphy of the Irish Midlands. Irish Association for Economic Geology, Dublin, 89 pp
- Philcox ME (1989) The mid-Dinantian unconformity at Navan, Ireland. In: Arthurton RS, Gutteridge P, Nolan SC (eds) The role of tectonics in Devonian and Carboniferous sedimentation in the British Isles. Yorkshire Geological Society, Bradford, pp 67–81

- Rizzi G (1992) The sedimentology and petrography of Lower Carboniferous limestones and dolomites; host rocks to the Navan zinc-lead deposit, Ireland. PhD Thesis, University of Glasgow, 369 pp
- Rizzi G, Braithwaite CJR (1997) Sedimentary cycles and selective dolomitization in limestones hosting the giant Navan zinc-lead ore deposit, Ireland. *Expl Miner Geol* 6:63–77
- Rizzi G, Braithwaite CJR, Ashton JH (1990) Dolomitization at Navan, County Meath, Ireland. Abstracts Volume, Mineral Deposits Studies Group, Dublin, 27 pp
- Sangster DF (1988) Breccia-hosted lead-zinc deposits in carbonate rocks. In: James NP, Choquette PW (eds) *Paleokarst*. Springer, Berlin Heidelberg New York, pp 102–116
- Sass-Gustkiewicz M (1996) Internal sediments as a key to understanding the hydrothermal karst origin of the Upper Silesian Zn-Pb ore deposits. In: Sangster DF (ed) *Carbonate-hosted lead-zinc deposits*. Society for Economic Geology Special Publication no 4, pp 171–181
- Sass-Gustkiewicz M, Dzulynski S, Ridge JD (1982) The emplacement of zinc-lead sulphide ores in the Upper Silesian district – a contribution to the understanding of Mississippi Valley-type deposits. *Econ Geol* 77:392–412
- Steed GM (1981) Reports for Tara Mines Ltd
- Strogen P, Jones GL, Somerville ID (1990) Stratigraphy and sedimentology of lower carboniferous (Dinantian) boreholes from West Co. Meath, Ireland. *Geol J* 25:103–137